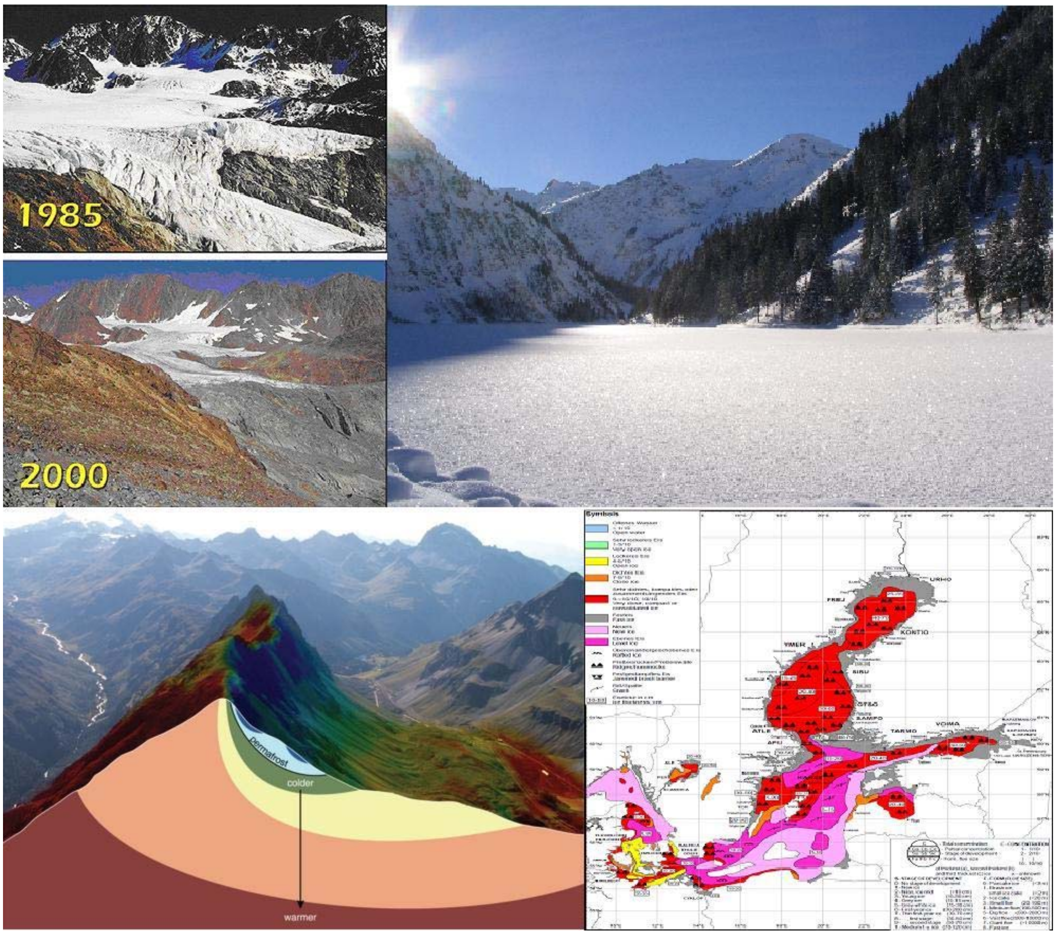
Impacts of climate change on snow, ice, and permafrost in Europe: Observed trends, future projections, and socio-economic relevance



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Front page picture:
European cryosphere– top left: shrinking glacier Vernagtferner (source: Weber; BAdW); top-right: snow-covered landscape in the Alps (source: Zebisch; EURAC); bottom left: temperature distribution within a mountain range containing permafrost (source: Gruber; Uni Zuerich); bottom right: Ice-map of the Baltic Sea (source: Schmelzer; BSH)

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1. Executive Summary

1.1. Purpose and scope

This Technical Paper presents an assessment of the impacts of recent and projected changes in climate on the cryosphere (ice, snow, and permafrost) in Europe, and of the societal relevance of these changes. This paper was prepared by the European Topic Centre on Air and Climate Change (UBA-D) in close cooperation with the Department of Geography of the University of Zuerich and the WSL Institute for Snow and Avalanche Research Davos (SLF) with important contributions by several experts across Europe. The aim of the paper is to provide short but comprehensive information covering the main components of the cryosphere across Europe. Information about cryospheric components of the Arctic region, in particular Arctic sea ice and the Greenland ice sheet, is not included because these systems are already covered by a large number of scientific publications, including those published as follow-up of the International Polar Year in 2007/2008.

This paper updates and completes information about the cryosphere as presented in previous EEA reports on climate change impacts in Europe (2004, 2008). It is intended to provide information on the European cryosphere to the European Environment Agency (EEA), which is deemed relevant for future EEA reports and for the Adaptation Clearinghouse for Europe. This paper is also intended to serve the information needs of a wider audience, including policy-makers at the European, national and sub-national level, non-governmental organisations, and the interested public.

1.2. Developments in science and policy

In its Fourth Assessment Report published in 2007, the Intergovernmental Panel on Climate Change (IPCC) confirmed and strengthened earlier scientific findings about key aspects of climate change. Increased monitoring and improved research have enhanced understanding of climate change, recent and projected impacts, and societal vulnerability to these impacts. European research on impacts and vulnerability in national and EU programmes has advanced considerably, making a major contribution to international assessments of the IPCC as well as the Arctic Climate Impact Assessment (2004), the UNEP Global Outlook for Ice and Snow (2007), the UNEP/wgms-report on global glacier changes (2008), and the UNEP Climate Change Science Compendium (2009). The international conference 'Climate Change –Global Risks, Challenges & Decisions' held in Copenhagen in March 2009 updated the available information on climate change and clearly demonstrated the urgency of political action.

The EU aims to limit global temperature increase to 2 °C above the pre-industrial level. This target is also referred to in the 'Copenhagen Accord', the key outcome of the UNFCCC COP-15 held in December 2009 and is fixed in the documents of the COP-16 as held in Cancun in December 2010. In addition, the EU has put in place a wide range of policy measures to address mitigation of climate change and adaptation to climate change within the EU.

1.3. Structure of this Technical Paper

The main part of this paper summarises the societal relevance of the 5 main components of Europe's cryosphere (glaciers, mountain permafrost, snow cover, Baltic Sea ice, lake and river ice), observed trends and future projections under the conditions of climate change, as well as selected 'secondary' impacts of climate change (e.g., avalanches and land slides) in the 7 most important regions with cryospheric components in EEA-Europe (the part of Europe as covered by EEA-member countries): the **Alps**, **Fenno-Scandinavia** (Sweden, Finland, and the Norwegian mainland), **Svalbaard**, **Iceland**, the **Tatra mountains,** the **Pyrenees,** and the **Baltic Sea**. Additional information on damages in the Alpine region and Scandinavia caused by extreme weather events in some of these regions is provided in attachments to this paper because this was regarded as beyond the scope of the main paper. An overview of European institutions and networks that are involved in observing the cryosphere or in protecting society from cryosphere-related hazards is also presented in Annexes.

1.4. Key messages

Climate change across Europe:

Global and European temperature have increased significantly since pre-industrial times. Global mean temperature in 2009 was 0.7–0.8 °C above the 1850–1899 average, and the annual average temperature for the European land area in 2009 was 1.3 °C above the 1850–1899 average. Annual precipitation increased in northern Europe by 10 to 40 % but decreased in some parts of southern Europe by up to 20 % in the 20th century. Mean winter precipitation has increased in most of western and northern Europe by 20 to 40 %, whereas southern Europe and parts of Central Europe were characterised by drier winters.

All regions considered in this report (the Alps, Fenno-Scandinavia, Svalbaard, Iceland, the Tatra Mountains, the Pyrenees, and the Baltic Sea) have experienced substantial warming as well as changes in the seasonality and intensity of precipitation, which varied across regions. Climate change projections (based on the outcome of ENSEMBLES) generally project a continuation and in most of the regions an acceleration of the observed trends, whereby these projections depend on the scenario of future greenhouse gas emissions.

Primary impacts of climate change on the cryosphere

Snow cover:

The observed climatic changes are already having a significant impact on the snow cover in Europe. Snow reliability has been reduced at low and medium altitude, which is mainly caused by warmer winter temperatures. Constant or stable snow amounts have only been observed at higher latitudes or at altitudes above 2000 m. The projected increase in temperature would shift this latitude and altitude limit even higher. Therefore, winter tourism will be restricted to a shorter time period and/or to regions at increasingly high altitude or latitude.
The declining snow reservoir will also cause longer periods of low river flow in summer in many parts of Europe. This can have severe consequences for several economic sectors, including agriculture, hydropower generation, water supply, and river navigation.

Glaciers and ice caps:

Due to their proximity to melting conditions, glaciers are one of the most reliable natural indicators for climatic changes. In the second half of the 20th century, European glaciers and ice caps (outside Greenland) covered a total of about 54,000 km^2 distributed in Svalbard (68%), Iceland (21%), Scandinavian Peninsula (6%), Alps (5%), and the Pyrenees (<1%).
The strong centennial retreat of glaciers from the Little Ice Age moraines is well documented and apparent in all European regions. In some regions, there have been intermittent periods of reduced glacier melting or even of glacier re-advance such as in the late 1970s in the Alps and Iceland and in the 1990s in coastal Scandinavia. In the European Alps, more than half of the ice-covered area disappeared since 1850.
European glacier changes since the Little Ice Age have been driven mainly by increased summer air temperatures but variations in winter precipitation also played a role. Both factors are influenced by atmospheric and oceanic circulation patterns. Further factors for the observed glacier melting in most regions are most probably the (re-) brightening of the atmosphere, extension of the ablation period, and reinforcing effects such as dust-related darkening or melt-induced elevation lowering of glacier surfaces.
Climate change scenarios for the 21st century project that glacier retreat will continue and they may totally disappear from some mountain ranges in the coming decades.

Permafrost:

In Europe, permafrost is a widespread phenomenon in the European sector of the Arctic as well as in the alpine high mountain environments (whereby the permafrost science in the mountains presents a relatively young field of research). In most of these regions the permafrost is “warm” (close to 0 °C) and is sensitive to air- and surface-temperature changes and/or to changes in local permafrost controlling conditions such as snow cover, ice content and vegetation.
Changes in spatial extent, thickness and temperature of permafrost are recognised as indications for climate change impacts.
A warming of the permafrost in the northernmost part of Europe of 0.5-1 °C was observed during the past decade. In the European Alps trends are less clear compared to northern Europe and masked by the high annual variations resulting from varying snow conditions and modulated by heat exchange in warm permafrost close to 0 °C.
The projected temperature increase (by 2100 up to 4° C in the Alps, 4-6° C in Svalbard) compared to the 1970-2000 period level will very likely lead to continued warming and thawing of permafrost.
Increasing permafrost temperatures are expected to contribute to enhanced destabilization of slopes. They can change the frequency and magnitude of rock falls, debris flows, and thus influence the safety and maintenance of constructions and infrastructure, especially in alpine permafrost environments with high ice contents.
Permafrost science is a relatively young research field and therefore only a limited number of continuous and long-term data series is available in Europe. The challenge lies in the ongoing and assured monitoring, for research purposes as well as for hazard assessments.

Lake and river ice:

The duration of ice cover on lakes and rivers in the northern hemisphere has shortened at a mean rate of about 12 days per century, resulting from an average of 5.7 days later for ice-on and 6.3 days earlier for ice-off. The changes are much more pronounced in lakes and rivers in the temperate region where the ice season is already short or ice cover only occurs in cold winters compared to lakes and rivers in colder regions such as northern Scandinavia. The inter-annual fluctuations in the timing of ice-off are highly correlated over large spatial scales, and are often dominated by climate modes such as the North Atlantic Oscillation.

Baltic Sea ice:

The maximum extent of Baltic Sea ice has decreased in the last century. The length of the ice season in the Baltic Sea has decreased by 14-44 days in the last century, depending on location.
Ice thickness data do not show clear trends during the 20th century but ice-thickness has decreased in the last 20 years (1990–2010). During the last ten years all ice winters have been average, mild, or extremely mild whereas none of them has been severe or extremely severe.
The maximum extent of Baltic Sea ice is projected to decrease by 57%-71% in the next 100 years, depending on the emissions scenario. The length of the ice season is projected to decrease by 1-2 months in the North and by 2-3 months in the central part of the Baltic Sea over this time period.

Secondary impacts of changes in the cryosphere

Avalanches:

The event history of avalanches is extremely difficult to relate to climate change aspects, due to two accompanying developments, the increase in snow sports and the large investments in technical avalanche defence measures.

However, the average avalanche activity in Europe has not changed during the last decade where reliable informations are available. The last winter with many large avalanches in Europe was 1998/1999. Several fatalities from avalanches occur every year, most of which occur in relation to snow sports.
High safety standards with respect to avalanches have been attained in Europe. Maintaining this safety level requires improving technical countermeasures, early warning systems and training of rescue-staff.
Climate change is expected to affect avalanche activity at lower altitudes only.

Landslides and rock slope failures

No significant change in frequency of shallow landslides and debris flows has been observed so far for European mountain regions. This is partly due to insufficient documentation because A synthesis perspective and area-wide information on landslides and mass movements in high-mountain regions of the Alps is missing. However, large rock slope failures in permafrost regions in the Alps have increased in the last two decades as compared to the previous 100 years.
The timing, frequency and magnitude of Alpine debris flows are likely to change in the coming decades, with a trend towards earlier initiation in the season and initiation from higher elevations. Climate change will likely create new areas being affected by landslide hazards. As a consequence, existing landslide, mass movement, and rock-fall hazard maps over many regions in Europe may become out dated.

Glacier floods

Glacier floods are rare events in Europe that only occur in some high mountain regions. However, glacier floods can be highly destructive, and the costs for measures to mitigate an imminent event or reconstruction after the event can amount to tens of millions EUR. While hundreds of people were killed by glacier floods in historic times, no fatalities have occurred in recent times.
Due to their rare occurrence it is difficult to detect any change in the frequency of glacier floods over the last several decades. Ongoing glacier retreat will result in further growth or new formation of glacier lakes. These changes may benefit hydropower production and tourism but they also entail considerable risks.
Of main concern are impacts from slope failures from destabilized high-mountain flanks that may increase with ongoing climate change. It is therefore important to timely assess such developments to be able to initiate prevention and adaptation measures.

2. Introduction

2.1. Purpose and scope

Recent decades have seen notable changes in global and European climate. This Technical Paper of the European Topic Centre on Air and Climate Change (ETC/ACC) presents an indicator-based assessment of recent and projected climate changes and their impacts on the European cryosphere. Its objectives are to:

- Present past and projected climate change and its impacts in the cryosphere through easily understandable, scientifically sound and policy-relevant indicators;
- Increase awareness of the need for global, EU and national policy action;
- Highlight the need to enhance cryosphere monitoring, data collection and dissemination, and reduce uncertainties in climate and impact modelling.

The paper is prepared by the ETC/ACC (here: by UBA-D) under close cooperation with several partners across Europe: the University of Zuerich (UZH), the WSL institute for snow and avalanche research in Switzerland (SLF), the Federal Research and Training Centre for Forests, natural Hazards and Landscape in Austria (BFW). Important and useful contributions from the University of Innsbruck, the Munich Reinsurance Company, and the Bavarian Academy of Sciences, the University of Oslo, the Norwegian Geotechnical Institute (NGI) and the Norwegian Institute for Water Research (NIVA) completed the paper. Several other European institutions kindly contributed information and figures to this paper.

The aim of the paper is to provide short but comprehensive information covering the main componets of the cryosphere across Europe. In some cases where Europe-wide data was not available, indicators for key regions in Europe have been presented as long as data was available for several countries.

This paper updates and completes information about the cryosphere as presented in previous EEA reports on climate change impacts in Europe (2004, 2008). It is intended to provide information on the European cryosphere to the European Environment Agency (EEA), which is deemed relevant for planned EEA reports and the Adaptation Clearinghouse for Europe. This paper is also intended to serve the information needs of a wider audience, including policy-makers at the European, national and sub-national level, non-governmental organisations, and the general public.

Cryospheric components of the Arctic region, in particular Arctic sea ice and the Greenland ice sheet are explicitly excluded because these systems are already covered by a large number of scientific publications, including those published as follow-up of the International Polar Year (IPY) in 2007/2008.

2.2. Background and policy framework

Anthropogenic climate change causes many different impacts on societies, economies, and ecosystems, most of which are considered adverse. The United Nations Framework Convention on Climate Change (UNFCCC) came into force in 1994 to limit the impacts of climate change, and to avoid dangerous consequences. The ultimate objective of the UNFCCC is “to achieve stabilisation of greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference with the climate system”. To avoid 'dangerous climate change' the EU has proposed to limit global temperature increase to 2° C above the pre-industrial level. Temperature stabilisation at this level will require global emissions to stop rising by 2020 and to be reduced to less than 50 % of 1990 levels by 2050. An international agreement on climate policy beyond 2012 is being negotiated under the UNFCCC, with the aim of reaching an effective agreement as soon as possible.

However, there is growing awareness that, even if GHG emissions were stabilised today, increases in temperature and associated impacts will continue for many decades.
The *cryosphere* as the frozen part of the world includes all permanent and seasonal snow and ice deposits on land, in the seas, rivers and lakes as well as frozen ground (permafrost). It is the second largest component of the climate system after the oceans with regard to mass and heat capacity. The cryosphere is on the one hand a very useful indicator to monitor changes climate, on the other hand it also plays a crucial role in many climate processes that directly affect human societies. Snow and ice play a key role in the earth's energy budget by reflecting heat from the sun. As melting replaces white surfaces with darker ones, more heat is absorbed (the albedo effect). Snow and ice play a key role in the water cycle and are essential for storing fresh water for hotter and often dryer seasons. Two thirds of the world's freshwater resources are frozen. Furthermore, the cryosphere is important for the exchange of gases between the ground and the atmosphere – these include several greenhouse gases, e.g. methane. Finally, ice and snow are defining components of ecosystems in the northern parts of the northern hemisphere and in high mountain areas. Many plants and animals have evolved to live under these conditions and can not live without. The cryosphere thus plays a major role in various dimensions of the climate system: it is affected if the climate changes, but its own changes in turn affect the climate system. Monitoring these changes therefore provides crucial knowledge about climate change.
Changes in the cryosphere provide a very visible expression of climate change because the cryosphere integrates climate variations over a wide range of time scales, from millennia to seasonal variations throughout the year. However, the interaction of processes with different time scales can also complicate detection, interpretation and attribution of observed changes as well as projection of future changes.
In the European Alps and other high mountain areas, temperatures have increased faster than the global average in the past few decades, which has lead to substantial decreases in the amount of ice and snow. European glaciers are shrinking, snow-covered areas are creeping higher up and further north, Baltic Sea ice is in retreat and mountain permafrost is starting to thaw. All these trends will accelerate with continued climate change.

2.3. Presentation of Indicators

The discussion of climate change impacts on the most important components of the European cryosphere is focussed on the regions as presented in Fig.2.1.
The various components of the European cryosphere as described in this paper play strong but different roles within the climate system:

- *Snow* covers a large area but has relatively small volume. It plays a key role in regional feedback processes with global importance due to large increases in the absorption of heat when snow melts (albedo effect).
- *Glaciers, ice caps and seasonal lake and river ice* react relatively quickly to changes in climate, influencing ecosystems and human activities on a local scale.
- *Baltic Sea ice* impacts the surface albedo and the energy transfer at the sea surface.
- *Mountain permafrost* influences mountain landscapes, affects human infrastructure, and contains valuable information on climate change.

The secondary impacts of climate change in the cryosphere, such as partly contributing to *avalanches, land slides, mass movements, rock fall, flash floods and outbursts of glacier lakes* have direct, and mostly adverse impacts, on human societies, which are described in a specific chapter.

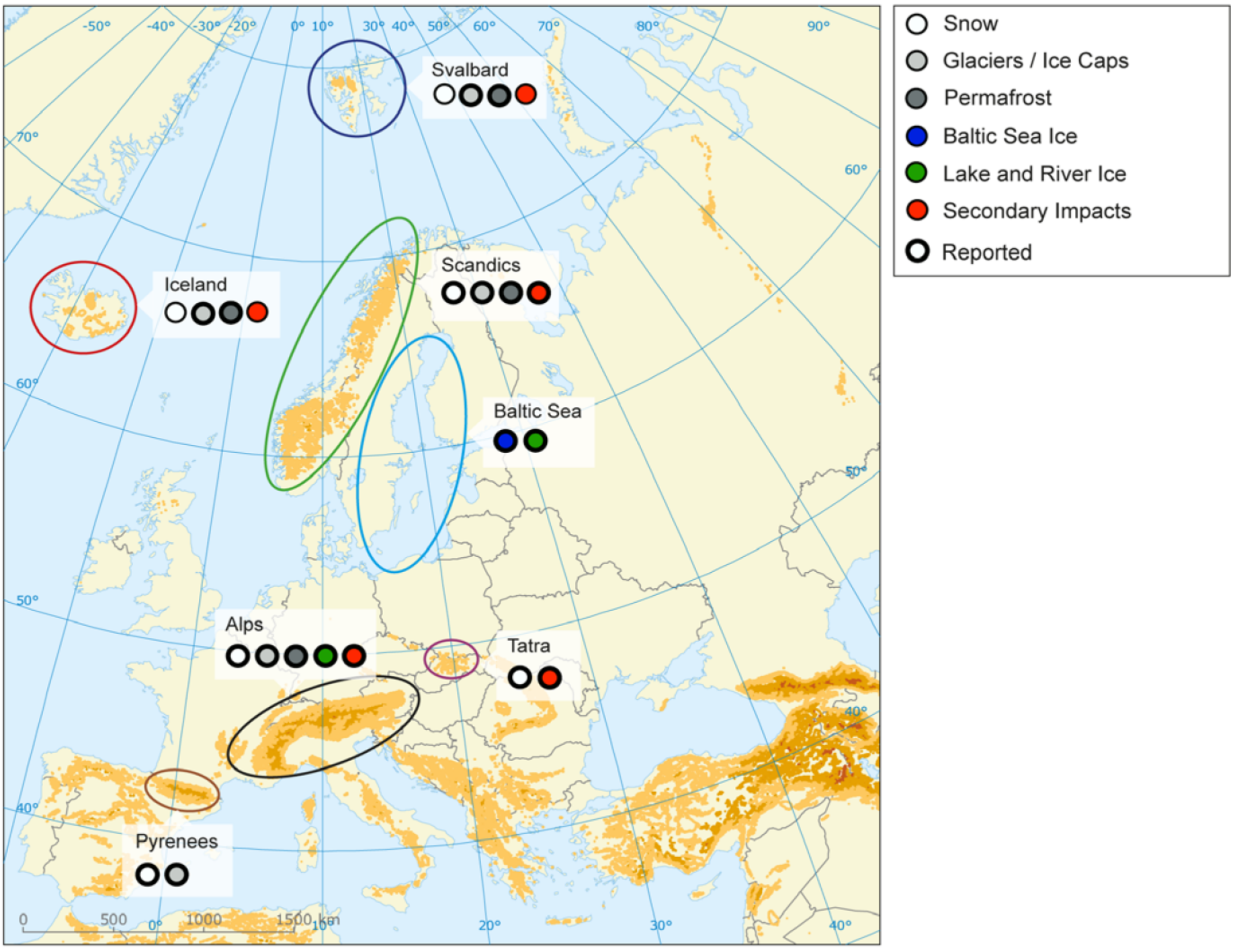


Figure 2.1: Overview on regions and components of the cryosphere as discussed in the paper

Source: Voigt/Woelfer ; UBA-D; 2010
Note: All existing cryosphere components in a region are characterized by coloured circles; those components covered in this paper are denoted by circles with thick black borders.

The cryosphere indicators in this paper cover strategic information from the following cryosphere components: glaciers and snow cover in Europe, mountain permafrost, Baltic Sea ice, as well as lake and river ice.
Each of the indicators is presented in a similar structure:

- Key messages that summarise observed and projected trends;
- Key graph to demonstrate observed trends;
- Environmental and socio-economic and relevance, societal response options to observed and/or projected changes and uncertainties related to the indicator and its data base;
- Past trends based mainly on analysis of long time series of reliable observations;
- Future projections, based mainly on results from existing global IPCC models and scenarios adapted to the European situation.

The structure of this paper is built around different components of the cryosphere such that the impacts of changes in these cryosphere components are considered separately. In the real world, however, changes in different components of the cryosphere, of other elements of the climate system, and non-climatic changes interact with one another, often in unexpected ways.

2.4. Data availability and quality of information

Data on the various components of the cryosphere vary significantly with regard to availability and quality. Long‑term data on glaciers for all glaciated areas in Europe (back to the 19^{th} century) are provided in good quality and quantity by the World Glacier Monitoring Service (WGMS). Similarly, long-term data on the Baltic Sea ice extent is provided by several national ice services. In contrast, time series for snow depth from in-situ measurements are rarely going back more than 100 years and are often much shorter because most regions have operational snow measurement networks only for the last 50 years. Data on snow cover have been measured with satellites since the 1970s and are available globally, for example from the Global Snow and Ice Data Centre (NSIDC) in Boulder, CO, USA. Data on permafrost from deep boreholes are often available for no longer than 15 years.
The gaps in the cryospheric data base are well recognised by the scientific community and many efforts are being made to improve monitoring and data collection.

2.5. Important background-information

This section explains four terms related to the science and policy of climate change to provide important background information for the main part of this paper.

(a) The North Atlantic Oscillation (NAO)

Europe's climate shows considerable regional and temporal variability. This is related to the continent's position in the northern hemisphere and the influence of neighbouring seas and continents, including the Arctic. Atmospheric circulation is an important driver of the temporal and regional variances (Box 2.1)

Box 2.1: Atmospheric circulation patterns in Europe and winter NAO-index 1864-2007

Box 2.1 Atmospheric circulation patterns in Europe

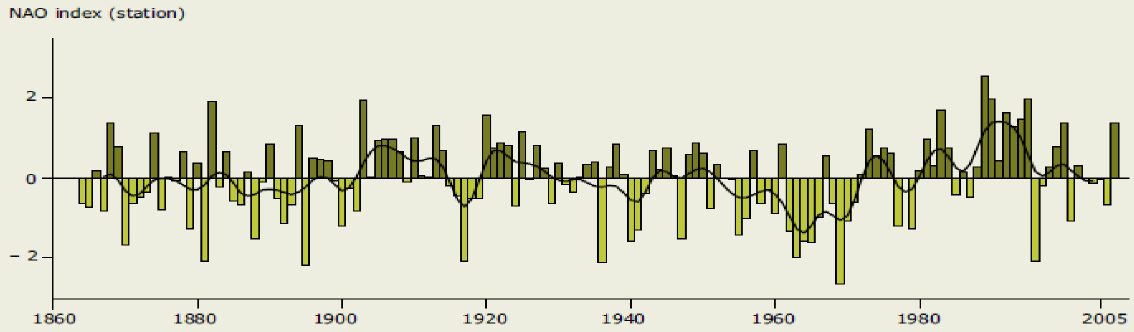
The atmospheric circulation moves air masses with their own specific characteristics, like temperature and humidity, over long distances. Important for the European climate is the prevailing western circulation at mid latitudes that directs the oceanic air masses inland over the continent. Stronger western advection brings milder and wetter weather and stronger winds to most of Europe, especially in winter. Weaker and blocked western circulation causes generally colder and drier winters and hotter and drier summers. Fluctuations in the behaviour of this circulation pattern are one of the main sources of variability in the European climate. The intensity of the western circulation in the European region is expressed by the North Atlantic Oscillation (NAO) index. NAO is the large-scale fluctuation in atmospheric pressure in the Atlantic ocean between the high-pressure system near the Azores and the low pressure system near Iceland (Figure 5.1).

The NAO is characterised by seasonal, inter-annual and inter-decadal variations. The driving mechanism of the short-term dynamics is connected with weather fluctuations. Longer time-scale variations are linked to atmosphere-ocean-ice interactions.

The seasonal anomalies have direct impacts on humans, often being associated with floods, heat-and cold-waves. The NAO appears to have been considerably more variable from year to year in the late 18th and early 19th centuries than in the 20th century. More recently, there was a large increase in the NAO index between 1970 and 1990, followed by a decrease back to about normal in 2005. The relationship with anthropogenic climate change is as yet unclear. Scenarios for future circulation patterns are very uncertain, because of the complexity of the processes and the limited ability to represent this in climate models.

The El Niño-Southern Oscillation (ENSO) in the Pacific Ocean has global impacts on decadal and longer-term variability and can cause precipitation and temperature changes over very large distances, including as far as Europe. Generally, for Europe, the effects of ENSO on precipitation and temperature are much weaker than those caused by variations in the NAO.

Figure 5.1 Mean winter (December–March) NAO index 1864–2007



Note: Positive indicates stronger western flow.

Source: Updated from Hurrell *et al.*, 2003.

Source: EEA; 2008

(b) The IPCC SRES-Scenarios

Most comprehensive assessments of climate change impacts globally and in Europe are based on the IPCC Special Report on Emissions Scenarios (SRES) (see Box 2.2). Four storylines have been developed to describe alternative options for the social, demographic, and economic development of the world throughout the 21st century. These storylines have been combined with assumptions on technological development to develop six 'marker scenarios' for future population, economic development, and greenhouse gas emissions in four world regions. None of these scenarios explicitly includes climate policies. The SRES emissions scenarios have been used as input to simulations of future climate change by about 20 general circulation models (GCMs, also known as global climate models). The projected changes in temperature and precipitation differ across GCMS as well as across SRES emissions scenarios. High emission scenarios generally lead to larger climatic changes than low emissions scenarios. From the SRES marker scenarios, A2 and A1T have the highest emissions, B1 and A1T have the lowest emissions, and A1B and B2 have intermediate emissions.

Box 2.2: The IPCC Special Report on Emission Scenarios (SRES)

Box 2.2 **e IPCC Special Report on Emissions Scenarios (SRES)**

A1. The A1 scenario family describes a future world of very rapid economic growth, global population that peaks in the mid-century and declines thereafter, and a rapid introduction of new and more efficient technologies. Major underlying themes are convergence among regions, capacity building and increased cultural and social interactions, with a substantial reduction in regional differences in per capita income. The A1 family develops into three groups that describe alternative directions of technological change in the energy system, distinguished by their technological emphasis: fossil-intensive (A1FI), non-fossil energy sources (A1T), or a balance across all sources (**A1B**) (where balanced is defined as not relying too heavily on one particular source, on the assumption that similar improvement rates apply to all energy-supply and end-use technologies).

A2. The A2 family describes a very heterogeneous world. The underlying theme is self-reliance and preservation of local identities. Fertility patterns across regions converge very slowly, which results in continuously increasing population. Economic development is primarily regionally oriented and per capita economic growth and technological change more fragmented and slower than in other scenarios.

B1. The B1 family describes a convergent world with the same global population, which peaks in the mid-century and declines thereafter, as in A1, but with rapid change in economic structures toward a service and information economy, with reductions in material intensity and the introduction of clean and resource-efficient technologies. The emphasis is on global solutions to economic, social and environmental sustainability, including improved equity, but without additional climate initiatives.

B2. The B2 family describes a world in which the emphasis is on local solutions to economic, social and environmental sustainability. It is a world with continuously increasing global population, at a rate lower than A2, intermediate levels of economic development, and less rapid and more diverse technological change than in B1 and A1. While these scenarios are also oriented towards environmental protection and social equity, they focus on local and regional levels.

Source: IPCC, 2001.

Source: EEA; 2008

(c) The "2 degree target" of the European Union

The EU has adopted in 1996, and confirmed at several occasions thereafter, a long-term climate goal of limiting global mean temperature increase to 2 °C above pre-industrial levels (or about 1.5 °C above 1990 levels) (Box 2.3). This target is also mentioned, although in somewhat vaguer terms, in the Copenhagen Accord, the main outcome of the 15th conference of the parties (COP-15) to the UNFCCC that was held in Copenhagen in December 2009.
In Cancun (Mexico) in December 2010 the EU 2°C-target became part of the formal COP-16 decisions.The "2 degree target" aims to limit the risks of climate change to an acceptable level but it will not avoid all impacts of climate change.

Box 2.3: EU-Target of limiting global temperature rise to 2°C above pre-industrial level

Box 2.3 **EU target of limiting global temperature rise to 2°C above pre-industrial level**

The EU first proposed a temperature limit of not more than 2 °C above pre-industrial levels in 1996, which was reaffirmed subsequently by a number of Environment Councils and European Councils (EU, 1996; EC, 2008; EU, 2010a). It was originally deduced from the evidence available at the time, including the temperature variation during the Holocene during which human civilization has developed, and from considerations of the adaptation rates of ecosystems. Significantly improved understanding of the vulnerability of societies and ecosystems to climate change has strengthened the scientific basis of this objective. However, the establishment of the political goal of limiting global warming to 2 °C also took into account technical feasibility and the cost of measures necessary to achieve the objective.
The EU has further stated in many Environment Council conclusions, for example in March 2010, that to stay below 2 °C requires GHG emissions to peak by 2020 at the latest and then be reduced by at least 50 % by 2050 compared with 1990 levels and continue to decline thereafter. In addition, the EU has stated that developed countries as a group, and the EU, should reduce their GHG emissions by 80 % to 95 % by 2050 below 1990 levels (see also the SOER 2010 mitigating climate change assessment (EEA, 2010; EU 2010b)).

Main sources: EEA 2010; EC 2008; EU 1996, 2010a, 2010b

Source: EEA; 2010

(d) Risks of climate tipping elements

A special kind of risks, particularly important but difficult to deal with from a policy point of view, are climate events that have a low or unknown likelihood of occurrence but potentially very large consequences for the world, including Europe. Many of these climate events involve positive feedbacks such that the process can no longer be stopped once a threshold is crossed. However, limited understanding of the physical mechanisms involved, as well as a lack of observational data, implies large uncertainty about the likelihoodof these events and about possible thresholds. These climate events are generally denoted in the scientific community as 'large scale discontinuities' (IPCC,2007), 'tipping elements and tipping points' (Lenton et al., 2008; Schellnhuber, 2009; Allison et al., 2009; UNEP, 2009), or 'climate eventualities'(Kattenberg et al.,2009; PBL,2009) (see Box 2.4 for examples).
Recent research suggests that several key components of the climate system could undergo irreversible change at significantly lower levels of global temperature increase than previously assessed (e.g. Levermann et al., 2010). The most important tipping elements for Europe all involve the cryosphere: the Greenland ice sheet and Arctic sea ice. Other potential tipping elements include the West Antarctic ice sheet, a rapid carbon dioxide and methane release from melting permafrost soils, a rapid release of sea bed methane, a collapse of the meridional overturning circulation in the North Atlantic, massive changes in ENSO and monsoon systems, and large-scaled dieback of the Amazon and the boreal forest.

Box 2.4: What are the risks of non-linear climate change?

Box 2.4 What are the risks of non-linear climate change?

The risk of large-scale discontinuities or non-inearities has been identified by IPCC as one of five 'reasons for concern' and deserves special attention, because of their potentially very large consequences for the world, including Europe. What is a non-linear, or abrupt change? If a system has more than one equilibrium state, transitions to structurally different states are possible. If and when a 'tipping point' is crossed, the development of the system is no longer determined by the time-scale of the forcing, but rather by its internal dynamics, which can be much faster than the forcing (IPCC, 2007a). A variety of different tipping points has been identified. Below we discuss a few with potentially large consequences for Europe.

One of the large-scale discontinuities relevant for Europe is the possible deglaciation of the West Antarctic Icesheet (WAIS) and Greenland. There is a medium confidence that 1–2 °C of sustained global warming above present temperatures (or 2–3 °C above pre-industrial) is a threshold beyond which there will be a commitment to a large sea-level contribution due to at least partial deglaciation of both ice sheets (IPCC, 2007a, 2007b). If so, the sea level may rise over the next 1 000 years or more on average by 7 m from Greenland and about 5 m from the WAIS (IPCC, 2007a). This would alter the world's coast lines completely. Note that the sea-level rise will not be evenly distributed over the globe, because of ocean circulation patterns, land movements, and density and gravitational factors.

There is less confidence about other non-linear effects, e.g. what may happen with the ocean circulation. A slow-down of the thermohaline circulation (THC), or equivalently, the meridional overturning circulation (MOC), may counteract global warming trends in Europe, but may have unexpected serious consequences for the behaviour of the world's climate system and exacerbated impacts elsewhere. Other examples of possible non-linear effects are the progressive emission of methane from permafrost melting and destabilisation of hydrates, and rapid climate-driven transitions from one ecosystem type to another (IPCC, 2007b). The understanding of these processes is as yet limited and the chance of major implications in the current century is generally considered to be low.

Source: EEA; 2008

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3. Important Ice and Snow Regions in Europe

Note: The preparation of this section relies on a range of regional and national data sources. Due to varying lengths of the available data records, the figures in this section cover somewhat different periods, thereby complicating comparability across regions.

3.1. The European Alps

(a) Key-facts and geographic structure

The **Alps** are the highest and one of the great mountain range systems of Europe, stretching from Austria and Slovenia in the east; through Italy, Switzerland, Liechtenstein and Germany; to France in the west. The Alps are generally divided into the Western Alps and the Eastern Alps. The division is along the line between Lake Constance and Lake Como, following the rivers Rhine, Liro and Mera. The Western Alps are higher, but their central chain is shorter and curved; they are located in Italy, France and Switzerland. The Eastern Alps (main ridge system elongated and broad) belong to Italy, Austria, Switzerland, Germany, Liechtenstein and Slovenia. The Alps-region is covering an area of about 190.000 km² and has about
13. 6 Mio. inhabitants.

(b) Observed climate trends

Climate change in the Alps over the past 250 years has been extensively studied by the project HISTALP (Auer *et al.*, 2007). The HISTALP database contains monthly homogenised records of temperature, pressure, precipitation, sunshine and cloudiness for time series dating back to 1760 for temperature and to 1800 for precipitation for the Greater Alpine Region (GAR). The average state of temperature and precipitation in the Alps for the period 1961-1990 is presented in Figure 3.1. The Alps have undergone an exceptionally high temperature increase of around + 2 °C between the late 19th and early 21st century, more than twice the rate of average warming of the Northern hemisphere (Fig.:3.2). Furthermore, a slight trend towards an increase in precipitation in the north alpine region and a decrease in the south has been recorded (Fig.:3.3), (EEA; 2009).

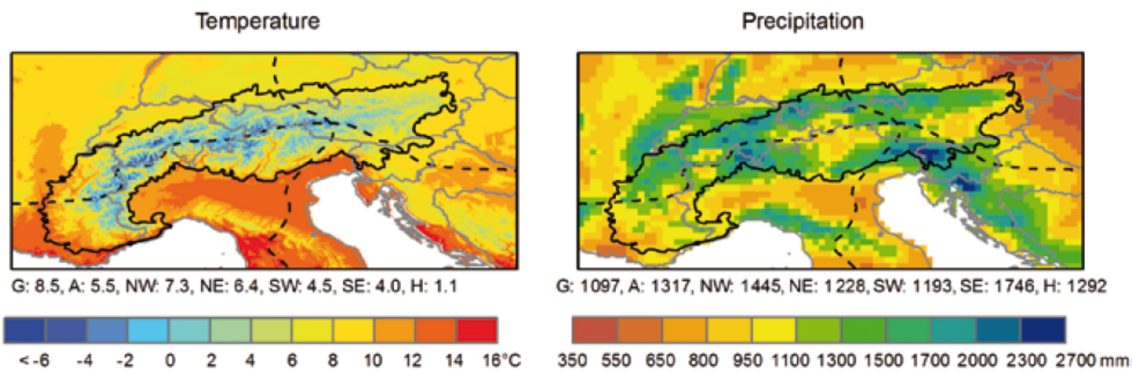


Fig. 3.1: Temperature and precipitation in the Alps for the period 1961-1990

Note: Regional statistics: G = Greater Alpine Region, A = Alps, NW = north-western Alps, NE = north-eastern Alps, SW = south-western Alps, SE = south-eastern Alps, H = higher than 1 500 m.

Source: Data for temperature, Auer *et al.* (2008); data for precipitation, Efthymiadis *et al.* (2006).

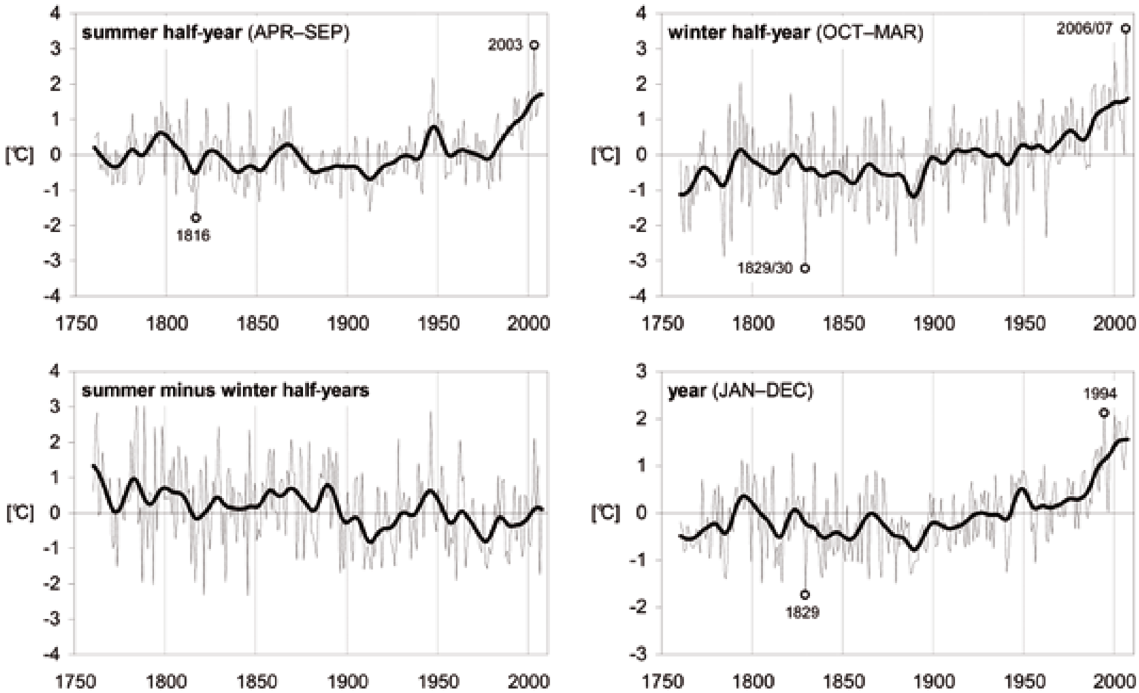


Fig. 3.2: Change in temperature 1760-2007 for the Greater Alpine Region (GAR).
(Single years and 20 years smoothed)

Note: Single years (thin lines) and 20-year smoothed mean (bold lines). All relative to 1851–2000 average, summer and winter half-years (first row), annual mean and annual range (second row).

Source: ZAMG-HISTALP database (version 2008, including the recent EI correction (EI = early instrumental period) described in Böhm *et al.*, 2008).

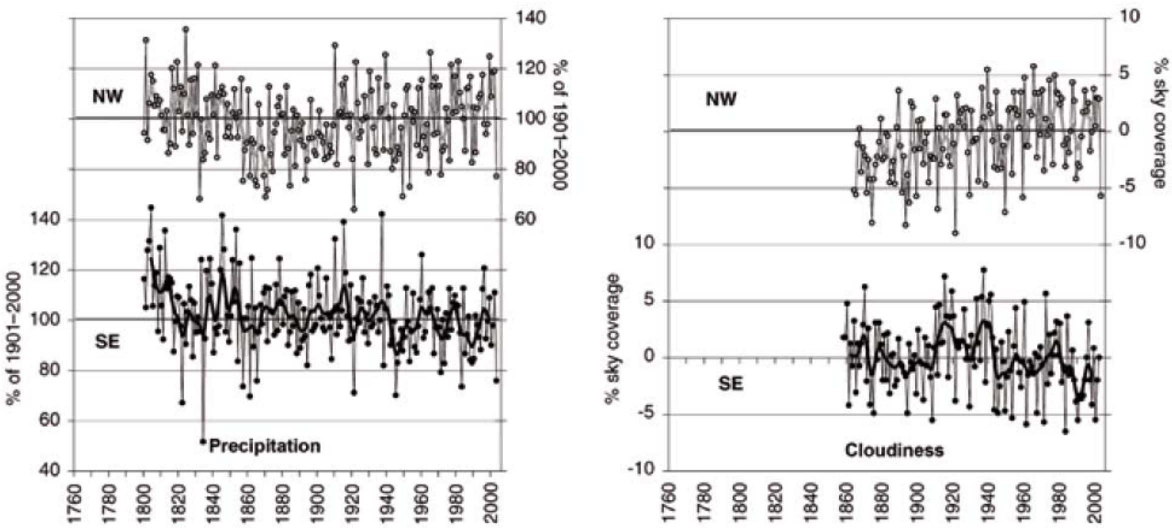


Fig. 3.3: Annual precipitation series (left graph) and annual cloudiness series (right graph)

Note: NW (top, grey) vs SE (bottom, black). All values relative to the 1901–2000 averages. Single years (thin lines) and 10-years smoothed (bold lines).

Source: ZAMG-HISTALP database (Auer *et al.*, 2007).

3.2. Scandinavia (Scandinavian Mountains)

(a) Key facts and geographic structure

The **Scandinavian Mountains** are a mountain range that runs through the Scandinavian Peninsula. The western sides of the mountains drop precipitously into the North Sea and Norwegian Sea, forming the famous fjords of Norway, while to the northeast they gradually curve towards Finland. To the north they form the border between Norway and Sweden, still reaching 2,000 m high (6,600 ft) at the Arctic Circle. The mountains are not very tall, but are at places very steep (highest peak: 2,469 meters). The combination of a northerly location and moisture from the North Atlantic Ocean has caused the formation of many ice fields and glaciers. The Scandinavian mountain system is geologically connected with the mountains of Scotland, Ireland and, crossing the Atlantic Ocean, the Appalachian Mountains of North America. The mountains are one of the oldest still extant mountain ranges in the world.
The Scandinavian Mountains are covering an area of about 325.000 km² (mostly in Norway) and have slightly more than 4.2 Mio. inhabitants.

(b) Observed climate trends

Long-term data records on temperature and precipitation has been published in 2009 by the Norwegian Met-Office (met.no) commonly with other Norwegian services (met-no; 2009).
According these records the mean annual temperature for the Norwegian mainland has increased by ca. 0.8 °C the latest 100 years, and with largest increase during spring.
There have periods with both increasing and decreasing temperatures, but since 1965 the annual temperature has increased by ca. 0.4 °C per decade (Fig.3.4a; b). The growing season has become longer over the whole country, and the heating degree-day sum has decreased. For the Norwegian mainland the annual precipitation has increased by almost 20% since the year 1900, with largest increase during winter and least during summer. The largest increase is found in western-Norway (Fig.3.5a; b). For frequencies of strong winds ($\geq$ 9 Beaufort) in the Norwegian ocean and coastal areas, there is no clear trend since 1880. The snow season has become shorter.

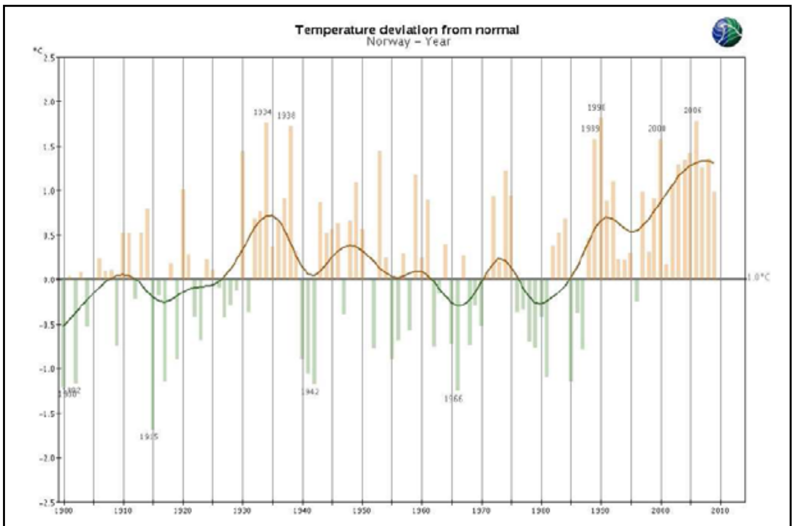


Fig.3.4a: Development of mean annual temperature for the Norwegian mainland (1900-2008)
Source : met.no; 2009
Note: bars show annual differences to the normal period 1961-1990; the curve is smoothed on a decadal scale

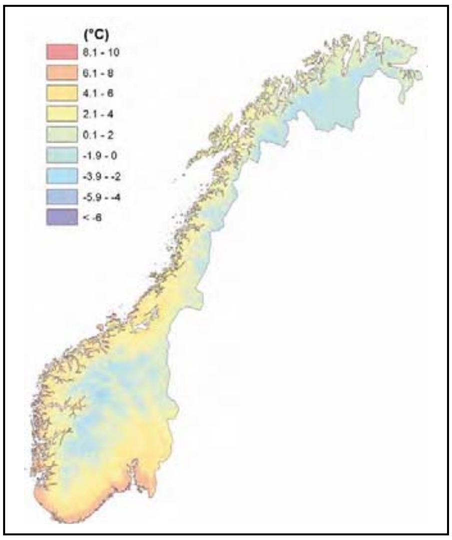
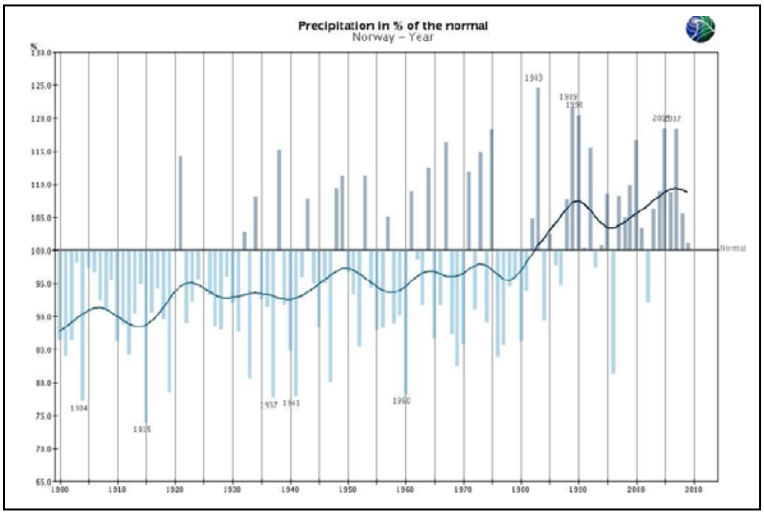


Fig.3.4b: Average annual air-temperature in Norway 1979-2008
Source : met.no; 2009



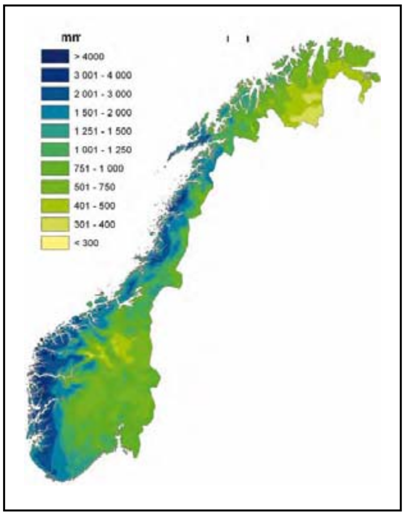


Fig.3.5a: Development of annual precipitation for Norwegian mainland (1900-2008)
Source : met.no; 2009
Note: Bars show annual differences from the normal period 1961-1990; the curve is smoothed on a decadal scale

Fig.3.5b: Average annual precipitation in Norway 1979-2008
Source : met.no; 2009

The regional climate-development in Norway-mainland (and Svalbard) during the previous decades as measured in the periods1978-2008 and 1961-90 is illustrated by changes in annual and seasonal mean temperature (upper map and table) and in the ration of annual and seasonal precipitation for both of the periods (lower map and table).
Source: met.no; 2009

Region	1	2	3	4	5	6	Norge	Svalbard*
År	0,63	0,47	0,55	0,51	0,57	0,53	0,57	1,35
Vinter	1,34	0,91	1,06	0,96	1,13	0,93	0,98	1,91
Vår	0,63	0,41	0,39	0,42	0,68	0,65	0,53	1,86
Sommer	0,34	0,37	0,44	0,39	0,30	0,26	0,37	0,66
Høst	0,35	0,30	0,39	0,29	0,22	0,32	0,33	0,97

Region	1	2	3	4	5	6	7	8	9	10	11	12	13	N	S
År	1,04	1,04	1,05	1,08	1,07	1,06	1,05	1,06	1,04	1,04	1,05	1,07	1,04	1,05	1,00
Vinter	1,19	1,08	1,11	1,25	1,25	1,23	1,09	1,16	1,12	1,15	1,15	1,18	1,05	1,17	0,96
Vår	1,05	1,09	1,08	1,09	1,09	1,14	1,09	1,08	1,02	1,08	1,09	1,15	1,07	1,10	0,90
Sommer	1,02	1,05	1,06	1,01	1,00	0,99	1,06	1,07	1,11	1,04	0,98	1,01	1,05	1,02	0,97
Høst	0,97	0,98	0,98	0,99	0,95	0,93	0,98	0,96	0,95	0,95	0,99	1,03	1,02	0,97	1,12

3.3. Svalbard

(a) Key-facts and geographic structure

Svalbard is an archipelago between the Arctic Ocean, Barents Sea, Greenland Sea and Norwegian Sea, constituting the northernmost part of Norway. The land area is 61,022 km^2 , and dominated by the island Spitsbergen, which constitutes more than half the archipelago, followed by Nordaustlandet and Edgeøya. All settlements are located on Spitsbergen. Glaciation covers 36,502 km^2 (14,094 sq mi) or 60% of Svalbard; 30% is barren rock while 10% is vegetated. The landforms of Svalbard were created through repeated ice ages, where glaciers cut the former plateau into fjords, valleys and mountains. The tallest peak is Newtontoppen (1,713 m/5,620 ft).

In 2009, Svalbard had a population of 2,753. Longyearbyen is the largest settlement on the archipelago, the seat of the governor and the only town to be incorporated.

(b) Observed climate trends

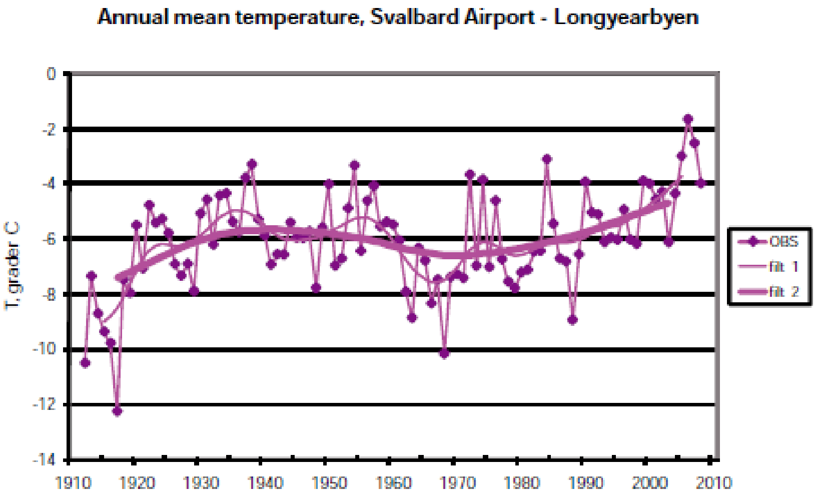
The climate for Svalbard-stations will be “maritime” (relatively mild and humid) in years (or periods) when the sea around the stations is ice-free. When the stations are surrounded by sea-ice, the climate will be “continental” (cold and dry) because the sea-ice isolates from the latent and sensible heat sources of the sea, and further reflects much of the solar radiation. Thus the high-Arctic temperatures show great inter-annual fluctuations, considering the high latitude.

In the Longyearbyen area the annual mean temperature has increased significantly from 1912 to present. The linear seasonal temperature trends at Svalbard Airport/Longyearbyen

(Fig. 3.6) from 1912 to 2007 are +0.22°C per decade (annual), +0.21°C per decade (winter), +0.45 (spring), +0.10 (summer) and +0.16 (autumn) (Fig. 3.7). Except for the winter season all seasonal trends are statistically significant at least at the 5%-level (NPI; 2009).

All Norwegian high-Arctic series show a positive trend in annual precipitation throughout the period of observations (Fig. 3.8). At Svalbard Airport the annual precipitation has in average increased by 2% per decade, while the increase on Bjornoya is 3% per decade (NPI; 2009).

Fig. 3.6: Annual mean temperature; Svalbard Airport- Longyarbyen



Source: NPI; 2009

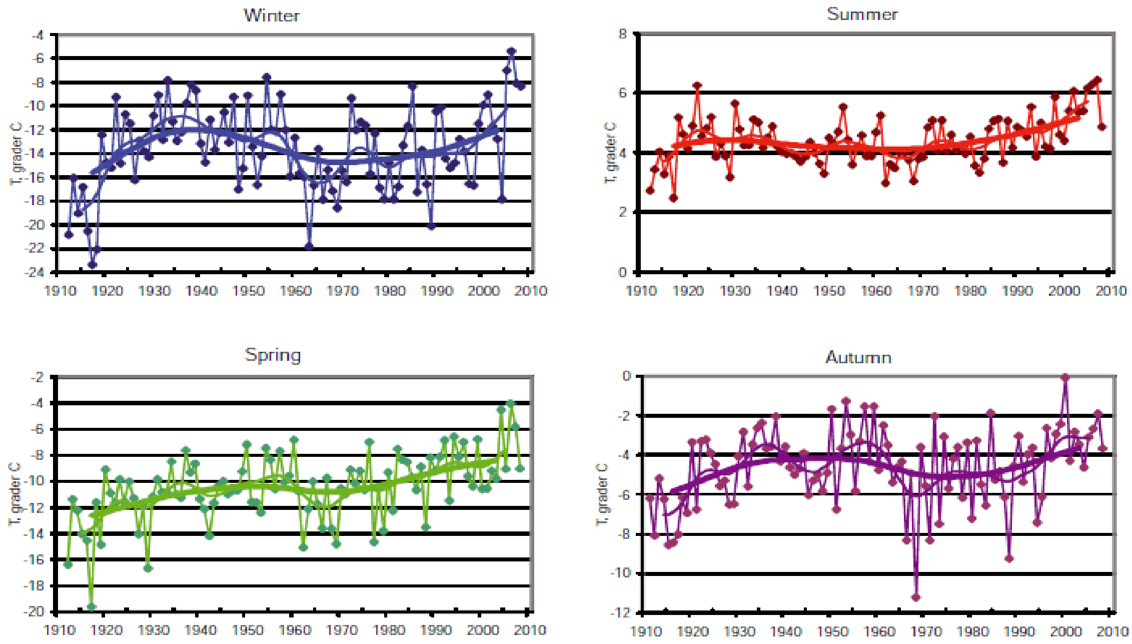


Fig.: 3.7. : Annual and seasonal temperatures at Svalbard Airport/ Longyearbyen 1911-2007.
(The smoothed curves (Filt. 1 and 2) show variations on a decadal resp. 30-years scale).
Source **:** NPI; 2009

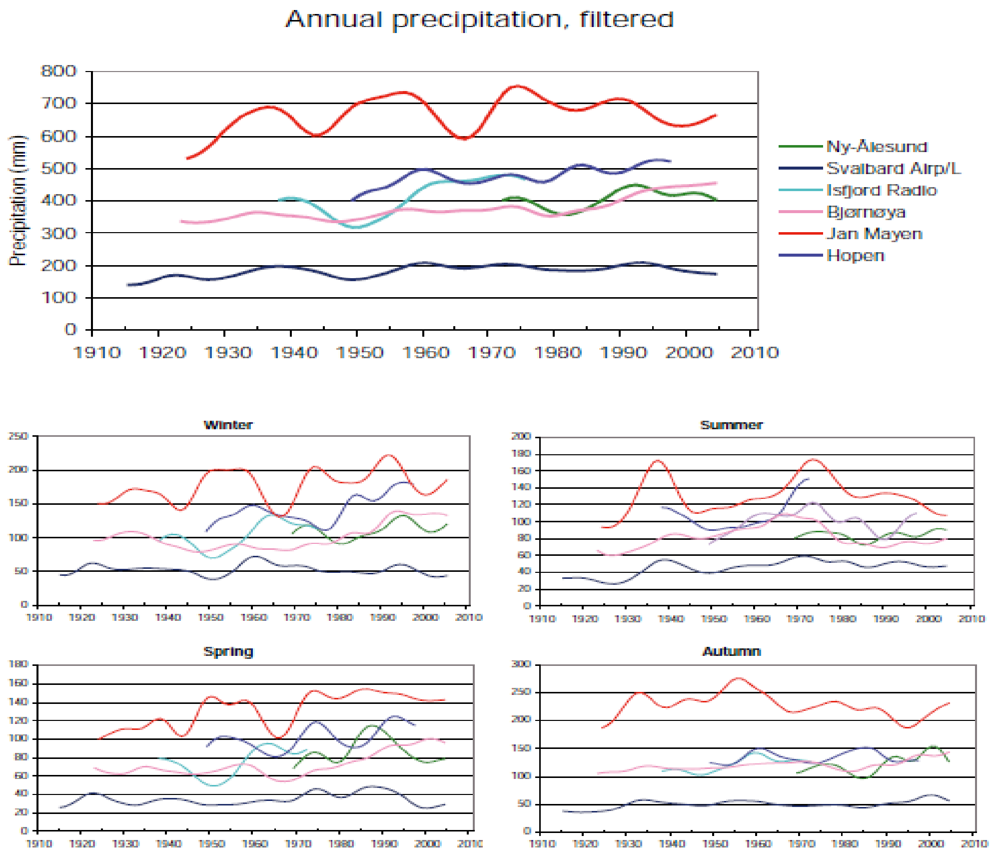


Fig. 3.8: Annual and seasonal precipitation at Norwegian high-Arctic stations 1911-2007
(The smoothed curves show variability on a decadal scale).
Source **:** NPI; 2009

3.4 Iceland

(a) Key-facts and geographic structure

Iceland is a European island country located in the North Atlantic Ocean on the Mid-Atlantic Ridge. It has a population of about 320,000 and a total area of 103,000 km^2. The capital and largest city is Reykjavík, with the surrounding area being home to some two-thirds of the national population. Iceland is volcanically and geologically active. The interior mainly consists of a plateau characterized by sand fields, mountains and glaciers, while many glacial rivers flow to the sea through the lowlands. Iceland is warmed by the Gulf Stream and has a temperate climate despite high latitude just outside the Arctic Circle.

(b) Observed climate trends

Iceland, located at 63-67°N and 18-23°W, has considerably milder climate than its location just south of the Arctic Circle would imply. A branch of the Gulf Stream, the Irminger Current, flows along the southern and the western coast greatly moderating the climate (Figure 3.9). The cold East Greenland Current flows west of Iceland, but a branch of that current, the East Icelandic Current, approaches Iceland's northeast- and east coasts. (http://www3.hi.is/~oi/climate_in_iceland.htm).
However, this brings mild Atlantic air in contact with colder Arctic air resulting in a climate that is marked by frequent changes in weather and storminess. Furthermore this leads to more rainfall in the southern and western part than in the northern part of the island.

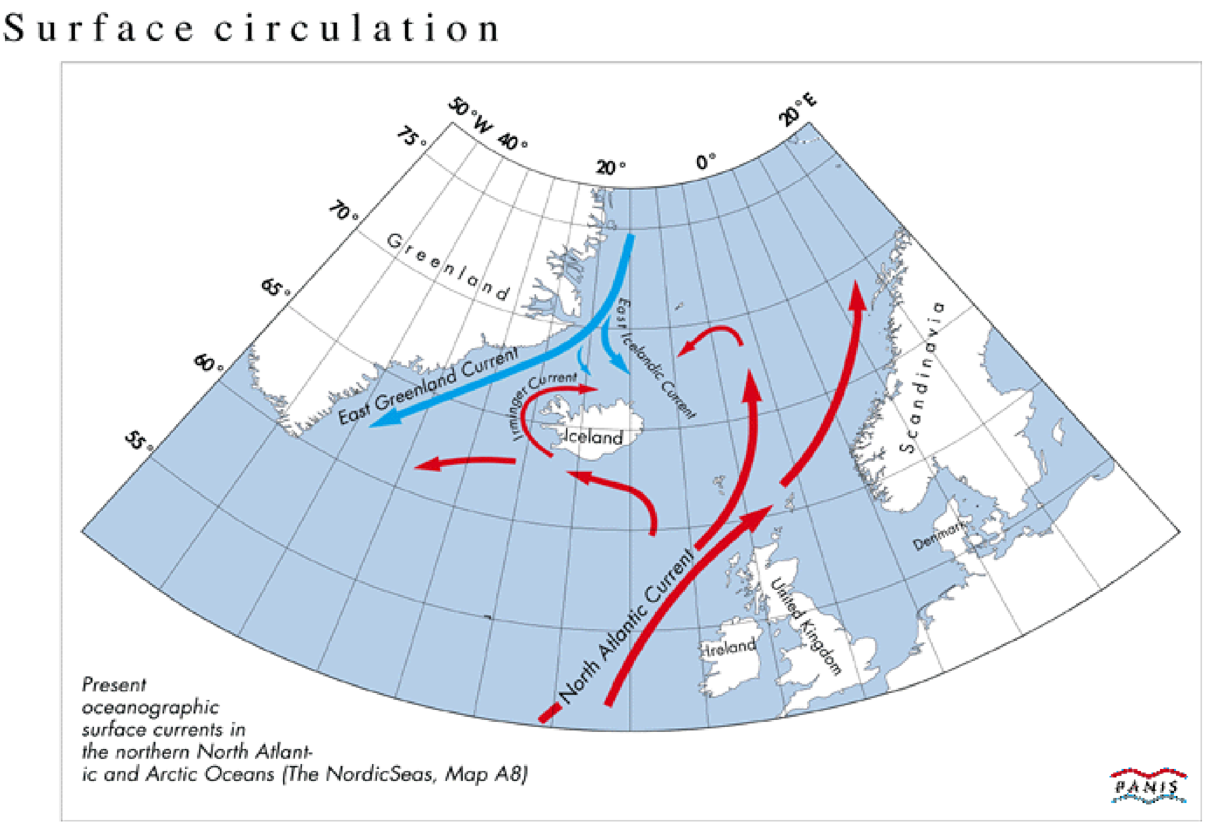


Figure 3.9: Present oceanographic surface currents around Iceland.

Image source: http://www.hi.is/~jeir/panis_currents.html

A simple classification of Icelandic climate puts it as cool temperate maritime, reflecting that it is very influenced by the cool ocean waters around Iceland. A map of the annual mean temperature (Figure 3.10) shows that only along the coasts of southern and southwestern Iceland do temperatures reach 4-6°C, but are lower in other parts of the island.

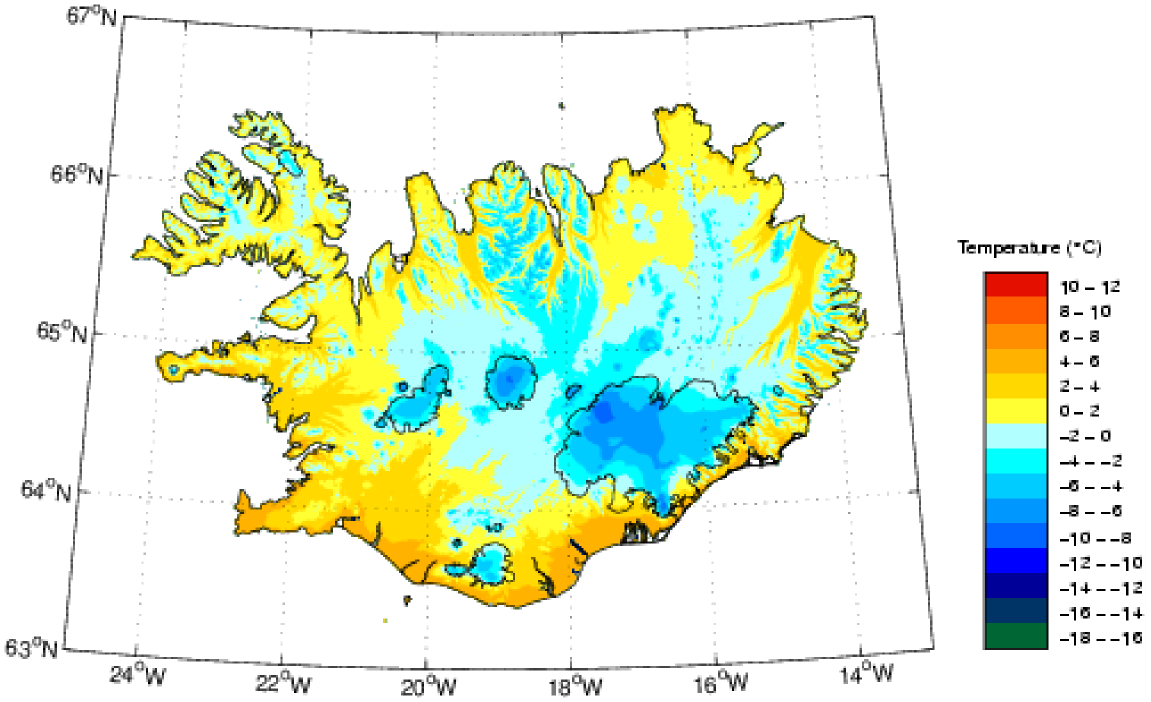


Figure 3.10: Annual mean temperatures in Iceland
Source : http://www.vedur.is/vedurfar/vedurfarsmyndir/EV_DTO/ann.html

The temperature (Fig. 3.11) has in the long run been increasing during the last 200 years at the rate of +0.7°C per century. This is similar to the general temperature increase in the whole Northern hemisphere during the same period. The warming has been very uneven, dominated by three cold periods and two warm ones (Fig. 3.12).

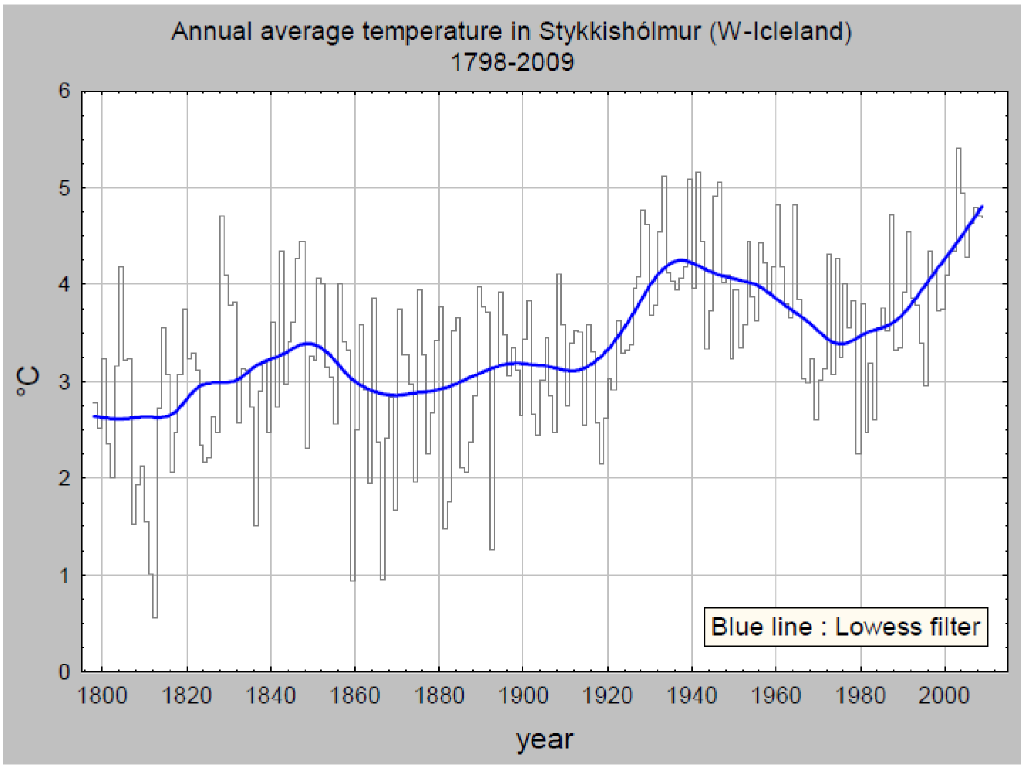


Fig. 3.11.: Annual temperature in Stykkishólmur 1798 to 2009.
Source **:** Trausti Jonsson (Icelandic Met-Office; 2010)

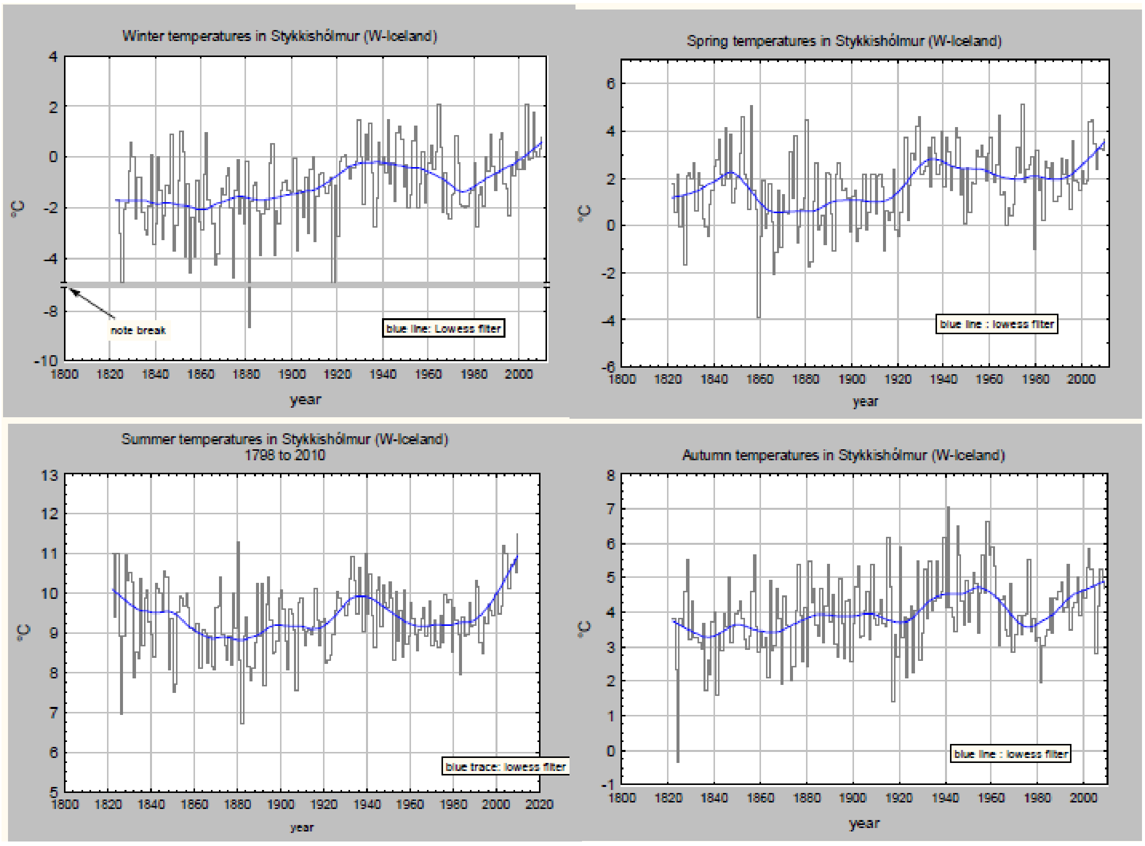


Fig. 3.12.: Seasonal temperature in Stykkishólmur 1798 to 2009.
Source : Trausti Jonsson (Icelandic Met-Office; 2010)

The annual temperature in Iceland is dominated by the large variability of the winter season. The graph showing the winter temperatures is very similar to the graph of annual temperatures. The trend is slightly larger than the annual temperature trend, 1.2°C per century. In the figure showing the summer (June to September) temperatures superficially similar variations as in the annual and winter figures are visible. The overall trend of the summer temperature is only 0.2°C per century. The long-term trend in the summer temperature is not significant as the summer variability is much less than the variability in winter. The temperature variations in the spring (April and May) are similar in timing to the winter. The overall warming trend is about 0.7°C per century. The autumn (October and November) shows slightly different variations. The recent warming has not reached the autumn and and the autumn temperatures remain below the warm autumns of the 1940 to 1960 period (http://en.vedur.is/climatology/clim/nr/1213; 2008).
The pattern of precipitation in Iceland reflects the passage of atmospheric low pressure cyclones across the North Atlantic Ocean from south-westerly directions, exposing the south coast to heavy precipitation (Figure 3.13), (Hanna et al., 2004). The long term record on precipitation-amounts (Fig. 3.14) shows a high variability with some periodic peaks.

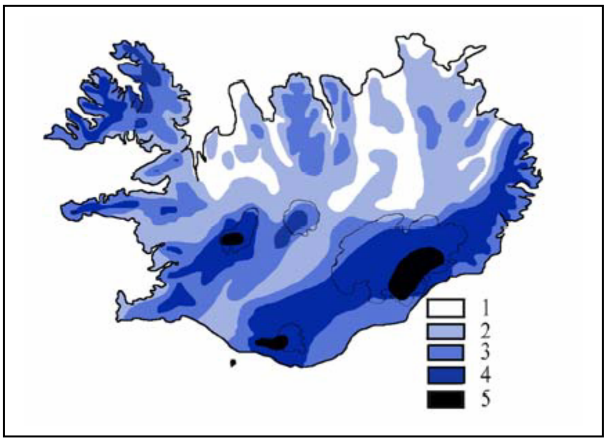


Fig. 3.13 : Mean annual precipitation in Iceland for the period 1931-1960
(1. < 600 mm; 2. 600 – 1199 mm; 3. 1200 – 1999 mm; 4. 2000 – 3999 mm; 5. > 4000 mm).
Source : http://www.nnv.is/skrar/DFHM03%20p167-178.pdf

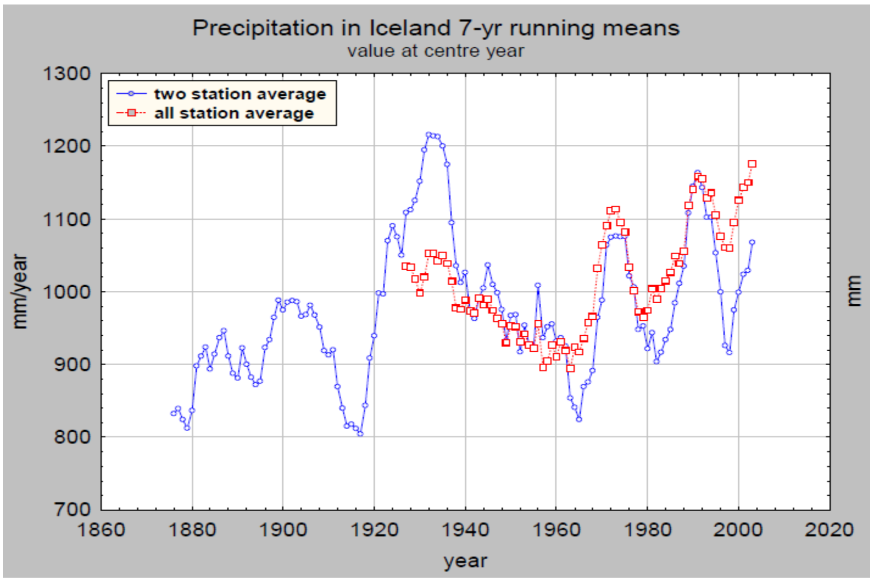


Fig. 3.14. : Annual precipitation in Iceland (7-yr running means)
Source : Trausti Jonsson (Icelandic Met-Office, 2010)

3.5 The Tatra mountains

(a) Key-facts and geographic structure

The Tatra Mountains, the highest part of the entire Carpathians, is the geographical unit with the alpine character, with an extent of 700 sq km. Major part (75 %) lies in the territory of Slovakia, the rest in Poland. The massif of the Tatra Mountains is markedly separated with fault cliffs in its southern part from Liptovská and Popradská kotlina basins. Along northern and eastern slopes the elongated erosion cut stretches away. It was eroded by water in less resistant flysch (Podtatranská brázda cut). From the geological point of view the Tatra Mountains consists of two zones, the southern one is predominantly crystalline and northern zone is built up by Mesozoic nappes. The Tatra Mountain region is composed of Western, High and Belianske mountains. The main ridge length of the Western Tatra is 37 km, High Tatra 26, 5 km and Belianske Tatra 14 km. The highest peak is Gerlachovský štít (Gerlach Peak) with altitude 2655 m a.s.l.. The Tatra´s ridge is watershed divide between Baltic and Black Seas.

(b) Observed climate trends

The monthly mean air temperature from May till October is above 0°C. The isotherm of the mean annual air temperature 0°C varies in height range 1600 – 1800 m a.s.l.(Fig.3.15a); Mean precipitation totals in surrounding basins are 600 – 700 mm, in the altitude above 2000 m a.s.l.; they reach more than 2000 mm (Fig.3.15b).

The annual averages of mean daily *air temperatures* tend to increase up to 0.35°C/10yr (Podbanské). This warming is significant at all stations, except for Ždiar-Javorina with increase of 0.1°C/10yr. However, there is a weak seasonal cooling (up to -0.06°C/10yr on average in September) or no trend in autumn months, while the most significant warming is most clearly pronounced in January (0.5°C/10yr), May (0.4°C/10yr) and July (0.35°C/10yr) (Figure 3.16, Stastny et al., 2010).

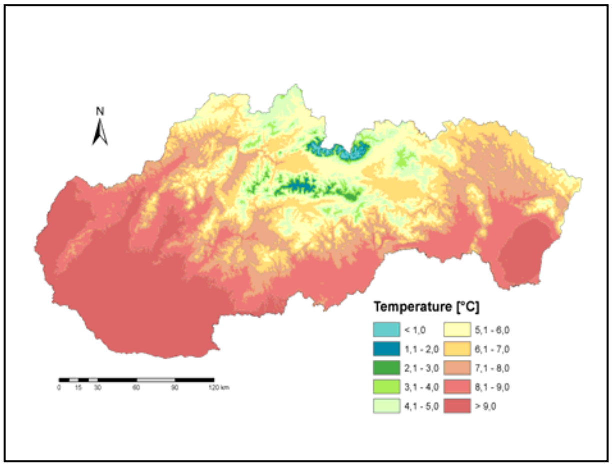


Fig.3.15a: Mean annual temperature 1961-1990
Source : SHMU, 2010

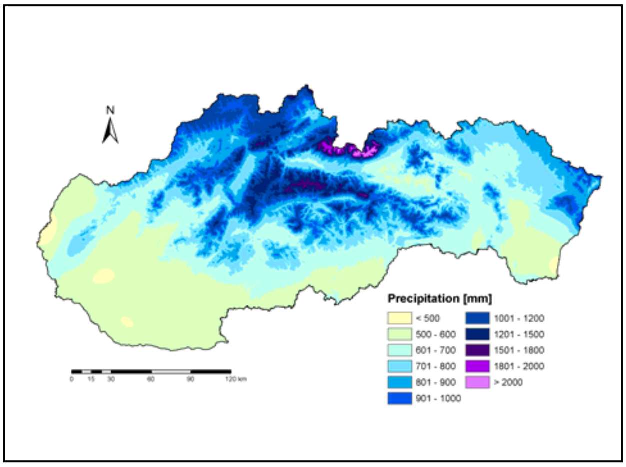


Fig.3.15b: Mean annual precipitation 1961-1990
Source : SHMU, 2010

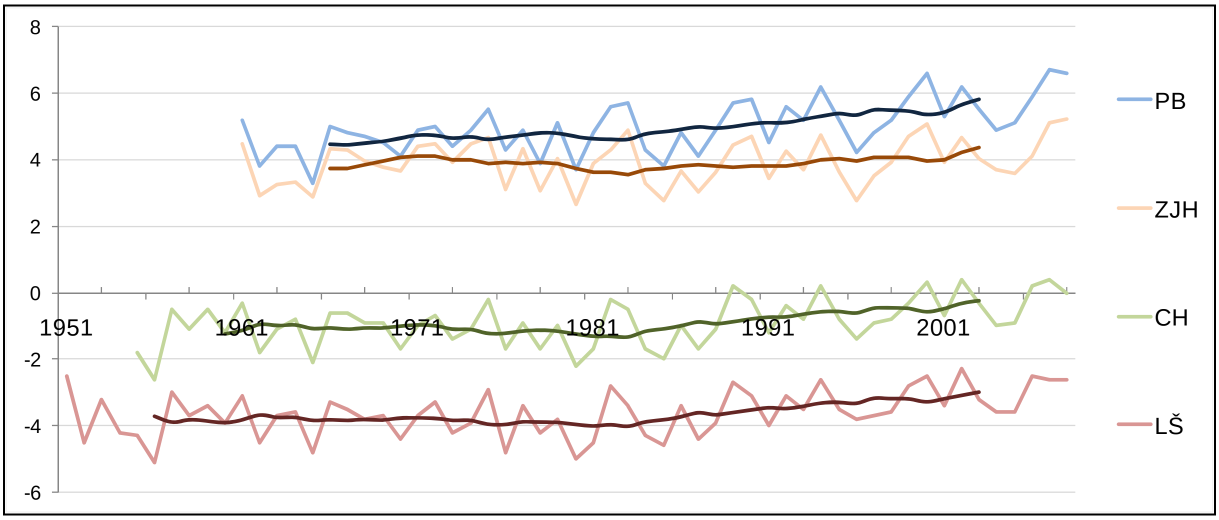


Figure 3.16: Yearly averages of mean daily temperatures (°C) and 11-year running averages from Podbanské (PB), Ždiar-Javorina (ZJH), Chopok (CH) and Lomnický štít (LŠ) stations
Source: Stastny et al. (SHMU; 2010)

The *Annual precipitation totals* at most of stations tend to increase, at more than half of stations even significantly. The greatest increase is 59mm/10yr (Ždiar-Javorina). Trends by months revealed the highest rise in July and March (4 mm/10yr on average), while the drop in June and November was on average -3 and -1 mm/10yr (Figure 3.17). The typical distribution of precipitation throughout the year tends to change most notably in June-August period.

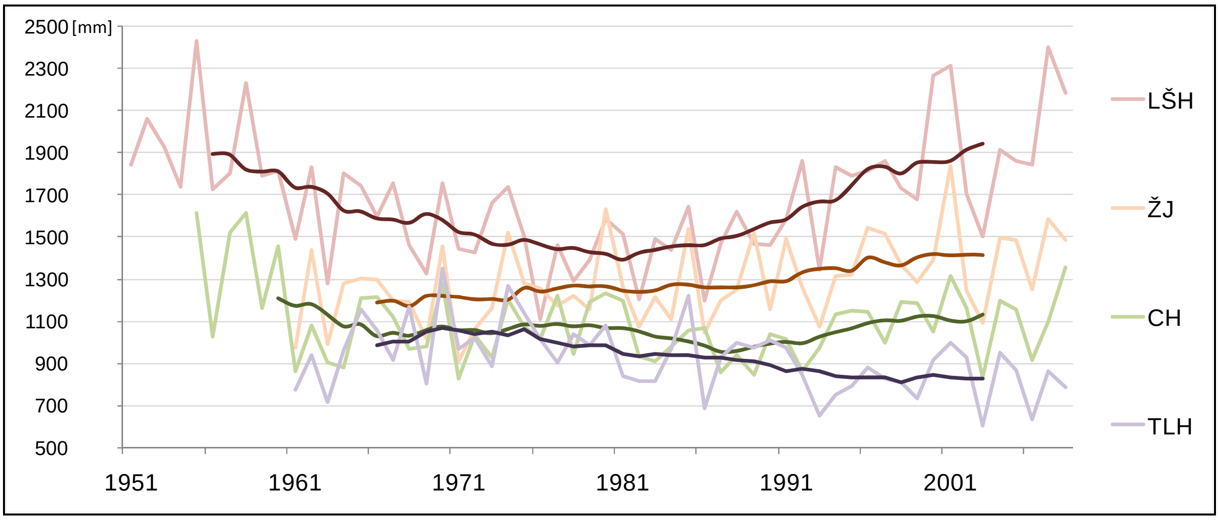


Figure 3.17: Yearly sums of daily precipitation totals and 11-year running averages from Lomnický štít (LŠH), Ždiar-Javorina (ŽJ), Chopok (CH) and Telgárt (TLH) stations
**Source : **Stastny et al. (SHMU; 2010)

3.6 The Pyrenees

(a) Key-facts and geographic structure

The Pyrenees mountain range extends over 450 km from the Mediterranean Sea to the Atlantic Ocean and forms the isthmus that links the Iberian Peninsula to the rest of the Eurasian continent. The high Pyrenees range between 2000 and more than 3000m in altitude reaching a maximum of 3404m (Pico de Aneto) and they are about 120 km wide in the middle of the chain. The Pyrenees cover an area of about 27,452 km² in France, 21,007 km² in Spain, the independent principality of Andorra (468 km²) and have about 2.2 million inhabitants.

(b) Observed climate trends

The peculiar geographical features that shape the Pyrenees play a major role in the climatic conditions affecting the whole chain. The zonal disposition of the axial range retains polar and arctic maritime air masses from north advections, and tropical maritime air masses from the south and southwest. The meridian valley configuration favors the penetration and the placement of unstable air masses, i.e. the forced lifts caused by the relief may sometimes result in heavy and persistent snowfalls. Because of the proximity of the Pyrenees to the Mediterranean Sea and the Atlantic Ocean, temperatures are less extreme than in inland ranges. Interestingly, there are extensive rain shadows close to the Mediterranean. Finally, the massif is a boundary between the humid oceanic climate and the subtropical dry climate due to its relatively low latitude.
The existence of strong climatic gradients in the region causes that evolution of precipitation and temperature and may exhibits noticeable differences in the magnitude of observed trends even at very short distances (López-moreno et al., 2010). Figure 3.18 shows the regional evolution of annual temperature in the Spanish Pyrenees for the period 1950-2005. It is observed that negative anomalies dominate from 1950 to 1980, whereas positive anomalies are recorded in practically all years of the two last decades. The warming rate during the analyzed period is 0.22ºC per decade. According to López-Moreno et al 2010, summer and winter are the seasons when temperature has exhibited a stronger increase, whilst autumn´s temperature had behaved in a rather stationary manner.

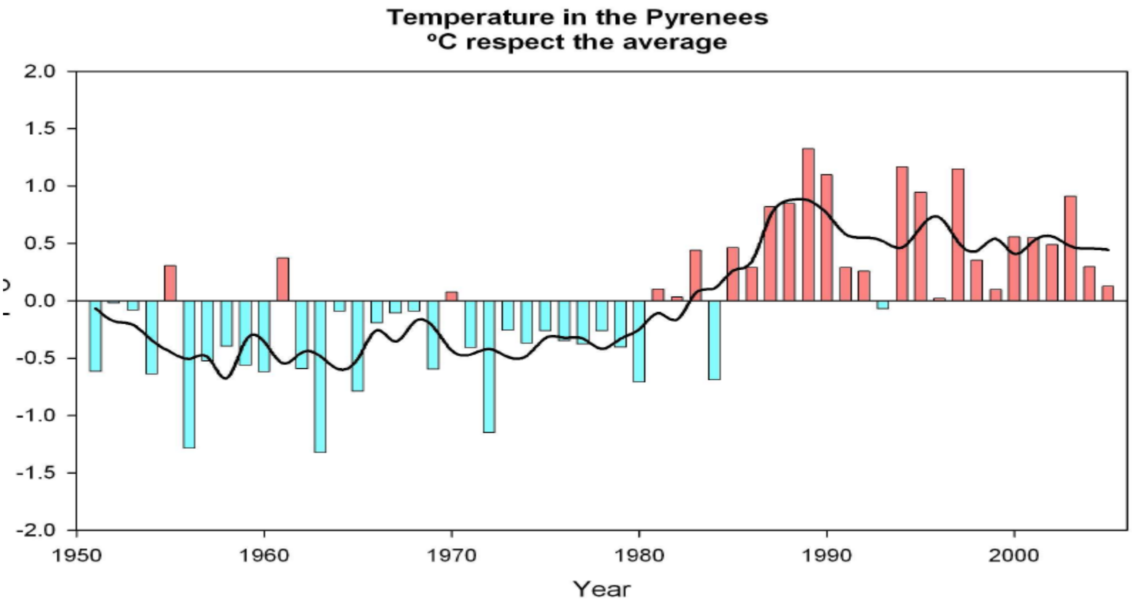


Fig. 3.18: Deviation of annual air-temperature in the Spanish Pyrenees with respect to the 1950- 2005 average

Source : J. Lopez-Moreno (IPE; 2010)

The annual precipitation in the Spanish Pyrenees (Fig.3.19) exhibits a large inter-annual variability, with a continuous alternation of positive and negative anomalies. However, in the last decades the frequency of negative anomalies exceeds clearly the occurrence of years above the long-term average. An increasing dominance of positive anomalies in the recent decades lead to an overall decrease of precipitation of -1.6% per decade. Trends in precipitation are specially subjected to spatial and seasonal variability (López-Moreno et al., 2010). Summer is the season when a major drying has undergone, followed by winter when negative coefficients have been detected in the majority of the territory. In spring evolution of the precipitation is stationary, and in autumn it has slightly increased in most of the region.

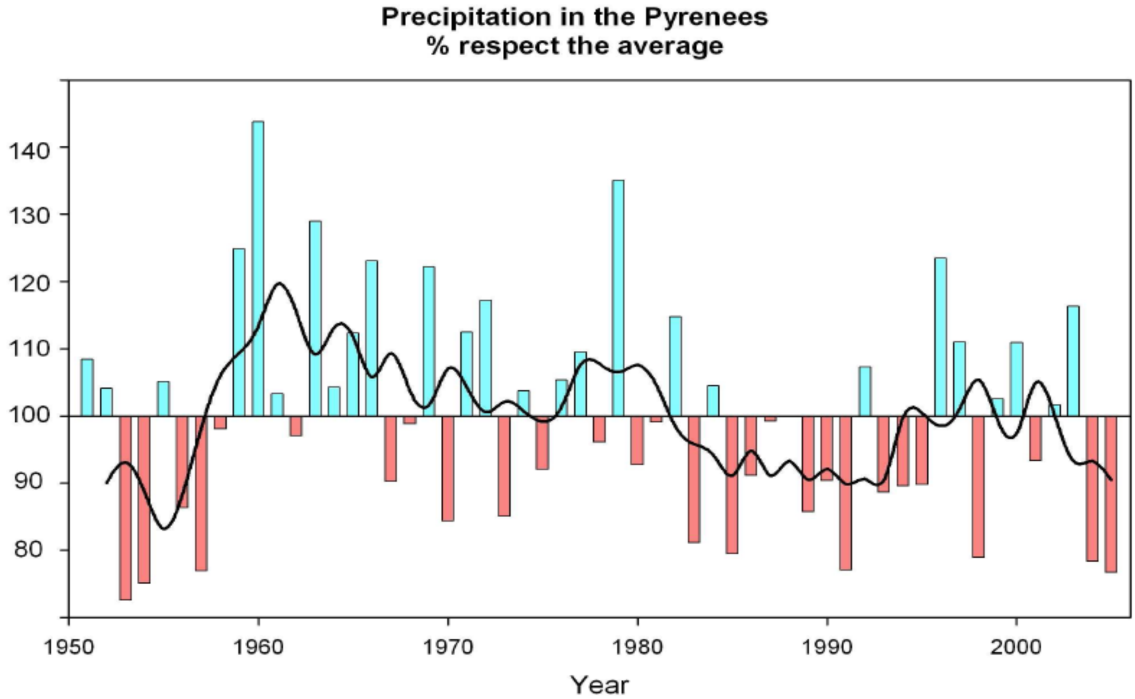


Fig. 3.19: Deviation of annual precipitation in the Spanish Pyrenees with respect to the 1950-2005 average

Source : J. Lopez-Moreno (IPE; 2010)

3.7 The Baltic Sea basin

(a) Key-facts and geographic structure

The Baltic Sea is a small, semi-enclosed brackish water basin. It is connected to the North Sea by a narrow entrance - the Danish Straits – that limits the exchange of water. The Baltic Sea is characterized by closed circulation, low salinity, and low biodiversity. Its waters are a mixture of sea water and fresh water from a catchment area that is four times larger than the sea itself (14 major river systems in the catchment). The Baltic Sea is about 1600 km long, an average of 193 km wide, and an average of 55 m deep. The maximum depth is 459 m (1506 ft), on the Swedish side of the center. The surface area is about 377,000 km² and about 85 million people live in its catchment area.

(b) Observed climate trends

The climate of the Baltic Sea is strongly influenced by the large scale atmospheric pressure systems that govern the air flow over the region: The Icelandic Low, the Azores High and the winter high/summer low over Russia. The westerly winds bring, despite the shelter provided by the Scandinavian Mountains, humid and mild air into the Baltic Sea Basin. Further the heat as brought through the Gulf Stream and the North Atlantic Drift is influencing the climate.

However, the climate in the south-western and southern parts of the basin is maritime, and in the eastern and northern parts is sub-arctic (BACC, 2008).

During the period 1871 – 2004 there were significant positive trends in the annual mean
temperature for the northern and southern Baltic Sea basin, being 0.10 °C/decade on average to the north of 60° and 0.07 °C/decade to the south of 60° N (Fig. 3.20) . The trends are larger than for the entire globe which amount to 0.05 °C/decade (1861 – 2000). In the annual mean temperatures there was an early 20th century warming that culminated in the 1930s. This was followed by a smaller cooling that finished in the 1960s, and then another strong warming until present days. Warming is characterised by a pattern where mean daily minimum temperatures have increased more than mean daily maximum temperatures. Spring is the season showing the most linear and strongest warming whereas wintertime temperature increase is intermittent but larger than in summer and autumn. A general tendency is that the start of the climatic seasons in the spring half-year (e.g. spring, growing season, summer) start earlier, whereas the climatic seasons in the autumn half-year (e.g. autumn, frost season, winter) start later (BALTEX, 2006).

Changes in precipitation are not spatially uniform. Within the Baltic Sea basin the largest increases have occurred in Sweden and eastern coast of the Baltic Sea. Seasonally largest increases have occurred in winter and spring. Changes in summer are characterised with increases in the northern and decreases in the southern parts of the Baltic Sea basin. In wintertime, there is an indication that number of heavy precipitation events has increased (Fig. 3.21).

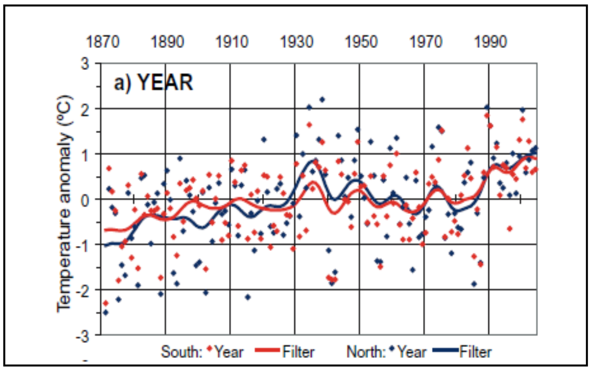


Fig. 3.20: Annual and seasonal mean surface air temperature for the Baltic Sea Basin 1871–2003,
Source: Jones and Moberg (CRU dataset; 2003)
Note:
Blue colour comprises the Baltic Sea Basin to the north of 60◦ N, and red colour to the south of that latitude. The dots represent individual years, and the smoothed curves highlight variability on timescales longer than 10 years (Gaussian filter, σ = 3)

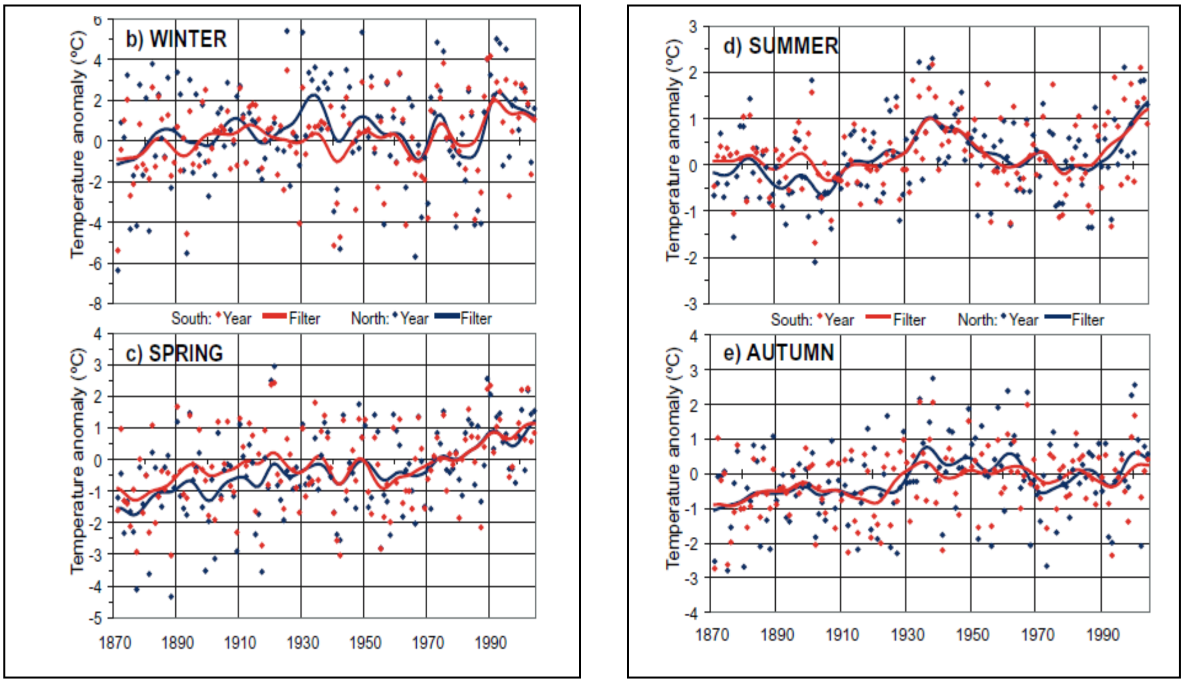
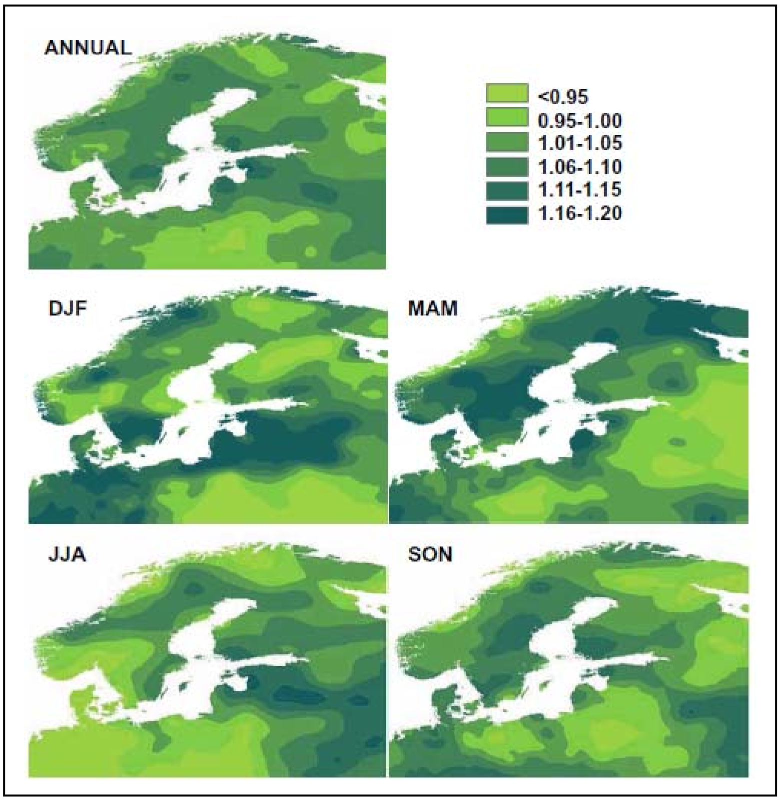


Fig. 3.21: Annual and seasonal precipitation ratios between the periods 1976−2000 and 1951−1975 based on VASClimO (Variability Analysis of Surface Climate Observations) data. **Source**: Beck et al. 2005.



3.8 Climate projections

Assessments of the potential impacts of climate change on Europe’s cryosphere require information on the climate change projections for the relevant regions. With the exception of Iceland and Svalbard, all regions considered in this paper are covered by high resolution climate projections from the ENSEMBLES project.

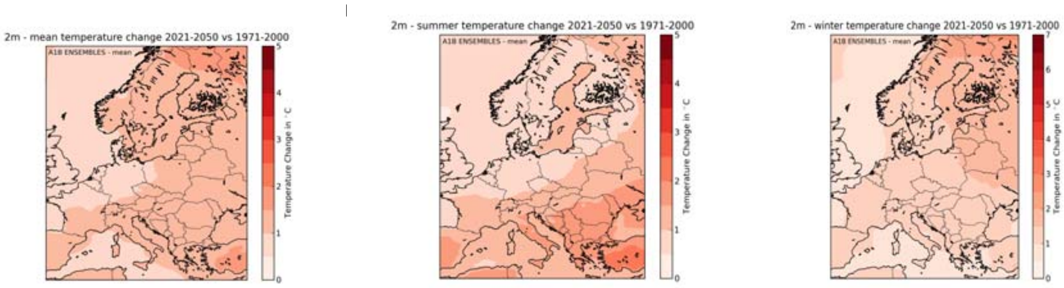
The European Commission initiated the ENSEMBLES project (2004-2009) under the Sixth Framework Programme for Research to provide researchers, decision makers, businesses and the public with future climate scenarios from a range of state-of-the art climate models. One of the project's principal objectives is to ‘allow the uncertainty in climate projections to be measured, so that a clearer picture of future climate can be formed’. To achieve this, the ENSEMBLES project developed ‘an ensemble prediction system for climate change, based on the principal state-of-the-art, high-resolution global and regional Earth system models developed in Europe’ (van der Linden and Mitchell (eds.) 2009).

This technical paper uses results of the regional component of the ENSEMBLES project (Research Topic 3), which provides an ensemble of high resolution regional climate change simulations for Europe from about 15 regional climate models. All models were driven by the IPCC A1B emissions scenario. For a detailed description of the models and the simulations set-up, see the final report of the ENSEMBLES project (van der Linden and Mitchell (eds.) 2009).

The maps and graphs as used in this section (Figures 3.22 (a)-(f)) have been provided by the Climate Service Centre (CSC) in Hamburg. They show emsemble-mean changes in climate based on a representative subset of the regional ENSEMBLES simulations, which are available from the ENSEMBLES RCM data portal (http://ensemblesrt3.dmi.dk/). More simulations were available to calculate the ensemble mean for temperature and precipitation, than for intensive snow fall and days with snow cover. Especially for days with snow cover, this could lead to a limitation of the robustness of the shown changes.

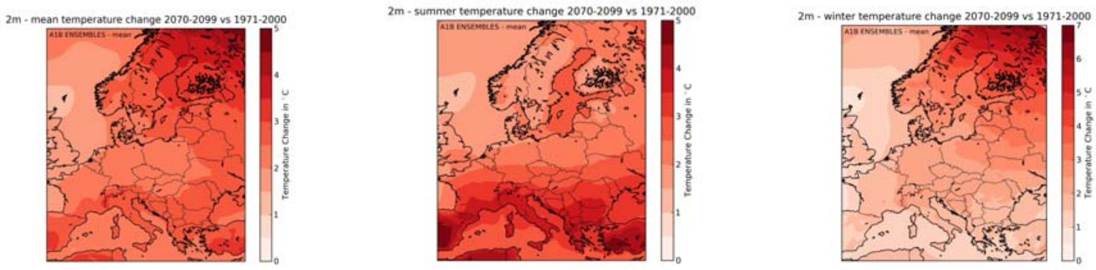
The ENSEMBLES final report states that “The evaluation of robustness in the regional RCM projections shows the mid-century signal for the multi-model mean temperature is one of warming in all of Europe and is much larger than the inter-model standard deviation. For precipitation the signal shows agreement in direction, projecting an increase in precipitation in the north and a decrease in the south, with all models agreeing in the north and twelve out of sixteen models agreeing in the south.” (van der Linden and Mitchell (eds.) 2009). Therefore the presented results are considered as robust indications for future changes.

Fig. 3.22
(a): Modelled changes in mean air temperature over Europe between 1971-2000 and 2021-2050



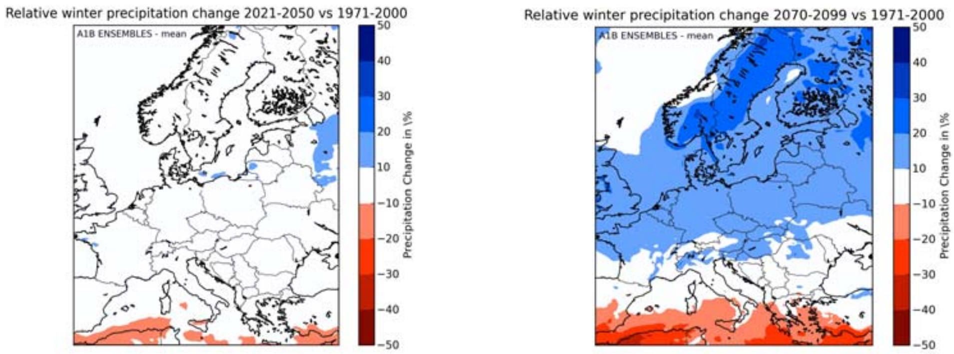
Remarks: *The average air temperature is projected to increase in the medium term (2021-2050) in all of the discussed regions in the annual mean as well as in the summer and winter season.*

(b): Modelled changes in mean air temperature over Europe between 1971-2000 and 2070-2099



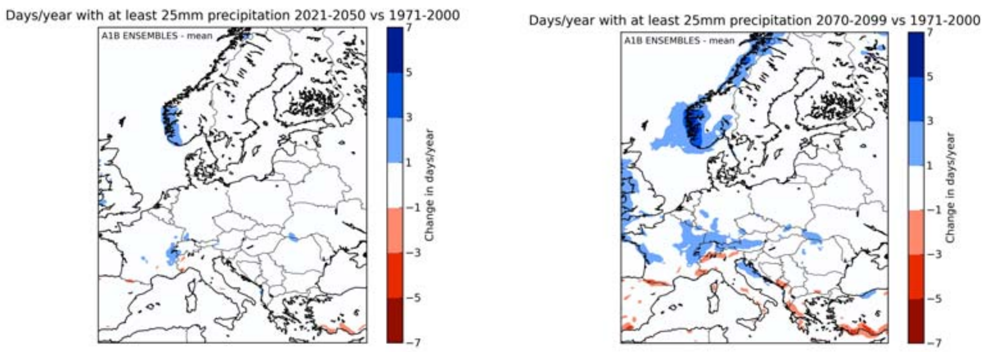
Remarks: *Modelled results project air-temperature increase to continue in the long-term (2070-2099). Particularly in the winter-season the more northern lying regions as Scandinavia and Svalbard (not covered by this map) are expected to experience more warming than the others (Alps, Pyrenees, Tatra).*

(c): Modelled changes in winter precipitation over Europe between 1971-2000 and 2021-2050 (left) and between 1971-2000 and 2070-2099 (right)



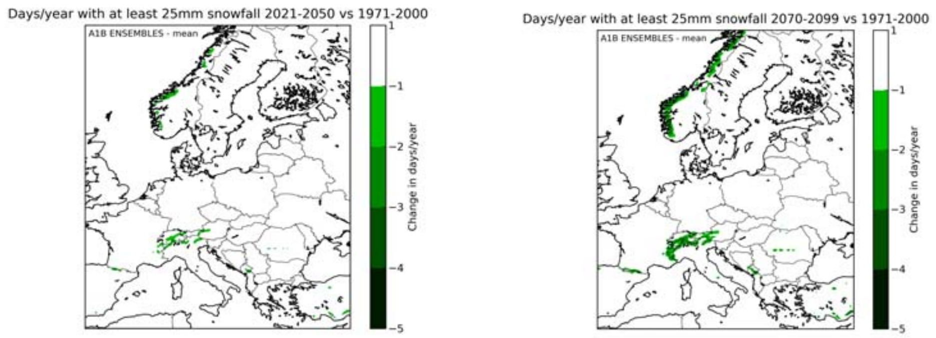
Remarks: *While the models don't project much change in winter precipitation in the medium term (left), long term projections indicate a general increase in mid and northern Europe and a decrease in the most southern parts. Due to increasing temperatures the fraction of snow is expected to decrease.*

(d): Modelled changes of days with intensive precipitation (>25mm) over Europe between 1971-2000 and 2021-2050 (left) and between 1971-2000 and 2070-2099 (right)



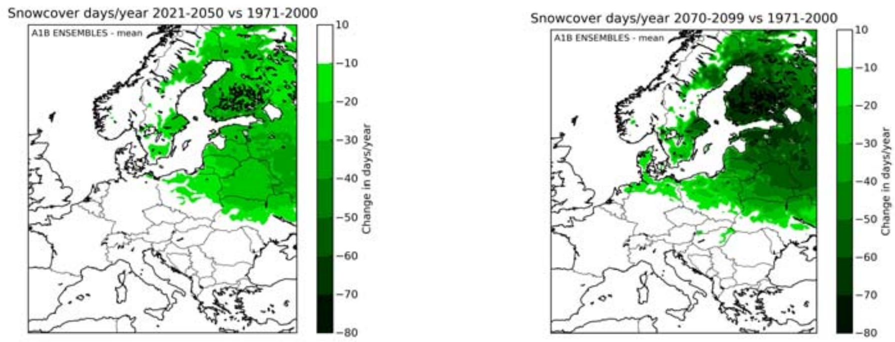
Remarks: *Some of the regions (parts of the Alps, coastal Norway) are expected to experience some more days with intensive precipitation, particularly in the long term, while the Pyrenees seem to experience a reduction.*

(e): Modelled changes of days with intensive snowfall (>25mm) over Europe between 1971-2000 and 2021-2050 (left) and between 1971-2000 and 2070-2099 (right)



Remarks: *According to the modelled results the number of days with intensive snowfall is expected to be slightly reduced in coastal Norway, parts of the Alps and the Pyrenees, particular in the long term.*

(f): Modelled changes of days with snow-cover in Europe between 1971-2000 and 2021-2050 (left) and between 1971-2000 and 2070-2099 (right)



Remarks: The number of days with snow-cover in north-eastern Europe will be clearly reduced, according to the ENSEMBLES-results. This phenomenon seems to be strengthened in the long-term.

Source: S. Pfeifer, R. Schmitt (CSC, 2010); based on results from the ENSEMBLES project

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4. Primary Impacts of Climate Change on the Cryosphere

4.1 Snow cover

Key messages:

- Observations show a clear reduction of snow reliability at low and medium altitude.
- Snow amounts are stable or increasing at high latitudes or altitudes (above 2000 m altitude in the Alps) because temperatures are too cold for melting and/or because winter precipitation has been increasing.
- Warmer winter temperatures are the main cause for the observed decrease in snowfall and snow depth in most of Europe.
- The altitude limit of snowfall is projected to shift even higher in the future as a consequence of increasing temperature. Simultaneous increases in winter precipitation at higher latitudes can no longer compensate for the effect of increasing winter temperatures.
- Winter tourism will be restricted to a shorter time period and/or to regions at increasingly high altitude.
- The declining snow reservoir will cause longer periods of low river flow in summer in many parts of Europe. This can have severe consequences for several economic sectors including agriculture, hydropower generation, water supply and river navigation.

Key graphs:

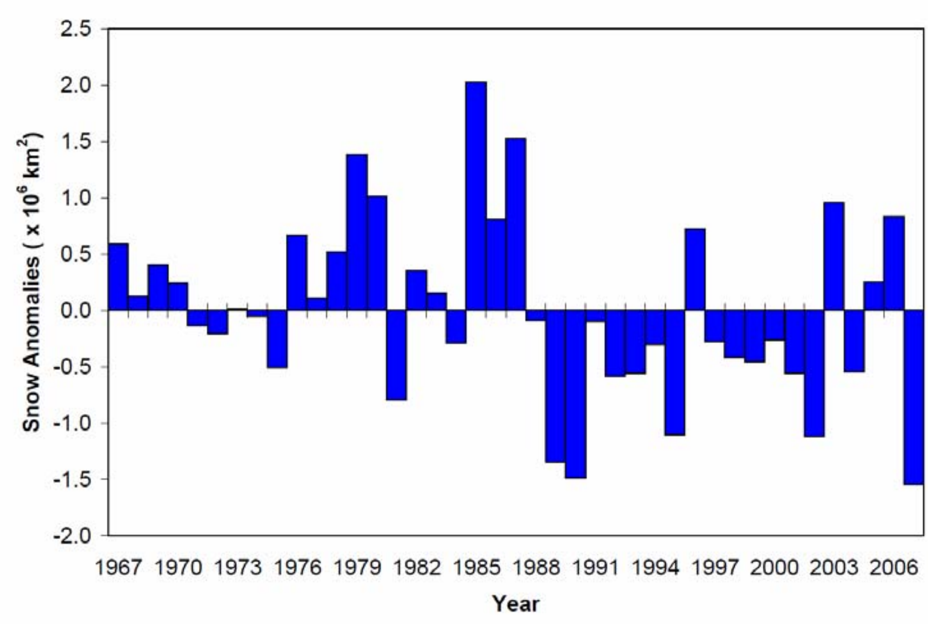

Fig. 4.1: Temporal trend of winter (JFM) snow cover extent over Europe from satellite data
Source : Henderson and Leathers, 2010

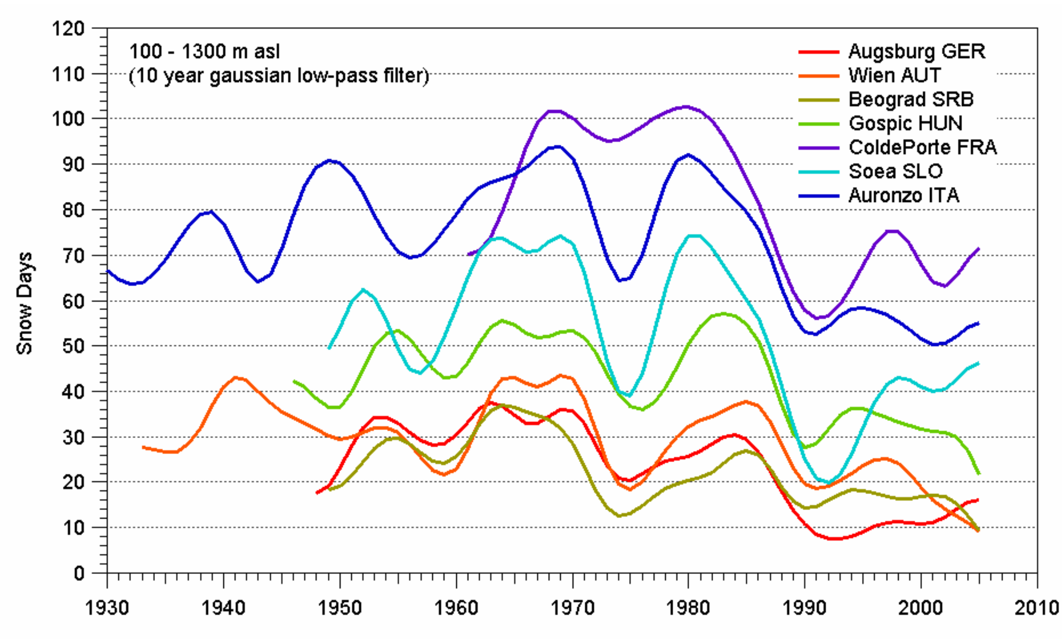

Fig. 4.2: Number of snow days at 8 locations in different European countries. All stations show a remarkable shift towards significantly less snow in the last 20 years. Shown are 10-year low-pass filtered values of annual snow days between December and March. The dashed line represents the median of the snow days from the individual stations. A snow day is defined as day with at least 5 cm snow on the ground
Source: Marty, 2009.

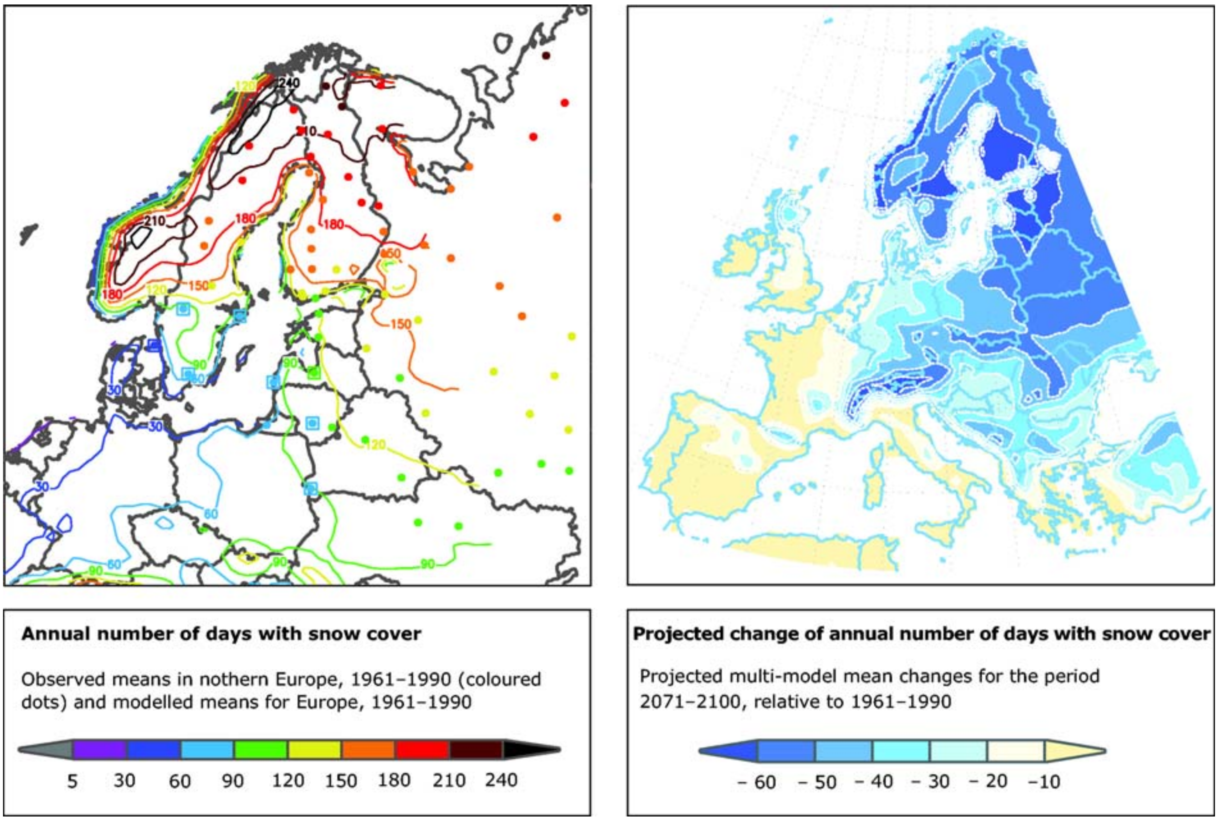

Fig. 4.3: Annual number of days with snow cover over European land areas 1961-1990 and projected change for 2071-2100
Source : Jylhä et al., 2008

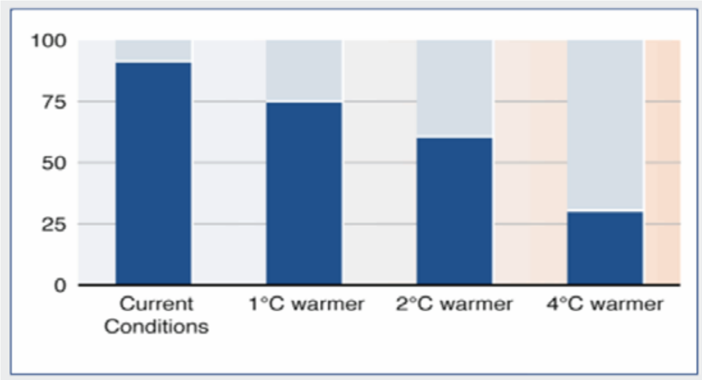

Fig. 4.4: Reliability of snow conditions in ski resorts in the European Alps. Dark blue shows the percentage of resorts with reliable snow conditions under present and future climate conditions **Source:** UNEP, 2007

Relevance

Snowiness in Europe is mainly dominated by elevation, but it also increases meridionally from south to north and latitudinally from west to east. Snow influences life and society in many ways. The amount and duration of snow in Europe has a high socio-economic significance in terms of both tourism and hydropower. Many mountain towns and villages heavily depend on snow, because their economy is dominated up to 90% by winter tourism (Abegg et al., 2007). The vast majority of customers of such ski areas live in the pre-alpine regions of Switzerland, Austria, Germany, Italy and France. A longer sequence of almost snow-less winters in these heavily populated regions, as was observed between the late 1980's and mid 1990's, caused serious troubles for some ski resorts. Winter tourism industry has already responded to the implications of these observed changes. A range of adaptation measures have been put into practice to offset the adverse impacts (Box 4.1). On the other hand, the European mountain systems are a huge water reservoir where winter precipitation is retained and stored in the form of snow and glaciers, which melt in the spring/summer months and provide a huge volume of water that replenishes surface and ground waters (Beniston, 2010). In Switzerland alone, seasonal snow cumulates to roughly 7 km^3 of water equivalent on average in March. Many large river systems throughout Europe, including the Rhine and Danube in Germany, the river Po in Italy, and the Rhone in France benefit from this natural storage. This often provides water when it is most needed, for instance, in the dry, precipitation-poor months of late summer. Norway, for example, where hydropower accounts for 99% of the electric power is highly dependent on the high elevation snow melt during summer. Therefore, periods of low water flow and droughts can have severe consequences throughout Europe for several economic sectors, particularly agriculture, river-navigation, energy production and drinking water provision (Barnett et al., 2005). Finally, snow also plays a key role in the ecology of many processes for a lot of plants and animals in Europe, especially in the high latitude and altitude regions (Jones et al., 2001).

Past trends

The importance of snow for hydrology and tourism in the Europe has led to quite a few studies, which investigated the past variability and trends of the snow cover mainly in the mountainous regions. Some of the countries have a relatively dense network of manual measurement stations available, where daily snow depth and snowfall is being measured with the help of a permanently mounted snow stake, respectively a new snow-board for 50 years or more. Remote sensing data of Europe's snow cover have only rarely been used for climatological purposes due to the lack of longer time series. A recent study (Henderson and Leathers, 2010) using satellite data between 1967 and 2007 revealed that the winter snow cover extent over Europe is slightly decreasing during the observed period (Fig.4.1). Moreover, it could be shown that especially in western and central Europe atypically small snow cover extents were strongly associated with the positive phase of the North Atlantic Oscillation (NAO) large scale atmospheric circulation pattern (Bednorz, 2010). In-situ observations in the different countries show large spatial variability of snow depth. The main factors contributing to this variability are the effects of altitude and latitude. Moreover, snow cover extent, which is closely controlled by air temperature, is less sensitive to changes in snowfall amount than the snow depth or snow water equivalent (SWE) as measured at the stations. Snow cover trends in the mountain regions, for example, are characterized by large regional and altitudinal variations (Brown and Mote, 2009) and the annual snow amount in the Alps is only weakly correlated with the NAO (Scherrer and Appenzeller, 2006). The following example from some European mountain regions will demonstrate the observed change dependence on altitude, region and local factors.

In the Alps the seasonal snow cover is primarily influenced by a high year-to-year variability due to anomalies in the large-scale weather patterns. Despite this fact several studies (Laternser and Schneebeli, 2003; Jurkovic, 2008; Durand et al., 2009; Valt and Cianfarra, 2010) have noted a general decrease of the snow depth, snow cover duration and snow fall days since the end of the 1980's for low-lying stations throughout the Alps (Fig.4.2). The decline could be linked to anomalous warm winter temperatures in the last twenty years (Scherrer et al., 2004; Marty, 2008), which seem to be unique for at least the last 500 years (Luterbacher et al., 2007). A trend towards less snow could mainly be detected at altitudes below about 1300 m asl, whereas no significant differences could be detected for high-altitude stations above 2000 m asl. The decreasing trend was generally stronger at southern slope of the Alps. In the Italian Alps a decrease in snow depth and snow duration in the last twenty years was even found for 20 stations situated between 2000 and 3000 m asl (Bocchiola and Diolaiuti, 2009). These examples demonstrate that the impact of climate change on snow cover is only clear and uniform for low altitudes. Above this limit the impact depends among others on altitude and region.

In the Tatra Mountains between Slovakia and Poland the changes of snow cover duration and thickness are very variable from region to region, as well as for different altitudes. Similar as in the Alps, a general decrease of snow cover duration as well as of solid precipitation was observed since winter 1921/1922. Delayed beginning and more intensive melting at the end of the winter due to warmer temperatures characterizes the last 20 years was observed at the lowest stations (Lapin et al., 2007). However, proportionally to altitude the number of stations with negative trend decreases and above the altitude of about 1000 m asl no significant negative trends are observed anymore. In contrast to the Alps the highest stations even show slight increases in snow duration. The estimated critical level, above which the snow cover duration trends become positive, lies at 1800 m asl on northern and at 2300 m asl on southern slopes (Vojtek et al., 2003). This increase at high altitudes is mainly caused by increasing spring snow falls. In agreement with other European regions, a substantial decrease in solid precipitation was detected at altitudes between below 1300 m asl, mainly on account of an increase of rain events during snow fall periods, which plays in favour of a denser snowpack. Thus, it is not surprising that the March snow water equivalent is slightly increasing in the last decades.

In the Pyrenees, snow is only measured operationally since 1986. Regional variability of the observed spring snow pack was related with precipitation and temperature. Highly significant correlations were found between snow depth in March and April-May and the climatic conditions in previous months. The good adjustment between predicted and observed series from 1985-2006 allowed to create predicted series of snow depth from 1950 to 2006, the time span for which climatic data was available. These constructed series shows an important inter-annual variability of snowpack in the region, and also a statistically significant decrease along the considered period. Thus, a clear dominance of years below the long-term average is observed after 1980. The snowpack decrease could mainly be explained by a significant depletion of precipitation during February and March (Lopez-Moreno, 2005), which is associated to the evolution of North Atlantic Oscillation index (NAO) in the last decades (Lopez-Moreno and Vicente-Serrano, 2007). Trends in low-elevation areas exhibited a sharper tendency, which suggests that warmer temperatures recorded since the early 80´s in the Pyrenees also contributes to the negative evolution of snow pack in the region.

Investigations in Scandinavian countries reveal for Norway a general decrease in snow depth and the length of the snow season at the majority of the stations during the last 100 years. This negative trend is more consistent at the southern most stations and more pronounced in the last few decades, reflecting the recent warming. The strongest trends are found for the end of the snow season and the number of days with snow. The fact that the no negative trends could be observed in maximum values and mean snow depth may be explained by the concurrent increase in winter precipitation (Dyrrdal and Vikhamar-Schuler, 2009). This theory is supported by an analysis of the maximum SWE during the last 100 years in the mountains of Norway, where a weak increase could be detected mostly at the snow rich northern stations (Stranden and Skaugen, 2009). Snow data from Finland, which is not exposed to the relatively warm westerly flows from the Atlantic Ocean, also demonstrate that the increasing winter precipitation has indeed an impact on the snow storage. Maximum SWE has been increasing in the eastern and northern part from 1946 to 2001 and decreasing in the southern and western part (Hyvärinen, 2003; Venäläinen et al., 2009). Despite large decadal variations increasing snow depth in the north of Scandinavia has also been confirmed by a long-term snow record form northern Sweden (Kohler et al., 2006).

Projections

Despite the fact that winter precipitation is projected to increase especially in northern and central Europe in the future (Christensen and Christensen, 2007), days with snow cover in Europe are projected to become rarer, because of less frequent occurrence of temperatures below zero. According to 7 Regional Climate Models (RCMs) decreases of more than 60 snow cover days were projected to occur around the northern Baltic Sea, on the western slope of the Scandinavian mountains and in the Alps. Conversely, the simulated percentage decrease in snow cover days (Jylhä et al., 2008) was most pronounced in the western and southern regions of Europe (Fig 4.3). These European wide scenarios may be compared with downscaled projections derived in the different mountain regions.

Regarding future snow cover in the Alps two different approaches, one based on the coupling physical models with RCMs (Beniston et al., 2003; Martin and Etchevers, 2005) and another based on the current snow-temperature sensitivity (Breiling and Charamza, 1999; Hantel and Hirtl-Wielke, 2007), both came to similar results, which, for a 2°C warming, point to a drastic decrease of snow depth of about 40-60% below 1800 m, a reduction of the snow cover duration of 4 to 6 weeks and a rise of the snow line by about 300-500 m. According to RCM projections the warming in the Alps will be accompanied by a small increase in winter precipitation. Some authors therefore concluded that higher altitudes, where the temperatures are still cold enough for snowfall might experience an increase in snow depth with climate warming. However, the outcomes of newer studies using the A2 and B2 scenarios revealed that the projected increase in winter precipitation over the Alps will not even in the higher resorts compensate for the projected increase in temperature (Uhlmann et al., 2008; Bavay et al., 2009) with important consequences for the accumulation zones of glaciers (Magnusson et al., 2010).

Based on GCM scenarios with a regional temperature increase of 2.5°C and a precipitation increase of 20% Lapin et al. (2007) estimated for the Tatra mountains that less frequent frost occurrence and more rain-on-snow events associated with higher temperatures will reduce the number of days with snow cover especially at altitudes below 1100 m asl.. On the other hand they expected the increase in winter precipitation totals will probably lead to increase of snow depth and occurrence, particularly at altitude above 1100 m asl.

The future Pyrenean snow accumulation and duration was simulated with a more sophisticated approach. A surface energy balance model including an explicit snow module was fed with hourly input data derived from daily outputs of the HIRHAM RCM (López-Moreno et al., 2009). Results for the end of the 21st century for two greenhouse gas emission scenarios (A2 and B2) indicated that snowpack in the Pyrenees will be strongly affected by projected climate change in the region. However, noticeable spatial differences in the magnitude of simulated changes in snowpack were detected. At 1500 m asl under SRES A2 and B2, SWE is predicted to decrease by up to 78% and 44%, respectively, and the duration of the snowpack by 70% and 32%. These results also show that two different greenhouse gas emission scenarios can lead to marked differences in the severity of expected changes in snowpack, being at least twice as pronounced under the A2 scenario compared with B2. These results are derived from the use of only one RCM. However, a comparison with 9 RCMs at one location revealed a large coherence between the different models (Lopez-Moreno et al., 2008).

In Scandinavia, the B2 scenario for 2071-2100 using two GCMs downscaled by the regional climate model HIRHAM has been applied to Norway. They predict a decrease in maximum SWE and a shorter snow accumulation season due to later snowfall and earlier snowmelt for the entire country (Vikhamar Schuler et al., 2006). The magnitude of the decrease in duration of the snow season,

however, diminishes with increasing altitude and distance from the coast. However, a more, topographically detailed hydrological study of the same scenario data set, suggests that for certain high elevation areas, annual maximum SWE might increase (Vikhamar-Schuler and Forland E., 2006). A study combing input from the RCM RCAO with a physical snow model at 5 locations in Finland using A2 and B2 scenarios found similar results (Rasmus et al., 2004). Four locations showed a clear decrease for snow depth, snow duration and maximum SWE. Only one location revealed a slightly higher snow depth and maximum SWE due to an increase in winter precipitation.
These projected changes in temperature and precipitation will most certainly be a challenge for the winter tourism (Fig. 4.4) and water resource management. Because of the sensitivity of the European snow cover to temperature, the depth, length and duration of the snow cover is highly influenced by climate change. As warming progresses in future, regions where snowfall is the current norm will increasingly experience rain and the snow on the ground will melt faster.

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4.2 Glaciers and ice caps

Key messages

- Due to their proximity to melting conditions, glaciers are one of the most reliable natural indicators for climatic changes.
- In the second half of the 20th century, European glaciers and ice caps (outside Greenland) covered a total of about 54,000 km^2 distributed in Svalbard (68%), Iceland (21%), Scandinavian Peninsula (6%), Alps (5%), and the Pyrenees (<1%).
- The total volume of glaciers can only be roughly approximated. Current estimates based on the above (area) data come to an ice volume of about 15,500 km^3. This corresponds to a potential sea level rise of about 40 mm, of which the vast majority is located in Svalbard (26 mm) and Iceland (12 mm).
- The Little Ice Age moraines that formed between the mid 18th and the mid 19th century mark the maximum glacier extents during the past 11,000 years (the Holocene).
- The strong centennial retreat of glaciers from these Little Ice Age moraines is well documented and apparent in all European regions. In some regions, there have been intermittent periods of reduced glacier melting or even of glacier re-advance such as in the late 1970s in the Alps and Iceland and in the 1990s in coastal Scandinavia.
- Glacier melt seems to be strongest in the European Alps. There, more than half of the ice-covered area and probably two third of the ice volume disappeared since 1850; the average annual ice thickness loss since 2000 has been above one meter.
- European glacier changes since the Little Ice Age have been driven mainly by increased summer air temperatures with secondary effects from variations in winter precipitation. Both are influenced by atmospheric and oceanic circulation patterns. Further factors for increasingly negative mass balances in most regions are most probably the (re-) brightening of the atmosphere, extension of the ablation period, and reinforcing effects such as dust-related darkening or melt-induced elevation lowering of glacier surfaces.
- Climate change scenarios for the 21st century suggest a continued increase in global mean air temperature by 1.4–5.8°C and 2.0–6.3°C in Europe (without policy measures). Corresponding projections of precipitation patterns show a more varied picture with seasonal change rates of 1-5% per decade. Under such scenarios, glaciers will continue to melt and may totally disappear in some mountain ranges in the coming decades.
- Available numerical model experiments of glaciers in Iceland, Scandinavia and in the Alps indicate that a further increase of regional summer air temperature by 2° C will reduce glacier area and volume by half or more of their present extents. The impact of a 1°C warming could only be offset if precipitation would increase by 20% or more. Potential re-growth of glaciers in these regions would require decades of cooler and wetter conditions.
- Glaciers changes in Europe already influence the local hazard situation, the runoff from alpine catchments, tourism and landscape, and – to a limited extent – global sea level. The anticipated marked changes for the 21st century might lead to impacts that are unprecedented during the last 11,000 years (the Holocene).

Key graphs

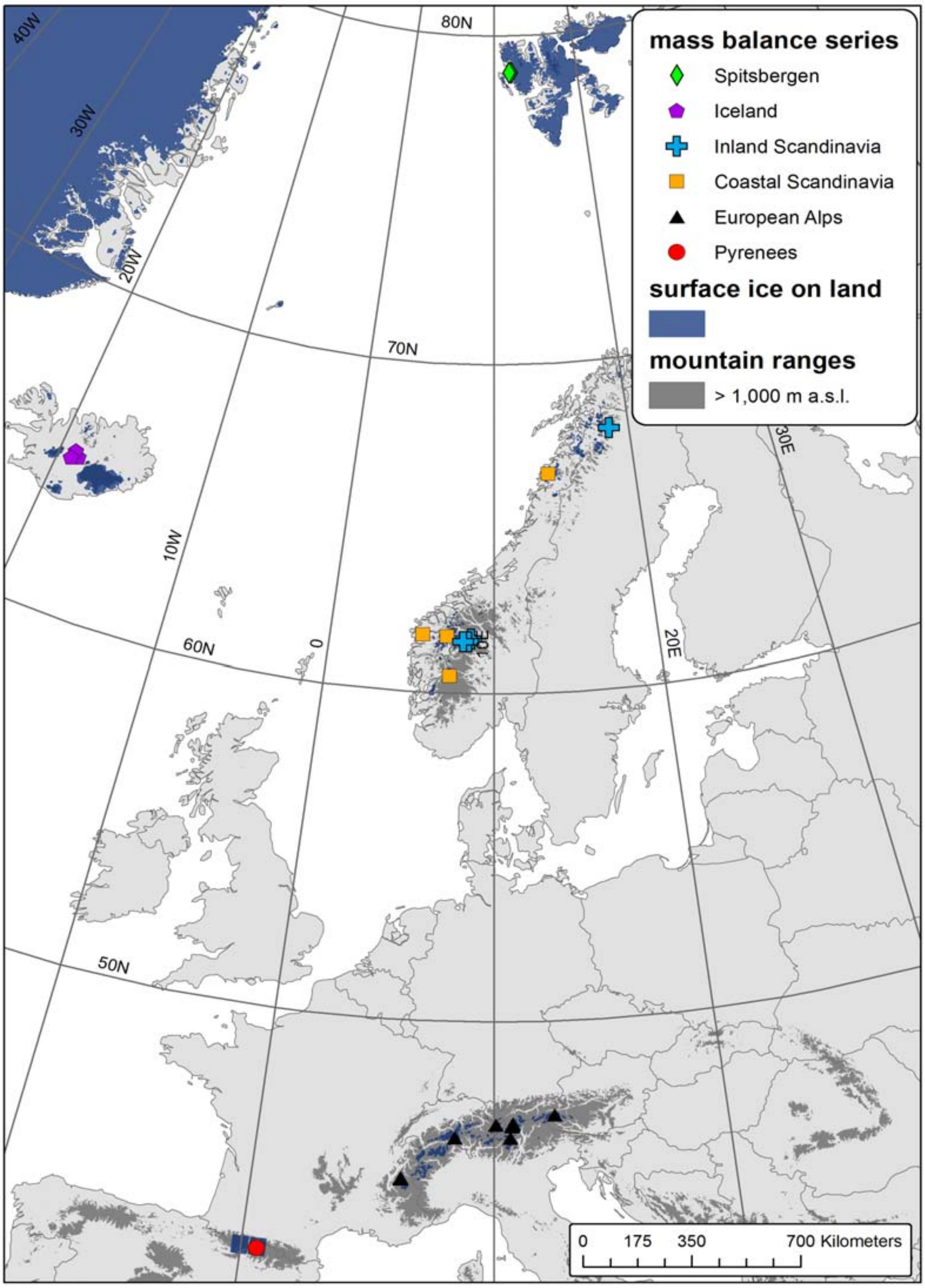

Figure 4.5: Glacier distribution in Europe. The map shows the distribution of glacier and ice caps as well as the Greenland Ice Sheet (upper left corner) together with the locations of long-term glacier mass balance observation programs (cf. Fig. 4.6). Note that the location of glaciers in the Pyrenees is marked with oversized squares which do not represent their real extents.

Source : Glacier information from the World Glacier Monitoring Service; country outlines and surface ice on land cover from ESRI's Digital Chart of the World.

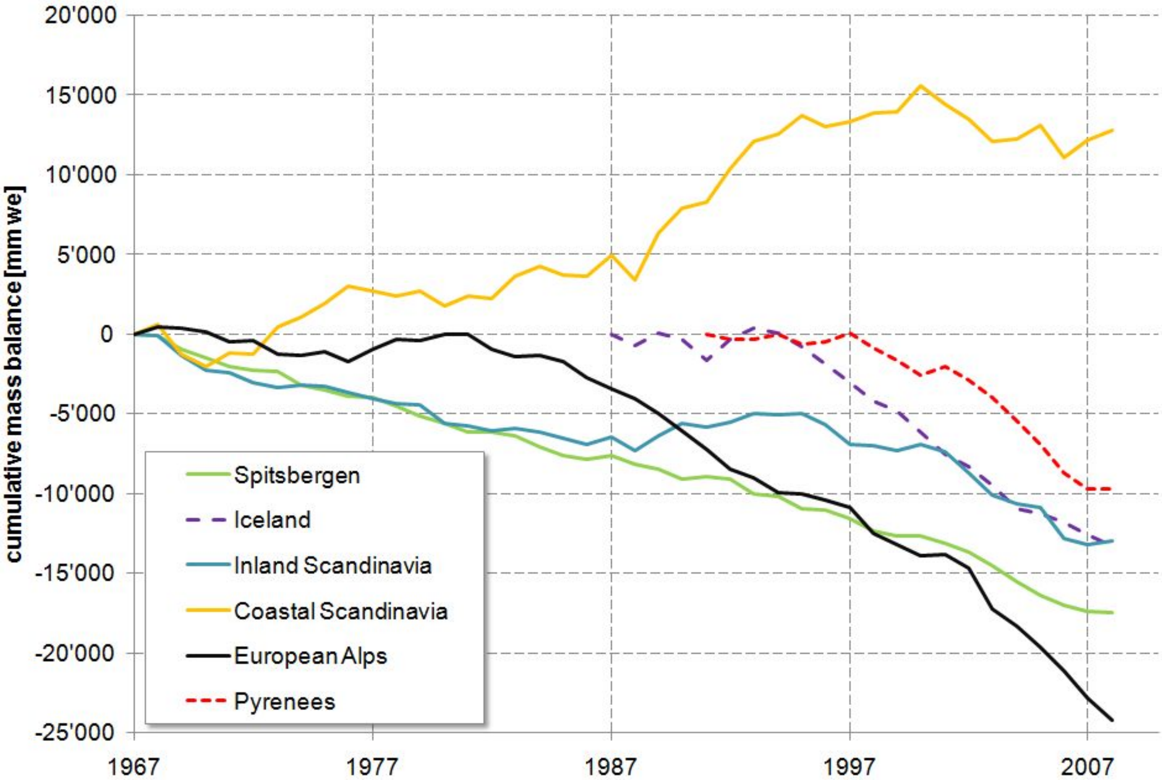

Figure 4.6: Glacier mass balance results in Europe. Long-term and continuous mass balance series are an excellent index of regional glacier changes. The absolute values might, however, not be representative for the total ice changes of the corresponding region. The index shows a cumulative loss in ice thickness from 1967 to 2008 in all regions except for coastal Scandinavia where glaciers gained mass until 2000. Note that data series from Iceland and the Pyrenees start later and, hence, their absolute values cannot directly be compared to the ones of the other series. The graph is based on data from the following glaciers: Spitsbergen: Midtre Lovénbreen, Austre Brøgerbreen; Iceland Hofsjökull E, N, ans SW; Inland Scandinavia: Gråsubreen, Hellstugubreen, Storbreen, Storglaciären; Coastal Scandinavia: Nigardsbreen, Engabreen, Hardangerjøkullen, Ålfotbreen; European Alps: Gries, Silvretta, Vernagtferner, Hintereisferner, Kesselwandferner, Sonnblickkees, Caresèr, Saint Sorlin, Sarennes; Pyrennes: Maladeta.

Source : World Glacier Monitoring Service.

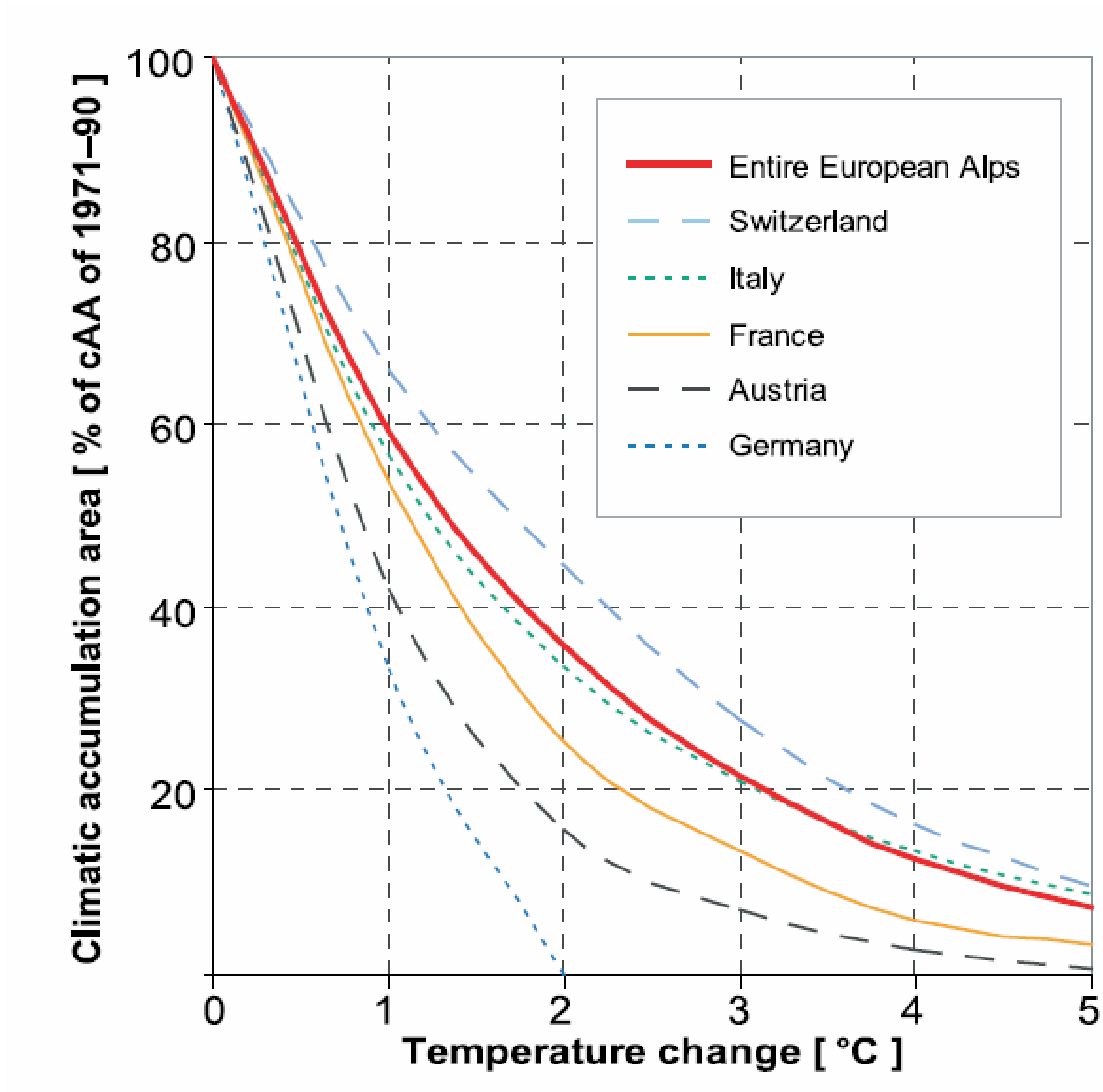

Figure 4.7: Alpine glacier scenarios for 21st century. The figure shows the estimated relative changes in glacier extent compared to the reference period (1971–90) for an increase in regional summer air temperature.

Source : Zemp et al. (2006).

Relevance

Glaciers are an inherent component of the culture, landscape and environment in Alpine and Nordic countries of Europe. They represent a unique resource of fresh water for domestic, agricultural and industrial use, an important economic component of tourism and hydro-power production, and a potential source of serious natural hazards. Glacier melt during summer is an important component of the hydrological water cycle in mountain areas and contributes to the discharge of river systems (BOX 4.2). In today's Norway, 15% of the used runoff comes from glacierized basins and 98% per cent of the electricity is generated by hydropower production (Andreassen et al. 2005). Globally, glaciers are one of the main contributors of present sea level rise. Their contribution has more than doubled over the past decades and is currently estimated to about one millimetre per year (Meier et al. 2007, Kaser et al. 2009). Moreover, glacier changes are recognized as high-confident indicators and as valuable elements in early detection strategies within the international climate monitoring programmes (GCOS 2004, GTOS 2008).

Box 4.1: Glacier contribution to the Alpine runoff
The Alps are widely known as the 'water towers' of Europe. The capacity to export water to the adjoining lowlands of its rivers is more dependable and specific runoff is higher than that of similar-sized lowland basins (Weber et al., 2010). Thereby, glacier ice-melt further enhances water yield from high mountain regions during dry and hot summer periods. Since 1973, the runoff from Vernagtferner in the Austrian Oetz Valley has been recorded continuously at the gauging station Vernagtbach at 2,635 m a.s.l. The runoff from this head watershed with an area of 11.4 km² and a present glacier area fraction of about 70% can be set in relation to basin precipitation and change in glacier storage. A recent analysis by Weber et al. (in press) shows that for the past decade that in glacierized head watersheds there is about an equal amount of runoff originating from ice-melt, snow-melt and rain (about 33% each). Further downstream the portion of ice-melt decreases sharply with the effect that ultimately 2% of annual runoff is of glacial origin in Passau/Achleiten (basin area of about 77,000 km², current glaciation 0.5 %), and about three quarter originates from rain and one quarter from snowmelt. Using climate scenarios from regional climate models (SRES A1B, cf. BOX 2.2) revealed that the contribution from ice-melt in the glaciated head watersheds will decrease sharply after some decades, the proportion of snowmelt will be about the same, and rain contribution will increase to about half of the annual runoff. In Passau, the portion from ice-melt will be negligible, and 80% of runoff will be from rain and 20% from snowmelt. With the anticipated warming over the whole year and the drying out of the summer season the Alps' capacity to export water will diminish, and water availability will be reduced mainly through the loss of summer precipitation and increased evaporation, and not so much due to the loss of glacier-melt.

Data compilation and process understanding

The observation of glaciers has been internationally coordinated since 1894 and is today lead by the World Glacier Monitoring Service (WGMS 2008; BOX 4.3). The fluctuations of a glacier, which is not influenced by thick debris covers, calving or surge instabilities, are a reaction to climatic forcing. Thereby, the glacier length change (i.e., the advance or retreat) is a delayed reaction to climatic changes over the past years to decades, or even centuries. The glacier mass balance (i.e., the change in thickness) is a more direct and un-delayed reaction to the annual atmospheric conditions (Haeberli and Hoelzle 1995). The mass balance variability of glaciers is well correlated over distances of several hundred kilometres and with air temperature (Lliboutry 1974, Letréguilly and Reynaud 1990, Schöner et al. 2000, Greene 2005). Glacier mass balance change provides an integrative climatic signal, however, and the quantitative attribution of the forcing to individual meteorological parameters is not straight forward. The energy and mass balance at the glacier surface is influenced by changes in atmospheric conditions (e.g., solar radiation, air temperature, precipitation, wind, cloudiness). Air temperature thereby plays a predominant role as it is related to the radiation balance, turbulent heat exchange and solid/liquid precipitation ratio (Kuhn 1981, Ohmura 2001). The climatic sensitivity of a glacier not only depends on regional climate variability but also on local topographic effects and the distribution of the glacier area with elevation, which can result in two adjacent glaciers featuring different specific mass balance responses (Kuhn et al. 1985).

Box 4.2: International Glacier Monitoring
Glacier research and monitoring have a long tradition in Europe. Already in 1894, the internationally coordinated collection of information about ongoing glacier changes was initiated at the 6th International Geological Congress in Zurich, Switzerland. Today, the World Glacier Monitoring Service (www.wgms.ch), in close collaboration with the US National Snow and Ice Data Center (www.nsidc.org) and the Global Land Ice Measurements from Space initiative (www.glims.org), continues the compilation and dissemination of standardised data and information on distribution and ongoing changes in glaciers worldwide (WGMS 2008). Together, these three bodies run the Global Terrestrial Network of Glaciers (www.gtn-g.org) which aims to combine field observations with remotely sensed data, process understanding with global coverage, and traditional measurements with new technologies within the global climate observation systems as a contribution to the United Nations Framework Convention on Climate Change (www.unfccc.int).
(details of the World Glacier Monitoring Service and its National Correspondents in Europe are given in Annex 6.4.).

Past Trends

20th century distribution in Europe

In Europe[1], glaciers (outside Greenland) are found on Svalbard, in Iceland, on the Scandinavian Peninsula, in the Alps, and in the Pyrenees. A few small glacierets and perennial snow fields are also found in the Apennines, the Tatras between Slovakia and Poland, and in the mountains of Albania, Bulgaria, and Slovenia. For the second half of the 20th century, an almost complete detailed inventory of Europe's glaciers and ice caps was compiled mainly based on topographic maps and aerial photographs for the World Glacier Inventory (WGMS 1989, Table 4.1). The largest ice masses are found on the Svalbard Archipelago which is situated in the Arctic Ocean north of mainland Europe. Its topography is more than half covered by ice (36,612 km^2), and is characterized by plateau mountains and fjords. Iceland is covered by 15 major ice caps and a few hundred other glaciers with a total area of 11,260 km^2. Icelandic ice caps and glaciers are often influenced by volcanic activities. Due to the combination of high latitude and the moisture from the North Atlantic, many glaciers and ice caps with an overall area of 3,058 km^2 developed on the Scandinavian Peninsula, all within 180 km of the west coast. The Alps host a total glacier cover of 2,909 km^2 distributed along the entire mountain range from the peaks above 4,000 m a.s.l. in France and western Switzerland, over Italy and Germany to Austria. The glaciers of the Pyrenees are situated in the Maladeta massif in Spain and around the peak Vignemale in France and sum up to about 12 km^2. All together, these inventories sum up to a total glacier cover of about 54,000 km^2 in Europe. Based on rough estimates by Radic and Hock (2010) this corresponds to an ice volume of about 15,500 km^3, or about 40 mm potential sea level rise, of which the vast majority is located in Svalbard (26 mm) and Iceland (12 mm).

[1] The European region covered in this report includes all member countries of the EEA

Table 4.1: Glacier distribution in Europe

Region	Glacier area [km^2]	Time period and standard deviation of data in WGI	Original sources
Svalbard total	**36,612**		
- Spitsbergen	21,871	1966 ±5	Hagen et al. 1993
- Nordaustlandet	11,309	1971 ±3	Hagen et al. 1993
- Edgeöya/Barentsöya	2,705	1971 ±3	Hagen et al. 1993
- Kong Karls Land	22	1970 ±0	Hagen et al. 1993
- Kvitöya	705	1977 ±0	Hagen et al. 1993
Iceland total	**11,260**		
- Vatnajökull	8,300	1960/73/77	Björnson 1980, Williams 1986
- Langjökull	953	1960/73/77	Björnson 1980, Williams 1986
- Hofsjökull	925	1960/73/77	Björnson 1980, Williams 1986
- Myrdalsjökull	596	1960/73/77	Björnson 1980, Williams 1986
- Drangajökull	160	1960/73/77	Björnson 1980, Williams 1986
- Eyjafjallajökull	78	1960/73/77	Björnson 1980, Williams 1986
- others	247	1960/73/77	Björnson 1980, Williams 1986
Scandinavian Peninsula	**3,058**		
- Northern Norway and Sweden	1,441	1961 ±6	Østrem et al. 1973
- Southern Norway	1,617	1966 ±1	Østrem et al. 1969
Alps total	**2,909**		
- Switzerland	1,342	1973 ±0	Müller et al. 1976
- Italy	607	1977 ±9	various, cf. WGMS 1989
- Austria	543	1969 ±0	Gross 1983, 1988; Patzelt 1980
- France	417	1971 ±4	Edouard and Vivian 1980
- Germany	1	1970/71	Finsterwalder and Rentsch 1973
Pyrenees total	**12**	1975 ±5	Höllermann 1968
Total Europe	**53,851**		

Source: WGMS (1989)

Changes from the Little Ice Age to present

The LIA moraines that were formed between the mid 18th and the mid 19th century mark a Holocene (i.e., the past 11,000 years) maximum extent of glaciers in Europe, as well as in many other regions of the world (Grove 2004, Solomina et al. 2008). The strong centennial retreat of glaciers from these moraines is apparent in all European regions and well documented in a wealth of studies. The following section on regional changes is based on WGMS (2008) and updated with a selection of more recent studies.

-Svalbard

During the LIA, most glaciers were close to their late Holocene maximum extents and remained there until the onset of the 20th century (Svendsen and Mangerud 1997). The western part of Svalbard is quite well represented with glacier observations. Front variation series span most of the 20th century and continuous mass balance measurements are available since the end of the 1960s from Austre Brøggerbreen and Midtre Lovénbreen. Annual mass balance records of small glaciers in western Spitsbergen indicate a negative mass balance regime since at least the mid 1960s (Jania and Hagen 1996, Hagen et al. 2003a,b, Sobota 2007; Fig. 4.6).In that region, comparisons of photogrammetric maps/DEMs, dating back to 1936, show substantial decreases of glacier area and volume (Nuth et al., 2007) with enhanced thinning rates after 1990 when compared to recent airborne LiDAR (Bamber et al. 2005, Kohler et al. 2007) and ICESat altimetry (Nuth et al. 2010). This 20th century mass loss has even been documented in an accelerated uplift of the rocky margins of the Island as indicated by high-precision global positioning system located in western Spitsbergen (Kierulf et al. 2009, Jiang et al. 2010). The mass balance of northeastern Spitsbergen glaciers has been less negative than the western ones (Bamber et al. 2005, Nuth et al. 2010). The Nordaustlandet ice caps, Austfonna and Vestfonna, have been close to balance over the last two decades (Pinglot et al. 2001, Moholdt et al. 2010, Nuth et al. 2010) if the calving front retreat losses are ignored (Dowdeswell et al. 2008). Based on repeat-track ICESat altimetry from 2003 to 2008, Moholdt et al. (in press) find that most glaciers have experienced low-elevation thinning combined with high-elevation thickening. Thereby, the largest ice losses have occurred in the west and south, while north-eastern Spitsbergen and the Austfonna Ice Cap have gained mass.

The climate and as such the fluctuations of glaciers and ice caps are much influenced by the extent and distribution of sea ice which in turn depends on ocean current and on the Arctic and North Atlantic Oscillations. Furthermore, dynamic processes such as calving and surges might dominate the fluctuations of some glaciers. About 60% of glacierized areas drain into tidewater glaciers (Blaszczyk et al. 2009), and surge activities have been observed over most of Svalbard (e.g., Lefauconnier and Hagen 1991; Hamilton and Dowdeswell 1996; Sund et al. 2009).

-Iceland

Most glaciers in Iceland reached their maximum postglacial extent around 1890 (Björnsson 1979, Kirkbride 2002, Sigurðsson 2007). During the first quarter of the 20th century, retreat was notable but not rapid (Björnsson 1998). The abrupt increase in temperature that occurred about 1925 was accompanied by a rapid retreat of glacier fronts all over Iceland (Eyþórsson 1931, 1963, Sigurðsson 1998). Regular front variation observations were started in 1930 (Eyþórsson 1931) and document the periods of glacier retreat (1930–60, after 1990) and intermittent re-advances (1970–85; Sigurdsson et al. 2007). Continuous glacier mass balance measurements started in 1988 (Björnsson et al. 2002, Sigurðsson et al. 2007; Fig. 4.6). Mass-balance measurements show alternating positive and negative mass balance of glaciers during the period 1987–95, but the mass balance has been predominantly negative since 1996 (Sigurdsson et al. 2007). As in Svalbard, the 20th century mass loss of the major ice caps, such as Vatnajøkull, has been documented in an accelerated uplift of the rocky margins of the Island (Jiang et al. 2010). The outlines of all 276 identified glaciers have been digitized using mainly satellite images, but also aerial photographs and direct field observations, for approximately the year 2000 (Sigurdsson et al. 2007). The total glacier area in 2000 was found to be 11,048 km^2 which is about 2% less than the area reported in the inventory of the 1970s (see Table 1). Changes in summer temperature seem to have been the predominant driving factor in variations of the mass balance of the non-surg-type glaciers in Iceland (Sigurðsson et al. 2007). However, some of the rapid glacier advances might have been related to volcanic activities and surges rather than to climatic events.

-Scandinavian Peninsula

After having probably disappeared in the early/mid Holocence (Nesje et al. 2008), Norwegian glaciers and ice caps reached their maximum extents in the mid-18th century (Grove 2004). These advanced glacier states are attributed both to lower summer temperatures and higher winter precipitation, due to a positive NAO-index, in the first half of the 18th century (Nesje and Dahl 2003). Blomsterskardsbreen, an outlet glacier of Folgefonna, is one of the known exceptions in southern Norway, reaching its maximum extent in the 20th century (Tvede and Liestøl 1977). In Lyngen in northern Norway the LIA glacier maximum is suggested to be about 1900–1910 (Ballantyne 1990, Bakke et al. 2005). In the Kebnekaise Mountains, Swedish Lapland, glaciers had their greatest Holocene extents probably in the period of 17th century and beginning of 18th century (Karlén 1973, 1988). Most glaciers in northern Sweden reached positions close to their Holocene maximum at the beginning of the 20th century (Holmlund and Jansson 1999).

At the turn to the 20th century, annual observations of glacier front variations began in Sweden (Holmlund 1996) and Norway (Andreassen et al. 2005). These observations reveal that Scandinavian glaciers experienced a general recession during the 20th century with intermittent periods of re-advances around 1910 and 1930, in the second half of the 1970s, and around 1990; the last advance stopped at the beginning of the 21st century (Grove 2004, Andreassen et al. 2005).Mass balance measurements began in 1946 at Storglaciären (Sweden; Schytt 1981, Jansson and Pettersson 2007) and in 1949 at Storbreen (Norway, Liestøl 1967,Andreassen et al. 2005), providing long and continuous series of winter and summer balances. In the 1960s and 1970s measurements began at many other glaciers in Scandinavia (e.g., Holmlund and Jansson 1999, Andreassen et al. 2005). In addition to the field observations, aerial photographs have been taken in about decadal intervals and are used for comparison of the glaciological and the geodetic mass balances (e.g., Andreassen 1999, Østrem and Haakensen 1999, Haug et al. 2009, Zemp et al. 2010). Results reveal cumulative mass surplus at the maritime glaciers (e.g., Hardangerjøkulen, Nigardsbreen, Ålfotbreen, Engabreen), whereas the more continental glaciers (e.g. Storglaciären, Gråsubreen, Hellstugubreen, Storbreen) continued their ice loss. All glaciers in Norway, except Langfjordjøkelen, had a transient mass surplus in the period 1989 to 1995 which was mainly a result of increased winter accumulation. Since 2001, all monitored glaciers have experienced an overall mass deficit (Kjøllmoen et al. 2010).

To gain an updated overview of the present state of overall glacier cover and its changes since the previous inventories, glaciers have been mapped in recent years based on Landsat data. Whereas glaciers in the Svartisen region in northern Norway showed an area reduction close to zero from 1968 to 1999 (Paul and Andreassen 2009), the glacier area reduced by 10% from about 1980 to 2003 in Jotunheimen, southern Norway (Andreassen et al. 2008). Analysis of aerial photographs from 1968 1985, and 2002 of the western Svartisen Ice Cap confirmed the limited glacier area reduction in that region and suggest a volume loss for the drainage basin of Engabreen (Haug et al. 2009). The latter is in contrast to mass gain over the same periods reported from direct glaciological measurements at Engabreen and might be caused by changing ice divides of the ice cap (Elvehøy et al 2009).

The mass balance of Scandinavian glaciers is strongly influenced by atmospheric and oceanic circulation changes over the North Atlantic (Hanssen-Bauer and Førland 1998, Chen and Hellström 1999, Nesje et al. 2000, Wanner et al. 2001, Uvo 2003). Summer air temperature – and thus glacier ablations – is strongly related to the position of the jet stream and the strength of high pressure areas as the corresponding circulation pattern determines the relative contribution of incoming air masses from the North Atlantic (wet and cold), the Arctic (dry and cold) or from the East (dry and warm). Winter precipitation – and thus glacier accumulation – is strongly related to the North Atlantic Oscillation (NAO) index: a positive index with strong westerly winds and increased cyclonic frequency across the North Atlantic leads to high amounts of winter precipitation, especially in the coastal areas of Southern Norway. Pronounced south-easterly airflows bring moisture from the Baltic Sea to glaciers east of the drainage divide of the Scandinavian mountain chain.

-Alps

The Alps are probably the densest populated mountain range with glaciers. It is hence not surprising that here the greatest number of available information about distribution and changes of glaciers as well as scientific studies are found. For several Alpine glaciers, fluctuations spanning time periods from centuries to millennia were reconstructed based on geomorphological and archaeological evidences, dendrochronology, as well as from historical documents and pictorial sources (Zumbühl 1976, Holzhauser and Zumbühl 1996, Pelfini 1999, Nicolussi and Patzelt 2000, Holzhauser et al. 2005, Nussbaumer et al. 2007, Nussbaumer and Zumbühl in press). Three main glacier advances are reported during the LIA period, i.e., in the 14th century, in the 17th century, and the last one culminating around 1850 in which most glaciers reached their Holocene maximum extent and destroyed earlier moraines (Gross 1987, Maisch et al. 2000, Grove 2004). However, in some regions the moraines from 1820 mark this maximum extent (e.g., Wetter 1987, Holzhauser and Zumbühl 2003).

From detailed repeat inventories in Switzerland (1850: Maisch et al. 2000; 1973: Müller et al. 1976; 1998/99: Kääb et al. 2002, Paul et al. 2002), the overall Alpine glacier cover is estimated to have reduced by 35% from 1850 to the 1970s (Paul et al. 2004; Zemp et al. 2008) and by another 22% from the 1970s to 1998/99 (Zemp et al. 2008). A second complete Alpine inventory was derived from Landsat scenes for 2003 and reveals another 9% since 1998/99 (Paul et al. in prep.). Analysis of early inventories in Austria (Gross 1987) and recent regional repeated inventories (Abermann 2009) confirm these estimates.

Annual front variation observations started in the second half of the 19th century. They document a general trend of glacier retreat over the past 150 years with intermittent Alpine glacier re-advances in the 1890s, 1920s, and 1970–1980s (Patzelt 1985, Pelfini and Smiraglia 1988, Citterio et al. 2007, Zemp et al. 2008).

Mass balance measurements show an accelerated ice loss after 1980 (Vincent 2002, Huss et al. 2008) culminating in an annual loss of 5 to 10 per cent of the estimated remaining ice volume (cf. Haeberli and Hoelzle 1995, Zemp et al. 2006, Farinotti et al. 2009) in the extraordinarily warm year of 2003 (Zemp et al. 2005). The mean annual mass balances of available long-term measurement series were slightly negative in the 1970s, roughly -0.5 m w.e. a^{-1} in the 1980s, about -0.75 m w.e. a^{-1} in the 1990s, and exceeded -1.0 m w.e. a^{-1} after the turn of the century (Fig 2). The measured rates of mass loss since 1980 are similar to modelled ones in the 1940s (Huss et al. 2008, Huss and Bauder 2009) but about two to four times the loss rates reconstructed from cumulative length changes for the time period after 1850 (Hoelzle et al. 2003, Steiner et al. 2005) and characteristic long-term mass changes during the past 2,000 years (Haeberli and Holzhauser 2003). Decadal volume changes back to the late 19th century from geodetic surveys confirm these change rates (Lang and Patzelt 1971, Kuhn et al. 1999, Bauder et al. 2007, Abermann et al. 2009) and are used to homogenize, validate and calibrate the direct mass balance measurements (Thibert et al. 2008, Huss et al. 2009, Fischer 2010).

The general centennial glacier retreat from the LIA moraines corresponds well with the observed warming trend over this period (e.g., Oerlemans 1994, Vincent et al. 2005). However, the onset of the retreat in the decades after 1850 might have been triggered by a negative winter precipitation anomaly (relating to the mean of 1901-2000; Wanner et al. 2005, Vincent et al. 2005). The intermittent periods of glacier re-advances in the 1890s, 1920s and 1970-1980s can be explained by earlier wetter and cooler periods, with reduced sunshine duration and increased winter precipitation (Patzelt 1987, Schöner et al. 2000, Laternser and Schneebeli 2003). In addition, the positive mass balance period between 1960 and 1980 was characterised by negative winter North Atlantic Oscillation index values, which caused an increase of the meridional circulation mode and more intense north-westerly to northerly precipitation regime (Hoinkes 1969, Wanner et al. 2005, Huss et al. 2010). The high mass loss rates of the 1940s are attributed to enhanced solar radiation whereas the dimming of solar radiation from the 1950s until the 1980s is in line with reduced melt rates and advancing glaciers (Ohmura et al. 2007, Huss et al. 2009). The observed trend of increasingly negative mass balances since 1980 seems to be consistent with strong warming (Schöner et al. 2000, Vincent et al. 2004, Bocchiola and Diolaiuti 2009), (re-) brightening of the atmosphere (Ohmura et al. 2007), extension of the ablation period (Vincent et al. 2004, Bocchiola and Diolaiuti 2009), and reinforcing effects such as dust-related darkening of glacier surfaces and corresponding surface albedo reduction (Paul et al. 2005, Oerlemans et al. 2009).

-Pyrenees

The moraines that mark the LIA maximum extension of glaciers in the Pyrenees are dated to around 1820–30 (Chueca and Julián 1996). At that time, overall glacier extent in the nine main Pyrenean mountainous massifs both in Spanish and French regions (Balaitús, Infiernos, Vignemale, Monte Perdido/Gavarnie, Pic Long, La Munia, Posets, Perdiguero and Maladeta) summed up to over 20 km^2 (Chueca et al. 2005). The continuous glacier retreat since then has been analyzed in several works (Martínez de Pisón and Arenillas 1988, Gellatly et al. 1995, Copons and Bordonau 1997, Julián and Chueca 1998, René 2000, Chueca and Julián 2002, Chueca et al. 2002, 2003) and revealed an overall loss of about two third of the LIA extent until the end of the 20th century. Extensive studies based on dated moraines, iconographic sources, topographic maps, aerial and terrestrial photographs have been carried out on the evolution of Maladeta Glacier in the central Spanish Pyrenees (Chueca et al. 2005, and references therein). Based on these works, the glacier lost about 36% of its LIA area (c. 1.5 km^2) until the year 2000, with an increase of its equilibrium line altitude of 255 m. The continuous glacier retreat was more pronounced in the second half of the 19th century and in the last two decades of the 20th century. Continuous mass balance measurements since 1991/92 show a close to zero balance until 1997 followed by a mean annual loss of about 1 m w.e. a^{-1} until present (Fig. 2). The glacier retreat since the LIA is generally attributed to the increase in (summer) temperature over this period and scarce (winter) precipitation during periods of accelerated glacier ice loss (Chueca et al. 2005, López Moreno 2005).

-Others, e.g. Apennines & Tatra Mountains

Small glaciers of the (French-Italian) Southern Maritime Alps (Gellatly et al. 1994a, Pappalardo 1999), the Central Apennines in Italy with the Ghiacciaio del Calderone (Gellatly et al. 1994b, D'Orefice et al. 2000), and of the Spanish Sierra Nevada with the Corral del Veleta Glacier (Messerli 1980, Gómez Ortiz and Salvador 1997) are reported to showing observed trends of constant retreat during the 19th and 20th century associated with prolonged periods of negative mass balances similar to the ones of glaciers in the Pyrenees (cf. Chueca et al. 2005). Perennial snow patches in the Tatra Mountains have been mentioned already back in the early 17th century with detailed descriptions and measurements starting in the early 20th century (Gadek 2008, and references therein). In the Slovak Tatras, the largest glacieret is situated in Medená Kotlina and is strongly influenced by avalanche accumulation and shading of the surrounding peaks (Gadek and Kotyrba 2007). From three of glacierets in the Polish Tatras (Mięguszowiecki, Pod Bulą and Pod Cubryną), annual measurements of their extents have been carried out since 1980 (Wiśliński 1985, 2002, Ciupak et al 2005). From these analyses, it was concluded that the fluctuations of these three mainly avalanche and snow-drift fed glacierets were usually not synchronous and do not show any trend, that changes in their dimensions are strongly connected to the destruction of subglacial tunnes, and that these inter-annual variations cannot be explained by changes in air temperature. Gadek (2008) showed that the fluctuations of these firn-ice patches depend most of all on the weather regime of the winter season and local topographic conditions. Unlike the firn-ice patches in the Tatras, the inter-annual fluctuations of the two investigated glacierets (Snezhnika and Banski suhodol) in the Bulgarian Pirin Mountains seem to be mainly related to variations in (summer) temperature, and precipitation (Gachev et al. 2009).

Projections *(21st century)*

Over the 20th century, glaciers have dramatically retreated from their LIA moraines which mark Holocene maximum extents in all European regions. Today, glaciers are close to the 'short end' of the Holocene variability, and may have already passed it in some regions of the Alps and in the Pyrenees. While coastal glaciers on the Scandinavian Peninsula were able to regain some mass in the last decade of the 20th century, the vast ice loss over the past few decades has already led to the disintegration of many glaciers (e.g., Carèser, IT) within the Alpine observation network (Carturan and Seppi 2007, Paul et al. 2007). The massive downwasting of many glaciers, rather than dynamic retreat, has decoupled the horizontal extent (i.e., length, area) of these glaciers from current climate. Under present climate change scenarios for the 21st century (IPCC 2007, Chapter 3.7), the ongoing trend of rapid, if not accelerated, glacier melting on the century time scale may lead to the deglaciation of large parts of many mountain ranges in the coming decade. However, while the process understanding of glacier reaction to (further) climatic changes is well developed, most available analysis are rather sensitivity studies of selected glaciers and forcing scenarios than standardised glacier ensemble projections for entire Europe.

From degree-day model and worldwide glacier samples (e.g., Oerlemans and Fortuin 1992, Gregory and Oerlemans 1998, Braithwaite and Zhang 1999), it is shown that the (static) sensitivity of mass balance to a +1°C temperature increase shows a global mean sensitivity of about -0.35 m w.e. a^{-1} K^{-1} (e.g., Raper and Braithwaite 2006) with a wide range from decimetres to a few meter water equivalent. Thereby, sub-polar glaciers (e.g., in Svalbard and northern Scandinavian Peninsula) have lower temperature sensitivities and more maritime (e.g., in Iceland and coastal Norway) and tropical glaciers have higher sensitivities.

De Woul and Hock (2005) use such a degree-day approach to investigate the sensitivity of 42 Arctic glaciers and ice caps. The authors confirm earlier findings (see above) and show that on average an increase in precipitation of about 20% is needed in order to offset the effect of a one degree warming. Thereby, much higher percentage increases in precipitation are needed for continental than for maritime glaciers. The mean (static) mass balance sensitivities to a one degree warming (and corresponding contributions to sea level rise) of investigated glaciers are -0.45 m w.e. a^{-1} (0.045 mm a^{-1}) for Svalbard, -0.74 m w.e. a^{-1} (0.006 mm a^{-1}) for the Scandinavian Peninsula, and -1.63 m w.e. a^{-1} (0.049 mm a^{-1}) for Iceland.

The sensitivity of the Vatnajökull Ice Cap to a warming over the 21st and 22nd centuries is examined by Flowers et al. (2005) using spatially distributed coupled models of ice dynamics and hydrology. For a prescribed warming rate of 2°C per century, simulated area and volume of the ice cap are reduced by 12–15% and 18–25%, respectively, by the end of the 21st century. As a consequence, glacier discharge from northern and northwestern Vatnajökull (distal from the coast) appears to be the most robust to climate warming, while discharge from Vatnajökull's southern margin (proximal to the coast) is particularly vulnerable and has implications for glacier flood routing and frequency. With a similar modelling experiment, Aðalgeirsdóttir et al. (2006) found an ice volume reduction by half of Hofsjökull and southern Vatnajökull by the first half of the 22nd century forcing their transient model with a warming rate of 1.5 and 3.0 °C per century in midsummer and midwinter, respectively.

The (flat) ice caps on the Scandinavian Peninsula are especially vulnerable to a rise in the equilibrium line altitude (ELA) due to their relatively small altitudinal difference between the present-day ELA and the maximum elevation of the individual (outlet) glacier. Oerlemans (1997) used a dynamic ice-flow model of Nigardsbreen and showed that a continued warming rate of 2°K a^{-1} would reduce the ice volume of this outlet glacier from Jostedalsbreen ice cap by more than 90%. Nesje et al. (2005) show that an increase in summer temperature of 2.3°C and an increase in winter precipitation of 16% by the end of the 21st century would cause the steady-state ELA to rise 260 m. As a result, about 98% of glaciers in Norway (including seven of the 34 largest) are likely to disappear and the overall glacier area may be reduced by c. 34%. The Hardangerjøkulen in southern Norway is expected to disappear for a linear temperature increase of 3°C until the end of the 21st century based on a spatially distributed mass balance model coupled to a vertically integrated ice-flow model by Giesen and Oerlemans (2010).

For Storglaciären in the Swedish Kebnekaise massif, typical (static) mass balance sensitivities to changes in temperatures and precipitation (see above) are found by Brugger (1997), Schneeberger et al. (2001), and Radic and Hock (2008). According to the latter, projections of volume change in the 21st century driven by the B2 emission scenario (cf. CHAPTER) from statistically downscaled regional and global climate model outputs result in a volume loss of 50–90% of the glacier's initial volume by end of the 21st century. From a model comparison, Hock et al. (2007) suggest that the total ice loss might be underestimated by temperature-index models as compared to detailed energy-balance approaches.

For the Alps, Zemp et al. (2006, 2007) used a distributed model of the ELA to simulate the glacierisation over the entire mountain range for the reference period of 1971-90 and climate change scenarios for the 21st century. They find that a summer temperature increase of +3°C (which corresponds to about +2°C from present days) would reduce the total Alpine glacier area of the reference period by 80% (Fig. 4.7). In the event of a 5°C warming, the Alps would become almost completely ice-free with only the thickest and highest reaching glaciers being able to survive into the 22nd century. In order to offset a 1°C warming, a precipitation increase of more than 25% would be needed. Earlier simple but robust estimates (Haeberli and Hoelzle 1995) as well as recent more sophisticated modeling approaches of individual glaciers (Huss et al. 2007, Le Meur et al. 2007, Jouvet et al. 2009, Huss et al. 2010b) and glaciated drainage basins (Huss et al. 2008) generally confirm these results, with some differences in the individual glacier melt rates. The processes of glacier ablation are modeled with high accuracy in most simple and more complex approaches, whereas the spatial distribution of glacier accumulation and spectral albedo require still improvement (Machguth et al. 2006, 2009).

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4.3 Permafrost

Key messages

- In Europe, permafrost is a widespread phenomenon in the European sector of the Arctic as well as in the alpine high mountain environments. In most of these regions the permafrost is “warm” (close to 0 °C). In general Permafrostis sensitive to temperature changes and/or to changes in local permafrost controlling conditions such as snow cover, ice content and vegetation.
- Changes in spatial extent, thickness and temperature of permafrost are recognised as indications for climate change.
- A warming of the permafrost in the northernmost part of Europe of 0.5-1 °C was observedin the periode 1998-2008. In the European Alps trends are less clear compared to northern Europe and masked by the high annual variations resulting from varying snow conditions and modulated by heat exchange in warm permafrost close to 0 °C.
- The projected temperature increase (by 2100 4° C in the Alps, 4-6° C in Svalbard) will very likely lead to continued warming and thawing of permafrost.
- Increasing permafrost temperatures are expected to contribute to enhanced destabilization of slopes. They can change the frequency and magnitude of rock falls, debris flows, and thus influence the safety and maintenance of constructions and infrastructure, especially in alpine permafrost environments with high ice contents.
- Permafrost science is a relatively young research field and therefore only a limited number of continuous and long-term data series is available in Europe. The challenge lies in the ongoing and assured monitoring, for research purposes as well as for hazard assessments.

Key graphs:

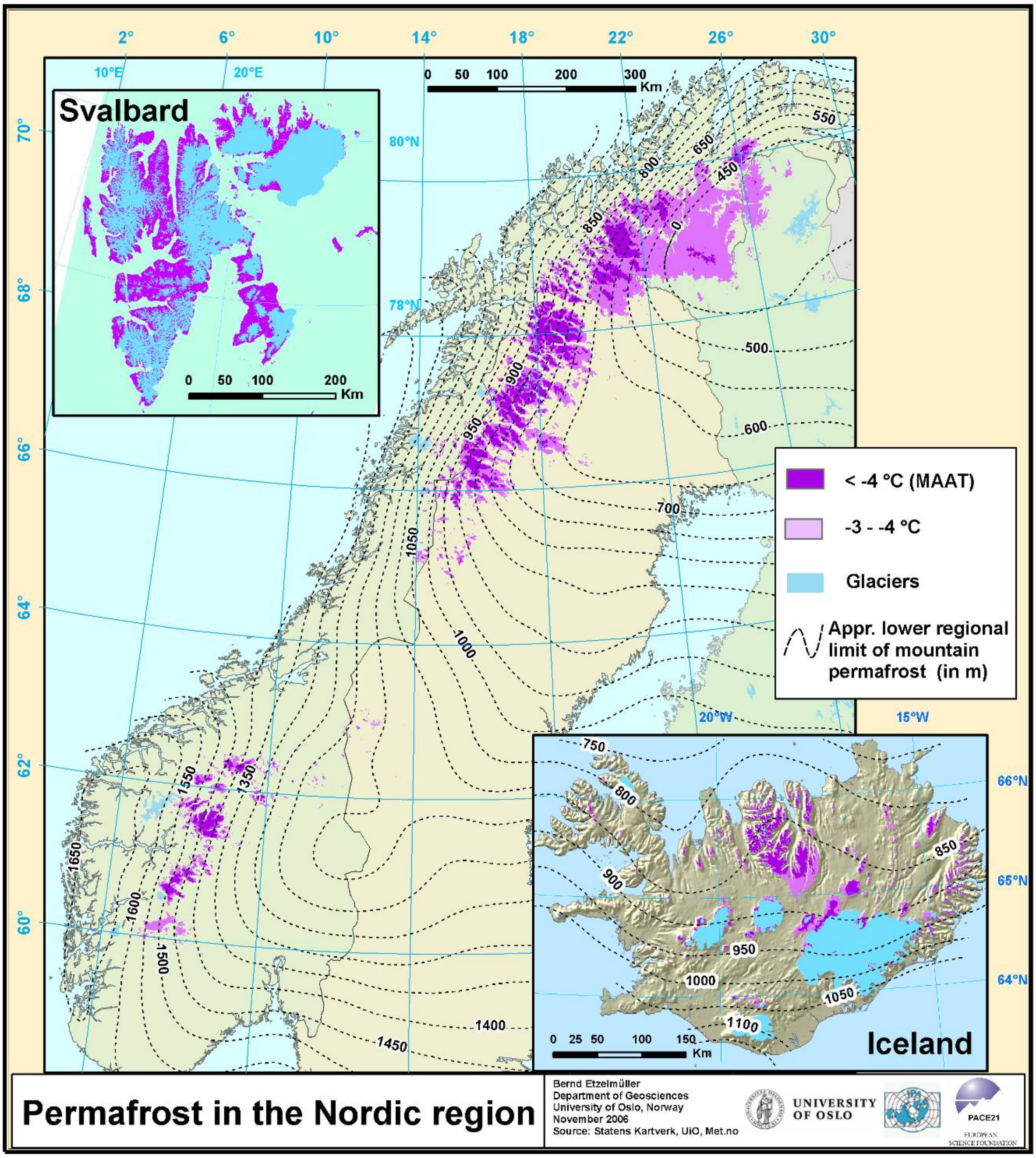

Figure 4.8: Modelled permafrost distribution in the Nordic region based on a relation between gridded air temperature data and permafrost existence, and not considering snow and topographic heterogeneity. The dashed contour lines indicate the regional lower limit of discontinuous mountain permafrost.

Source **:** Harris et al., 2009.

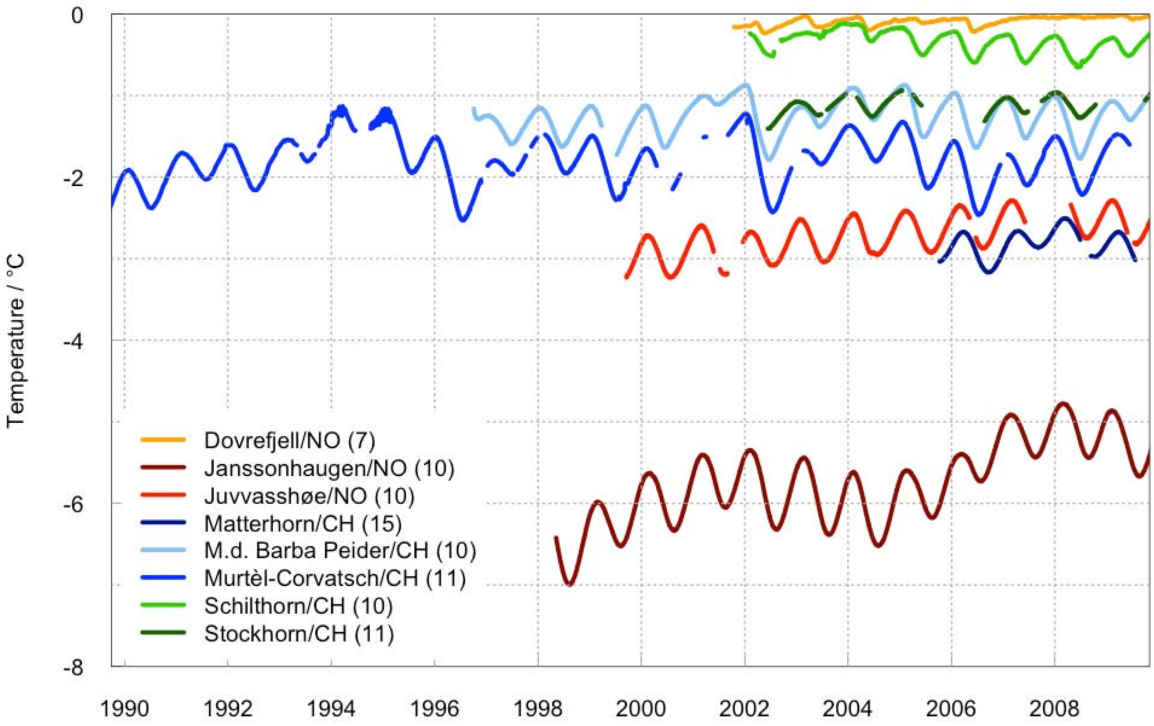

Figure 4.9: Permafrost borehole temperatures at around 10m depth from the Alps (Switzerland), Norway and Svalbard (Janssonhaugen).
Source : Haeberli et al., 2010; in press; data: PERMOS and met.no.

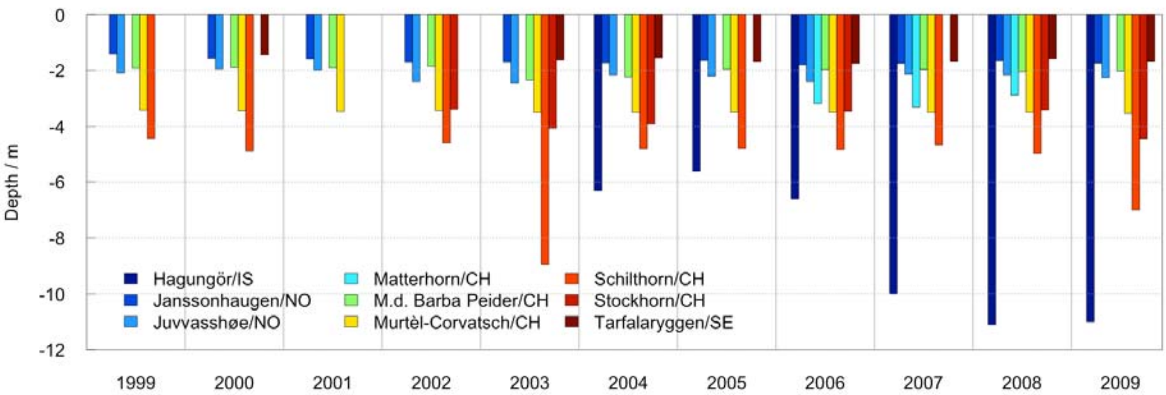

Figure 4.10: Maximum active layer thickness from different borehole sites in the Alps (Switzerland), Iceland, Norway and Svalbard (Janssonhaugen). The active layer is the upper layer of the ground, which is thawing in summer (and refreezing in winter).
Source : J. Nötzli, 2010; data: PERMOS, met.no and B. Etzelmüller.

Relevance

Permafrost is a thermally defined phenomenon describing ground (bedrock or sediment) that remains at or below 0°C for at least 2 years, irrespective of the presence of water or ice (ipa.arcticportal.org). Besides the rock matrix, permafrost involves ice, air and often unfrozen water in pore spaces or rock discontinuities. Permafrost is influenced by climate, topography and ground conditions. While climatic conditions determine the pattern of thermal subsurface conditions on a global scale, topography becomes more important on a regional to local scale affecting e.g. air temperature by elevation and solar radiation by insulation angle and shading, whereas on the local scale, surface (e.g. snow cover) and subsurface conditions (e.g., ice content) influence the transition of the climate signal into the ground temperatures (e.g., Harris et al. 2009). As permafrost is a thermal system with slow response to changing atmospheric conditions, the present state of permafrost is in part a function of former climatic conditions, and present day climate changes will affect the future thermal state of permafrost (e.g., Noetzli & Gruber 2009).

The currently most comprehensive permafrost map covering entire Europe is the "Circum-Arctic Map of Permafrost and Ground-Ice Conditions" of the International Permafrost Association (Brown et al. 2001). Permafrost regions are traditionally divided into several zones based on estimated geographic continuity in the landscape. A typical classification recognizes continuous permafrost (underlying 90-100% of the landscape); discontinuous permafrost (50-90%); and sporadic permafrost (0-50%) (ipa.arcticportal.org). In addition, permafrost distribution maps exist for sub-regions or single nations (e.g., Scandinavia, Iceland, Switzerland), based on diverse regional modeling and mapping approaches. Here, an example is given for the Nordic countries, which is based on the relation between gridded air temperature data and permafrost existence and does not consider snow and topographic heterogeneity (Figure 4.8; Harris et al. 2009).

Detailed information on permafrost existence, ground surface temperatures, subsurface temperatures, and active layer thickness is derived from a number of deep boreholes (of at least 20m depth). The active layer is the upper layer of the ground, which is thawing in summer. The thickness of this layer is changing from year to year, depending on snow and temperature conditions, as well as on subsurface characteristics (PERMOS 2009). Summer surface temperatures at borehole sites in the Nordic countries are significantly lower than those in the Alps, so that active layers are thinnerin the former than in the latter, whilst other factors remain equal (Harris et al. 2009; see Figure 4.10). The first permafrost borehole in blocky material in Europe was drilled in 1987 in the Murtèl rockglacier in Switzerland, and is the longest data record in the relatively young field of permafrost research in Europe. A regional approach was initiated at the end of the 1990ies by the PACE (Permafrost and Climate in Europe) EU project, where a north-south transect, from Svalbard to southern Spain, of boreholes was drilled and instrumented for continuous monitoring (Harris et al. 2001, 2009). In addition to the thermal characteristics, geomorphological processes related to the existence of permafrost (e.g. permafrost creep, rock fall) are documented and monitored systematically in some regions, e.g. in Switzerland (von der Mühll et al., 2008; PERMOS 2010). The importance of a systematic, long-term documentation and investigation of permafrost is supported by the first results (Figure 4.9), showing that a regional warming in Svalbard and Scandinavia of 0.5-1.0 °C is recorded from the boreholes during the last decade (Isaksen et al. 2007b; Christiansen et al. 2010). Measurements in the Alps also show a slight warming trend at several measurement sites (Figure 4.9) with more or less immediate response in active-layer thickness to extreme temperature events such as in summer 2003 (Figure 4.10). Other sites, however, show no warming trends or even reverse signals at different depths (Zenklusen Mutter et al., 2010; Haeberli et al.in press). Trends are, therefore, less clear compared to Svalbard and Scandinavia and masked by the high annual variations resulting from varying snow conditions. In addition, several sites are likely modulated by latent heat exchange in warm permafrost close to 0 °C. The importance of short-term extreme thawing events is also emphasized (Harris & Isaksen 2008). While direct responses to extreme annual temperature variations is anticipated in the near surface temperatures and active layer thickness, thermal profiles extending to several decameters below the depth of zero annual amplitude reflect changes over many decades or centuries (Harris & Isaksen 2008, PERMOS 2010). Permafrost will last for centuries to millennia, where it is continuous and thick, while it may get reduced significantly where it is thin and where the geothermal heat flux is high as is the case in Iceland. Increasing permafrost temperatures approaching 0° C are likely influencing the stability of the ground and the potential for natural hazards such as rock falls, debris flows and rock slides (Noetzli et al. 2003, Gruber & Haeberli 2007, Harris et al. 2009). This is especially important in densely populated permafrost areas within Europe, e.g. the Alps and the fjord areas of north-western and northern Norway.

Results from the International Polar Year (IPY) circumarctic permafrost thermal snapshot show that the warmest high arctic permafrost exist in the European sector of the Arctic in Svalbard, where permafrost temperatures at around 10 m depth is as high as -2.3 °C on the west coast. In e.g. Eastern Greenland temperatures at 3 m depth is -8 °C representing most other high arctic sites, where permafrost temperatures range from -5 °C to -15 °C (Romanovsky et al. 2010).

Past Trends

Permafrost in the Nordic Countries

In the Nordic countries (Norway, Svalbard, Sweden, Finland and Iceland) permafrost is widespread, and ranges from continuous in Svalbard to wide-spread discontinuous permafrost in high-mountain regions of Iceland and the Scandinavia to sporadic patches related to palsas and peat plateaus, especially in Iceland and northern Scandinavia (Christiansen et al. 2010, Harris et al. 2009; Figure 4.8). Compared to other regions in the Arctic at similar latitudes, less permafrost is present due to the influence of the North Atlantic Ocean, which also explains the present relatively warm permafrost in this region (Christiansen et al. 2010). Several studies exist, and during the International Polar Year (IPY) 2007-2009 under the international project Permafrost Observatory Project: a contribution to the Thermal State of Permafrost (TSP), the distribution and thermal state of permafrost in the Nordic countries was studied in several national projects and a thermal snapshot presented (Christiansen et al. 2010).

In general the lower limit of mountain permafrost in Scandinavia decreases from the western coast towards the more continental conditions in eastern Norway and western Sweden, and rises again towards the Baltic Sea (see Figure 4.8). In northern Norway the lower permafrost limit is located at 800-900 m a.s.l. in the coastal areas of the Troms region and decreases to 600-700 m a.s.l. in the more continental parts (Isaksen et al. 2004), while it is at about 400-500m a.s.l. in coastal Finnmark (Farbrot et al. 2008, Isaksen et al. 2008). In southern Norway the transition zone of mountain permafrost decreases from 1300-1550 m a.s.l. in western parts (Jotunheimen, Dovrefjell) to 900-1100 m a.s.l. in eastern parts (Sølen and Elgåhogna) (e.g. Heggem et al. 2005, Isaksen et al. 2002, Sollid et al. 2003). In Jotunheimen, continuous permafrost monitoring was initiated with the PACE borehole at Juvasshøe (1894m a.s.l., 129 m deep). The borehole data indicate a permafrost thickness of more than 300 m and show low-temperature gradients (Isaksen et al. 2001). The temperature at 9 m depth was about -2.4 °C in 2008/2009, with a mean temperature about 0.5 °C higher than at the start of monitoring 10 years earlier (Isaksen et al. in revision). The active layer depths at this site were 20% greater in the 2003, 2004 and 2006 summers than in the previous years. On the northernmost PACE drill site on Janssonhaugen, Svalbard (275 m a.s.l., 102 m deep) snow cover is usually thin or absent, surfaces are normally dry and the soil water content low, so that the active layer thickness is well correlated with local summer air temperatures (Isaksen et al. 2007b).

In the Tarfala valley, northern Sweden, discontinuous permafrost was found up to 1200 m a.s.l., above which permafrost is continuous (Hauck et al. 2001, King 1984). Data from the PACE borehole here (Tarfalaryggen, 1550 m a.s.l., 100 m deep) indicate a permafrost depth of more than 300 m (Isaksen et al. 2001); the permafrost temperature at the depth of the zero annual amplitude (18-19m) was around -2.4 °C between 2007 and 2009 (Christiansen et al. 2010).

In Svalbard, permafrost is continuous outside the 60 % glacier-covered areas (Humlum et al. 2003). In general the mountain permafrost in Svalbard is of Weichselian age (the Weichselian glaciation started 115 ka BP (thousands of years before present) and ended at the transition to the Holocene 11.5 ka BP) and is about 400-500 m thick, while that in the lowlands dates from the Holocene with a thickness of about 100 m (Humlum et al. 2003, Liestøl 1976). Permafrost temperatures at 10-15 m depth in the Adventdalen area vary from -3.2 °C in a solifluction sheet, to -5.2 °C at the mountain top of Janssonhaugen and -5.6 °C in a loess terrace (Christiansen et al. 2010). Borehole monitoring at Janssonhaugen indicate that the permafrost has warmed considerably during the last decade. Significant warming is detectable down to at least 60 m depth and present decadal warming rate at the permafrost table are in the order of 0.07 °C yr^{-1},(Isaksen et al. 2007b, see also Figure 4.9).

In Iceland, permafrost is probable above elevations of 800-1000 m a.s.l.; the lower limit of mountain permafrost decreases from southeast to northwest (see Figure 4.8). Below this limit, widespread palsa fields at elevations above 600 m a.s.l. indicate the presence of sporadic permafrost (Etzelmüller et al. 2007, Farbrot et al. 2007). The occurrence of relict rockglaciers reaching low elevations, demonstrate that permafrost had a wider spread during the Holocene. Local studies on the interaction of volcanoes and a climate that favours permafrost show that a thin layer of tephra is sufficient to reduce the sub-tephra snow ablation substantially, possibly even to zero, causing aggradation of the surface and preserving massive ground ice and permafrost (Kellerer-Pirklbauer et al. 2007).

Permafrost in the Alps

Several studies carried out in the Alps have shown permafrost distribution above approximately 2500m a.s.l. In Austria, up to 2000 km^2 are potentially underlain by permafrost (Lieb 1998). A large number of rockglaciers, typical permafrost landforms, are mapped in central and eastern Austria, underlining the past and present permafrost relevance in that region. In Switzerland, about 5 % of the land surface (about 2100 km^2) is underlain by permafrost (www.bafu.ch).

In general, ground temperatures are only a few degrees below zero and permafrost may be thin near the lower permafrost boundary (Harris & Isaksen 2008, PERMOS 2010). Due to complex topography and spatially highly variable substrate properties, a much larger spatial and temporal variability of permafrost temperatures is recorded in the Alps, compared to the Nordic area (Gruber & Haeberli 2009). The ground thermal field is strongly three-dimensional (Noetzli et al. 2007) and the thermal offset between the mean ground surface temperature and mean permafrost table temperature reflects spatial heterogeneity in the active layer composition and snow distribution (e.g. Gruber & Hoelzle 2008). Beside a number of modeling approaches, systematic measuring and monitoring of permafrost started about 10 years ago with initiating the "Swiss Permafrost Monitoring Network" (PERMOS) and later on also similar projects in other Alpine countries (→ see BOX 4.4).

Permafrost temperatures at depths around 10 m in Swiss boreholes are given in Figure 4.9. The permafrost temperatures show a great range between -0.2 °C at Schafberg (2750 m a.s.l.) and -3 °C at the Matterhorn (3295 m a.s.l.). At Murtèl-Corvatsch, a marked warming was recorded between 1987 and 1995, but this was reversed in the following winters mainly due to the influence of the snow cover (Harris & Isaksen 2008). Since then, no marked overall permafrost temperature trend is apparent at Murtèl and other Swiss boreholes, revealing the importance of both air temperatures and snow depths as regulators of ground temperatures at the regional scale (PERMOS 2010). However, the impact of short-term extreme events became visible. Such events, as e.g. the hot summer of 2003 (with air temperatures 3 °C higher, than the average of 1961-1990), showed considerably different reactions in the Alpine boreholes, due to the diverse ground conditions. At the Murtèl-Corvatsch borehole in ice-rich frozen debris, the active layer thickness ranges between 3.1 and 3.5m over the entire period, with a slight trend towards increasing depth (Figure 4.10). In contrast to that, the data from ice-poor bedrock of the Stockhorn and Schilthorn mountains show a clear thickening in 2003 (and also in 2009), indicating heat conduction coupled with possible advective heat transfer by water (Gruber et al. 2004). The depth of thaw penetration at Schilthorn in 2003 was around 9 m and twice the average of the previous years (PERMOS 2009). In this context, geophysical soundings at the end of the summer (August – September) are of great importance in order to detect and monitor changes in ice- and water-content in the ground (Hilbich et al. 2008).

In addition to the thermal monitoring, geomorphological processes related to the existence of permafrost (e.g. permafrost creep, rock fall) are documented and monitored systematically in several Alpine regions (PERMOS 2010). This was emphasized with the detection of strong temporal variations in creep of rockglaciers, which are permafrost landforms (Barsch 1996, Haeberli et al. 2006). Observations showed that most surveyed Alpine rockglaciers, whatever their size and velocity, respond sensitively and almost synchronously to interannual and decadal ground temperature changes (Roer et al. 2005, Kääb et al. 2007, Delaloye et al. 2010). Many Alpine rock glaciers are located in the vicinity of the lower limit of discontinuous permafrost and display temperatures close to 0° C (between -2 and 0 °C). Such "warm" rockglaciers are expected to react much more to even small temperature changes compared to "colder" rock glaciers. In addition, some landforms indicate destabilization by the formation of deep trenches or the collapse of entire rockglacier tongues (Roer et al. 2008).

Permafrost in other mountain regions in Europe

Beside the main permafrost areas in the Arctic and in the Alps, some marginal permafrost exists in other high mountain regions, such as the Tatra Mountains and the Sierra Nevada. When permafrost has a marginal character, it may be especially susceptible to react to slight climatic changes. In the Slovak Tatras, small glaciers are studied (see also chapter4.2) and in that context buried ground ice was investigated at one site, evidencing the occurrence of permafrost, which was not determined in the Carpathian-Balcanic region before (Gadek & Kotyrba 2007). Europe's southernmost permafrost remnant was detected in south-east Spain in the Sierra Nevada; also here, buried ice was detected and permafrost existence was verified by temperature measurements (Gómez et al. 2001).

Significance of observed changes

The presented data show that relatively warm permafrost dominates in Europe, from the discontinuous alpine mountain permafrost in the Alps to the continuous permafrost in the Arctic north in Svalbard. This indicates the potential for the permafrost to respond to climate change and corresponding impacts in surrounding landscapes, but the response is often rather complex. Hence, interpretations have to be made with caution, due to the lack of long data series and the extreme heterogeneity at the different sites.

Changing active-layer thicknesses are a direct response to annual climate conditions and show interannual variations between +20 % in Svalbard (Janssonhaugen) and +100 % in the Alps (Schilthorn), depending on site conditions. On Svalbard significant near-surface warming was reported by Isaksen and others (2007a); it resulted from a remarkable temperature anomaly during winter and spring 2005–2006. Mean ground temperature at the permafrost table during 2006 were 1.8 °C higher than the mean for the previous six years. In the Alps, extreme events such as the summer 2003 had a major impact. A relatively direct response was recently also observed in the spatio-temporal variation of rockglacier velocities in the Alps (Delaloye et al. 2010).

Permafrost temperature records from the three northernmost PACE boreholes in Sweden, Norway and Svalbard suggest rapid recent warming, rates at the permafrost table being estimated as 0.04–0.07 °C yr^{-1} (Isaksen et al. 2007b). In the Swiss Alps bedrock borehole time series are strongly affected by short term seasonal extremes such as the cold winter of 2001–2002 and the hot summer of 2003. However, at depth there is evidence for warming, at Stockhorn, for instance, rates are around 0.01 °C yr^{-1} at 48.3m depth, and at Schilthorn, similar or slightly higher warming rates are indicated (Harris et al. 2009, Haeberli et al. in revision, PERMOS 2010). Trends observed in the ice-rich frozen debris at Murtèl–Corvatsch indicate significant warming over the past 20 years, but the record is largely affected by the influence of snow depth and duration rather than atmospheric temperatures (Harris et al. 2009).

Projections

According to recent model calculations based on the regional climate model REMO and following the IPCC SRES-Scenarios A1B, A2 and B1, a warming of up to 4 °C by 2100 compared to the 1970-2000 period is projected for the Alpine region in Europe (Jacob et al. 2007, see also introduction and chapter 3.8). For Svalbard, a 4-6 °C warming and +5 % precipitations increase is projected by 2100 according to the SRES A1B emission scenario (Benestad 2005). A limited number of specific studies on past and projected permafrost conditions exist, based on measurements and modeling approaches. For mountain permafrost, Noetzli & Gruber (2009) showed that the past climate variations that essentially influence present-day permafrost temperatures at depth are the last glacial period and the major fluctuations in the past millennium. Projected future warming, however, is likely larger than that from past climate conditions because larger temperature changes at the surface occur in shorter time periods (Noetzli & Gruber 2009). A modeling approach performed for Svalbard showed a gradually permafrost warming by 1° C between 1850 and 1990, and since then by 0.5° to 1°C (Etzelmüller et al. 2010). For the future, a significant increase in ground temperatures and an increase of active layer thickness, depending on soil characteristics, is predicted (Etzelmüller et al. 2010). They suggest that a major degradation of the continuous permafrost on Svalbard is not expected during the next c. 50

years. However, permafrost degradation can be expected at low elevation, e.g. close to the coast. In general, changes in Europe's permafrost are likely to continue in the near future and the majority of permafrost bodies will presumably experience warming anddegradating.
The rise in temperature and thawing permafrost could increasingly destabilize mountain slopes and increase the frequency of rock falls, posing problems to mountain infrastructure and communities (Gruber et al. 2004). The warming and thawing of permafrost in bedrock can sometimes be rapid and failure along ice-filled joints can occur even at temperatures below 0 °C (Davies et al. 2001). Water flowing along linear structures and the advection of heat along joints in the ground will further accelerate destabilization (Gruber & Haeberli 2007). Over much longer timescales, permafrost warming may lead to a rise in the lower permafrost boundary, decreasing permafrost thickness by bottom-upward thawing, and hence increase the risk of large, deep-seated landslides (Harris et al. 2009). In order to develop risk assessment strategies, long and continuous data series are to be combined with process-based modeling approaches. Thus, there is a need for continued integrative research of permafrost scientists, climatologists, geomorphologists and engineers.

BOX 4.4: Permafrost Monitoring and archiving

→ PERMOS (www.permos.ch)
The main goal of the Swiss Permafrost Monitoring Network PERMOS is the systematic, long-term documentation and investigation of permafrost in the Swiss Alps with suitable parameters and techniques. The network was initiated as a research-oriented network in the 1990ies, and officially started with a 6-year pilot phase in 2000. Since 2007 the networks is formally implemented with a coordination office and secured long-term funding by the Swiss GCOS Office and the Federal Office for the Environment and the Swiss Academy of Sciences. The network is based on three types of observations: (1) ground temperatures measured in boreholes and at the surface near to the drill site, (2) changes in subsurface ice and water content at the drill sites by geo-electrical surveys, and (3) velocities of permafrost creep determined by geodetic surveys and/or photogrammetry. In addition, standardized documentation of fast mass movements from permafrost areas (e.g., rock fall) is being established. Detailed information on the monitoring strategy, key test sites, partners, and funding institutions is available on the PERMOS website (www.permos.ch).
→ NORPERM (www.ngu.no/norperm)
NORPERM – The Norwegian Permafrost Database was developed at the Geological Survey of Norway during the International Polar Year (IPY) 2007–2009 as the main data legacy of the IPY research project "Permafrost Observatory Project: A Contribution to the Thermal State of Permafrost in Norway and Svalbard" (TSP NORWAY). NORPERM follows the IPY data policy of open, free, full and timely release of IPY data, and the borehole metadata description follows the Global Terrestrial Network for Permafrost (GTN-P) standard. The purpose of NORPERM is to store ground temperature data safely and in a standard format for use in future research. NORPERM stores temperature time series from various depths in boreholes and observations on air temperature, snow cover, ground-surface or near-surface temperatures recorded by miniature temperature data-loggers, and temperature profiles with depth in boreholes obtained by occasional manual logging. It contains all the temperature data from the TSP NORWAY research project and from some other and pre-IPY permafrost research projects in Norway and Svalbard, totalling 32 boreholes and 98 sites with miniature temperature data-loggers for continuous monitoring of micrometeorological conditions, and 6 temperature depth profiles obtained by manual borehole logging (→ More details in Juliussen et al. 2010).
→ GTN-P (www.gtnp.org)
Together with other monitoring and research initiatives, data are implemented into the Global Terrestrial Network of Permafrost GTN-P, which aims at organizing and managing a global network of permafrost observatories for detecting, monitoring, and predicting climate change.

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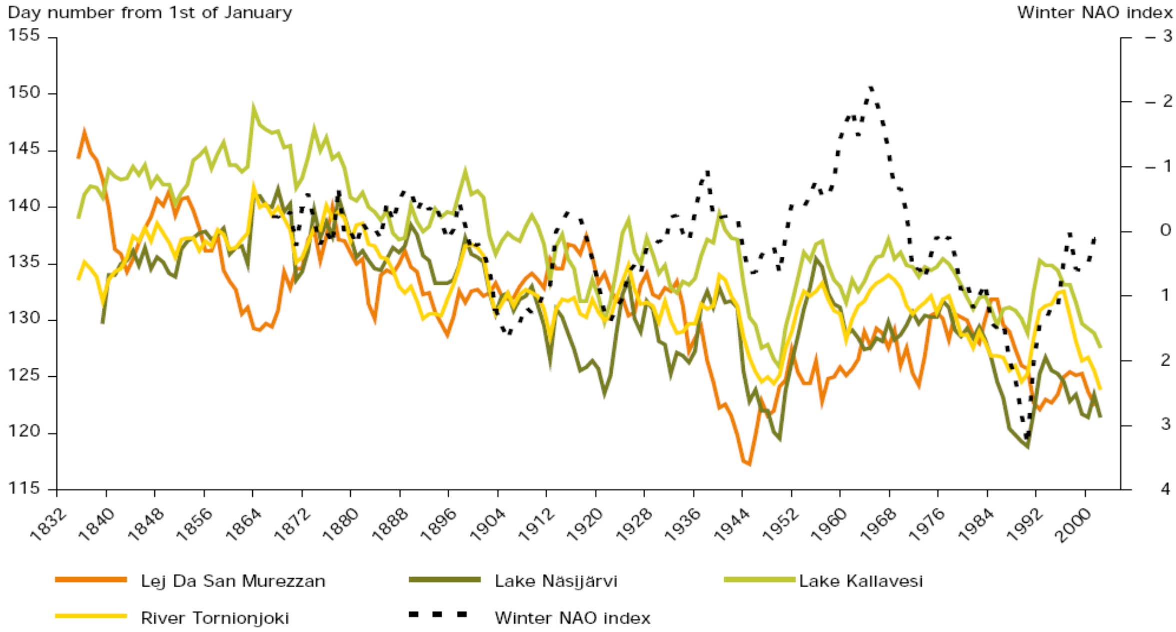
4.4 Lake and River ice

Key messages:

- The duration of ice cover in the northern hemisphere has shortened at a mean rate of 12 days per century, resulting from an average 5.7 days later for ice-on and 6.3 days earlier for ice-off.
- The ice cover of lakes in the temperate region where the ice season is already short or ice cover only occurs in cold winters is much more affected by the observed changes in temperature than the ice cover of lakes in colder regions such as northern Scandinavia.
- Inter-annual fluctuations in the timing of ice-off are highly correlated over very large spatial scales, and are often dominated by climate modes such as the North Atlantic Oscillation.

Key graph:

Figure 4.11: Ice break-up dates from selected European lakes and rivers (1835-2006)



Note: Data smoothed with a 7-year moving average. See Box 5.1 'Atmospheric circulation patterns in Europe'.

Source: Benson and Magnuson 2000 (Updated to 2006 by J. Korhonen, Finnish Environment Institute (SYKE) and D. Livingstone, Water Resources Department, Swiss Federal Institute of Environmental Science and Technology (EAWAG)).

Relevance

The appearance of ice on lakes and rivers requires prolonged periods with air temperatures below 0 °C. Hendricks Franssen and Scherrer (2008) found good correspondence between the sum of negative degree days (NDD) and the probability of lake freezing.
Higher temperatures will affect the duration of ice cover, the freezing and thawing dates and the thickness of the ice cover. Air temperature is the key variable determining the timing of ice break-up (Palecki and Barry, 1986; Livingstone, 1997). Wind has generally a strong impact on timing of the ice formation on lakes and a very limited impact on lake ice break up. Other meteorological forcing variables, such as precipitation and solar radiation, generally have a more important influence than winds on the timing of lake ice break up. For rivers, the impact of wind onthe timing of freezing and thawing, is negligible (Livingstone, 2010). Here the riverdischarge affects the ice cover, giving retarded freeze-up and accelerated break-up with increasing flow and vice versa (Beltaos and Prowse, 2009).

Climatic conditions not only influence the timing and duration of the ice cover, but also the thickness of the ice cover and the nature of break-up. In years with low snowpack and/or protracted spring melt, the ice-break at rivers will mainly be thermal, characterised by extensive ablation and weakening of the ice cover prior to an increase in flow, if any. With a thick snowpack and rapid melt, break-up will be mainly mechanical, characterised by a rapidly rising flow driven into a thick and mechanically strong ice cover. This process increases the chance of ice jamming and flooding (Beltaos and Prowse, 2009).
Variations in lake and river ice phenology are relevant in terms of freshwater biology and hydrology and human activities such as winter transportation, bridge and pipeline crossings, and winter sports (IPCC, 2007).
Changes in ice cover are of critical ecological importance for lakes because of their effect on the underwater light climate (Leppäranta *et al*., 2003), nutrient recycling (Järvinen *et al*., 2002) and oxygen conditions (Stewart, 1976; Livingstone, 1993), which influence the production and biodiversity of phytoplankton (Rodhe, 1955; Phillips and Fawley, 2002; Weyhenmeyer *et al*., 1999) and the occurrence of winter fish kills (Greenbank, 1945; Barica and Mathias, 1979). There are only limited options for helping the freshwater ecosystem to adapt to changes in ice cover, however, reducing other human pressures will generally make the ecosystems more robust to cope with changing climate conditions.
In remote areas frozen rivers and lakes are often used as transportation corridors and longer ice-free periods mean reduced or more expensive access to communities and industrial facilities.
The greatest impacts of freshwater ice on human society are associated with the ice-induced floods that accompany dynamic freeze-up and break-up events. Damages caused by severe ice jams can be costly. However, in Europe there is some evidence for a reduction of ice-jam floods due to reduced freshwater freezing during the last century (Svensson *et al.,* 2006).

Large-scale, comprehensive records of river and lake-ice thickness are relatively rare.
Limited by the availability of detailed observations, most historical evaluations of changes in freshwater ice have focused on relatively simple parameters, such as the timing of freeze-up and break-up, and maximum ice-cover thickness.
In Europe particularly in Finland and Switzerland a few very long-lasting data-records exist.

Past Trends

An analysis of long (more than 150 year) ice records from lakes and rivers throughout the northern hemisphere by Magnuson et al. (2000) indicated that for a 100 year period, ice cover has been occurring on average 5.7 ± 2.9 days later (± 95 % confidence interval), while ice break-up has been occurring on average 6.3 ± 1.4 days earlier, implying an overall decrease in the duration of ice cover at a mean rate of 12 days per 100 years (Fig.4.11). These results do not appear to change with latitude, or between North America and Eurasia, or between rivers and lakes.
A few longer time-series reveal reduced ice cover (a warming trend) beginning as early as the 16th century, with increasing rates of change after about 1850. The early and long-term decreasing trend in the ice break-up dates is the result of the end of the Little Ice Age, which lasted from about 1400 to 1900 (Kerr, 1999). Increasing winter air temperatures in Europe during the 20th century were partly a result of global warming and partly the result of an anomalously positive NAO during the last part of the century. Therefore fluctuations of the NAO are definitely reflected in the European lake ice phenology (Livingstone, 2000; Livingstone et al., 2010).
Beltaos and Prowse (2009) reviewed various studies of river ice in the northern hemisphere/Arctic, showing an overall trend towards earlier spring break-up. However, there was considerable spatial variability in freeze-up date trends. In most cases, the ice season decreased, but there were several exceptions. The changes were often most pronounced in the last few decades of the 20th century. As an overall approximation, the authors suggest that the autumn and spring warming occurring in the 20th century warming has produced a 10-15 days delay in freeze-up and advance in break-up in these areas.
Studying ice cover information from 11 Swiss lakes over the last century, Hendricks Franssen and Scherrer (2008) found that the freezing frequency of lakes was significantly reduced in the past 40 years, and especially during the past two decades. The frequency of winters that produce enough cold

to freeze rarely freezing lakes has declined strongly; while there have been fewer declines in the frequency of winters with a cold sufficient to freeze frequently freezing lakes.
Ice cover of lakes in southern Sweden is more sensitive to climate change than those in the north, where mean winter temperatures are below 0°C most of the winter. A study of 196 Swedish lakes along a latitudinal temperature gradient revealed that a 1 °C air temperature increase caused an up to 35 days earlier ice break-up in Sweden's warmest southern regions with annual mean air temperatures around 7 °C. It caused only about 5 days earlier break-up in Sweden's coldest northern regions where annual mean air temperatures are around − 2 °C (Weyhenmeyer et al., 2004; Weyhenmeyer, 2007). Ice break-up in Finland has also become significantly earlier from the late 19th century to the present time, except in the very north (Korhonen, 2006).
Karetnikov and Naumenko (2008) studied long-term (1943-2006) ice cover data from Lake Ladoga, north-west Russia, the largest dimictic lake in Europe. In this case the trend analysis revealed only minor changes in ice cover. In Lake Baikal, the ice-free season has lengthened and the ice thickness decreased in the last century (Hampton et al., 2008; Shimaraev et al., 2002).
There are very few studies on historical trends in river ice-cover thickness, but there are reports on decreasing trends (Beltaos and Prowse, 2009). Even less reported is the trend in river ice jams, and the results point in different directions, underscoring the complexity of ice-jam processes (Beltaos and Prowse, 2009). However, there is some evidence of a reduction in ice-jam floods in Europe due to reduced freshwater freezing during the last century (Svensson et al., 2006).

Projections

Future increases in air temperature associated with climate change are likely to result in generally shorter periods of ice cover on lakes and rivers. The most rapid decrease in the duration of ice cover will occur in the temperate region where the ice season is already short or only occurs in cold winters, because of the non-linear dependence of the duration of ice cover on mean annual air temperature (Weyhenmeyer et al., 2004; Livingstone and Adrian, 2009). As a result, some of the lakes that now freeze in winter and that mix from top to bottom during two mixing periods each year (dimictic lakes) will potentially change into monomictic (mixing only once) open-water lakes.
By the end of the century, the ice cover is predicted to shorten dramatically in lake Baikal (Shimaraev et al., 2002; Todd and Mackay, 2003). Climate change may also change the ice quality, but it is difficult to predict the effects; a shift towards more rain rather than snow in the spring may give more cloudy ice if the rain falls on snow, but a great deal of rain produces clear ice (Moore et al., 2009).
Borsch et al. (2001) calculated the expected changes in dates of freeze-up and break-up of river ice for regions in Russia based on simple correlation with air temperature. Although rough, the calculation shows that the largest changes will occur in the most westerly parts of Russia.
Warmer autumns and higher flows would give the largest potential for river freeze-up jamming. The likelihood of break-up jamming depends both on the nature of the established ice cover and the climatic conditions during break-up. A very important factor is the snowpack in the catchment, as more intense snowmelt increases the probability of a mechanical breakup (Beltaos and Prowse, 2009). According to the IPCC (Meehl et al., 2007) winter precipitation in high-latitude regions will increase, but the corresponding effect on snowpack will depend on the present and future temperature, determining whether the additional precipitation will fall mainly as rain or snow. At present, more research is needed to be able to predict the trends in spring ice jams with a changing climate (Beltaos and Prowse, 2009).
Mid-winter melts are likely to become more frequent in a warmer climate. Intensified mid-winter thaws would enhance the severity of mid-winter breakups, but would also reduce the potential for spring jamming. More importantly, rivers that currently have a permanent ice cover will be susceptible to mid-winter ice-break-ups, which may have severe consequences, as e.g. on regional transportation (Beltaos and Prowse, 2009).

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4.5 Baltic Sea Ice

Key Messages:

- The maximum annual extent of Baltic Sea ice has decreased since the end of the 1980s.
- Since 1987 all ice winters have been average, mild, or extremely mild whereas none of them has been severe or extremely severe.
- The length of the ice season in the Baltic Sea has decreased by 14-44 days over the20th century, depending on the location.
- Ice thickness data do not show clear trends during most of the 20th century but ice thickness in the period 1990–2010 was about one third lower as compared to the previous hundred years.
- The ice extent, the date of ice break-up, and the length of the ice season aresignificantly influenced by the North Atlantic oscillation and the Arctic oscillation.
- The maximum extent of Baltic Sea ice is projected to decrease by 57%-71% in the next 100 years (depending on the emission scenario). The length of the ice season is projected to decrease by 1-2 months in the North and by 2-3 months in the central part of the Baltic in the next 100 years.

Key graph:

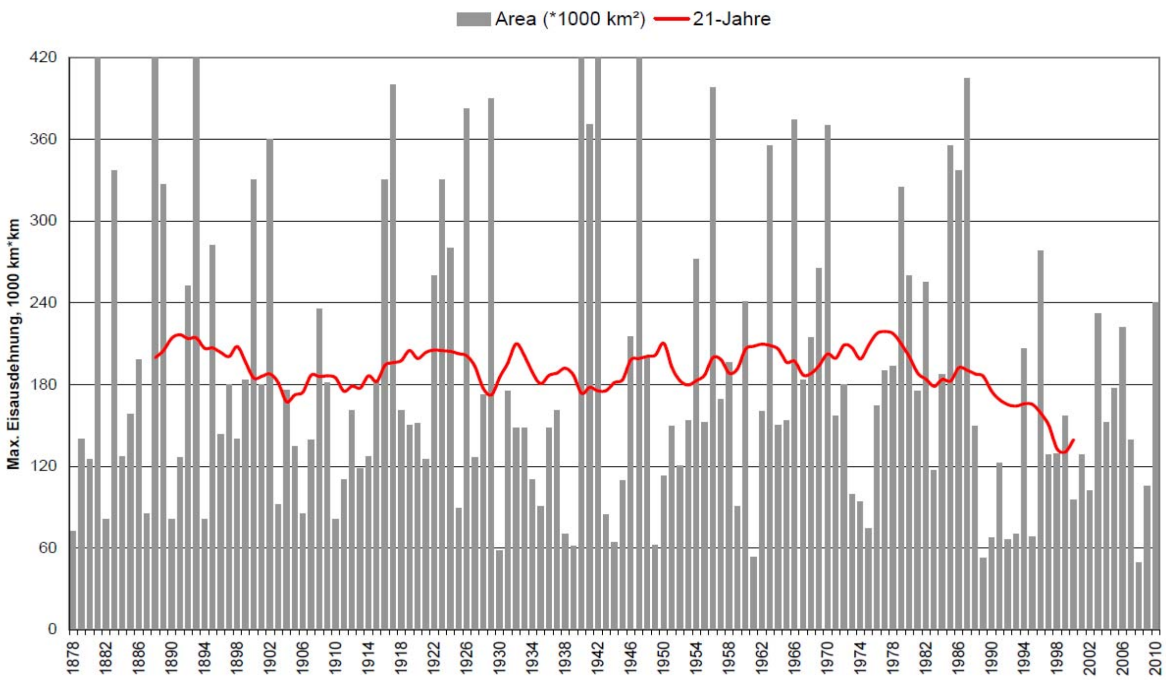

Fig.: 4.12: Annual maximum ice-extent in the Baltic Sea (1878-2010; 21-year running average)
Source: Seinä and Palosuo, 1996 and Finnish Ice Service http://www.itameriportaali.fi/en/tietoa/jaa/jaatalvi/en_GB/jaatalvi/ (updated by Schmelzer, 2010)

Relevance

Sea ice, on one hand, poses a risk to human activities in the marine environment, such as navigation, fishery and offshore operations. In areas with heavy sea marine transportation, which are abound in the Baltic Sea (in 2005 about 800 Mio tons/ year), ice may hinder navigation considerably and sometimes poses a severe threat. Also port operations and facilities may be affected by ice.

On the other hand, ice has a major impact on global climate, especially by influencing interactions between the ocean and the atmosphere. Sea-ice conditions also affect the Baltic Sea as an ecological system by influencing significantly the whole spectrum of biological circumstances, as e.g. light conditions, levels of oxygen dissolved in water, salinity, and water temperatures.
However, despite its importance, a detailed understanding of the relation between the degree of ice-coverage and the above-mentioned factors is often still an open problem in Baltic Sea oceanography.
Ice formation reacts very sensitively to local climate conditions, and hence to climatic changes. Particularly the annual extent of the Baltic Sea ice coverage correlates well with air temperatures. For this reason, long term series of ice data play an important role in monitoring climatic changes. Because of the crucial importance of ice conditions to navigation in the Baltic, ice services in the Baltic Sea region were established rather early, almost all in the 19th century. Today, there are more than 500 observation stations located in the Baltic Sea region. The data recorded include the amount and arrangement of ice, stages of ice development (ice thickness), the topography or form of ice, and navigation conditions in ice-covered sea areas. (An overview on locations of important observation stations is given in Figure 4.13).

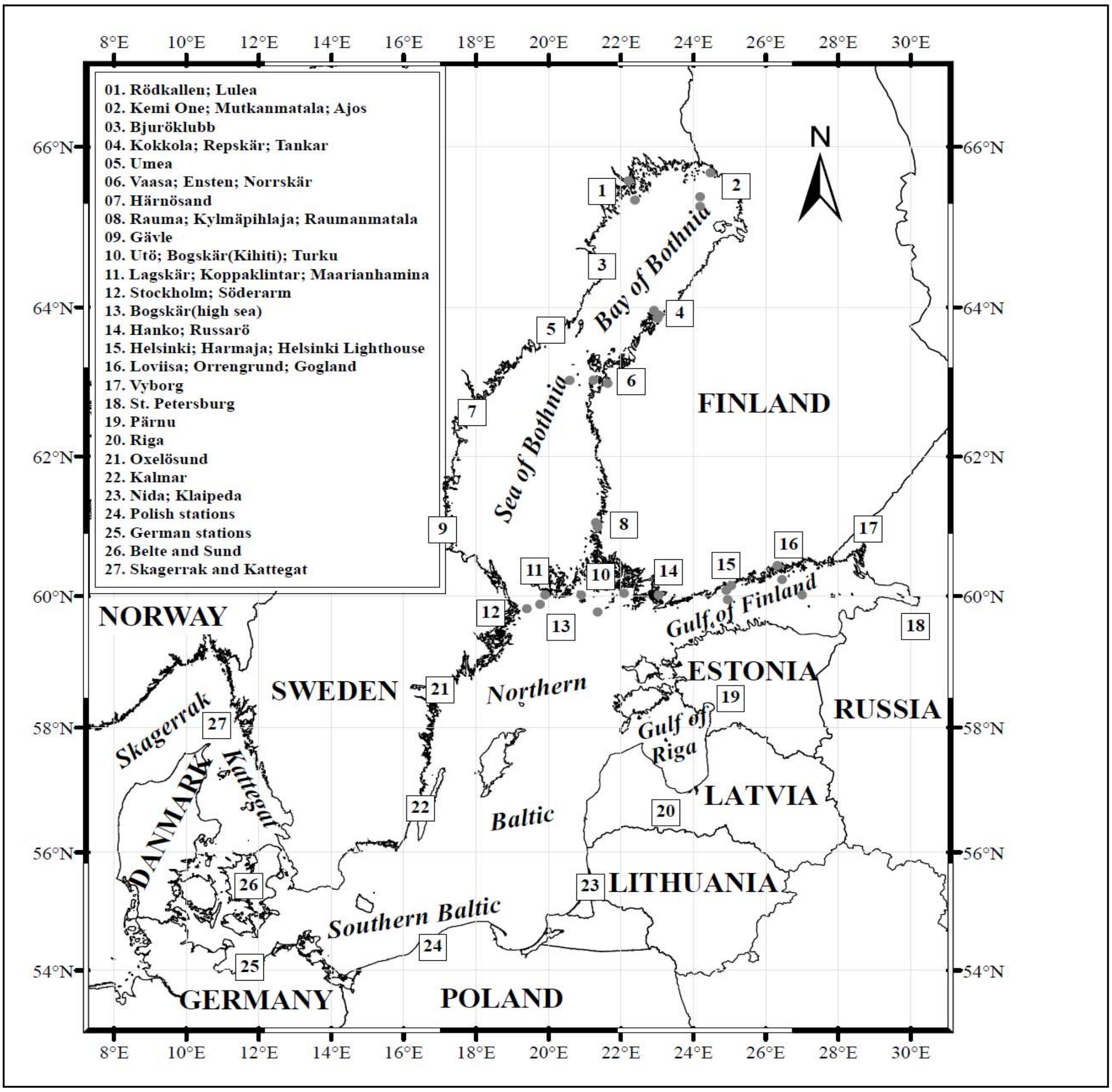

Fig. 4.13: Map of the Baltic Sea showing the location of important observation stations
Source: Schmelzer et al., 2008

The Baltic Sea freezes over every year and reaches the maximum extent of ice coverage between January and mid-March. Although the ***maximum ice extent*** may last just for 1 day, it is a very good indicator of the severity of the ice season. But the ice climate of the Baltic Sea can be characterised by several variables, including the concentration (percentage of ice-coverage) and thickness of the ice and the duration of the ice season.
However, a synoptic characterization of the entire area of the Baltic Sea is hardly possible because conditions vary considerably at the local level, which could be well illustrated by Figure 4.14 showing the relative frequency of ice occurrence in the different regions of the Baltic Sea.

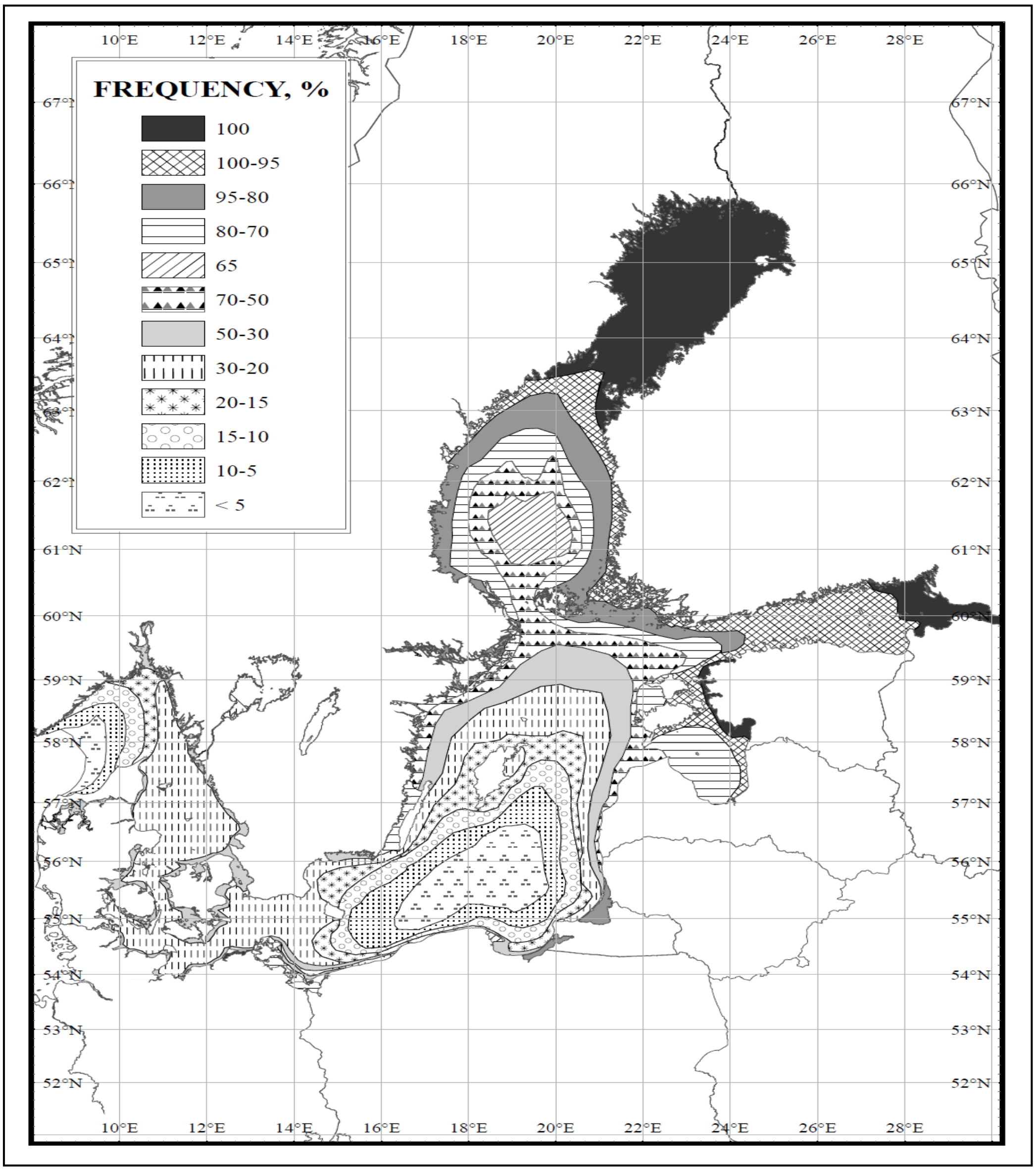

Fig. 4.14: Relative frequency of ice occurrence in the Baltic Sea in the periode from 1956-2005
Source: Schmelzer et al.; 2008

Past trends

Seinä (1994) and Seinä and Palosuo (1996) have summarised the annual maximum ice extent in the Baltic Sea utilizing the material of the Finnish operational ice service from the winters of 1945-1995 and information collected by Prof. Jurva from the winters of 1720-1940. The latter originated from various sources, including observations at lighthouses, old newspapers, records on travel on ice, scientific articles, and air temperature data from Stockholm and Helsinki. As stated by him, from about the year 1880 onwards the extension of ice cover was basing on notes made on board ships navigation during many winters, or may be rather easily and sufficiently accurately estimated on the basis of the time analysis of ice winters.

An overview of the data-record on annual maximum sea ice extent from 1878 to 2010 (updated by Finnish Ice Service http://www.itameriportaali.fi/en/tietoa/jaa/jaatalvi/en_GB/jaatalvi/ and Schmelzer) is given in figure 4.12.

Since the extent of the sea ice cover varies a lot from year to year Seina and Palosuo (1996) classified the ice winters in extremely mild *[<81.000 km²]*, mild *[81.001-139.000 km²]*, average *[139.001-279.000 km²]*, severe *[279.001-383.000 km²]* and extremely severe *[>383.000 km²]* ones. According to that classification, during the last ten years all ice winters have been average, mild or extremely mild. The latest extremely severe ice winter occurred in 1986/1987. In contrast the winter 2007/2008 was an extremely mild one (Fig. 4.15 (a), (b)).

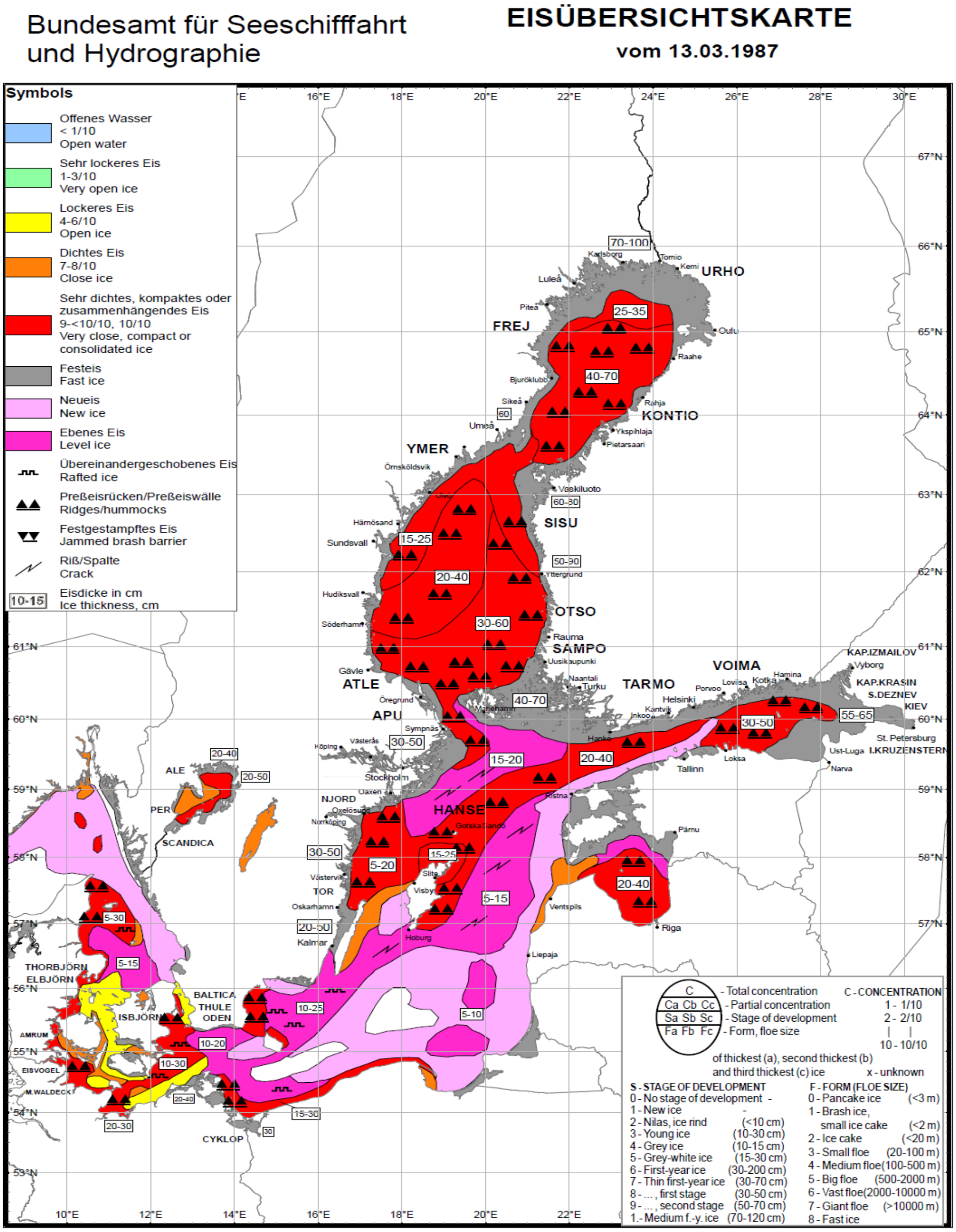

Fig. 4.14(a): Ice covering of the Baltic Sea in the extremely severe ice winter 1986/1987
Source: Schmelzer (BSH; 2010)

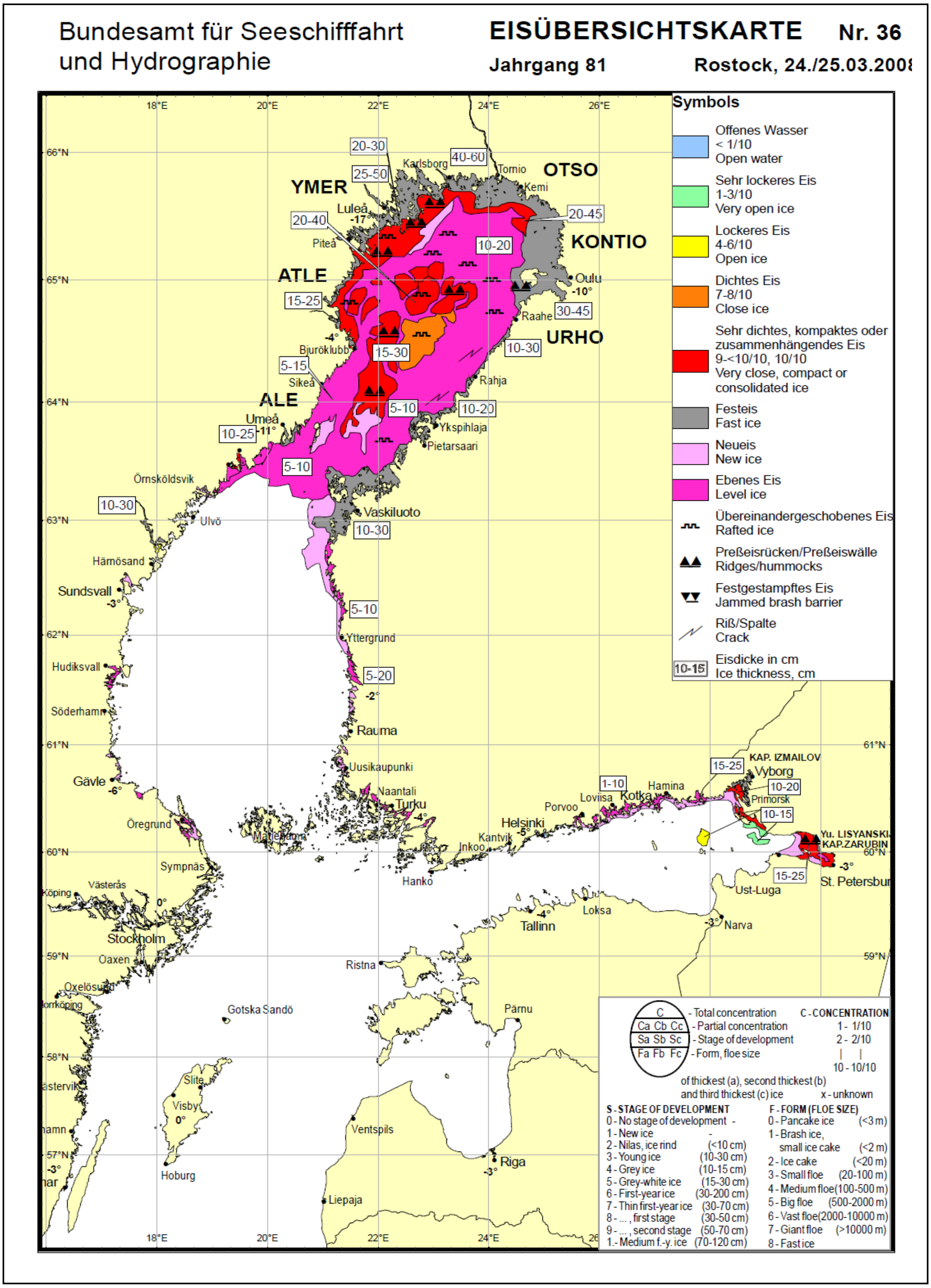

Fig. 4.14(b): Ice covering of the Baltic Sea in the extremely mild ice winter 2007/2008
Source: Schmelzer (BSH; 2010)

According to Haapala and Leppäranta (1997) the maximum annual ice extent in the Baltic Sea did not show clear trends during the 20th century. In the updated graph (Fig.) we see, however, a decreasing trend of ice extent (BACC, 2008), particularly in the data on the previous 3 decades.
Considering the whole Baltic Sea, Jevrejeva et al. (2004) did a comprehensive analysis of 20th century time series at 37 coastal stations around the sea. In general, the observations show a tendency towards milder ice conditions, where the largest change is in the ***length of the ice season*** (ice cover duration), which has decreased by 14-44 days in a century, which, in turn, is largely due to the earlier ice break up.
In their analysis of 37 time series from the coastal stations around the Baltic Sea, Jevrejeva et al. (2004) did not find any consistent change in the ***annual maximum ice thickness***. According to Haapala and Leppäranta (1997), the level-ice thickness in the Baltic Sea did not show clear trends during the 20th century. However, in all stations decreasing trends have been observed since the 1980s (BACC, 2008).
From the view of climatology, the most relevant parameter for describing ice conditions is the total mass of ice, but there is a lack of data on it over large regions. Koslowski and Loewe (1994) calculated the areal ice volume over a small region (coastal area of Schleswig-Holstein; Germany) and showed that in the period from 1879-1992 it was negatively correlated with the NAO winter index. Further, in a more recent publication, Jaagus (2006) analysed the freezing and break-up dates near the Estonian coast in relation to large-scale atmospheric circulation.
Generally, no correlation was found between the NAO and AO indexes (as meteorological patterns characterising the circulation over the North Atlantic and the polar region; see also box X in the Introduction-chapter) and the date of the first appearance of ice in the Baltic. But stronger correlations were found with the date of ice break-up and the length of the ice season.

Projections

Since anthropogenic climate change might affect the ice season in the Baltic Sea considerably, the fate of the Baltic Sea ice in changing climate has been investigated in several studies (e.g. Tinz 1996,1998; Haapala and Leppäranta 1997; Omstedt et al. 2000; Haapala et al. 2001; Meier 2002, 2006; Meier et al. 2004). These authors have applied different methods, based upon either statistical or dynamical downscaling of GCM results.
The main conclusion from these studies is that the projected decrease of ice cover over the next 100 years is dramatic, independent of the applied models and scenarios. Rossby Centre Atmoosphere Ocean climate model (RCAO) results from Sweden suggest that the Baltic Sea ice extent may decrease by 57% or 71% towards the end of the 21st century in the B2 and A2 scenarios respectively (Meier et al.2004). The Bothnian Sea, large areas of the Gulf of Finland and Gulf of Riga, and the outer parts of the south-western archipelago of Finland would become ice-free in the mean. The length of the ice season would decrease by 1-2 month in the northern parts and 2-3 month in the central parts of the Baltic Sea (Meier et al. 2004). None of the simulated winters in 2071-2100 are completely ice- free due to a non-linear sensitivity of the simulated ice cover and the winter mean temperature. Severe ice winters are projected to be more sensitive to anthropogenic climate change than mild winters. However, based upon the variability of the entire 20th century , an ice-free winter was found assuming changes of atmospheric surface variables corresponding to an A2-scenario (Meier 2006). Omstedt et al. (2000) found that the scenario simulation indicates a maximum ice extent close to the observed long-term minimum and that there is no ice during 3 out of ten winters. In addition to scenarios, sensitivity studies were performed (e.g. Omstedt and Nyberg 1996; Omstedt et al. 1997; Meier 2002, 2006). These studies show that the summer heat content may affect only the subsequent ice season. The time scale of the upper layer heat content amounts to a few month at the maximum.

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5. Secondary Impacts of Climate Change on the Cryosphere

5.1 Avalanches

Key messages:

- Average avalanche activity has not changed during the last decade, and climate change will only affect the lower altitudes
- The event history of avalanches is extremely difficult to relate to climate change aspects, due to two accompanying developments, the increasing role of snow sports and the large investments in technical avalanche defence measures
- The last winter with many large avalanches in Europe was 1998/1999, but there are still several fatalities every year, however, most of which occur in relation to snow-sports.
- High safety standards with respect to avalanches have been attained in Europe. Maintaining this safety level requires improving technical countermeasures, early warning systems and training of rescue staff.
- It is currently still difficult to make a clear forecast for the long term development of avalanche hazards under a changing climate

Key graph:

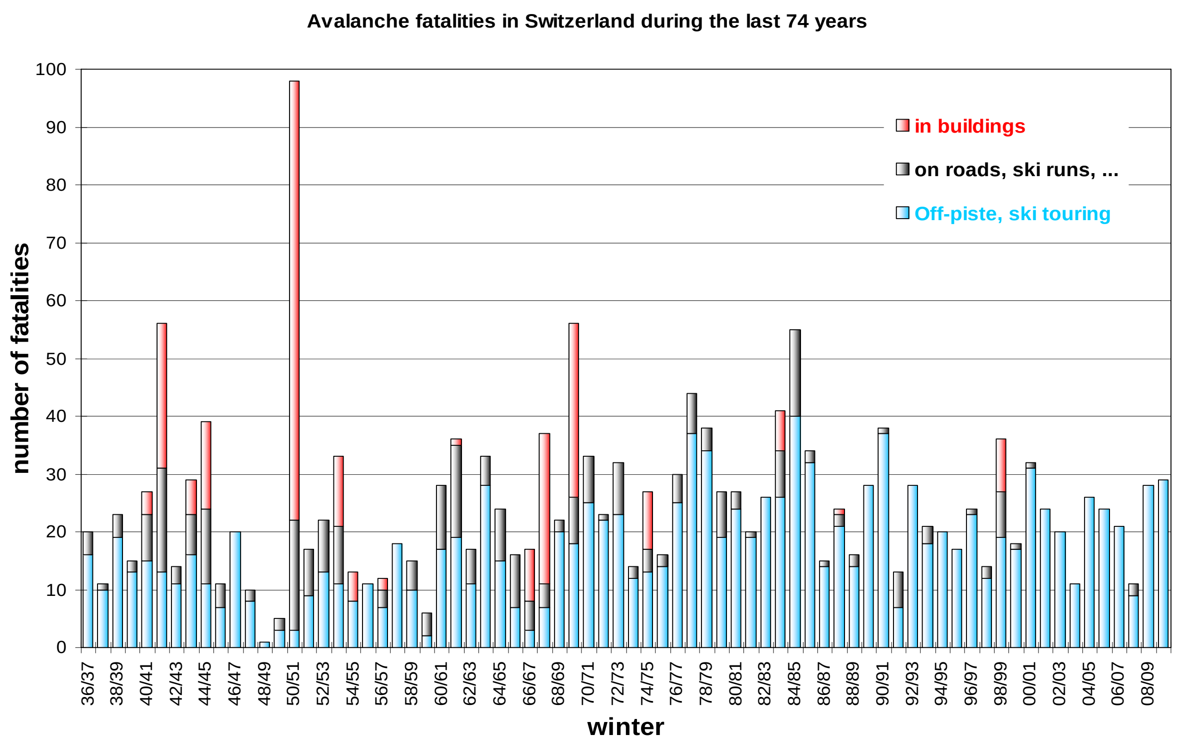

Figure 5.1: Number of casualties due to avalanches in Switzerland (1936-2009)

Source: SLF; 2010

Note: The number of casualties on roads and in buildings is clearly decreasing due to technical prevention-measures

Relevance*:*

An avalanche is “a snow mass with typically a volume greater than 100 m3 and a minimum length of 50 meters, that slides rapidly downhill” (EAWS). Avalanches range from small slides barely harming skiers, up to catastrophic events endangering mountain settlements or traffic routes. Avalanche formation is the result of a complex interaction between terrain, snow pack and meteorological conditions. Avalanches are generally natural events and the majority occur without causing damage or even being noticed. However, alpine avalanches kill around 100 people every winter (average for the past 30 years) whereas the large majority of these fatalities occur in relation to snow-sports, where the avalanches are mostly triggered by the involved persons. The number of fatalities has stayed at a constant level in all Alpine countries, in spite of a steady increase of the number of snow sport avalanche accidents. This is very likely due to the improved warning systems, the good education of of the people at risk, the fast reaction of rescue-teams and the highly developed avalanche beacons, often permitting fast search and rescue of people buried under the snow.

In environmental terms the (mostly non-human triggered) large avalanches are a part of the dynamic of a mountain ecosystem and can cause soil erosion, break trees or even destroy whole forests. This disturbance can have a beneficial influence on several aspects of the ecosystem, as a study of the WSL Institute for Snow and Avalanche Research (SLF) shows (Brugger et al., 2004). When an avalanche starts above the forest, large trees can break off, increasing the amount of light reaching the ground. Levels of nutrients and water also rise in the absence of the dominant trees using these resources. These changes can create the conditions that many plant species need for growth, thereby allowing a different plant population to develop. The seedlings and saplings are sheltered by the snow cover or are flexible enough not to be destroyed by subsequent avalanches.

The biodiversity in avalanche tracks is therefore often high, up to three times higher than in the surrounding forests (Rixen et al. 2004). The frequency of avalanches is highest in the centre of an avalanche track. Also, there are areas where the snow accumulates and others where it is eroded. Because of these factors, a variety of habitats develop within a small area.

In economic terms, the direct losses due to avalanche impacts in most parts of Europe have been small. However, the tourist agencies are still concerned with the so-called indirect losses, because Alpine tourism is a very important economic factor for the Alpine regions and in some areas the only source of income for the local population in winter. According to a study following the avalanche winter 1999 (Nöthiger et al. 2004) the short term reactions by tourists to avalanche events is substantial. Reductions in overnight stays in the alpine region are still noticeable one year after a disaster.

In Norway several communities along the coast depend on roads through hazardous terrain, and road closures are common. Closures of main transport routes have large economic consequences because it disrupts regularity in transport of goods. Approximately 50.000 properties with a net value of more than 100 billion NOK are located within areas susceptible for landslides and avalanches. Therefore large annual budgets are used to secure transportation routes and other infrastructure (Solheim et al.; 2010).

Important sources of information include the official avalanche warning services in the different European countries and regions (see below), as organized by the European Avalanche Warning Services (EAWS), and the International Committee for Alpine Rescue (ICAR).

The quality of the information concerning both, human and economic losses is variable throughout Europe, and a common data base to collect such information is highly desirable.

A different problem occurs in respect of data on injuries. Accidents with non-fatal injuries are not registered in a dedicated data base, and often, in respect of minor injuries, are not even registered at all. Therefore it is most probably impossible to establish high quality statistics on non-fatal accidents.

Past Trends

A long-term overview on the numbers of documented damaging avalanches per decade in Austria since the 1880s is given in Fig. 5.2. It has to be considered the incompleteness of the graph as caused by several reasons, e.g. by the lack of staff in the beginning of registration and during the time of wars. Since the 1950s the data-gathering has been intensified. The very intensive avalanche-winters in 1950/51 and in 1953/54 which caused about 278 casualties and immense damages in Austria are covered by the big column presenting the 1950s. The reduced number of registered events following the 1980s could be caused, among other reasons, by successful implemented measures on avalanche-control.

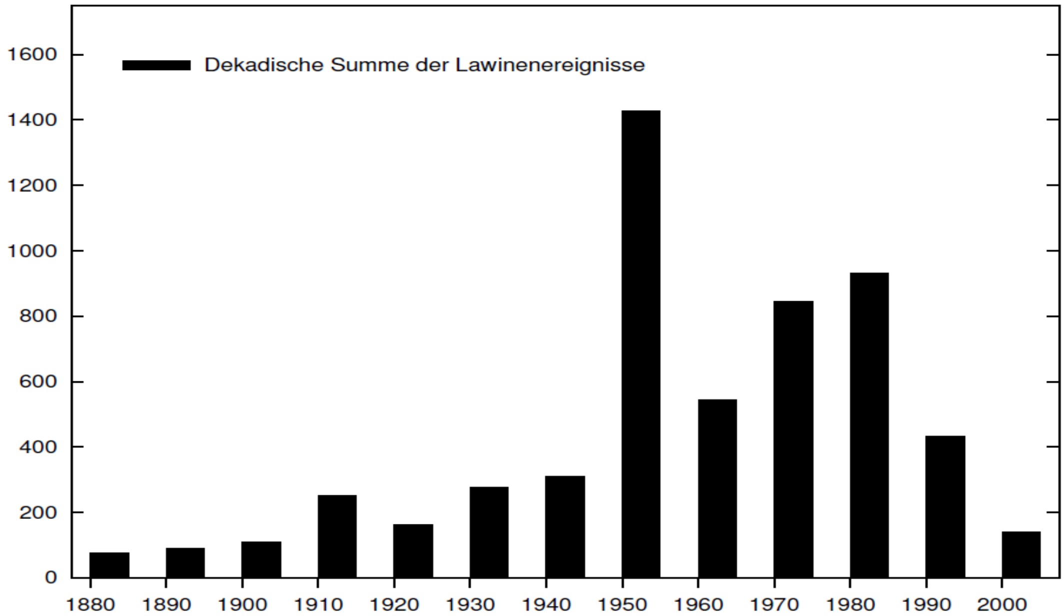

Fig.:5.2: Number of damaging avalanche-events per decade in Austria
Source: Institut für Alpine Naturgefahren, BOKU-Wien, published by INTERPREVENT (2009)

An analysis of the avalanche records in the Swiss Alps shows that natural avalanche activity has not changed during the last 70 years (Laternser et al. 1997). A more detailed analysis of the fatality-structure of all registered avalanche events in recent decades in Switzerland (Fig. 5.1) documents a clear decrease of casualties on roads and in buildings (as mostly due to big natural avalanches) and a quite constant number of deaths in the open country (as mostly due to snow-sport triggered events). This confirms a European wide trend caused by successfully implemented organisational and technical prevention measures (Laternser, 2002).
Climate change is, however, having a more and more pronounced effect on the snow cover at altitudes below 1000m a.s.l., where a significant temporal as well as spatial reduction of snow coverage is already taking place (e.g. Vojtek, 2010; Pecho, 2009; Marty, 2008; Scherrer, 2004). In contrast, no trend is visible at higher altitudes. Further increases of temperature obviously reduce the period during which large avalanches can occur at all. However, the occurrence of large avalanches is not governed by the general climatic trends, but rather by short term weather events, e.g. particularly intense snow falls during a couple of days, possibly linked with strong winds, or a rapid temperature increase with rainfall at high altitudes.
However, another (minor, but climate related) factor possibly triggering avalanches in lower altitudes was recently identified in Slovakia. Since global warming influences the forest composition in favour of deciduous trees the frequency of avalanches from forested slopes might increase, due to the reduced friction between the (dropped) leave covered ground and snow cover. In addition, the snow is bonded to the trunks of coniferous trees better than to the deciduous ones (Kolacny, 2009; Pet’o and Kyzek, 2004). An increase of these atypical avalanches has already been observed in some lower mountain regions (e.g. Horny Jelenec; Vel’ka Fatra) in recent years (APC, 2006).

In Norway the key weather elements that trigger avalanches may, however, vary from region to region. Heavy snow fall is the main triggering mechanism along the South-West coast, while strong winds are more important in northern Norway (Solheim et al.; 2010).
The last catastrophic winter in Europe with a large number of fatalities in secured areas (i.e. settlements and traffic routes) was 1998/99. The heaviest snowfall period in the Alpine region for 50 years triggered numerous fatal avalanches in particular in Austria, France, Switzerland, Italy and Germany. Table 5.1 shows the major avalanche accidents for the period 1997/98 to 2007/08 in France, Italy, Austria, and Switzerland. With the exception of the winter of 1999, almost all fatalities occurred in relation to snow sports. In Norway, very similar, has been a marked shift over the last decades from avalanche related fatalities on roads, in houses or other secured areas, to deaths related in winter sports. During the last 10 years, 95% of the fatalities in snow avalanches were skiers or snowmobilers or related to other leisure activities (Solheim et al.; 2010).
In Iceland snow avalanches and landslides have caused both death and injury and done great damage to infrastructure and property. In the twentieth century, 193 persons died, thereof 69 persons after 1974. Financial costs resulting between 1974 and 2000 amount to 3.3 billion IKR (Jóhannesson and Arnalds, 2001).
A profound land-use and economic change in the Catalanian Pyrenees (Spain) has undergone a transformation from traditional rural society to a growing leisure industry related to winter sports and mountain recreation. Due to the rapid urbanization and the resulting population densities the number of people at risk in these areas has increased considerably. In the winter 1995/1996 a variety of meteorological situations produced several episodes of major avalanches (as well as in the winters 1971/1972 and 2002/2003). An avalanche warning system (avalanche forecast) prevented human casualties, however there was considerable damage to forests and infrastructure (E. Muntan et al.: 2009).

Date of the event	Location (Country)	Number of fatalities	Area
28.1.1998	Les Orres (France)	11	sports area
22.3.1998	Tuncely (Turkey)	12	Military
9.2.1999	Montroc (France)	12	secured area
21.2.1999	Evolène (Switzerland)	12	secured area
23.2.1999	Galtür (Austria)	31	secured area
24.2.1999	Valzur (Austria)	7	secured area
28.12.1999	Jamtal (Austria)	9	sports area
19.1.2000	Lyngen (Norway)	5	secured area
28.3.2000	Kitzsteinhorn (Austria)	12	sports area
12.7.2007	Jungfrau (Switzerland)	6	Military
25.8.2008	Mt. Blanc (France)	8	sports area
25.1.2009	Mt. Zigana (Turkey)	10	sports area

Table 5.1: Major avalanche accidents 1998-2009
Source : EAWS and ICAR

There have been very few non-snow sports triggered avalanches that caused casualties in the period 2003-2007. However, avalanche risk in these secured areas has not become negligible. Despite the intense efforts of the avalanche safety services, there are several cases each winter where avalanches e.g. reach public roads that had not been closed.
However, several useful attempts to reduce the risk by avalanches are undertaken on national and international level, e.g. by the European Avalanche Warning Services (EAWS). This organisation improves regularly methods and strategies to inform and warn people efficiently in regions

endangered by avalanches (as discussed and decided e.g. at their 15th conference in June 2010 in Innsbruck).
In Iceland, for example, the Meteorological Office is responsible for avalanche warnings and hazard zoning and advices the government on avalanche protective measures. The office employs snow observers in the most important villages in avalanche-prone areas and maintains a database for avalanches.

Projections

The occurrence of large avalanches is often governed by short term weather events, e.g. particularly intense snow falls during a couple of days, possibly linked with strong winds, or a rapid temperature increase with rainfall at high altitudes. Such marked weather periods may possibly become more frequent in many regions of the Alps with climate change. The percentage of wet snow avalanches is expected to increase, as compared to dry snow avalanches. An increase or decrease in the size of the avalanches should not be expected, as the avalanche size is governed by the release height and release area, which are hardly influenced by climatologically developments, but mainly by the topography and shear strength of the snowpack, respectively. From the above counteracting tendencies - reduced snow coverage, but possibly more heavy precipitation events - it is currently still difficult make a clear forecast for the long term development of avalanche hazards under a changing climate (Marty, 2009).
Most climate change scenarios for Norway indicate that there will be an increase in overall precipitation for most of the country in the coming years. In addition, both the frequency and the intensity of extreme events are expected to increase. With regards to snow avalanches, more precipitation may lead to higher avalanche frequency in the high mountains, whereas an anticipated increase in the level of both the tree line and the snow line, may lead to a lower frequency in lower areas, where most infrastructure exists. More frequent periods with temperatures around zero, and more rain-on-snow events is expected to lead to increased problems from slush flows and wet snow avalanches (Kronholm et al., 2006).

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5.2 Landslides and rock slope failures

Key messages

- A synthesis perspective and area-wide information on landslides and mass movements in high-mountain regions of the Alps is missing.
- No significant change in the frequency of shallow landslides and debris flows has been observed so far for European mountain regions. This is partly due to insufficient documentation.
- Large rock slope failures in permafrost regions in the Alps have increased since the 1980s as compared to the previous 100 years.
- The timing, frequency and magnitude of Alpine debris flows are likely to change in the coming decades, with a trend towards earlier initiation in the season and initiation from higher elevations.
- Climate change will likely lead to new areas being affected by landslide hazards. As a consequence, existing landslide, mass movement, and rock-fall hazard maps over many regions in Europe may become outdated.

Relevance

Landslides and rock slope failures are widespread over Europe's mountain regions. Single events can cause millions of Euro damage or more if critical infrastructure is affected (e.g. transnational traffic routes). Very large events (e.g. such as the 2002 Caucasus rock-ice avalanche) might cause thousands of casualties and billions of EUR damage in case they impact populated and developed mountain areas.

In Norway, landslides together with snow avalanches represent the most severe natural hazard in terms of number of fatalities. Landslides have caused more than 2000 fatalities during the last 150 years. The largest tragedies include one huge quick clay slide (more than 100 fatalities in 1895), and tsunamis caused by rock slides into deep fjords and lakes (3 events in the 20th century causing in total 175 fatalities). Closures of main transport routes have large economic consequences because it disrupts transport of goods both to domestic and international markets. Large annual budgets are used to secure transportation routes and other infrastructure. Approximately 50.000 properties with a net value of more than 100 billion NOK are located within areas susceptible to landslides or avalanches.

In the Alps, landslides and debris flows have impacted new locations in recent years, where historically no such impacts were known. An example are debris flows from permafrost areas at Guttannen, Central Swiss Alps, where formerly unobserved activity from Ritzlihorn has started in 2009 and 2010 and had severe impacts on important traffic corridors and transnational energy lines.

Past trends

Shallow landslides and debris flows typically occur after prolonged and/or intense rainfall. In addition, in high-mountain environments with glacier, permafrost and snow seasonal and more long-term cumulative developments can be important, e.g. snowmelt, snowfall line or increased availability of sediment due to glacier retreat or permafrost degradation.

In permafrost areas potential effects on slope stability by recent warming may currently have penetrated to depths of several tens of meters but will continue to reach increasingly greater depths with future warming. If water penetrates rock slopes, heat can be rapidly transported into the ground and contribute to permafrost thaw (S. Gruber and W. Haeberli, 2007). However, studies on the impacts of climatic extreme events on steep permafrost bedrock have only very recently been initiated. Case studies of exceptionally warm periods of weeks to months duration indicate that both small-scale and large-scale slope failures can be triggered (Gruber et al., 2004a; Fischer, 2010; Huggel et al., 2010). Warming of firn and glacier ice can produce more melt water and transform steep, formerly cold and stable glaciers into temperate and potentially unstable glaciers.

Altough area-wide information on high-mountain landslides is missing (possibly with an exception in Valle d'Aosta, Italy),, evidence of permafrost degradation and slope destabilization comes from a number of recent slope failures in permafrost areas, including a range of volumes of ~10^2 to 10^7 m^3. The European Alps are among the best observed mountains worldwide but evidence also comes from other mountain regions (Gruber and Haeberli, 2007; Huggel, 2009; Allen et al., 2010). Examples from the Alps are the 1997 Brenva rock avalanche in the Mont Blanc region (Barla et al., 2000), the 2004 Thurwieser rock avalanche, Italy (Sosio et al., 2008), rock slides from Dents du Midi and Dents Blanches, Switzerland, in 2006, or from Monte Rosa, Italy, in 2007 (Huggel et al., 2010), with volumes of a few millions of cubic meters. Very large rock and ice avalanches with volumes of 50 to over 100 million m^3 have occurred in 2002 in the Caucasus (Kolka avalanche, Kotlyakov et al., 2004; Haeberli et al., 2004) and 2005 in south-central Alaska (Mt. Steller, (Huggel et al., 2008)), and cannot be excluded in the future in the Alps.
Extraction of trends of occurrence of such events over time is difficult due to incomplete documentation. Nevertheless, compared to the 20th century an increase of large rock slides during the past two decades, and especially during the first years of the 21st century could be observed for the European Alps (Fischer, 2010), correlating with strong temperature increase, glacier shrinkage and permafrost degradation.
Massive reduction of glacier area and thickness since the mid-19th century, considerably accelerated in the last two decades, has an impact on the stability of adjacent rock slopes. Large rock slope failures such as at Lower Grindelwald glacier (Oppikofer et al., 2008) are a direct response to 20th glacier shrinking.

With respect to shallow landslids and debris flows, observations indicate that the initiation zones move upwards as glaciers retreat and new poorly consolidated sediment becomes exposed (Zimmermann and Haeberli, 1993; Rickenmann and Zimmermann, 1993; Haeberli and Beniston, 1998). Concerning the frequency of debris flows research has so far not provided any clear indications of change. In the Swiss Alps it was found that debris flow activity on a local site was higher during the 19th century than today (Stoffel et al., 2005) while in the French Alps no significant variation of debris flow frequency could be observed since the 1950s in high-mountain terrain above 2200 m asl (Jomelli et al., 2004). Indirect climate effects such as increase of available sediment or changing seasonal snow patterns can also influence debris flow activity (Rebetez et al., 1997; Beniston, 2006). Statistics are not completely clear but there could be an increase of debris flow activity in alpine regions during the past decades due to more extreme rainfall events and rainfall occurring at higher elevations. Observations show that large debris flow events in the past 20 years, triggered by intensive rainfall, and affecting extensive areas of the Alps, occurred in summer or fall, and were typically characterized by a high elevation of the snow fall limit (Rickenmann and Zimmermann, 1993; Chiarle et al., 2007).

Due to its combination of geological, topographical and climatic conditions, Norway has historically suffered from severe landslides. The steep terrain and high precipitation particularly along the Western and Northern coasts facilitate rock and debris slides, debris flows, and rock falls, whereas the flatter parts below the post glacial marine limit in south-eastern and middle part of Norway sre more susceptible to slides in sensitive clays (quick clay). A recent inventory of slides and avalanches in Norway includes more than 33,000 slides (Jaedicke et al., 2009). The slide database has been coupled with a meteorological database, and statistical analyses of the two data sets show correlations between various weather elements, or combination of these, and slide events in Norway. The key weather elements that trigger landslides may, however, vary from region to region. Debris slides are mainly triggered by heavy rainfall, often in combination with snow melting and a prolonged wet period prior to the event. For rock falls, however, the statistical analyses show no apparent correlation between slide events and weather conditions, although one may suspect such correlation to exist, due to more frequent freeze – thaw cycles in combination with increased precipitation. Quick clay slides, specific for Norway and Sweden, have been triggered in the last decades mainly by human activity, but can also be caused by erosion in rivers and streams during periods of flooding, and may therefore also be somewhat susceptible to climate change (Resseguier, 2006).

The granitoid High Tatras, the highest mountains in Slovakia and Poland, exceed 2600 m asl. The High Tatras had been deglaciated after the last Pleistocene Ice Age, and gravitational processes such as debris flows and rock falls are now overprinting the landscape. Intense rainfall events have repeatedly caused flash-floods and debris flows, depending on factors such as runoff characteristics, topography or sediment availability (Lukniš, 1973, Midriak, 1983). As in other mountain regions, air temperature, snowmelt and snowline also play an important role for flash-flood and debris flow frequency and magnitude in the Tatras. Intermediate storage, blockage and subsequent rupture can cause particularly high peak discharge. For the High Tatras there is currently no sufficient flash flood or debris flow database, which would allow toderive significant trends, but some recent events are remarkable:

- The flash-flood recorded on the Štrbský Creek (southward of High Tatras) on July 24th, 2001. The flash-flood was local and impacted only a small area (the catchment area of Šrbský Creek is about 12 km^2). Precipitation intensity was very high with up to 90 mm in less than 1 hour. Peak discharge was estimated as 65 m^3s^{-1} (26 $m^3s^{-1}km^{-2}$) and 120 m^3s^{-1}, for the upper and lower part of the catchment, respectively. The return period of the event was evaluated as more than 1000-years (Šťastný and Majerčáková, 2003).
- At Biela and Ždiarsky Creek (North-East from High Tatras) a flash flood occurred on June 30th, 2006. The measured precipitation in Ždiar was 153 mm in 24 hours. Peak discharge was between 75 and 100 m^3s^{-1}, with an estimated return period of more than 100-years (Blaškovičová, 2007).
- Based on multi-temporal remote sensing data the trend of the activity and geomorphic effectiveness of debris flows of two valleys in the central and eastern parts of the Tatra was analyzed for the period 1949-2006 (Kapusta et al., 2010). During the period 1949-1986 no clear trend in debris flow activity could be observed. However, between 1986 and 2006 an increased activity was detected in both valleys. Yet, in terms of extreme rainfall events, the data does not indicate any discernible trend for Slovakia over the same period (Fig. 5.3).

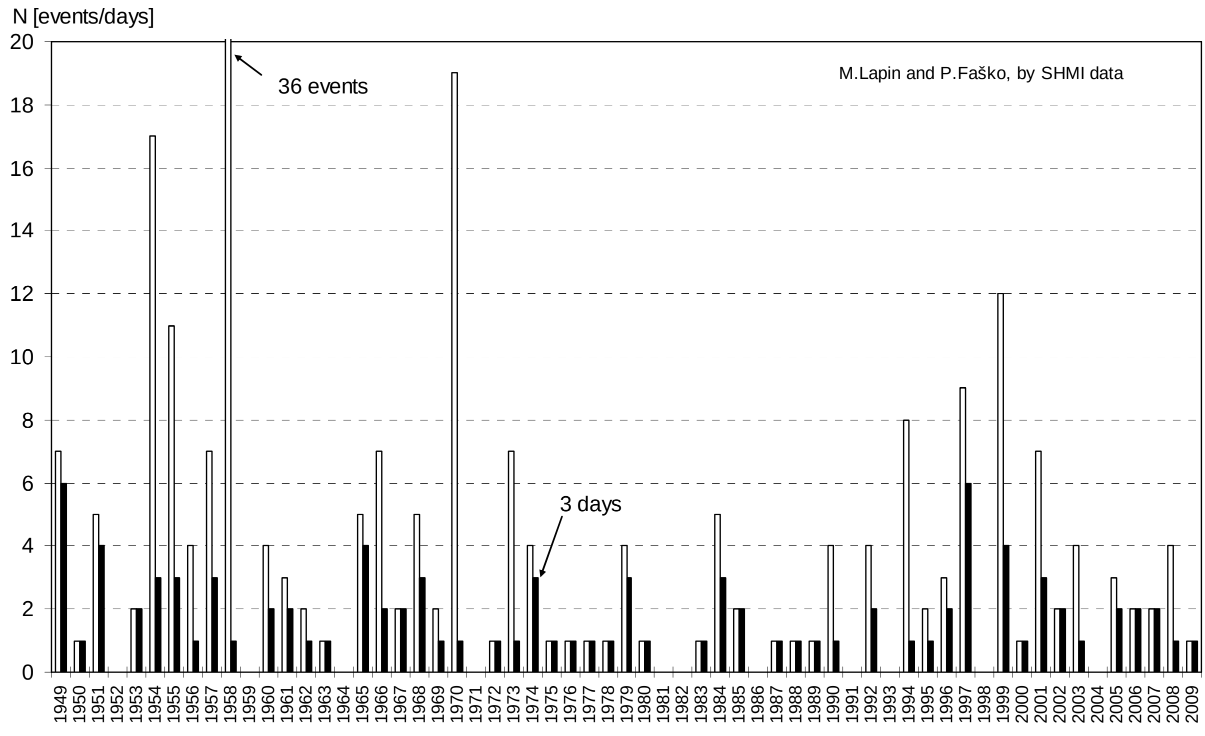

Figure 5.3.: Number of events and number of days with daily precipitation sum of 100 mm and more, recorded at least at one station *(data compiled from ca. 600 meteorological stations of Slovakia)*
Source: SHMI.

Projections

The magnitude of shallow landslides and debris flows from recently deglaciated terrain could increase due to higher availability of unconsolidated sediment (Haeberli and Beniston, 1998), depending on future rainfall volumes and intensities. Frequency of debris flows depends on the future frequency of debris flow triggering rainstorms but changes in the availability of erodible sediment can also have important effects on debris flow frequency and magnitude. Earlier snow melt in the future will result in earlier onset of high-mountain debris flows, and shallow landslides in lower mountain ranges may increase with projected higher precipitation intensities in winter over central Europe (Beniston et al., 2007).

The probability of large rock slope failures is increasing in the Alps but still small. Landslides that transform into highly mobile debris flows or impact natural or artificial lakes and cause outburst floods can considerably increase the reach of destruction.

In general, an extension of source zones of different types of mass movements to higher elevations can be expected, resulting in a change of hazard zones as they used to be known or mapped. New extreme events with higher impact and longer downstream reach than historically known are possible. The probability of occurrence of very large events such as rock-ice avalanches, as recently observed in the Caucasus and Alaska, will increase but still remain low. However, the occurrence of such an event in highly populated mountains such as the European Alps would result in hundreds to thousands of casualties and billions of EUR damage.

In general it is likely that continued permafrost degradation leads to a general decrease of rock slope stability. Future location and timing of large rock avalanches are extremely difficult to predict, as they depend on a multitude of factors, including local geological conditions. It is more likely than not that the probability of large, combined events, such as landslides impacting lakes and generating large outburst floods, will increase. Appropriate monitoring is highly recommended.

For Norway most climate change scenarios indicate that there will be an increase in overall precipitation for most of the country in the coming years. In addition, both the frequency and the intensity of extreme precipitation events are expected to increase. This may have an effect on the event frequency for mass movements involving abundant water, such as debris flows. For other event types, for instance rock falls or large rock slides, the effect on slide frequency from the climate change is less apparent. The anticipated change in debris slide frequency due to climate change is illustrated in Figure 5.4. As shown, the largest change is anticipated for the middle and Northern part of Norway (Kronholm et al., 2007). It is, however, important to stress that the statistical base for the analyses is rather incomplete, and this may greatly affect the outcome, such as the distribution shown in Figure 5.4. Current hazard assessments of flash-floods and partly debris flows are based on statistics of the past, and, as the system changes, may no longer be appropriate for assessing hazard and risks in the future. Over all regions in Europe existing landslide, mass movement, and rock-fall hazard maps may become out dated due to climatic and anthropogenic changes, but inherent uncertainty may be higher than the assumed changes.

Whether the estimated increase in slide frequency will cause an increase in the number of fatalities is uncertain, however. Changes in exposure and vulnerability are probably more uncertain than changes in the landslide hazard. It is expected that cost of damage and investments in prevention and mitigation measures will increase in the coming years.

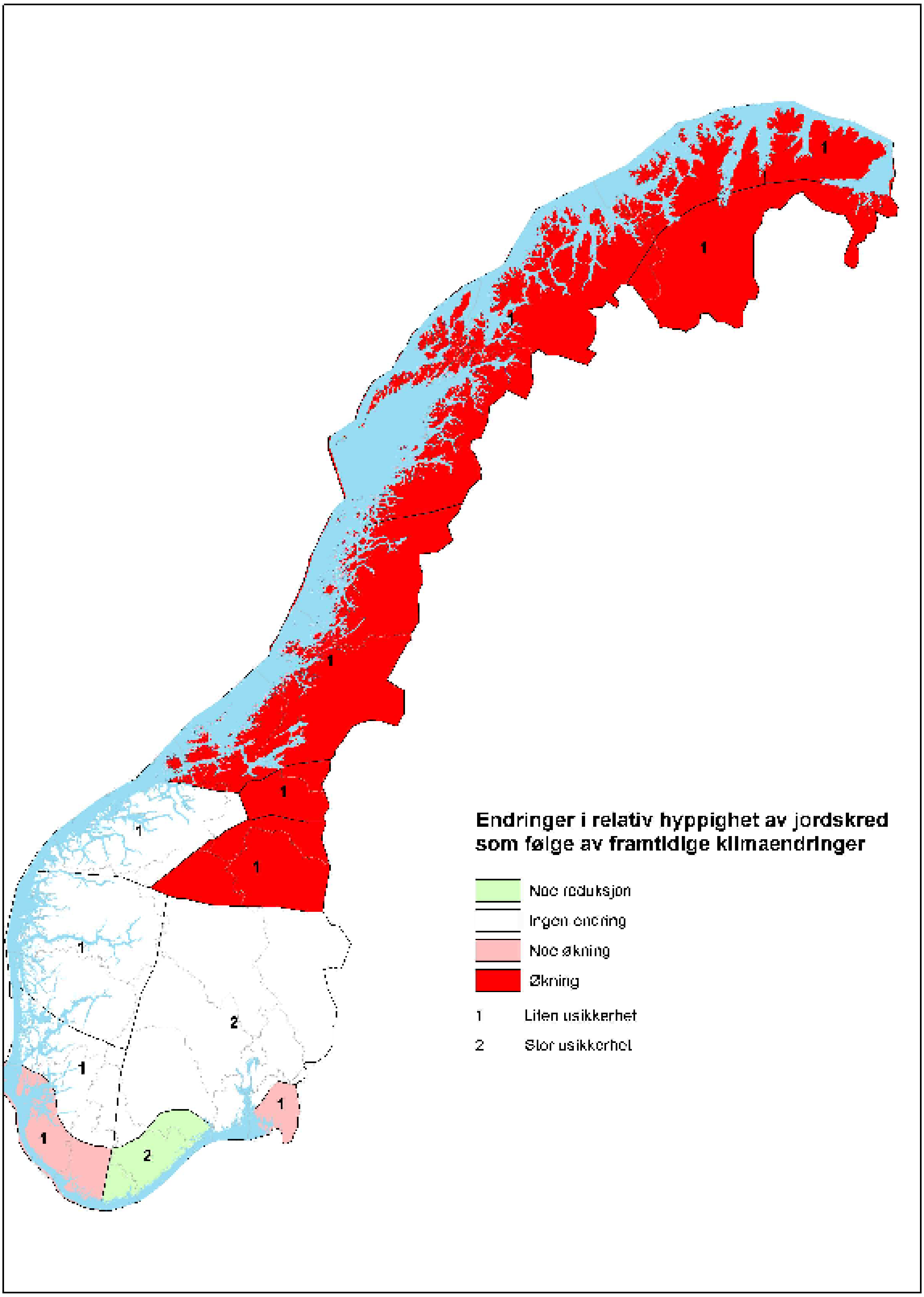

Figure 5.4.: Anticipated changes in relative frequency of shallow landslides as a consequence of expected climate changes.

The classes are: green=some reduction, white=no change, pink=some increase, red=increase. The uncertainties in the anticipated changes within each region are either 1=small uncertainty, 2=large uncertainty.

SOURCE: NGI, 2010

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5.3 Glacier floods

Key messages

- Glacier floods are rare events in Europe that only occur in some high-mountain regions. However, glacier floods can be highly destructive, and the costs of measures to mitigate an imminent event or reconstruction after an event can amount to tens of millions EUR.
- Hundreds of people were killed by glacier floods since the 16th century but no fatalities have occurred in the 20th and 21 century.
- Due to their rare occurrence it is difficult to detect any change in the frequency of glacier floods over the last several decades.
- The likelihood of glacier floods in the future is difficult to estimate due to a multitude of determinant factors.
- Ongoing glacier retreat will result in further growth or new formation of glacier lakes. These changes may be benecial for hydropower production and tourism but they also entail considerable risks.
- Of main concern are impacts from slope failures from destabilized high-mountain flanks that may increase with ongoing climate change. It is therefore important to timely assess such developments to be able to initiate prevention and adaptation measures.

Relevance

In European mountain regions, glacier lake outburst floods (GLOF) have caused hundreds of fatalities and tens of millions EUR of damage both in historic and recent times. For example, about 200 people were killed in two events in 1595 and 1818 at Giétro/Mauvoisin (Valais, Swiss Alps), or in the French Mont Blanc region a GLOF from Tête-Rousse glacier killed about 175 people in St. Gervais in 1892 (Vincent et al., 2010 (a), (b)). Since the beginning of the 20th century few people were killed by GLOFs but damage was high. In 2001 in Täsch (Valais, Swiss Alps) a GLOF caused a damage of about EUR 12 millions. In Macugnaga (Italy; Fig. 5.5) emergency actions due to imminent risk of a GLOF implied a cost of about EUR 6 million, while mitigation measures at Unterer Grindelwald glacier since 2006 amounted to more than EUR 10 million (including a new 2 km long drainage tunnel completed in late 2009), and at Tête-Rousse a subglacial lake (detected in 2009) has been drained in 2010 to avoid a new disaster (3000 people threatened) to 2.5 million EUR. Due to the enormous magnitude of glacier floods in Iceland (also termed Jökulhlaups) widespread damage and destruction of important infrastructure (mainly national highways), involving high reconstruction cost, is common with such events.
In Norway, many of the 18 known glacial lakes are located in unpopulated areas. The risk level for settlement and infrastructure is therefore generally low and is not considered as threatening. However, in the western part of Norway there are settlements close to outlet glaciers and the latest GLOF in Norway, in Fjærland 2004, caused inundation and damage to 250,000 m^2 of farmland. Controlled drainage through artificial spillways has been done at two glacier lakes in Norway to minimize the risk (e.g. Elvehøy et al., 1997; Fig. 5.6).

Past Trends

Glacier lake outburst floods (GLOF) are typically a result of cumulative developments of climate change as a result of glacier retreat. Glacier lakes may be dammed by bedrock or moraine dams, or by an ice dam. Presently much less frequent, yet quite common in historic times, glacier lakes may also form as a result of glacier advance, with glacier ice damming tributary rivers. GLOFs may be triggered by impact waves caused by mass movements (landslides, rock or ice avalanches), high lake levels and damming of the dam overflow, progressive failure of the moraine dam or a combination thereof (Haeberli, 1983; Huggel et al., 2004; Kershaw et al., 2005). GLOFs may occur (i) only once (e.g., full-breach failure of moraine-dammed lakes), (ii) for the first time (e.g., new formation and outburst of glacial lakes), and (iii) repeatedly (e.g., ice-dammed lakes with regular drainage cycles) (Clarke, 1982; Clague and Evans, 2000; Huggel et al., 2004; Dussaillant et al., 2010). GLOFs may impact areas tens or even hundreds of kilometers downstream of the lake. In Europe severe impacts from GLOFs were so far limited to a few tens of kilometers of downstream areas. The reach of the event thereby depends on the mobility of the flood, including factors as water volume and sediment concentration.

In the Alps, Switzerland has been the country most affected by GLOFs. About one GLOF event in 4 years was recorded for the period 1550 to present, whereas 15% did not cause any damage. Since the mid-20th century about one event every 1.5 years was recorded in the Swiss Alps (Haeberli, 1983; Raymond et al., 2003; www.glacierhazards.ch). Due to the relatively rare occurrence of GLOFs, it is difficult to derive any trends of frequency of occurrence of such events, at least for the recent past. In recent decades, important events (both actual GLOF disasters and emergencies due to imminent risk of GLOF) occurred in the Alps: Gruben glacier – Saas Balen 1968/70 (Haeberli et al., 2001); Belevedere glacier - Macugnaga 1979 and 2002-03 (Kääb et al., 2004; Chiarle et al., 2007; Fig. 2); Weingarten glacier - Täsch 2001 (Huggel et al., 2003); Rochemelon glacier 2004/05 (Vincent et al., 2010) or Unterer Grindelwald glacier 2006-2010 (Werder et al., 2010).

Figure 5.5: Pumps and tubes employed at Lago Effimero, Belvedere Glacier, Macugnaga (Italy) in July/August 2002 *(an emergency action in view of the imminent risk of a GLOF from the supraglacial lake which rapidly formed in June 2002 with a maximum water volume of ~3.5 million m^3. In June 2003 a moderately sized GLOF occurred but important damage could be avoided thanks to previous mitigation measures.)*
Photo by C. Huggel.

In Norway, following a general glacier retreat trend after the 1930's several glacier lakes have formed within or at the margin of glaciers. These lakes can regularly or unexpectedly drain, producing GLOFs of potentially large discharges. During the retreat of the last ice sheet from

Norway, the 250 m deep canyon ‘Jutulhogget’ was formed due to the outburst of a large glacier dammed lake.
18 glacial lakes are known to exist in Norway, some of them also regularly emptied by GLOFs. For details on these events, see Liestøl (1956), Elvehøy et al. (1997), and Kjøllmoen (2000). One of the most recent GLOFs in Norway occurred in Fjærland, 2004, when the lake at Supphellebreen drained through the glacier’s end moraine and grew into a major debris flow (Breien et al. 2008) (Fig. 3).

Glacier floods in Iceland are predominantly related to volcano-ice interactions. Melting of ice due to enhanced volcanic heat flow and/or eruptions can result in the formation of very large lakes under the ice, impounding tens or hundreds of million of m^3 of water. Sudden drainage in case critical pressure thresholds are overcome can lead to extremely large floods with discharge in the order of tens of thousands m^3/s (Björnsson, 2003; Roberts, 2005). Several GLOFs have occurred in historic times in Iceland. One of the largest ones in recent years was observed in 1996. The 2010 eruption of Eyjafjallajökull also produced a jökulhlaup in April 2010. There is an ongoing debate whether climate related glacier changes have an impact on volcanic activity and thus Icelandic GLOFs. The involved time scales, however, are rather in the millennia (Tuffen, 2010).

Figure 5.6.: The glacial lake at Flatbreen/Supphellebreen *(it drained suddenly through the end moraine in 2004 and caused a debris flow which ended in the populated Supphelle valley. A retreated outlet glacier is seen in the middle of the photo.)*
Photo by A. Elverhøi 2008

Projections

There is a very high likelihood of further glacier retreat. In glacially overdeepened sections that become exposed due to glacier retreat new glacier lakes may form. An assessment and mapping of potential new glacier lakes has recently been developed for the Swiss Alps (Frey et al., 2010). Such lakes also have considerable potential importance for tourism and hydropower. The likelihood of GLOFs from future lakes is difficult to estimate at the moment because of a multitude of determinant factors.

For Norway, most climate change scenarios indicate a future increase in temperature and winter precipitation for most of the country. In particular increased summer temperatures can make many of the glaciers in Norway disappear within 100 to 200 years (Johannesson et al., 2004; Nesje et al 2008), possibly providing an increase in meltwater from glaciers during this period. Withdrawal of glacier fronts may generate new glacier lakes, and rapid melting may result in higher potential for sudden outbursts of stored water (Nesje et al. 2008).

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6. Annexes

6.1 Damages and losses caused by natural hazards

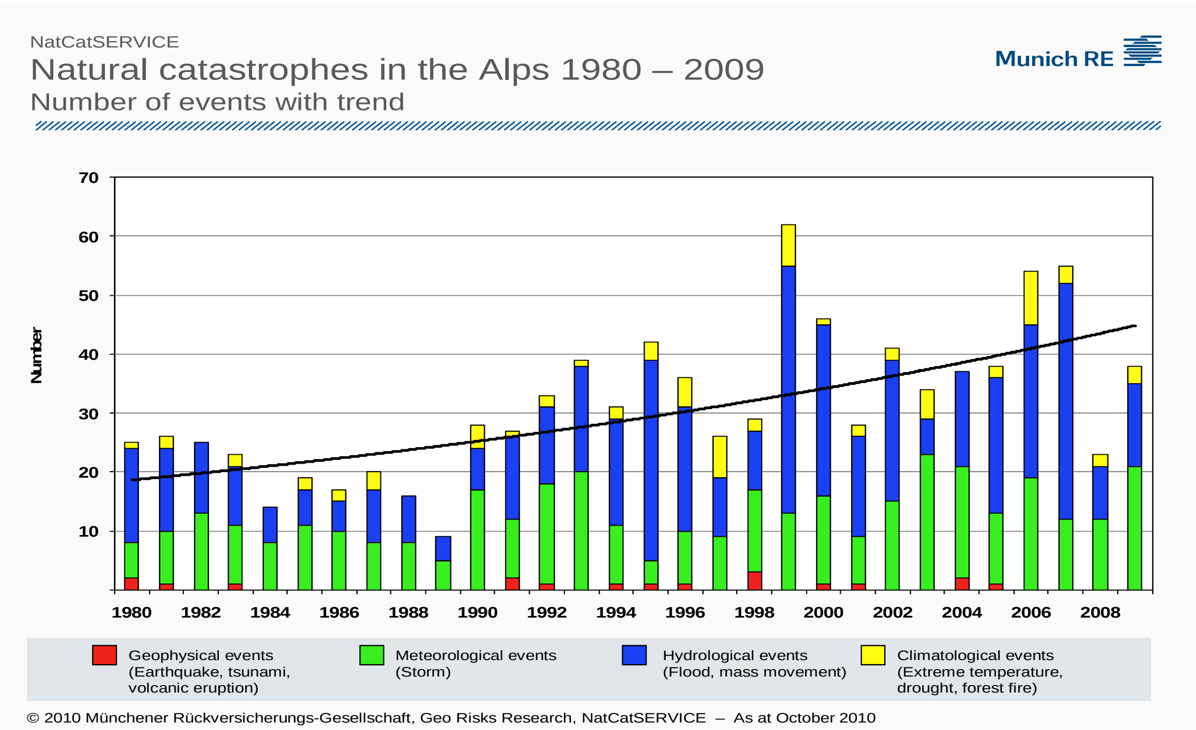
(Direct losses from weather disasters in the Alps and in Scandinavia)

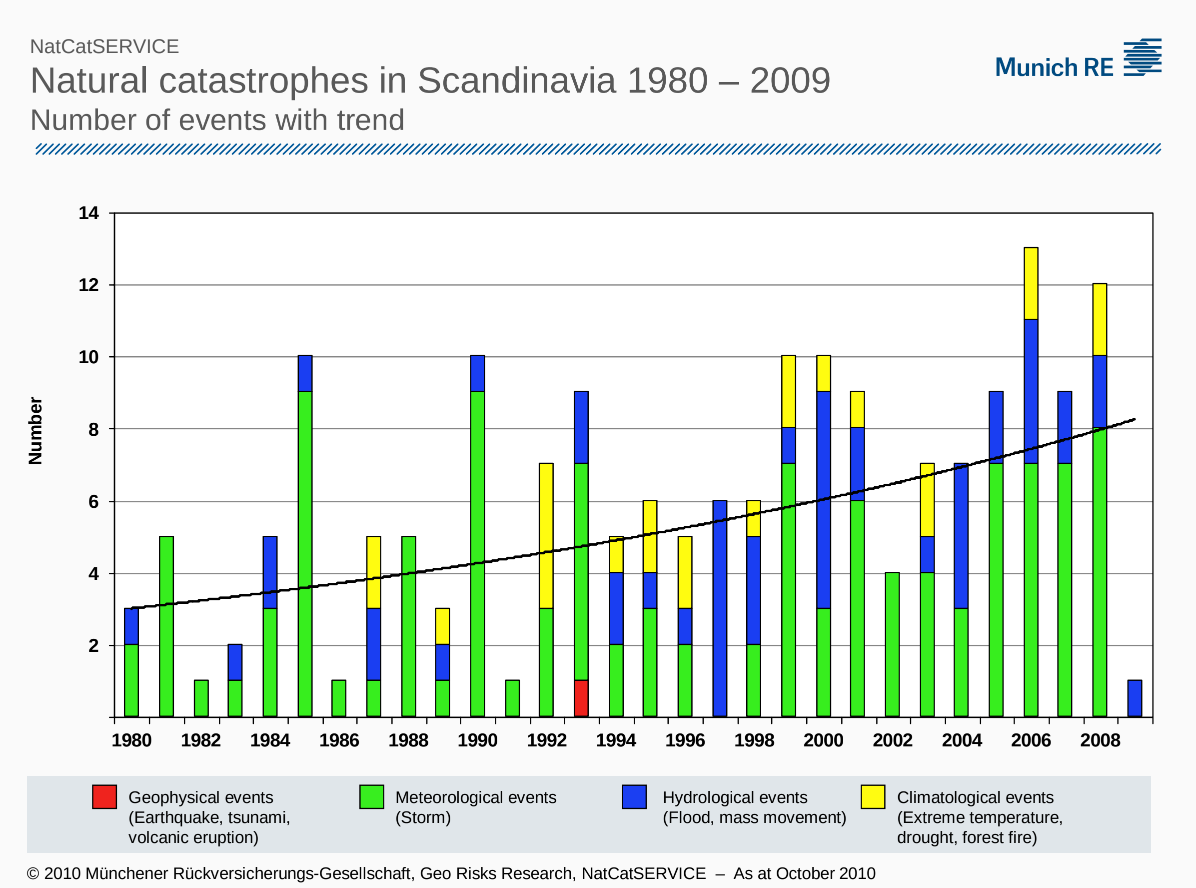
Key messages:

- About 98 % of all natural disasters in the Alps and 99% in Scandinavia that have occurred since 1980 are directly or indirectly attributable to weather and climate. About 99 % of economic losses and 99 % of the fatalities caused by catastrophic events in the Alps-region and almost all economic losses and fatalities caused by these events in Scandinavia have resulted from weather and climate-related disasters.
- The number of annual weather and climate–related events registered with significant losses in both of the regions, the Alps and Scandinavia, nearly doubled over 2000-2009 compared with the 1980s, while non-weather events (e.g. earthquakes) remained stable.
- While in Scandinavia storms, in particular winter storms, represent the clear majority of catastrophic events causing most of the fatalities and losses, in the Alps hydrological events as floods and mass movements dominate the disasters and generate the majority of fatalities and losses.
- Specific damages resulting from events related to snow and ice in Europe are not separately registered by re-insurance companies.

Key graphs

Figure A 1: Number of natural disasters causing significant loss and/or fatalities for the Alps and Scandinavia. Documentation is reasonably consistent and complete for the period 1980-2009.





Relevance

Changes in the frequency and intensity of storms, floods and extreme temperatures combined with changes in the exposure and vulnerability of human/societal systems affect the financial sector, including the insurance sector, through the amount of compensation payments. Examining insurance claims related to weather disasters can help to identify the sectors (e.g. agriculture, forestry, infrastructure, industry or private households) that are most affected by damage and/or could be most affected in future (EEA, 2008).

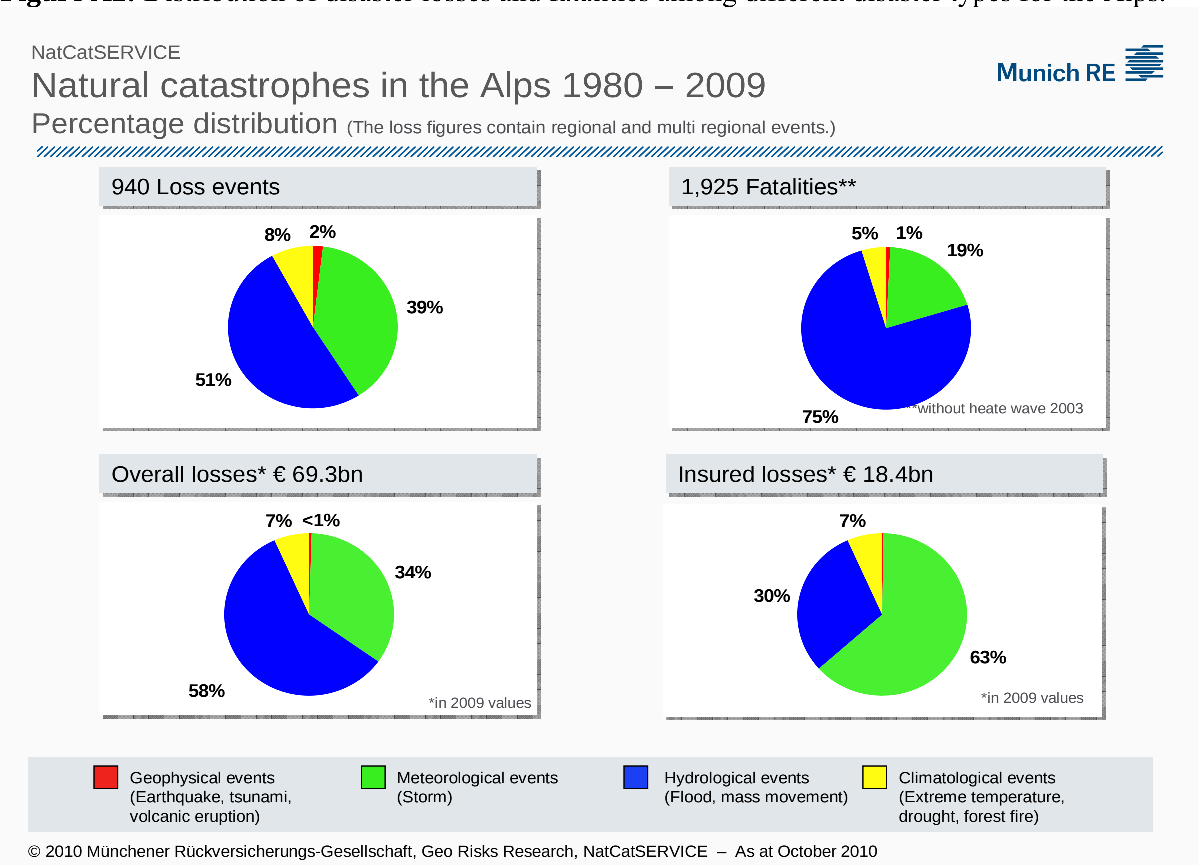
Even though the observed increase in losses is dominated by socio-economic factors (such as population growth, increased number of habitations in vulnerable areas, increased wealth, increased amount and value of vulnerable infrastructure), there is evidence that changing patterns of natural disasters are also drivers (Fig. A1). It is however not known how much of this increase in losses can be attributed to anthropogenic climate change (Höppe *et al.*, 2006). Insurance mechanisms are an increasingly important component in risk management and hence can play an important role in adapting to climate change by covering the residual risks and providing incentives for risk reduction. Through their underwriting policy, the (re)insurance companies can indeed increase risk awareness and provide incentives for risk reduction. Insurance companies have inherent interests in minimising the impacts of climate change in order to maintain residual risks insurable. Through their investment policy and asset management, the financial sector as a whole (savings, loans and insurance companies as well as other institutional investors) has great influence on companies' investment decisions. They can therefore ensure that any investments made are more climate-resilient and channel money into projects related to adaptation and mitigation of climate change. On the other hand the industries with greatest exposures will have to respond increasingly with innovative products, e.g. catastrophe bonds (Bouwer *et al.*, 2007).

Important sources of information include the insurance-companies in the different European countries and regions, in particular the services of the big European Re-Insurance companies as the Munich-Re (e.g. its NatCatSERVICE) and the Swiss-Re (Sigma).
The quality of the information concerning both, human and economic losses is variable throughout Europe.
However, a system to subdivide and specify losses due to impacts of climate change on Europe's snow and ice doesn't exist yet.

Past trends

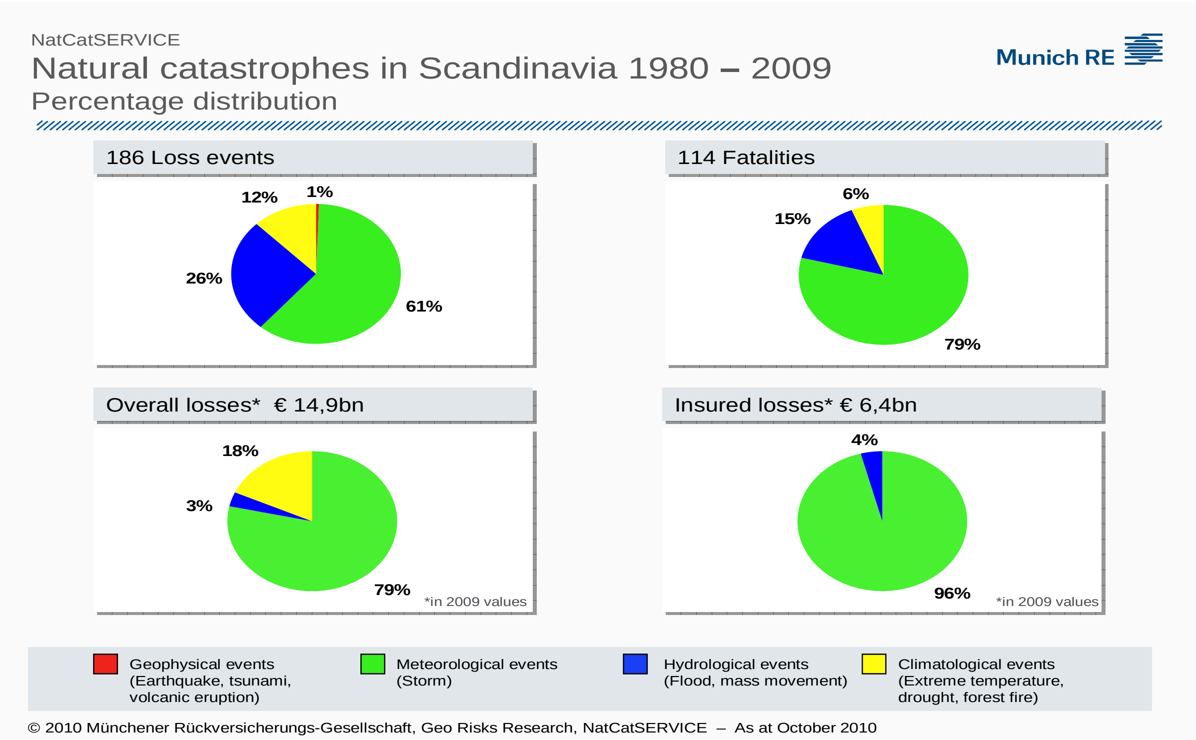
Munich-Re maintains a database on disaster events that is consistent over the past decades and records events that cause significant loss and/or fatalities. Events that do not cause damage are not documented. For the Alps events are recorded that struck large parts of Europe, including the Alps (e.g. large wind storms), as well as events whose effects were restricted to the Alpine region (e.g. snow avalanches). Munich-Re applies an inflation correction to the loss records to avoid distortion of trends over decadal time periods.
Based on the Munich-Re database, about 98% of all loss events in the Alps since 1980 are directly attributable to weather and climate (meteorologically, climatologically and hydrological events) and only 2% have a geophysical background. About 99 % of the overall losses and 99 % of all deaths caused by disastrous events result from such weather and climate-related events (Figure A2).The annual average number of these weather- and climate-related events in the Alps nearly doubled during the period 2000-2009 compared with the 1980s, while non-climatic events, such as earthquakes, remained stable (Fig. A1)).

Figure A2: Distribution of disaster losses and fatalities among different disaster types for the Alps.



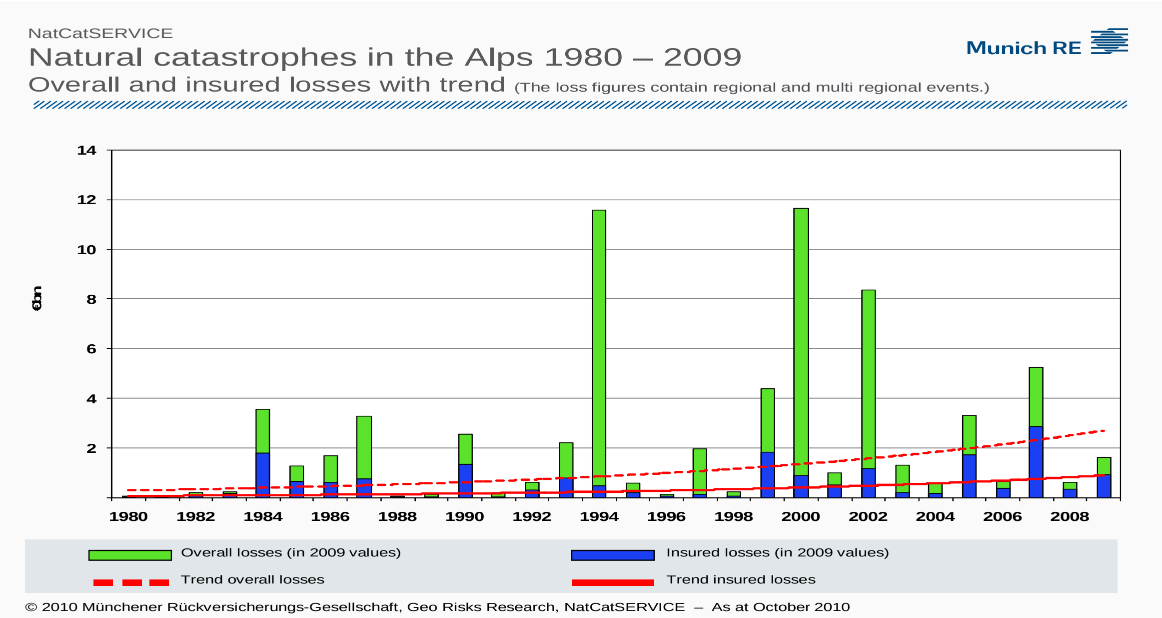
In (Fenno)-Scandinavia (Norway, Sweden, Finland and Denmark) about 99% of the loss events have a background in climate and weather and only 1% is attributable to geophysics. Deviating to the Alpine region, where the hydrological events as floods and mass-movements are dominating, in Scandinavia meteorological events as storms, in particular winter-storms, play an outstanding role (Fig.A2 and A3).

Figure A3: Distribution of disaster losses and fatalities among different disaster types for Scandinavia.



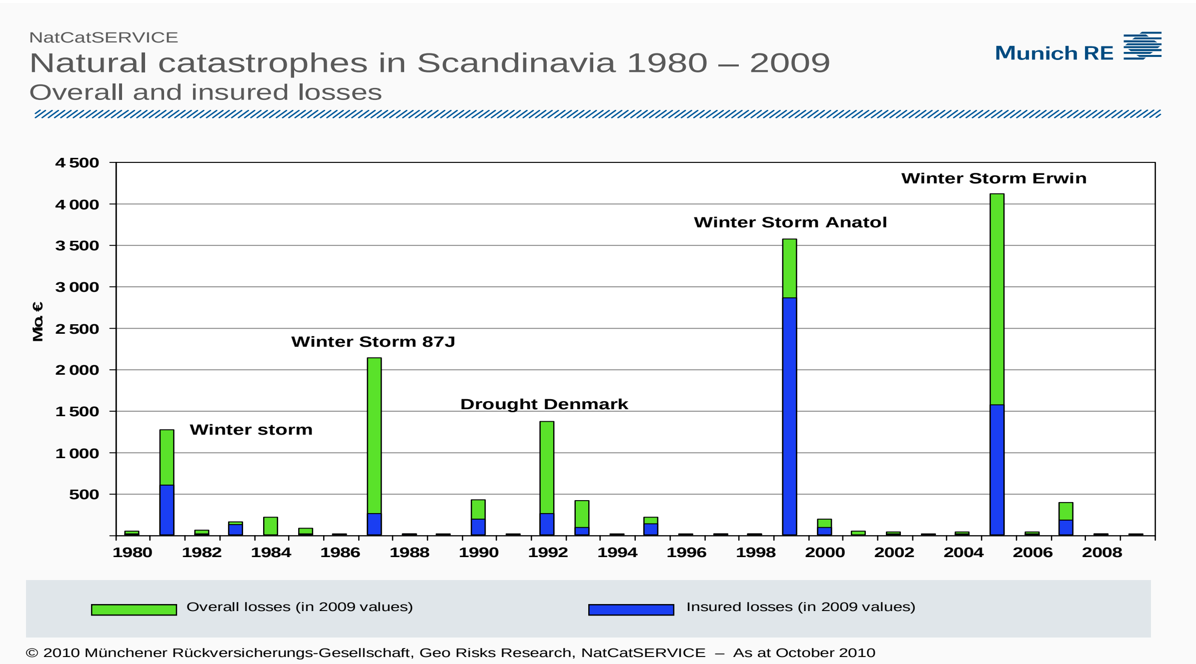
In the Alps, overall losses caused by weather and climate-related events more than tripled during the period 1980–2009 from a decadal average of about EUR 1,05 billion overall losses (0,41 billion insured) in the decade 1980–1989 to about EUR 3,43 billion overall losses (0,91 billion insured) in the decade 2000–2009. In this region four of the seven years with the largest overall losses in the period from 1980-2009 have occurred in the last decade (Figure A4). The insured portion of the losses generally rose, although with great year-to-year variability. It is currently not known whether any share of this increase can be attributed to climate change. Most studies conclude that for most disaster types a climate change signal cannot be detected so far, and that the dominant portion of the increase is due to changes in exposure and sensitivity of societal systems, in particular the increase of human activities in hazard-prone areas (e.g. Pielke et al., 2005; Barredo, 2010).

Figure A4: Overall and insured losses for the Alps and the period 1980-2009. Losses are inflation corrected.



In Scandinavia the losses rose from EUR 0,40 billion (0,1 billion insured) in the decade 1980-1989 up to EUR 0,48 billion (0,19 billion insured) in the decade 2000-2009. Particularly disastrous extreme events in Scandinavia in recent years were clearly dominated by the severe winter storms Erwin (2005) and Anatol (1999) (Fig. A5).

Figure A5: Overall and insured losses for Scandinavia and the period 1980-2009. Losses are inflation corrected.



Projections

Extreme weather events such as heat waves, heavy precipitation and with some less likelihood severe storms are projected to increase in frequency and intensity in Europe (IPCC, 2007a). However, the

associated time scale and hazard over the next 20 years remains uncertain. The most severe effects of anthropogenic climate change are expected in the second half of the century (EEA, 2008).
Predicting the future effects of extreme events also remains difficult because of increasing exposure caused by changes in economic development (e.g. tourism), which increases the value and density of human and physical capital. Disaster losses are expected to rise more rapidly than average economic growth, stressing the importance of risk reduction (Bouwer *et al.*, 2007).
It has to be noted here, that direct damages on human infrastructure as caused by reacting components of the cryosphere (e.g. damages on lift-systems or buildings caused by thawing ground) are currently not separately registered by the big insurance-companies. However, considering the climate projections regarded to these regions, increasing damages have to be expected.
The possible future increases of damages in general will enhance the vulnerability of the insurance sector and have important implications for the role of financial services under climate change (IPCC, 2007b). In high-risk areas people will experience increasing difficulty or costs in getting adequate insurance. This is likely to lead to greater levels of uninsured assets, particularly to socially-deprived groups, hence exacerbating inequalities. Thus governments may need to consider new ways of ensuring that especially poorer and more vulnerable people will still be able to have insurance and/or may be compensated for possibly increasing losses in future (e.g. through public-private insurance schemes such as those introduced in Belgium and proposed in the Netherlands (Bouwer *et al.*,2007)). Nevertheless, the noticeable differences in the climate predictions on the different regions show that there is no one-size-fits-all solution and suggest, more specifically, that European countries might need to implement different insurance schemes to secure sustainable and flexible loss-compensation systems (EEA, 2008).

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6.2 Ice services and ice products on the Baltic Sea region

(a) Ice products on the Baltic Sea region

Product/ Project	***Institution***	***Country***	***Home-page***	***Comments***
Baltic Sea Ice Services	*Baltic Sea Ice Meeting*	*Baltic countries*	*http://www.bsis-ice.de/index.shtml*	*- Actual Ice-Reports from all BS-countries* *-Actual Ice-Maps and forecasts* *-Satellite images* *-Summarising ice- reports*
Marine Weather/ Ice conditions	*Finnish Meteorological Institute (FMI)*	*Finland*	*http://www.fmi.fi/weather/index_9.html*	*- Actual Ice charts of the BS* *- Ice-Thickness charts and sea ice forecasts from POLAR-VIEW*
Sea Ice Services (Sweden)	*Swedish Meteoroogical and Hydrological Institute (SMHI)*	*Sweden*	*http://www.smhi.se/oceanografi/iceservice/is_prod_en.php*	*- Actual Ice charts* *-Swedish ice-report* *-Traffic restrictions*
Navy-Admiral Danish Fleet	*Navy- Admiral Danish Fleet*	*Denmark*	*http://forsvaret.dk/SOK/eng/National/Ice/Ice_Reports/Pages/default.aspx*	*- Danish Ice charts of the Baltic Sea* *- Danish Ice Bulletin*

(b) National ice services in the Baltic Sea Region

Country	***Name***	***Address***	***E-Mail/ Home-page***
Denmark	*Admiral Danish Fleet-Danish Ice Service*	*Sumatravej 3* *Postboks 483* *8100 Aarhus C*	istjensten@sok.dk *www.sok.dk*
Estonia	*Estonian Meteorological and Hydrological Institute(EMHI)*	*Toompuiestee 24* *EE-10149 Tallinn*	mere@emhi.ee *www.emhi.ee*
Finland	*Finnish Institute of Marine Research - FIMR -Finnish Ice Service*	*Merentutkimuslaitos* *PO Box 304, (Porkkalankatu 5)* *00181 Helsinki*	ice@fmi.fi *www.fmi.fi*
Germany	*Federal Maritime and Hydrographic Agency (BSH)*	*Neptunallee 5* *D-18057 Rostock*	ice@bsh.de *www.bsh.de*
Latvia	*Latvian Hydrometeorological Agency (LMHA)*	*Maskavas Str. 165* *LV-1019 Riga*	marine@lvgma.gov.lv *www.meteo.lv*
Lithuania	*LHMS Klaipeda Department*	*Taikos Str. 26* *5802 Klaipeda*	khmo@klaipeda.aiva.lt *www.meteo.lt/english*
Poland	*Instytut Meteorologii i Gospodarki Wodnej (IMGW)*	*Oddzial Morski* *Ul. Waszyngtona 42* *PL-81-342 Gdynia*	hydrologia.gdynia@imgw.pl *www.imgw.pl*
Russia	*North-Western Regional Administration for Hydrometeorology and Environmental Monitoring*	*23 Line, 2-a Basil Island* *199026 St. Petersburg*	sea@meteo.nw.ru *www.adm.meteo.nw.ru*
Sweden	*Swedish Ice Service, SMHI*	*S-601 76 Norrköping*	ice@prod.smhi.se *www.smhi.se*

6.3 Services and products on avalanches in Europe

(a) Products on avalanches in Europe

Product/ project	***Institution***	***Country***	***Home-page***	***Comments***
European Avalanche Warning Services (EAWS)	*Lawinenwarndienst Tirol*	*Austria*	*http://www.avalanches.org/*	*-glossary on snow and avalanches* *-adresses of national surviwes* *- degree of hazards*
International Committee of Alpine Rescue (ICAR-CISA)	*International Committee of Alpine Rescue*	*Alpine Countries*	*http://www.ikar-cisa.org/*	*-recommandations* *-statistics*
Intercantonal Early Warning and Crisis Information System (IFKIS)	*Institute for Snow and Avalanche Research SLF*	*Switzerland*	*http://www.ifkis.ch/*	*-informationplatform for the avalanche safety services* *-education and training programmes*
Avalanche Information	*Institute for Snow and Avalanche Research SLF*	*Switzerland*	*http://www.slf.ch/lawineninfo/index_EN?-C=&*	*-avalanche bullitins*
Avalanche Information	*Norwegian Geotechnical Institute (NGI)*	*Norway*	*http://www.ngi.no/en/Contentboxes-and-structures/Main-page/Feature-articles-/Avalanches-in-close-up/*	*-information on avalanches in Norway*
Avalanche Info	*Icelandic Met-Office; Civil Protection Dept. of Police*	*Iceland*	*http://en.vedur.is*	*Hazard zoning; Evacuation maps*
Avalanche Path Maps (ICC); Catalanya's Snow Hazard Database	*Geological Hazards Unit, Geological Institute of Catalonia ; Barcelona; Catographic Institute of Catalonia*	*Spain*	*http://www.icc.cat/msbdac/?lang=en_UK* *http://www.igc.cat/web/en/allaus.html*	*Avalanche Path Maps (APM); Avalanche Database of Catalonia (ADBC) containing Avalanche Pathes (AP), Avalanch Inquiries (AI) and Avalanche Observation (AO)*

(b) National Avalanche Services

Country	Name	Address	Home-page
Austria	*Amt der **Tiroler** Landesregierung - Abt. Zivil- u. Katastrophenschutz (+ 6 further Federal Countries in Austria)*	*6020 Innsbruck, Herrengasse 1-3*	*http://lawine.**tirol**.gv.at/* *http://www.lawinenwarndienst-**niederoesterreich**.at/* *www.lwz-**salzburg**.org* *www.**ktn**.gv.at* *www.lawine-**steiermark**.at* *www.land-**oberoesterreich**.gv.at/lawinenwarndienst/* *www.**vorarlberg**.at/lawine*
Czech Republic	*Horská služba*	*54351 Špindlerův Mlýn, č.260*	*www.hscr.cz*
France	*8 mountain regions*	*Info currently not available*	*Info currently not available*
Germany	*Lawinenwarndienst Bayern*	*80636 München, Lazarettstraße 67*	*www.lawinenwarndienst-bayern.de*
Iceland	*Icelandic Met-Office;*		*http://en.vedur.is*
Italy	*Regione autonoma Friuli Venezia Giulia - Direzione centrale risorse agricole, naturali e forestali - Servizio gestione territorio rurale e irrigazione (+ 6 further Provinces)*	*Via Sabbadini 31 33100 Udine Italy*	*www.regione.fvg.it/asp/newvalanghe/welcome.asp* *www.provincia.bz.it/valanghe* *www.aineva.it* *http://www.regione.vda.it, then look for Bollettino valanghe* *www.arpa.veneto.it*
Norway	*Norwegian Water Resources and Energy Directorate (NVI), with assistance of the Norwegian Geotechnical institute (NGI)*	*NO-0301 Oslo; P.O. Box 5091 Majorstua* *0806 Oslo, Sognsveien 72*	*www.nve.no* *www.ngi.no*
Poland	*Info currently not available*	*Info currently not available*	*Info currently not available*
Romania	*Regional Forecasting Center, Sibiu, Romania*	*550003 Sibiu, Somesului 49*	
Scotland	*Info currently not available*	*Info currently not available*	*Info currently not available*
Slovakia	*Horská záchranná služba - Stredisko lavínovej prevencie*	*Jasná 84, 032 51 Demänovská dolina, Slovak Republic*	*www.laviny.sk*
Slovenia	*Environmental Agency of the Republic of Slovenia, Snow Avalanche Service*	*1000 Ljubljana, Vojkova 1B*	*www.arso.gov.si/vreme/napovedi%20in%20podatki/snegraz.html*
Spain	*Agencia Estatal de Meteorolgia (+ 2 further provinces)*	*c/Arquitecte Sert n°1. 08005. Barcelona. Spain*	*www.aemet.es/es/eltiempo/prediccion/montana* *www.igc.cat* *www.aemet.es/es/eltiempo/prediccion/montana*
Switzer-land	*WSL-Institut für Schnee- und Lawinenforschung SLF*	*7260 Davos, Flüelastrasse 11*	*www.slf.ch*

6.4 Address details of World Glacier Monitoring Service and its National Correspondents in Europe

An updated and complete list National Correspondents of all countries that are active in glacier monitoring is found on the website of the World Glacier Monitoring Service:
http://www.wgms.ch/nc.html

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