5 An indicator-based assessment

5.1 Introduction

General

For this report, about 40 indicators have been selected to describe the state of the climate and the impacts of climate change on various natural and societal systems in Europe. These indicators were divided into nine separate categories which are presented in this chapter:

- Atmosphere and climate;
- Cryosphere (glaciers, snow and ice);
- Marine biodiversity and ecosystems;
- Water quantity;
- Freshwater quality and biodiversity;
- Terrestrial ecosystems and biodiversity;
- Soil;
- Agriculture and forestry;
- Human health.

The indicators were selected because of their measurability, their causal link to climate change, their policy relevance, the availability of historic time series (in most cases at least about 20 years), data availability over a large part of Europe (ideally they should cover all of Europe), and their transparency, i.e. they can be easily understood by policy-makers and the general interested audience.

Many other impact indicators were considered for inclusion but were rejected, often because of the difficulty of attributing an observed trend to climate change or insufficient data availability. If more information becomes available, some of these indicators might be reconsidered for inclusion in a future report, to achieve a more comprehensive picture of climate change impacts on the environment and society (see also Chapter 8).

Indicators from existing national indicator sets have been integrated where feasible. Others have been rejected because of missing data for the whole of Europe or because their relevance is limited to national issues.

Links to other EEA indicators

The indicators presented in this report can be regarded as part of a broader set of indicators that the EEA uses to present the key relationships in the causality chain for environmental and sustainability issues; from socio-economic driving forces, to pressures, state of the environment, impacts and societal response actions.

EEA has established a core set of indicators, for three main purposes: to provide a manageable and stable basis for indicator-based reporting, to prioritise improvements in the quality and geographical coverage of data flows, especially Eionet priority data flows, and to streamline EEA/Eionet's contributions to other European and global indicator initiatives, for example EU structural indicators and EU sustainable development indicators. The EEA core set of indicators (CSI) comprises 37 indicators representing 10 different categories. The 'climate change' category contains two relevant impact-related indicators (global/European temperature and greenhouse gas concentration) which are fully consistent with the corresponding indicators in this report (for more information, see http://themes.eea.europa.eu/IMS/CSI).

Other specifically relevant indicator sets are those related to biodiversity, inland water and marine. Various indicators within these themes are related to the indicators presented in this report. For such cases the indicators included in this report have been made as consistent as feasible regarding the data sources, methodologies and key messages.

For biodiversity the key process is SEBI 2010 (Streamlining European 2010 Biodiversity Indicators). This process aims to measure and help achieve progress towards the target of halting biodiversity loss by 2010 and has compiled a first set of 26 indicators. An assessment report on Europe's progress towards the 2010 target based on these indicators will be published by the EEA in 2009 (for more information, see http://www.eea.europa.eu/themes/biodiversity/eea-activities).

For inland water, reliable, high quality information about the environmental state of surface waters is essential for water management and for improving the environmental quality of Europe's waters, especially in relation to the Water Framework Directive. EEA is preparing various state of the environment (SOE) assessments of Europe's waters: assessment of the state and trends in relation to the Water Framework Directive, using indicators like the EEA Core Set of Indicators and other more specific indicators; broader assessment of specific water-related issues, such as eutrophication, hazardous substances, water abstraction and use, hydro-morphological impacts as well as goods and services deriving from aquatic ecosystems; and assessment of the impact on water resources of specific sectors, such as agriculture, hydropower, industry, navigation, tourism and water management (for more information, see http://www.eea.europa.eu/themes/water.

For the marine topic, the EEA is leading the process for developing a common pan-European set of indicators for the marine environment which was started under the European Marine Monitoring and Assessments (EMMA) Working Group ([1]). This work will support the implementation of the Marine Strategy Framework Directive and the further development of the EEA's pan-European marine assessments. In addition to this work, the EEA has also developed indicators based on operational oceanography, such as indicators on sea-level rise and sea surface temperature.

Data and information sources for this report

This report uses recorded data and model results to assess past and future climate change and its impact. While recorded data are a good source for the description of past trends of measurable factors, models are needed for the assessment of complex parameters which cannot be measured directly and for the assessment of future trends. All information on indicators presented in this report is subject to various types of uncertainty. These can result from gaps in knowledge of climate-change processes, insufficient data availability, difficulties in attributing an observed change to climate change, and a wide range of possible future socio-economic developments and levels of emission of greenhouse gases. Data sources, projections and uncertainties are briefly addressed in the description of each indicator and explained in more detail in Chapter 8.

Presentation of indicators

The presentation of each indicator comprises:

- key messages that summarise observed and projected trends;
- a relevance section that explains the policy, socio-economic and environmental relevance, possible adaptation options and uncertainties related to the indicator;
- past trends based mainly on analysis of long time series of reliable observations;
- projections (future trends), based mainly on results from existing global IPCC models and scenarios adapted to the European situation.

([1]) More information on the work done under EMMA and follow-up can be found at: http://circa.europa.eu/Public/irc/env/marine/library?l=/workingsgroups/europeansmarinesmonitori/emma_30-31_2007/3_-_report/emma_2007_070810doc/_EN_1.0_&a=d.

5.2 Atmosphere and climate

5.2.1 Introduction

Europe's climate shows considerable regional 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 (see Box 5.1).

This section describes the changing climatic and atmospheric conditions. The indicators are global and European temperature, precipitation, temperature and precipitation extremes, storms and storm surges, and atmospheric ozone concentration. Whereas most indicators focus on Europe, global temperature has been included because of the EU policy target to limit the global average temperature increase to a maximum of 2 °C above pre-industrial levels, in order to keep climate change at a manageable level and reduce the likelihood of irreversible disruptions.

Box 5.1 Atmospheric circulation patterns in Europe

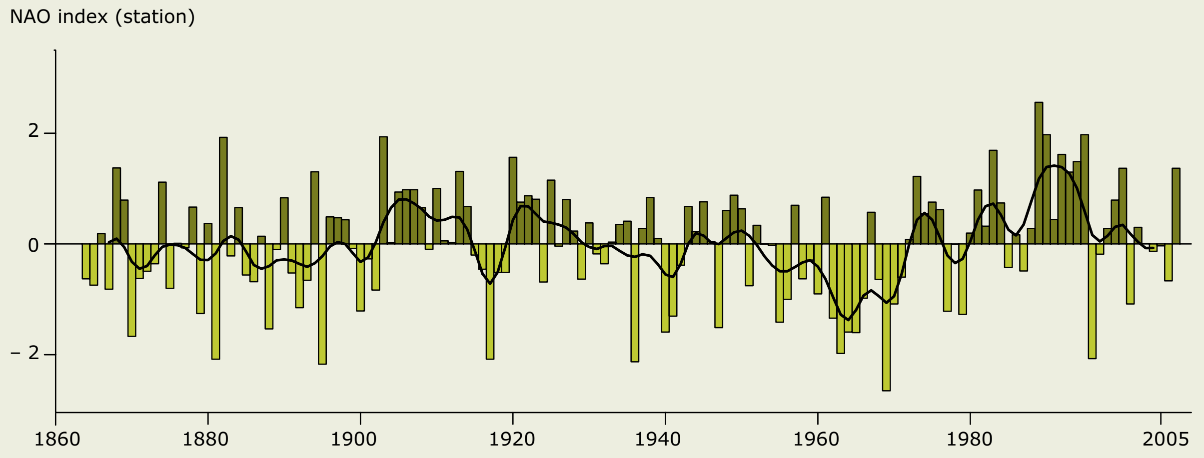
The atmospheric circulation moves air masses with their own specific characteristics, like temperature and humidity, over long distances. Important for the European climate is the prevailing western circulation at mid latitudes that directs the oceanic air masses inland over the continent. Stronger western 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.

The indicators represent different characteristics of the climate system that have diverse impacts on physical and biological systems and on human society; these can be independent of each other or, more often, have combined effects. High temperatures and reduced precipitation, for example, may lead to more intense droughts. Droughts are addressed in Section 5.2.5 (precipitation extremes in Europe), which gives different definitions of drought that have different consequences for the sectors involved. Temperature and precipitation extremes are both included because they have the largest impacts on society and the environment. Coastal areas are most vulnerable to storms, and damage caused by storms may be aggravated by floods related to storm surges. Within limits and at a cost, adaptation options are available for many of the consequences of changes in these extremes. If applicable, these will be mentioned in the sections on the individual indicators. Surface ozone concentrations can increase as a result of temperature increases and chemical reactions of air pollutants emitted by human activities in the lower atmosphere. Ozone is not only a greenhouse gas, higher ozone concentrations have adverse health effects — in particular the elderly — and on the environment.

Data availability and accuracy have been two of the important criteria for the selection of indicators. In general, the data availability of climate indicators is good compared with other indicators, although reliable data, particularly on the regional scale, can be scarcer. Air temperature data are the most available and reliable; precipitation and wind data are less available and — if available — more variable across different regions in Europe. Projections for precipitation and wind are more uncertain than those for temperature, and depend on future, still uncertain, atmospheric circulation patterns.

5.2.2 *Global and European temperature*

Key messages

Global

- The global (land and ocean) average temperature up to 2007 was 0.8 °C higher than pre-industrial levels (1850–1899 average). For land only, the average was 1 °C higher.

- The rate of increase of global average temperature has increased from 0.1 °C per decade over the past 100 years to 0.2 °C per decade in the past decade.

- The best estimates for projected global warming during this century are a further rise in average temperature of between 1.8 and 4.0 °C for different scenarios that assume no further/additional action to limit emissions.

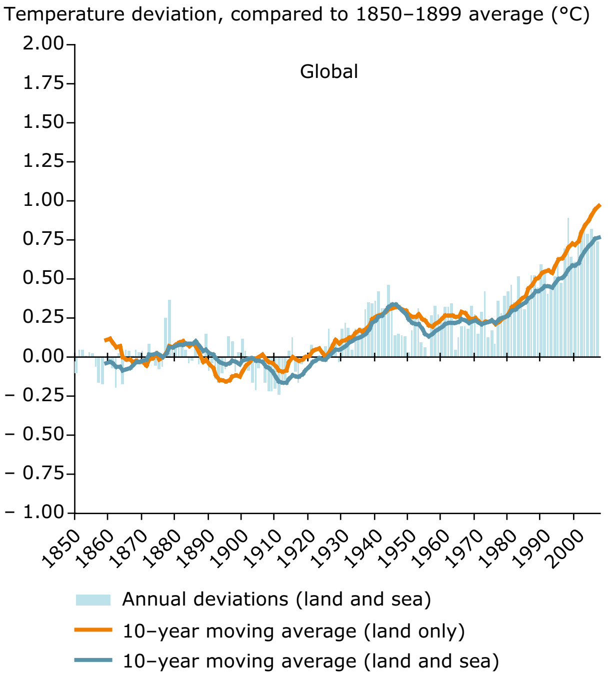
Europe (*)

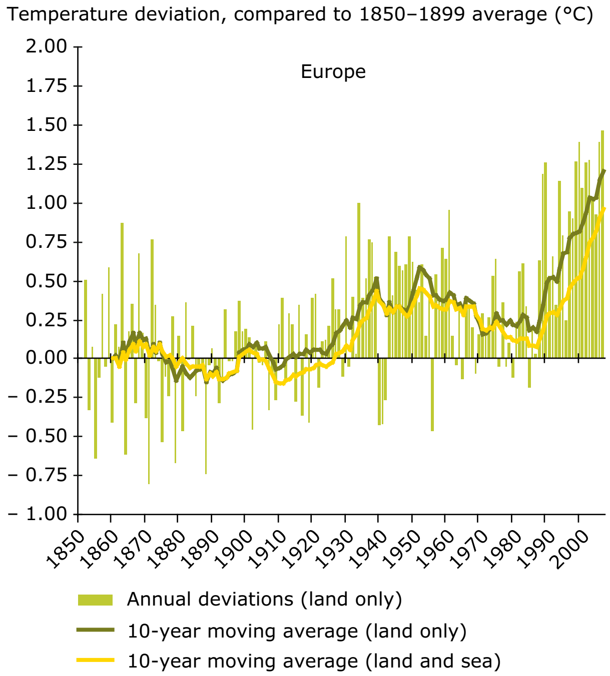
- Europe has warmed more than the global average. The annual average temperature for the European land area up to 2007 was 1.2 °C above pre-industrial levels, and for the combined land and ocean area 1 °C above. Eight of the 12 years between 1996 and 2007 were among the 12 warmest years since 1850.

- The annual average temperature is projected to rise this century by 1–5.5 °C (best estimate) with the largest warming over eastern and northern Europe in winter, and over south-western and Mediterranean Europe in summer.

(*) Europe is defined as the area between 35 and 70°N, – 25 and 30°E, plus Turkey (35–40°N, 30–45°E).

Figure 5.2 Observed global and European annual average temperature deviations, 1850–2007





Note: The source of the original data is the Climatic Research Unit of the University of East Anglia. The global mean annual temperature deviations are in the source in relation to the base period 1961–1990. The annual deviations shown in the chart have been adjusted to be relative to the period 1850–1899 to better monitor the EU objective not to exceed 2 °C above pre-industrial values. Over Europe average annual temperatures during the real pre-industrial period (1750–1799) were very similar to those during 1850–1899.

Sources: Climate Research Unit (http://www.cru.uea.ac.uk/cru/data/temperature/) (left); KNMI (http://climexp.knmi.nl/) (right).

Relevance

Of all the parameters used in monitoring and projecting climate change, air temperature is the nearest to our perception of 'climate'. Fortunately, air temperature data are also the most reliable climate data. There is a dense network of stations across the world — especially in Europe — with standardised measurements and often sophisticated quality control systems and homogeneity procedures. These provide high-resolution temperature information. Monthly information is available for long time-series (standardised from 1850 onwards). Time series with daily data generally start later. Furthermore, climate models have become increasingly sophisticated over recent years, with the uncertainty in the longer-term temperature projections being especially related to uncertainties in future emissions of greenhouse gases rather than uncertainties in modelling the climate system.

Temperature changes affect almost all the indicators included in this report, directly or indirectly. The projected temperature rise may have some beneficial impacts in the northern part of Europe (at least for a limited period), but the impacts in most parts of Europe are and will be adverse. In relation to climate change policy, the global temperature rise is relevant because of the EU objective of limiting global average temperature increase to a maximum of 2 °C above pre-industrial levels, as described in Section 5.1. Monitoring temperature change is thus also relevant for comparing actual developments with this target. Some other studies have proposed additional 'sustainable' targets of limiting the rate of temperature change, ranging from 0.1 °C to 0.2 °C per decade, based on the limited capability of ecosystems to adapt (Rijsberman and Swart, 1990; WBGU, 2003; van Vliet and Leemans, 2006). In this context it is important to note that land warms faster then the oceans.

Past trends

Global warming is accelerating, most probably as a result of the observed increase in anthropogenic greenhouse gas concentrations (IPCC, 2007a). Global average temperature (land and ocean) in 2007 was 0.8 °C above the pre-industrial level (defined as the 1850–1899 average (IPCC, 2007a)); the increase over land only was 1.0 °C. Eleven of the last 12 years (1996–2007) rank among the 12 warmest years (the exception being 1996). The warmest two years on record were 2005 and 1998 (see Figure 5.2). The rate of increase of global average temperature has increased from an average of 0.1 °C per decade over the past 100 years to 0.2 °C per decade over the past 10 years (all values represent land and ocean area) (IPCC, 2007a).

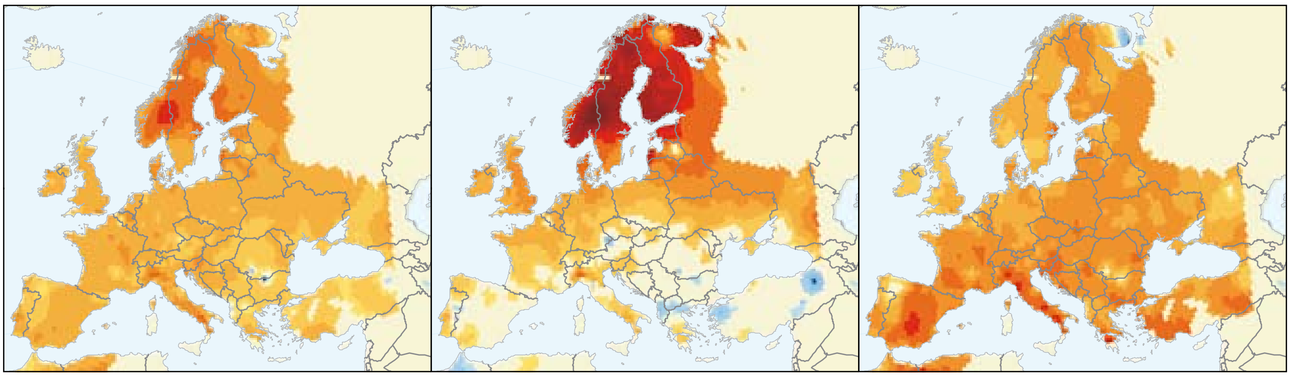
Photo: © Stockxpert

Europe has warmed slightly more than the global average. The annual average temperature for the European land area in 2007 was 1.2 °C above pre-industrial levels, and for the combined land and ocean area 1 °C above. Eight of the 12 years between 1996 and 2007 were among the 12 warmest years since 1850s in Europe. Seasonally, Europe warmed most in spring and summer. Remarkably, autumn saw almost no warming. Geographically, particularly significant warming has been observed in the past 50 years over the Iberian Peninsula, in central and north-eastern Europe and in mountainous regions (Böhm *et al.*, 2001; Klein Tank, 2004). In the past 30 years, warming was strongest over Scandinavia, especially in winter, whereas the Iberian Peninsula warmed in summer (Map 5.1).

Projections

The global and European average temperature is projected to continue to increase. Globally, the projected increase in this century is between 1.8 and 4.0 °C (best estimate), and is considered likely (66 % probability) to be between 1.1 and 6.4 °C for the six IPCC SRES scenarios and multiple climate models (see Chapter 4), comparing the 2080–2100 average with the 1980–1999 average. These scenarios assume that no additional policies to limit greenhouse gas emissions are implemented (IPCC, 2007a). The range results from the uncertainties in future socio-economic development and in climate models. The EU 'sustainable' target of limiting global average warming to not more than 2.0 °C above pre-industrial level is projected to be exceeded between 2040 and 2060, for the all six IPCC scenarios.

Map 5.1 Observed temperature change over Europe 1976–2006



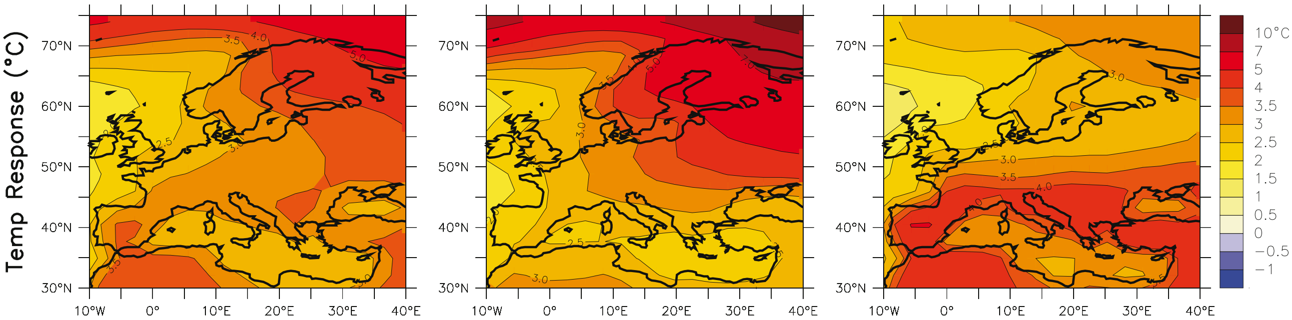
Observed temperature change over Europe during the period 1976–2006
Left: annual mean; middle: winter (DJF); right: summer (JJA)

°C per decade

−2 −1.8 −1.6 −1.4 −1.2 −1 −0.8 −0.6 −0.4 −0.2 0 0.2 0.4 0.6 0.8 1 1.2 1.4 1.6 1.8 2

Source: The climate dataset is from the EU-FP6 project ENSEMBLES (http://www.ensembles-eu.org) and the data providers in the ECA&D project (http://eca.knmi.nl).

Map 5.2 Modelled change in mean temperature over Europe between 1980–1999 and 2080–2099



Note: Left: annual; middle: winter (DJF); right: summer (JJA) changes in °C for the IPCC-SRES A1B emission scenario averaged over 21 models (MMD-A1B simulations).

Source: Christensen *et al.*, 2007. Published with the permission of the Intergovernmental Panel on Climate Change.

The annual average temperature for Europe is projected to increase by 1.0–5.5 °C (comparing 2080–2100 with the 1961–1990 average). This range takes into account the uncertainties in future socio-economic development by including two of the IPCC-SRES scenarios (the high emissions A2 and the medium emissions A1b), and the uncertainties in the climate models (Christensen *et al.*, 2007) (Map 5.2). The warming is projected to be greatest over eastern Europe, Scandinavia and the Arctic in winter (December to February), and over south-western and Mediterranean Europe in summer (June to August) (Giorgi *et al.*, 2004; IPCC, 2007a). The temperature rise in parts of France and the Iberian Peninsula may exceed 6 °C, while the Arctic could become on average 6 °C and possibly 8 °C warmer than the 1961–1990 average (IPCC, 2007a, 2007b; ACIA, 2004).

Box 5.2 Climate reanalysis

Climate change can be monitored in different ways. In addition to the use of direct surface observations, climate variables such as near-surface temperature can also be estimated using the so-called reanalysis approach. Reanalysis uses a modern data assimilation system to combine historic data from different sources, including satellites, radiosondes, aircraft, surface data, and ships. This approach produces a comprehensive atmospheric data set, including parameters that are not well observed. The results can be used to assess climate change and climate variability.

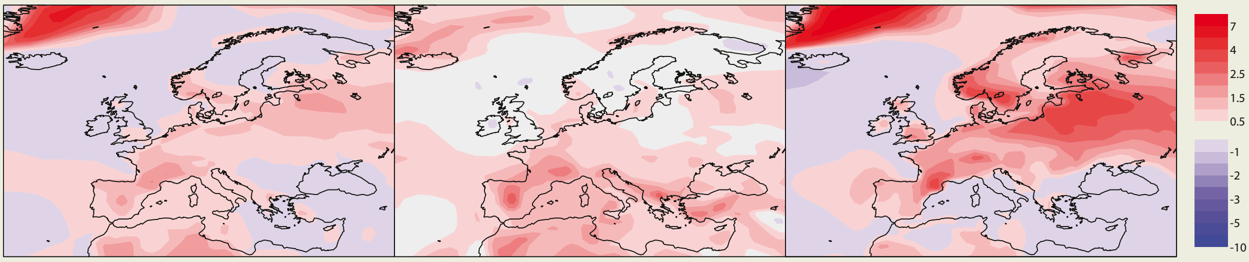
Map 5.3 is an example of the product of the latest European reanalysis, ERA-40, carried out by the European Centre for Medium-Range Weather Forecasts (ECMWF), with support from Europe's National Weather Services and the European Commission's Fifth Framework Programme. Estimated temperature changes across Europe in ERA-40 are consistent with observed changes over a similar period based on the station data (Map 5.1).

Global reanalysis provides a basis for regional reanalysis and downscaling projects (e.g. the BALTEX regional reanalysis project for the Baltic Sea). These projects can provide spatially detailed climate change data to support studies of local climate change and climate impacts.

An advantage of the reanalysis methodology is its complete temporal and spatial coverage, and the additional information that a model-based data assimilation system can provide. As such it is very useful for estimating sparsely observed climate variables such as wind speed/direction. However when a reanalysis is extended over long periods that include major changes (e.g. in the observation system), caution must be exercised when interpreting the analysed data sets, in particular in regions with insufficient observational coverage (IPCC, 2007a; Bengtsson *et al.*, 2004; Third WCRP International Conference on Reanalysis Conference Statement, 2008). For the purposes of climate analysis in particular, ERA-40 estimates of trends and low-frequency variability are more accurate for the post-satellite era from 1979 onwards than for the earlier period (Simmons *et al.*, 2004).

Overall, with the continuing development of analysis and reanalysis of climate data for the oceans, land and sea ice, and the initiatives towards providing more detailed information through regional reanalysis, there is huge potential for further progress and improved knowledge of the past climate in Europe.

Map 5.3 The linear trend in surface temperature over Europe 1958–2001



Note: Linear trend (°C/50 years) calculated from ERA-40 data for the period 1958 to 2001. The left panel is for the annual change, the central panel for summer (JJA) and the right panel for winter (DJF).

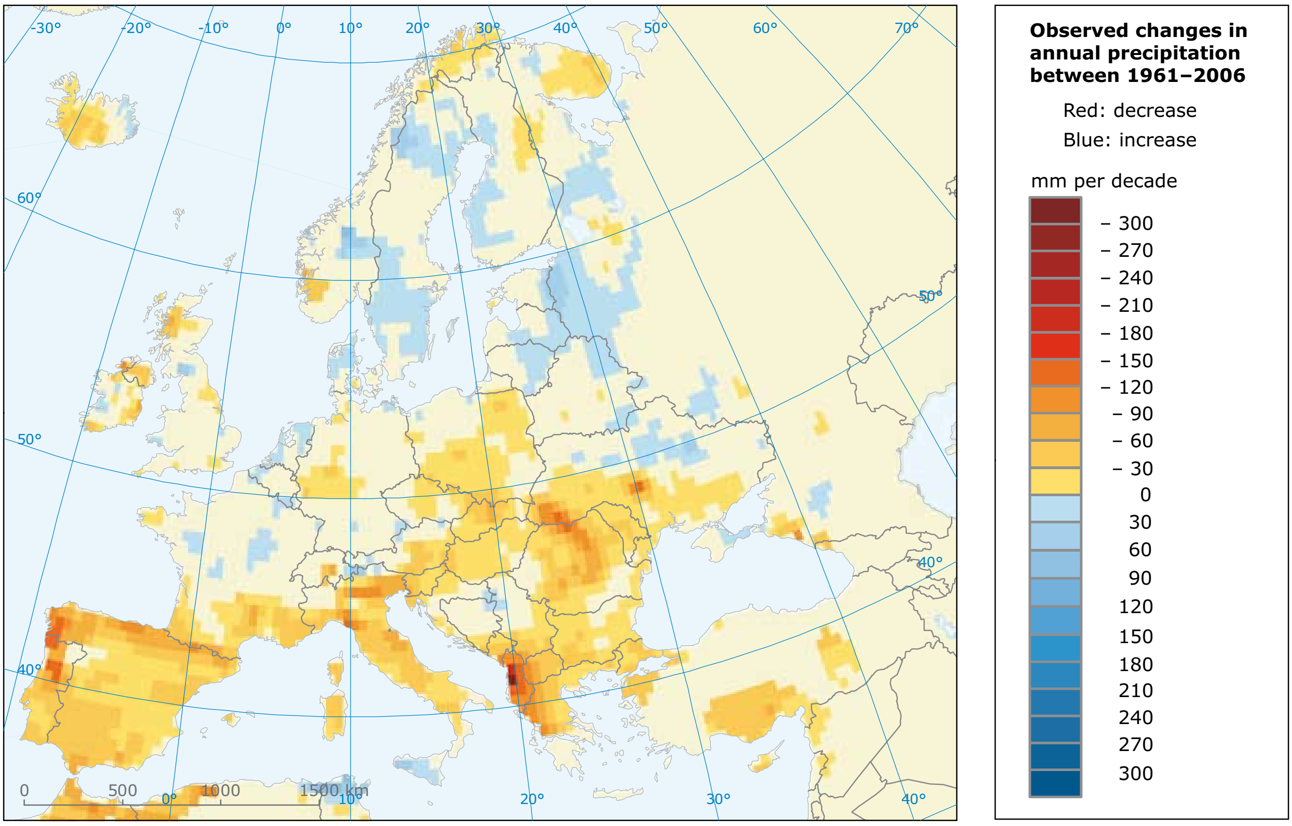
Source: European Centre for Medium-Range Weather Forecasts (ECMWF), ERA-40 Global Atmospheric Reanalysis, (http://www.ecmwf.int).

5.2.3 European precipitation

Key messages

- Annual precipitation trends in the 20th century showed an increase in northern Europe (10–40 %) and a decrease in some parts of southern Europe (up to 20 %).
- Mean winter precipitation has increased in most of western and northern Europe (20 to 40 %), whereas southern Europe and parts of central Europe were characterized by drier winters.
- Models project an increase in winter precipitation in northern Europe, whereas many parts of Europe may experience dryer summers. But there are uncertainties in the magnitude and geographical details of the changes.

Map 5.4 Observed changes in annual precipitation 1961–2006



Note: Data are in mm per decade, blue means an increase, red a decrease. The observations indicate that large decadal scale variability in precipitation amount is superposed on the long time scale trends described above. This variability is partly related to the decadal scale variability in atmospheric circulation anomalies (see Box 5.1). Calculating trends over shorter time periods may therefore lead to different results. The yellow color indicates that the trend 1961–2006 is not significant at level 25 %.

Source: The climate dataset is from the EU-FP6 project ENSEMBLES (http://www.ensembles-eu.org) and the data providers in the ECA&D project (http://eca.knmi.nl).

Relevance

Precipitation is a major component of the hydrological cycle. The amount and spatial distribution of European precipitation is strongly influenced by circulation patterns (see Box 5.1). Most precipitation over Europe is connected with the advection of maritime air masses from the

Atlantic and the Mediterranean. The combination of changes in the precipitation regime and increases in air temperature can lead to extreme hydrological evens such a flooding and droughts (see e.g. Sections 5.2.5, 5.5.3 and 5.5.4). Some systems or sectors, closely connected with the hydrological cycle, are very sensitive to the combined effects of higher temperatures and changed precipitation characteristics. Within limits and at a cost, adaptation to many of the impacts is possible. These options will be briefly mentioned in the individual indicator sections.

Homogenous time series of monthly precipitation data and interpolation and gridding methods enable analysis of various periods from 1901 on various temporal and spatial scales. However, differences between climate models for future precipitation projections indicate higher uncertainty for regional and seasonal results than for temperature projections and observed precipitation trends.

Past trends

Precipitation in Europe generally increased over the 20th century, on average 6–8 % between 1901 and 2005. Geographically, there is a variation (see Map 5.4); an increase in north-west Europe, partly due to stronger advection of wet Atlantic air masses to this part of the continent. Drying has been observed in the Mediterranean and eastern Europe and no clear trends have been observed in western Europe (Norrant and Douguédroit, 2006).

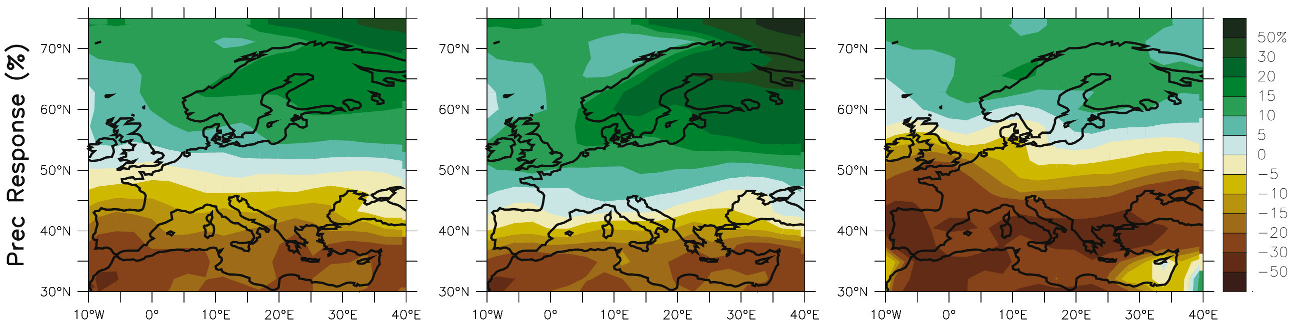
Mean winter (December–February) precipitation is increasing 20–40 % in most of western and northern Europe (Klein Tank *et al.*, 2002), because western circulation was stronger in winter. Conversely, southern Europe and parts of central Europe were characterized by a drier winter. Trends in spring and autumn were not significant.

Projections

Climate models project changes in precipitation that vary considerably from season to season and across regions. Geographically, projections indicate a general precipitation increase in northern Europe and a decrease in southern Europe. The change in annual mean between 1980–1999 and 2080–2099 for the intermediate IPCC SRES A1B projections varies from 5 to 20 % in northern Europe and from – 5 to – 30 % in southern Europe and the Mediterranean (Map 5.5). Many impact studies (see other indicators) use the high emission A2 scenario. Under this scenario the projected changes are mostly larger.

Seasonally, models project a large-scale increase in winter precipitation in mid and northern Europe. Many parts of Europe are projected to experience dryer summers (Map 5.5). Relatively small precipitation changes are projected for spring and autumn (Räisänen *et al.*, 2004; Kjellström, 2004).

Map 5.5 Modelled precipitation change between 1980–1999 and 2080–2099



Note: Left: annual; middle: winter (DJF); right summer (JJA) changes % for the IPCC-SRES A1B emission scenario averaged over 21 models (MMD-A1B simulations).

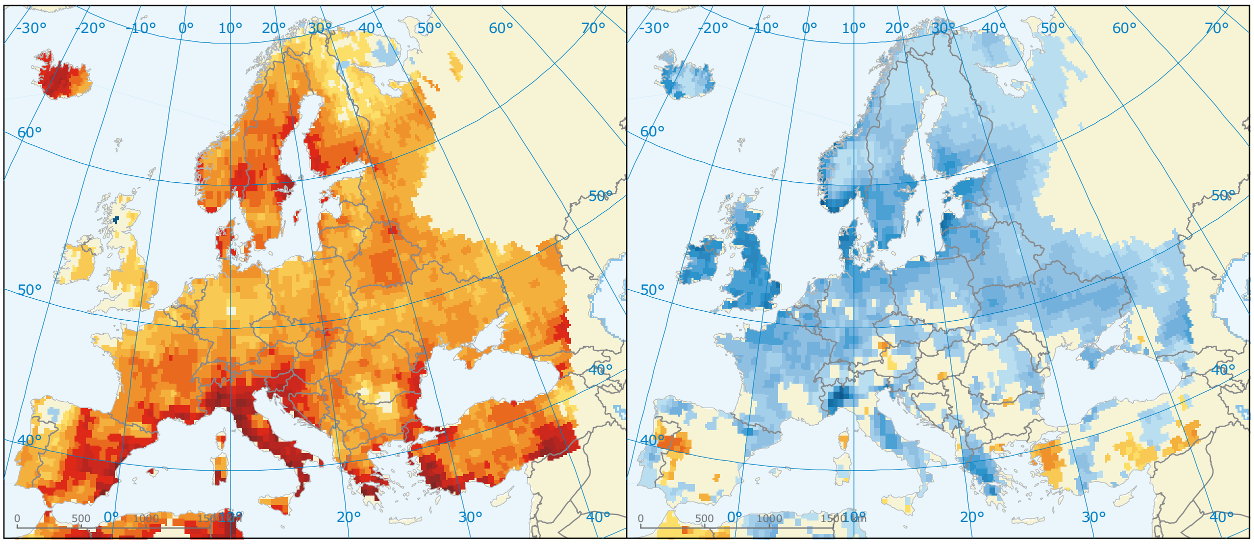
Source: Christensen *et al.*, 2007. Published with the permission of the Intergovernmental Panel on Climate Change.

5.2.4 *Temperature extremes in Europe*

Key messages

- Extremes of cold have become less frequent in Europe while warm extremes have become more frequent. The frequency of hot days almost tripled between 1880 and 2005.
- For Europe as a whole heat waves are projected to increase in frequency, intensity and duration, whereas winter temperature variability and the number of cold and frost extremes are projected to decrease further. The European regions projected to be most affected are the Iberian Peninsula, central Europe including the Alps, the eastern Adriatic seaboard, and southern Greece.

Map 5.6 Observed changes in warm spells and frost days indices 1976–2006



Observed changes in duration of warm spells in summer (left) and frequency of frost days in winter (right), in the period 1976–2006

Days per decade: − 10, − 9, − 8, − 7, − 6, − 5, − 4, − 3, − 2, − 1, 0, 1, 2, 3, 4, 5, 6, 7, 8, 9, 10

Source: The climate dataset is from the EU-FP6 project ENSEMBLES (http://www.ensembles-eu.org) and the data providers in the ECA&D project (http://eca.knmi.nl).

Relevance

As seen by the public, climate change manifests itself most clearly through changes in the frequency of weather extremes and their impacts. Nearly all adaptation measures relate to changes in climate extremes. Extreme temperature events may lead to heat waves and intensive and long-lasting droughts, having, in turn, many impacts on natural ecosystems and society (e.g. agriculture, public health).

The time series for studying temperature extremes are based on daily data. More than 50-year of European time-series data allow detailed assessment of extreme events.

Past trends

High-temperature extremes like hot days, tropical nights, and heat waves ([2]) have become more frequent, while low-temperature extremes (e.g. cold spells, frost days) have become less frequent (Klein Tank *et al.*, 2002; IPCC, 2007a; Map 5.6). The average length of summer heat waves over Western Europe doubled over the period 1880 to 2005 and the frequency of hot days almost tripled (Della-Marta *et al.*, 2007).

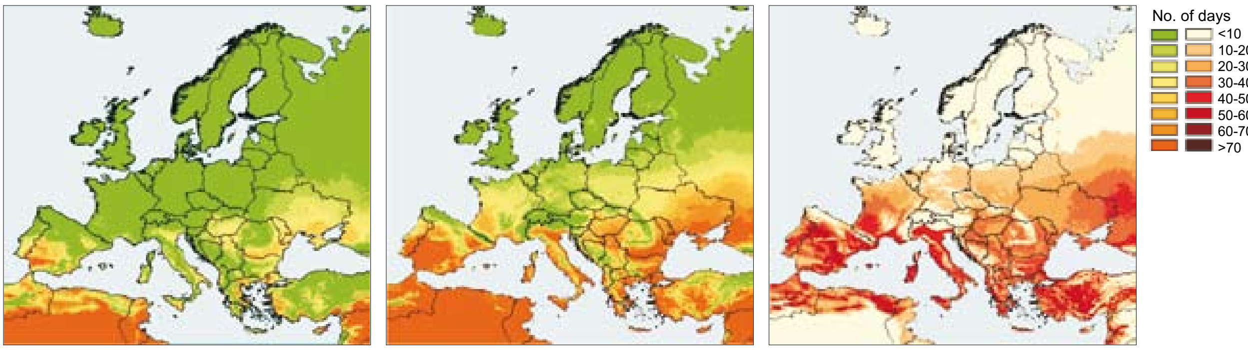
Photo: © Stockxpert

Projections

Extreme high temperature events across Europe, along with the overall warming, are projected to become more frequent, intense and longer this century (Schär *et al.*, 2004; Tebaldi *et al.*, 2006; IPCC, 2007a, 2007b; Beniston *et al.*, 2007). Likewise, night temperatures are projected to increase considerably (Map 5.7), possibly leading to additional health problems and even mortality (Halsnæs *et al.*, 2007; Sillman and Roekner, 2008), at least partly compensated by reduced mortality in winter (see Section 5.10.2).

Geographically, the maximum temperature during summer is projected to increase far more in southern and central Europe than in northern Europe, whereas the largest reduction in the occurrence of cold extremes is projected for northern Europe (Kjelström *et al.*, 2007; Sillman and Roekner, 2008). Under the A2 scenario, central Europe, for example, is projected to experience the same number of hot days as are currently experienced in Spain and Sicily by the end of the 21st century (Beniston *et al.*, 2007).

Map 5.7 Modelled number of tropical nights over Europe during summer (June–August) 1961–1990 and 2071–2100



Note: Reference period (1961–1990) (left), scenario period (2071–2100) (centre) and change between periods (right). Data were used from the Danish Meteorological Institute (DMI) with the HIRHAM4 regional climate model with boundary conditions of the HadCM3 model and the IPCC-SRES A2 emission scenario.

Source: Dankers and Hiederer, 2008.

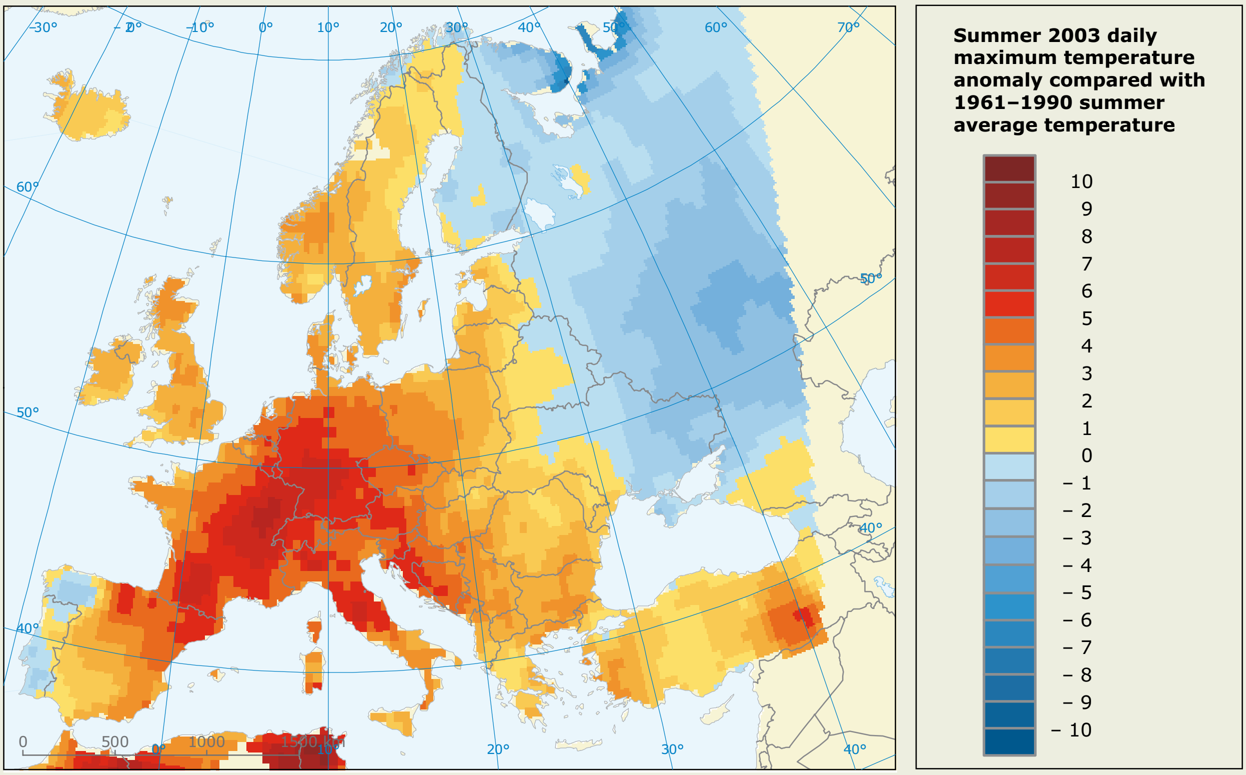
([2]) A hot day is defined as one where the daily maximum temperature exceeds the long-term daily 95th percentile of daily maximum temperature; a tropical night is one with minimum temperature > 20 °C, a heat wave is a period of at least six consecutive days with maximum temperature > 30 °C); a cold spell is a period of at least six consecutive days with minimum temperature below the 10th percentile of daily minimum temperature (e.g. for the period 1961–1990); frost days are defined as days with daily minimum temperature below 0 °C).

Box 5.3 The heat wave of summer 2003

Much of Europe was affected by a heat wave during the summer of 2003 (June, July and August). It is estimated that this was the hottest summer since at least 1500 (Luterbacher *et al.*, 2004). Seasonal temperatures were the highest on record in Germany, Switzerland, France and Spain (Map 5.8). Average summer (June–August) temperatures were far above the long-term mean, by up to five standard deviations, implying that this was an extremely unlikely event under current climatic conditions (Schär and Jendritzky, 2004). Hot summers like 2003 may, however, become much more frequent during the second part of the 21st century (Beniston, 2007; Dankers and Hiederer, 2008).

The 2003 heat wave was associated with a particular air pressure field pattern over Europe, leading to an advection of hot air from the south which reinforced the strength and persistence of the heat waves. Nearly all radiation from the sun was converted to heat because of the soil and vegetation dryness. At many locations, day-time temperatures rose to more than 40 °C. In the European Alps, the average thickness loss of glaciers reached about 3 m water equivalent, nearly twice as much as during the previous record year of 1998 (WMO, 2004; see Section 5.3.2). Annual precipitation deficits up to 300 mm caused droughts in many areas which resulted in reduced agricultural production (Section 5.6.2), more extensive forest fires (Portugal, Section 5.6.6), and record low levels of many major rivers (e.g. Po, Rhine, Loire and Danube; Section 5.7.2). In all the affected countries together, more than 70 000 additional deaths were related to the 2003 heat waves (Section 5.10.2).

Map 5.8 Summer 2003 (June–August) daily maximum temperature anomaly



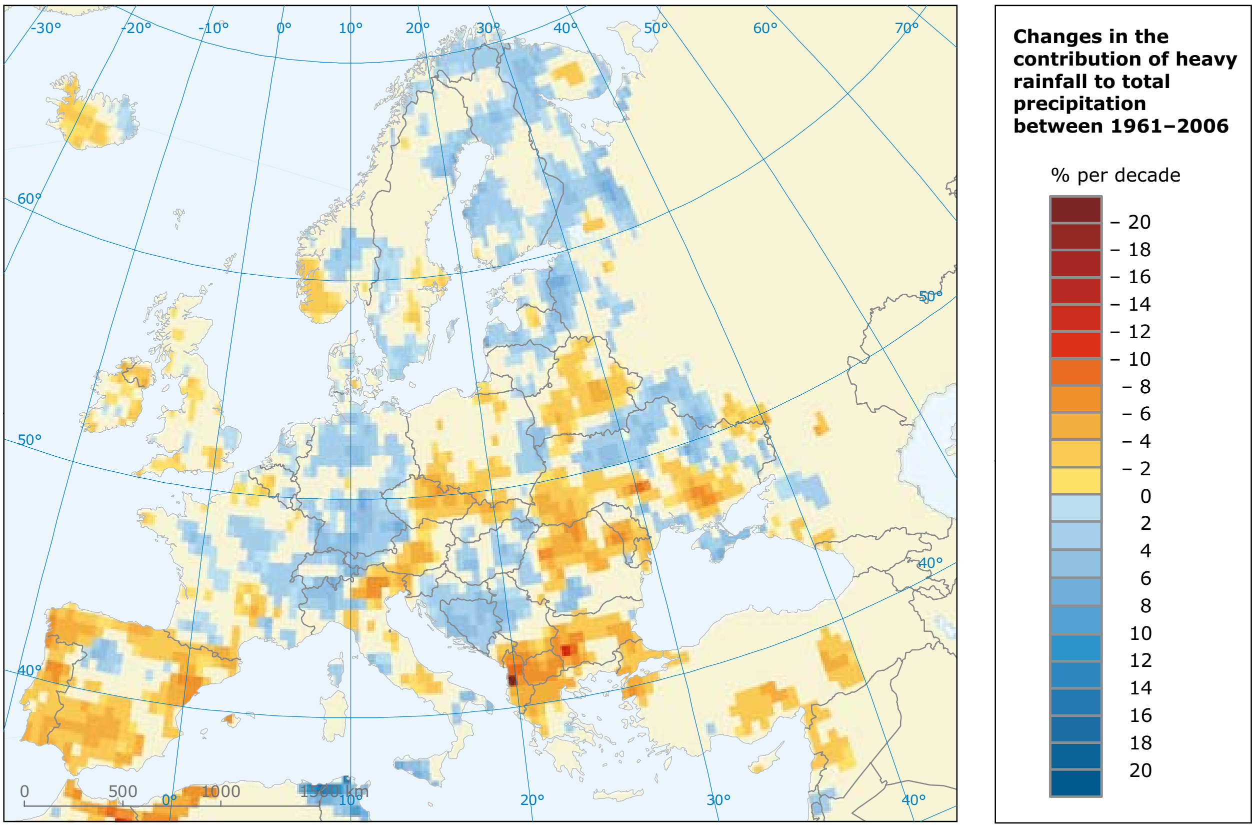
Source: The climate dataset is from the EU-FP6 project ENSEMBLES (http://www.ensembles-eu.org) and the data providers in the ECA&D project (http://eca.knmi.nl).

5.2.5 *Precipitation extremes in Europe*

Key messages

- For Europe as a whole, the intensity of precipitation extremes such as heavy rain events has increased in the past 50 years, even for areas with a decrease in mean precipitation such as central Europe and the Mediterranean.
- The proportion of Europe experiencing meteorological drought conditions did not change significantly during the 20th century.
- For Europe as whole, heavy precipitation events are projected to continue to become more frequent.
- Dry periods are projected to increase in length and frequency, especially in southern Europe.

Map 5.9 Changes in the contribution of heavy rainfall to total precipitation 1961–2006



Source: The climate dataset is from the EU-FP6 project ENSEMBLES (http://www.ensembles-eu.org) and the data providers in the ECA&D project (http://eca.knmi.nl).

Relevance

Both high and low precipitation extremes (high intensity or long-lasting rain and droughts, respectively) can lead to periods with a high amount of total precipitation or with precipitation deficit. The periods can range from minutes (e.g. in case of intense showers) to days, weeks or even months (with long-lasting rain events or absence of precipitation). Low precipitation extremes can lead to droughts (see Box 5.4). High precipitation extremes can result in fast flash floods, sewerage system failure and land-slides, or devastating floods, affecting large catchments and having longer duration.

Precipitation extremes can be described in different ways. Precipitation deficits are often expressed as the number and duration of dry periods (e.g. the number of consecutive dry days) and high precipitation events as the number of wet days, consecutive wet days, and the frequency and intensity of heavy precipitation events (Klein Tank and Können, 2003).

The time series for studying precipitation extremes are based on daily data. As for temperature extremes, there is more than 50 years of European time series data available for statistical analyses.

Past trends

The number of extreme precipitation events has increased over most of the European land area, linked to warming and increases of atmospheric water vapour. For Europe as a whole, also the intensity of extreme precipitation such as heavy rain has increased in the past 30 years, even for areas with a decrease in mean precipitation, such as central Europe and the Mediterranean. In particular, the contribution of heavy rain to total precipitation has increased (Map 5.9).

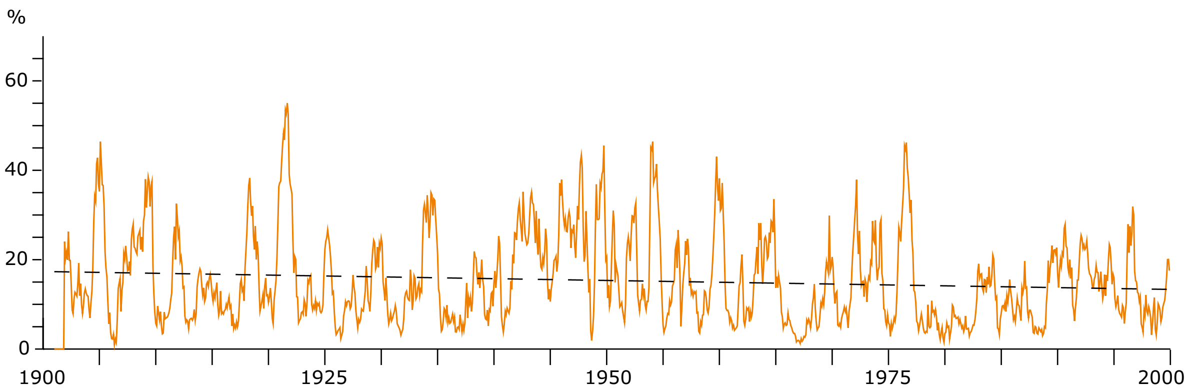
The proportion of Europe that has experienced extreme and/or moderate meteorological drought conditions did not change significantly during the 20th century (Figure 5.3) (Lloyd-Hughes and Saunders, 2002). Some drying trends were observed over central and eastern Europe, and western Russia. Similar, some trends were observed in winter/spring. Summer droughts showed no statistically significant trends in the period 1901–2002 (Robock *et al.*, 2005; van der Schrier *et al.*, 2006).

Projections

For Europe as whole it is likely (66 % probability) that heavy precipitation events will continue to become more frequent (IPCC, 2007a). In summer, the frequency of wet days is projected to decrease, but the intensity of extreme rain showers may increase. In addition, the frequency of several-day precipitation episodes is projected to increase. Geographically, there is considerable regional differentiation in the projections. Extreme precipitation events are projected to increase by 17 % in northern and 13 % in central Europe during the 21st century, with no changes projected in southern Europe (for the ECHAM 4 climate model, A1B scenario, Figure 5.4, Sillmann and Roeckner, 2008).

The combination of higher temperatures and reduced mean summer precipitation is expected to enhance the frequency and intensity of droughts across Europe. This can be illustrated, for example, by the projected number of consecutive dry days, defined as days with precipitation below 1 mm (Figure 5.5). In southern Europe, the maximum number of these days is projected to increase substantially during the 21st century. The longest dry period within a year may be prolonged here by one month at the end of 21st century. In central Europe, prolongation of longest dry period is by one week, and no prolongation is projected for northern Europe. Thus regions in Europe that are now dry are projected to become even more vulnerable.

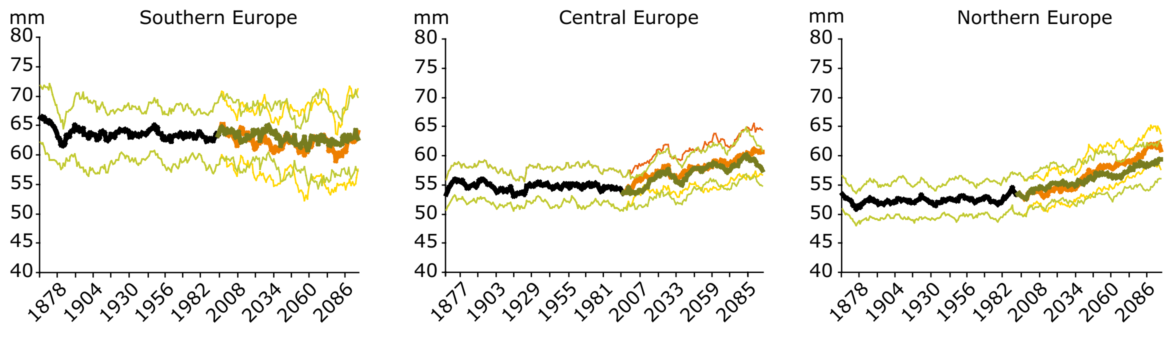
Figure 5.3 Percentage of Europe experiencing moderate drought conditions during the 20th century



Note: Expressed as standardized precipitation indices (SPI) for time scales of 12 months. The dashed line shows the linear trend. Errors are ± 2 standard errors in the gradient.

Source: Lloyd-Hughes and Saunders, 2002.

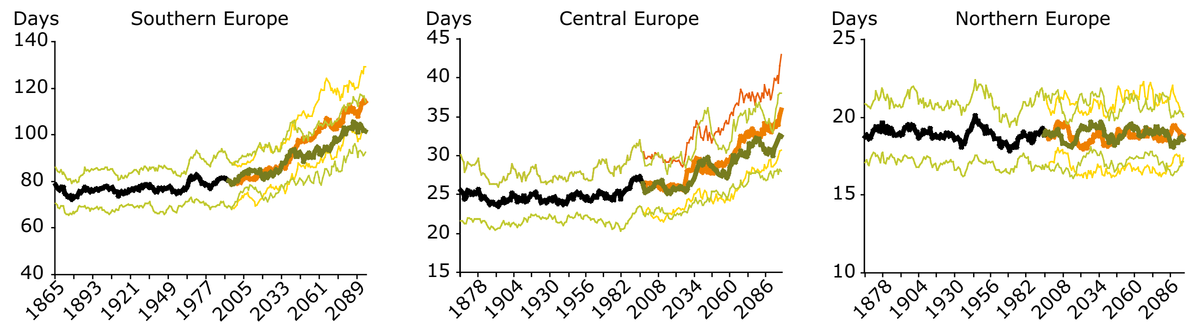
Figure 5.4 Simulated land average maximum 5-day total precipitation for different European regions (1860–2100)



Note: The 20th century (black), models simulations for IPCC SRES intermediate A1B (orange) and low B1 (green) emission scenarios. The respective ensemble means are displayed. The minimum and maximum of the ensemble members are indicated by thin green (B1) and yellow (A1B), respectively. Data are smoothed by 10-year running means.

Source: Sillmann and Roeckner, 2008.

Figure 5.5 Simulated land average maximum number of consecutive dry days for different European regions (1860–2100)



Note: The 20th century (black), models simulations for IPCC SRES intermediate A1B (orange) and low B1 (green) emission scenarios. The respective ensemble means are displayed. The minimum and maximum of the ensemble members are indicated by thin green (B1) and yellow (A1B), respectively. Data are smoothed by 10-year running means.

Source: Sillmann and Roeckner, 2008.

Box 5.4 Drought

Drought is a natural phenomenon, defined as sustained and extensive occurrence of below-average water availability. Drought should not be confused with aridity, which is a long-term average feature of a dry climate. Nevertheless, the most severe human consequences of drought can be found in arid regions, where water availability is naturally lower. Likewise, drought should not be confused with water scarcity, which reflects conditions of long-term imbalances between water supply and demands (e.g. van Lannen *et al.*, 2007).

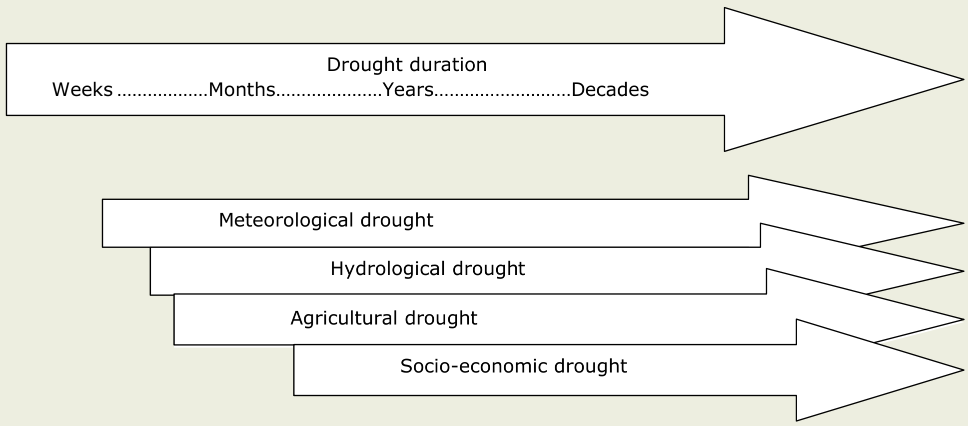
Droughts can affect both high and low rainfall areas of Europe and can develop over short periods of weeks and months or much longer periods of several seasons, years and even decades. In many cases drought develops gradually, making it difficult to identify and predict. The most common definitions and types of drought are:

- Meteorological drought: departure of precipitation from normal values for an extended period of time, the primary cause of the other types of drought.
- Hydrological drought: deficiencies in surface and subsurface water supplies, reflecting effects and impacts of meteorological droughts.
- Agricultural drought: a deficit of soil moisture affecting a particular crop at a particular time.
- Socio-economic drought: imbalance between supply and demand for an economic good, capturing both drought condition and human activities.

Precipitation is the primary factor controlling the origin and persistence of drought conditions for all types of drought. Deficiency of precipitation results in water shortage for some activity or for some group. The impacts of droughts on people and the environment result from a combination of the intensity and duration of drought events and the vulnerability of agricultural or water resources systems, including water management policies, the characteristics of regional and local water infrastructure, and social responses to drought situations. Drought is a phenomenon that is not constrained by international boundaries and can therefore grow to afflict many countries simultaneously and may stress relationships between them.

Climate projections indicate that in warmer conditions droughts may become longer-lasting and more severe in current drought-prone regions because of decreased rainfall and enhanced evaporation.

Figure 5.6 Diagram of drought types



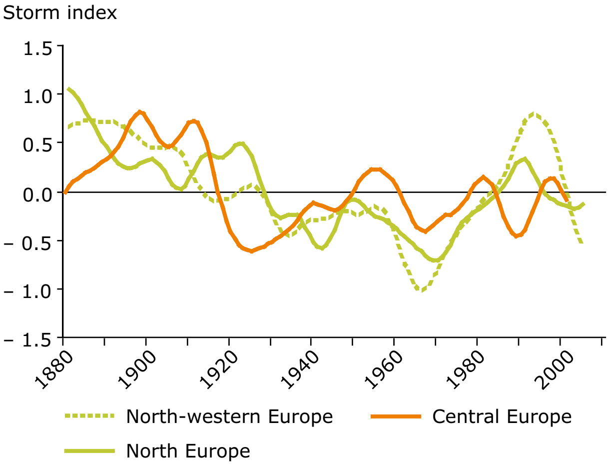
Source: Wade, 2007.

5.2.6 *Storms and storm surges in Europe*

Key messages

- There has been considerable variation, but no clear long term trend in storminess in Europe. Storm frequency was relatively high during the late 19th and early 20th century; then decreased in central and northern Europe. The recent high level is similar to the late 19th century level of storminess.
- Despite the variation in storminess, water levels along most vulnerable European coastlines of the North Sea and Mediterranean Sea have shown no significant storm-related variation.
- Extra-tropical storm tracks are projected to move pole-wards, with consequent changes in wind, precipitation, and temperature patterns, continuing the broad pattern of observed trends over the past half-century.
- Climate models indicate a slight decrease in the number of storms and an increase of the strength of the heaviest storms.
- Projections to the end of the 21st century show a significant increase in storm surge elevation for the continental North Sea and south-east England.

Figure 5.7 Storm index for various parts of Europe 1881–2005



Note: Positive values of the index mean higher storminess.

Source: Matulla *et al.*, 2007.

Relevance

Storms in Europe consist of extreme, near-surface damage-causing winds, associated with the passage of intense extra-tropical cyclones (Pinto *et al.*, 2007). Storms occur, in general, in north or north-western Europe all year, but in central Europe mainly between November and February. Storm surges are temporary increases in sea level, above the level of the tide, often causing coastal flooding. Storm events can have large impacts on vulnerable systems such as transport, forestry and energy infrastructures, and also on human safety.

Storm activity in Europe and the neighbouring part of the Atlantic is closely connected with atmospheric circulation (see Box 5.1). But the correlation between the NAO index and storminess across Europe varies with space and time. Direct wind observation data of sufficient quality are often lacking. Instead, storm intensity and frequency can be indirectly assessed through changes in the air pressure fields. Note that projections of changes in wind conditions are highly uncertain, mainly because of the uncertainty in atmospheric circulation projections.

Storm surges result from the combined action of atmospheric pressure and strong wind on the sea surface and occur mostly in shallow water. An increase in mean sea level will directly affect extreme levels. Changes in water depth can also influence the tidal component, modifying the extent of flooded areas. Future storm surge extremes are related both to storminess and to sea level changes.

Past trends

Storminess in Europe has shown considerable variation over the past century, but with no clear long-term trend. This is illustrated by means of the storm-index time series (Figure 5.7), based on air-pressure data. These series show that storminess in north-western, northern and central Europe was relatively high during the late 19th and early 20th century; then decreased in central Europe and northern Europe. The subsequent rise in the late 20th century was most pronounced in north-western Europe, while slow and steady in central Europe. Most recent years have shown average or calm conditions (Matulla *et al.*, 2007). On the local scale, station wind data can show different behaviour. Decreases and increases continuing over several decades can be seen at particular locations. For example a strong decrease in wind storms has been observed over the Netherlands during the past 40 years (Smits *et al.*, 2005).

Evaluating high tide levels along the North Sea in the past century showed clear changes in mean levels (related to sea-level rise) but no storm-related variations (von Storch *et al.*, 2002). Similarly, in the northern Adriatic Sea the trends for high sea levels and the subsequent occurrence of storm surges can not be associated with any trends in storminess (Lionello, 2005).

Photo: © Pavel Šťastný

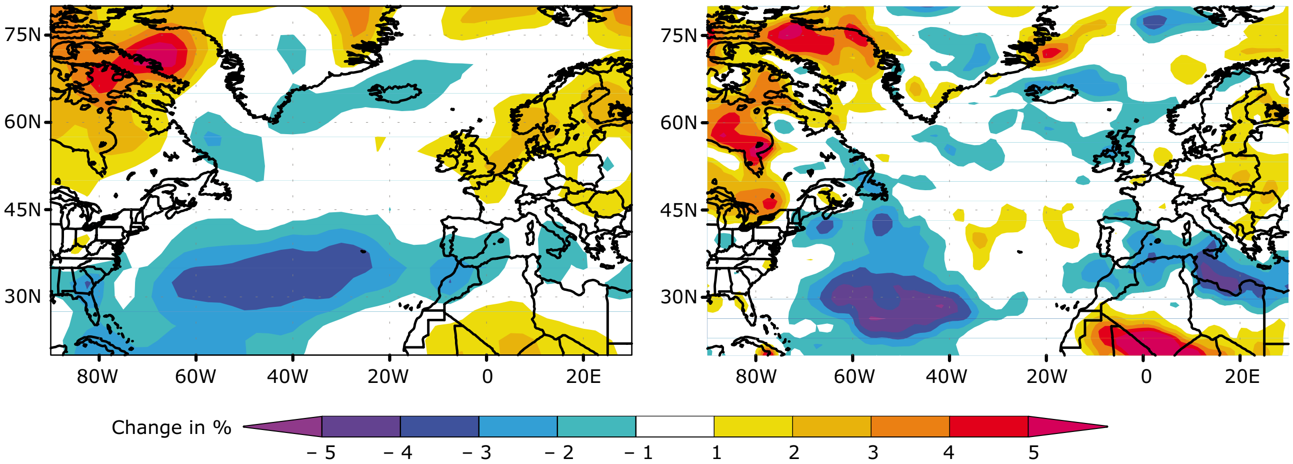
Projections

Extra-tropical storm tracks are projected to move pole-ward, with consequent changes in wind, precipitation, and temperature patterns, continuing the observed trends over the last half-century (IPCC, 2007a). The total number of storms is projected to decrease, but the strengths of the heaviest storms may increase, depending on the model used (see Map 5.10 showing different regional maximum wind distributions with different models). Note that these projections are still very uncertain and model-dependent.

During historic times, storminess and large-scale temperature variations were mostly decoupled, but the projections show a closer relationship. Some projections, based on the high emissions IPCC SRES A2 scenario, show a related increase in temperature and the frequency of heavy storms in the North Atlantic Ocean. The future storminess in this region depends on projections of sea surface temperature, retreat of Arctic ice and changes in the air pressure field (Fischer-Bruns *et al.*, 2005).

Projections of storm surges are closely connected with future storminess. The projections for the end of 21st century show a significant increase of storm surge elevations for the continental North Sea coast, by between 15 and almost 25 cm (Woth, 2005). For the UK coastline, a large increase in

Map 5.10 Projected relative change of annual maximum daily mean wind speed between 1961–2000 and 2050 using different models

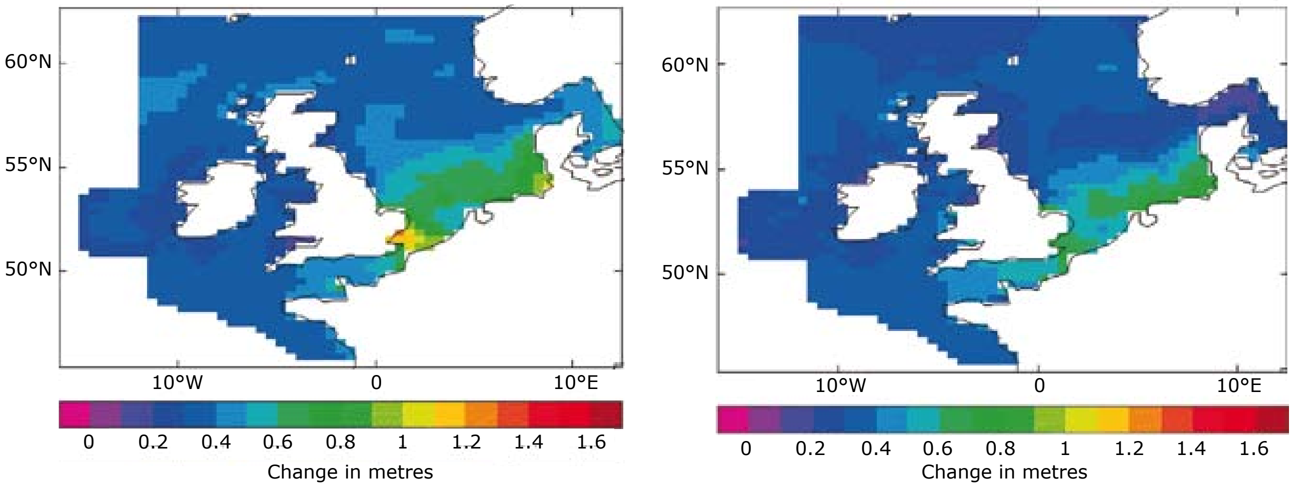


Note: Data are calculated for 10 m height using the + 2 °C scenario for 2050 (IPCC-SRES A1B emission scenarios) and the reference climate (1961–2000) from three similar models (left) and one different model, MIROCHi (right).

Source: van der Hurk *et al.*, 2006.

relative surge height is projected for the high IPCC SRES scenario A2 and the intermediate scenario B2, especially along the south-east coast of England, where the changes in storminess will have their largest effect and where the land is sinking most rapidly (Lowe and Gregory, 2005; Map 5.11).

Map 5.11 Change in the height of a 50-year return period extreme water level event for the end of 21st century for different scenarios



Note: The water level is measured relative to the present day tide, due to changes in atmospheric storminess, an increase in mean sea level and vertical land movements. Results are shown for the (left) A2 scenario and (right) B2 scenario, used model HadRM3H. A 50-year return period means the average probability of occurrence of two events in 100 years.

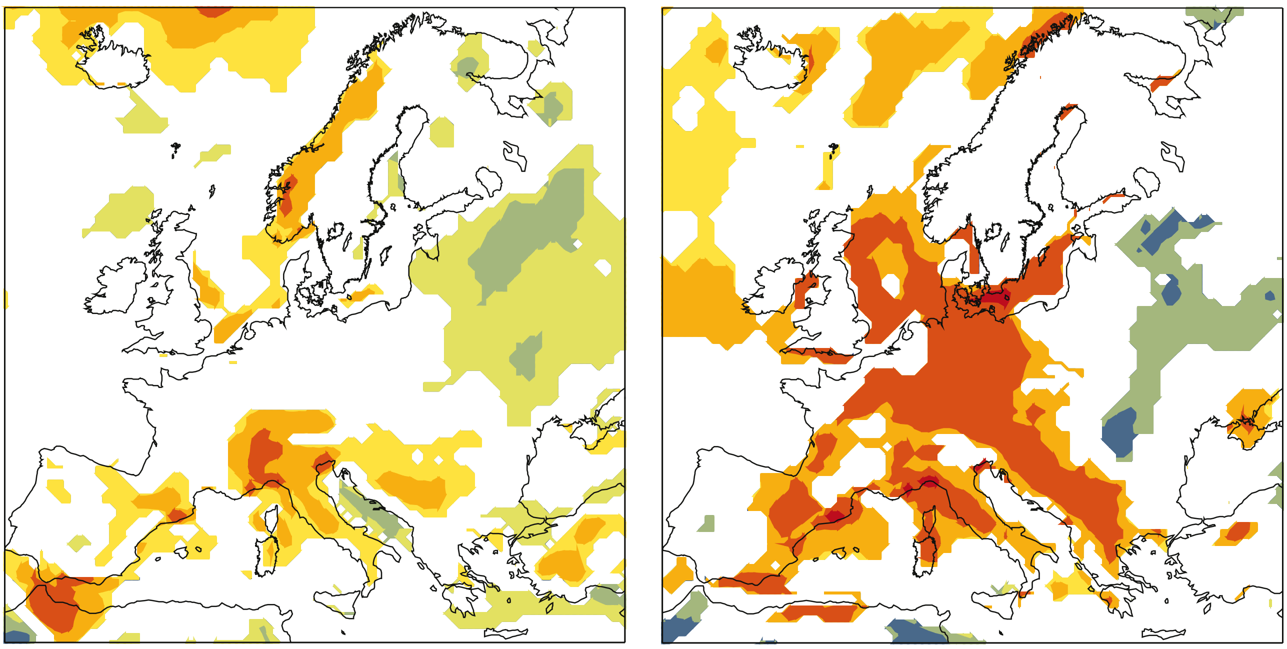
Source: Lowe and Gregory, 2005.

5.2.7 *Air pollution by ozone*

Key messages

- Climate variability and change has contributed to an increase in average ozone concentrations in central and South-Western Europe (1–2 % per decade).
- During the summer of 2003, exceptionally long-lasting and spatially extensive episodes of high ozone concentrations occurred, mainly in the first half of August. These episodes appear to have been associated with the extraordinarily high temperatures over wide areas of Europe and illustrate the expected more frequent exceedances of the ozone information threshold under projected climate change.
- The projected climate-induced increase in ozone levels may result in current ozone abatement policies becoming inadequate.

Map 5.12 Modelled change in tropospheric ozone concentrations over Europe 1958–2001 and 1978–2001



Trend in tropospheric ozone concentrations due to climate variability and climate change

Left: 1958–2001; right: 1979–2001

% per decade

– 10 – 5 – 2 –1 0 1 2 5 10

Note: The modelled changes shown are only due to climate variability and climate change. White areas have no significant trend.

Source: Andersson *et al.*, 2007.

Relevance

Tropospheric ozone is one of the air pollutants of most concern in Europe. Ozone is estimated to cause about 20 000 acute mortalities each year (European Commission, Clean Air for Europe impact assessment, 2006) and economic damage due to crop loss of EUR 4 625 million per year (Holland *et al.*, 2006). Ozone is formed in the lower troposphere as a result of complex chemical reactions between volatile organic compounds and nitrogen oxides, in the presence of sunlight. EU legislation has established ozone exceedance thresholds and national emission ceilings for ozone precursor emissions to protect human health and prevent damage to ecosystems, agricultural crops and materials.

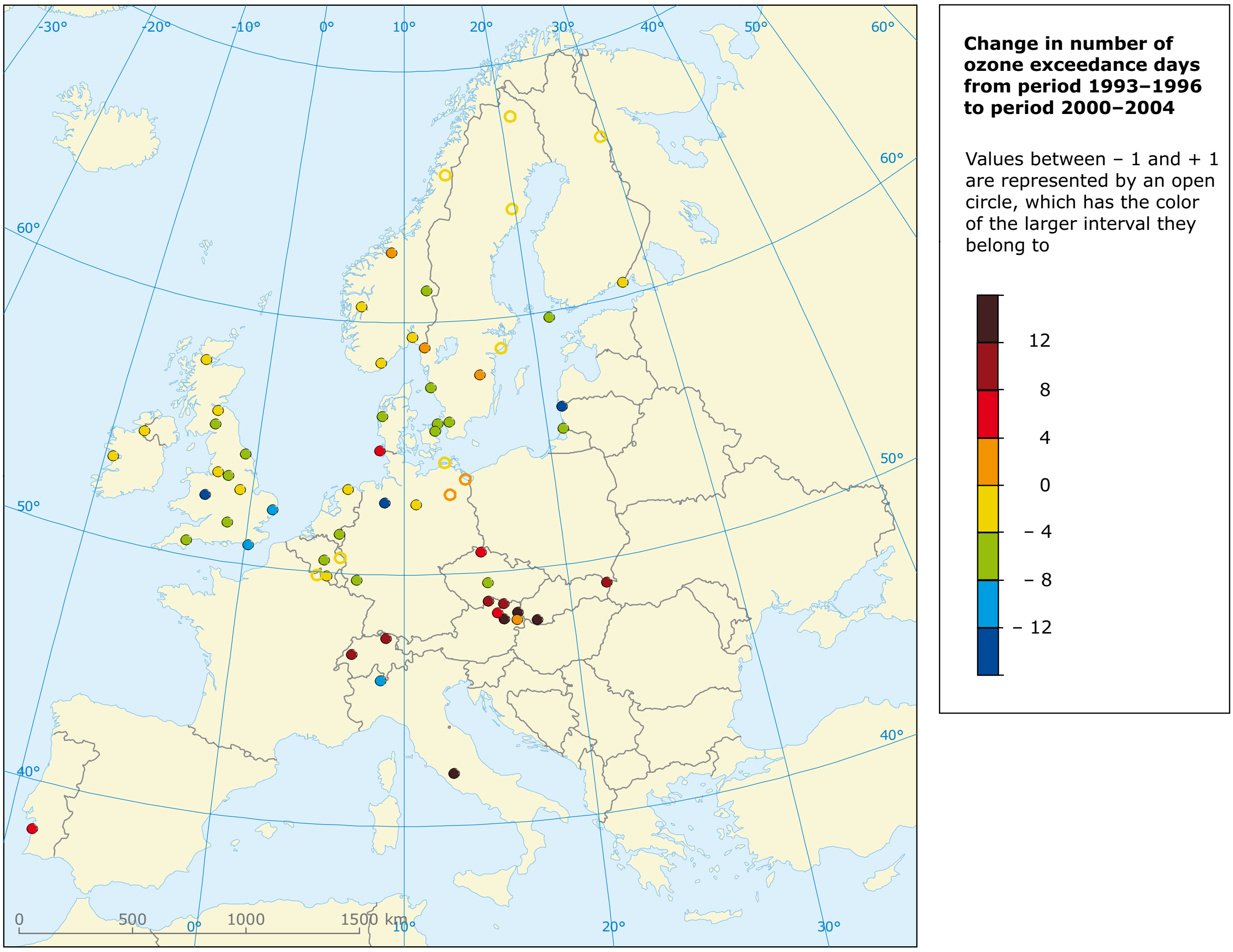
Episodes of elevated ozone levels occur mainly during periods of warm sunny weather (Schichtel and Husar, 2001; Rao *et al.*, 2003). The projected increase in hot extremes in Europe (see Section 5.2.4) is therefore expected to result in ozone episodes that require more vigorous emission reduction measures and the use of the available adaptation measures such as improved public information and health care services.

Past trends

A modelling study from 1958 to 2001 (Andersson *et al.*, 2007) shows that climate variability and change contributed to increased ozone concentrations during the period 1979–2001 over south-central and south-western Europe, and a decrease in north-eastern Europe (Map 5.12). The reason for this is a combination of changes in temperature, wind patterns, cloud cover and stability. Further, temperature plays a role in various processes which directly affect the formation of ozone, like the emission of biogenic organic compounds (e.g. isoprene), and the photo-dissociation of NO_2.

A link between temperature and ozone concentration is also evident from observations. A statistical analysis of ozone and temperature measurements in Europe for 1993–2004 shows that in central-western Europe and the Mediterranean area, a change the increase in the daily maximum temperature in 2000–2004 compared with 1993–1996 contributed to extra ozone exceedences (Maps 5.13 and 5.14). In south and central Europe, the temperature trend was responsible for an average of 8 extra annual exceedence days of 120 µg/m³, i.e. 17 % of the total number of exceedences observed in that region.

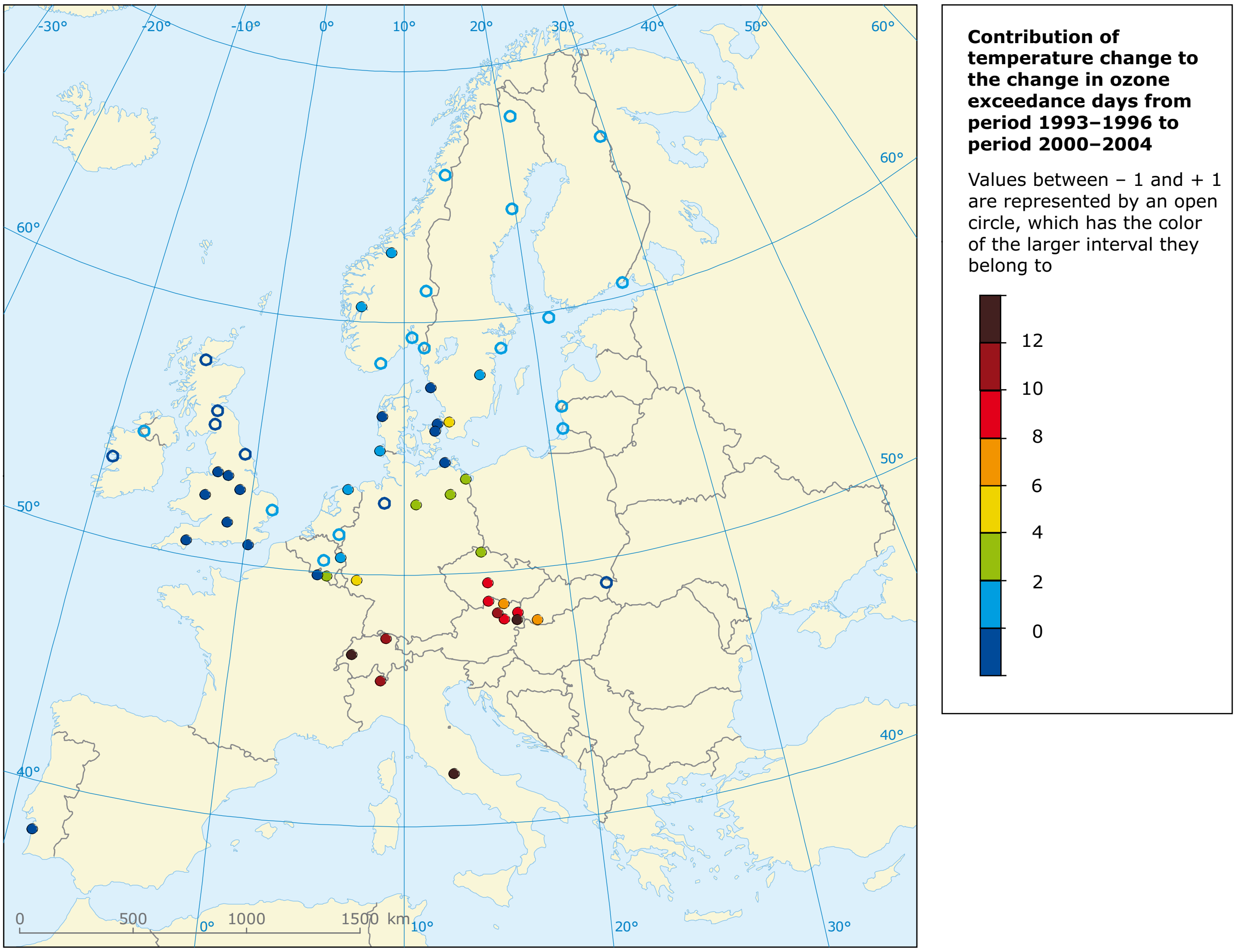
Map 5.13 Change in number of ozone exceedance days between 1993–1996 and 2000–2004



Note: Ozone exceedance days meaning days where the maximal 8 hr average ozone concentration exceeds 120 µg per m³ (year 2003 excluded).

Source: Van Dingenen *et al.*, 2008.

Map 5.14 Contribution of temperature increase to the change in ozone exceedance days between 1993–1996 and 2000–2004



Source: Van Dingenen *et al.*, 2008.

An analysis of trends over the past twelve years indicates that in the EU the average number of hours when ozone concentration exceeded the information threshold of 180 μg/m³ was higher in summer 2003 than in all previous years (Fiala *et al.*, 2003).

Projections

The projected trends for ozone and other air pollutants are closely linked to projections for radiation, temperature, cloudiness, and precipitation. On a global scale, the effect of climate change alone on tropospheric ozone concentrations is expected to be small, because of a reduction in ozone lifetime as a consequence of higher humidity (Stevenson *et al.*, 2006). However, regional differences can be large. Regions where climate change is expected to result in an increased frequency of stable anticyclonic conditions with associated high temperatures, large solar inputs and little boundary layer ventilation may experience a deterioration of air quality (Hogrefe *et al.*, 2004; Sousounis *et al.*, 2002). A 30-year model study for the period 2071–2100, based on the IPCC A2 and B2 scenarios for CO_2 emissions (but with otherwise constant emissions of pollutants) shows that daily peak ozone amounts as well as average ozone concentrations will increase substantially during the summer in future climate conditions (Meleux *et al.*, 2007), in particular in central and western Europe, in line with observed trends from the past. The study also finds that summer ozone levels in future climate conditions are similar to those found during the exceptionally hot summer of 2003. The expected impact on human health may be exacerbated by the aging of the population, the elderly being more susceptible to air pollution than the average population (OECD, 2008).

5.3 Cryosphere

5.3.1 Introduction

The cryosphere is the frozen part of the world. It includes all permanent or seasonal snow and ice deposits on land, in the seas, rivers and lakes, and in the ground (permafrost). It is the second largest component of the climate system after the oceans with regard to mass and heat capacity. Snow and ice play a key role in the earth's energy budget by reflecting heat from the sun because of its light surfaces. As melting replaces white surfaces with darker ones, more heat is absorbed (the albedo effect). Two thirds of the world's freshwater resources are frozen. Snow and ice play a key role in the water cycle and are essential for storing fresh water for hotter and often dryer seasons.

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.

Selection of indicators

The various components of the cryosphere play strong but different roles within the climate system.

- Because of their large volumes and areas, the continental *ice sheets* of Greenland and Antarctica actively influence the global climate over very long time-scales. Sea levels can however be affected more rapidly.
- *Snow* also covers a large area, but has relatively small volume. It is important for key global interactions and feedbacks like increased absorption of heat (albedo effect).
- *Sea ice* also covers a large area. It is important due to its albedo and its impacts on ocean circulation, which transports heat from the equator to the poles.
- Melting *permafrost* releases the strong greenhouse gas methane from frozen organic material. Together with seasonal snow, it influences the water content of the soil and the vegetation.
- *Glaciers, ice caps* and *seasonal lake ice*, with their smaller areas and volumes, react relatively quickly to changes in climate, influencing ecosystems and human activities on a local scale. They are good indicators of climate change.

The selected indicators cover strategic information from all these compartments of the cryosphere: glaciers and snow cover in Europe, the Greenland ice sheet, the Arctic sea ice and mountain permafrost in central Europe. Lake and river ice conditions are presented in the water chapter.

Indicators and vulnerability

The cryosphere is vulnerable to global warming. It is a very visible expression of climate change. It integrates climate variations over a wide range of time scales, from millennia to seasonal variations throughout the year. This can complicate interpretation of why changes happen.

In Europe, the most vulnerable areas are the high mountain areas and the Arctic. In both the Arctic and the European Alps, temperatures have increased at more than the global rate in the past few decades, in the Arctic as a whole, twice as much. The amount of ice and snow, especially in the northern hemisphere, has decreased substantially over the last few decades due to increased temperatures. European glaciers are shrinking, snow-covered areas are creeping higher up and further north, sea ice in the Arctic is melting and getting thinner, permafrost is starting to thaw, and the Greenland ice sheet shows increasing signs of disintegration and thawing at its borders. These trends will accelerate as climate change is projected to continue. However, there are large uncertainties in the fate of key components of the cryosphere: it is not yet possible to make reliable predictions of when the Arctic sea ice may melt completely in summer, neither is it yet possible to predict the future of the Greenland ice sheet with any confidence.

Data and info sources

Data on the components of the cryosphere vary significantly with regard to availability and quality. Long-term data on glaciers in all European glaciated areas are provided in good quality and quantity by the World Glacier Monitoring Service (WGMS) in Zürich. Data on snow cover and Arctic sea ice have been measured globally since satellite measurements started in the 1970s and are available for example at the Global Snow and Ice Data Centre (NSIDC) in Boulder/USA. However sensors in the satellites have gradually improved, allowing for new observations. Hence area-wide data on the Greenland ice-sheet are often available for not longer than 15 years. This is also the situation for data on mountain permafrost measured in boreholes which are drilled into frozen rock walls.

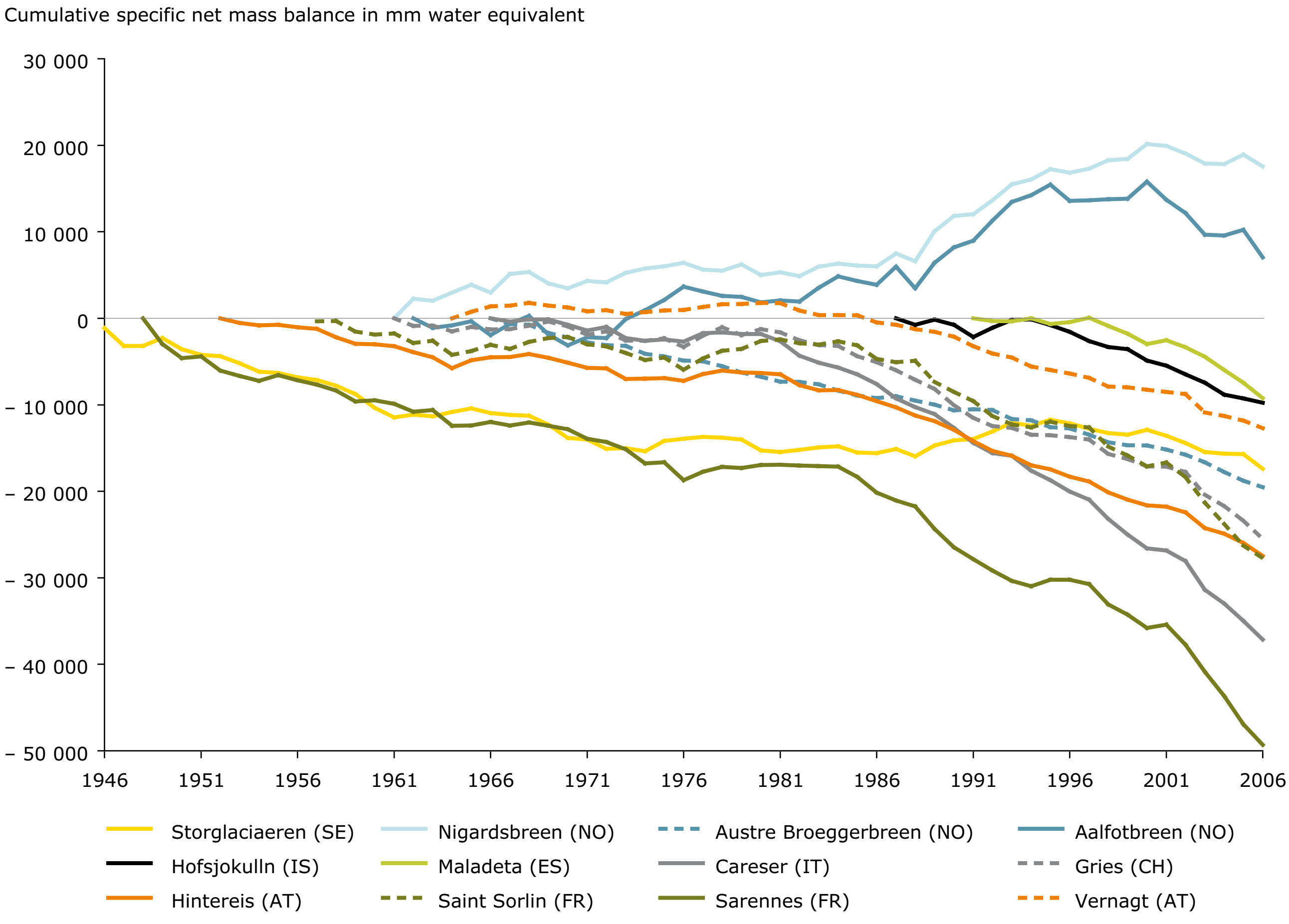
The gaps in the cryospheric data base are well recognised by the scientific community and many efforts are being made to improve the knowledge, e.g. during the International Polar Year (IPY) 2007–2008.

5.3.2 *Glaciers*

Key messages

- The vast majority of glaciers in the European glacial regions are in retreat.
- Since 1850, glaciers in the European Alps have lost approximately two thirds of their volume, with clear acceleration since the 1980s.
- Glacier retreat is projected to continue. A 3 °C increase in average summer air temperature could reduce the existing glacier cover of the European Alps by some 80 %. With continuing climate change nearly all the smaller glaciers and one third of the overall glacier area in Norway are projected to disappear by 2100.
- Glacier retreat has serious consequences for river flow. It affects freshwater supply, river navigation, irrigation and power generation. It could cause natural hazards and damage to infrastructure.

Figure 5.8 Cumulative specific net mass balance of glaciers from all European glaciated regions 1946–2006



Note: Positive values mean ice growth, negative values mean ice loss.

Source: Fluctuation of Glaciers Database (FoG), World Glacier Monitoring Service (www.wgms.ch), 2007.

Relevance

Glacier changes are among the most visible indications of the effects of climate change. Glaciers are particularly sensitive to changes in the global climate because their surface temperature is close to the freezing/melting point (Zemp *et al.*, 2006). Glacier fluctuations showed a strong relation to air temperature throughout the 20th century (Greene, 2005). Therefore the change in the mass balance of glaciers is considered to be an immediate signal of global warming trends. A negative mass balance indicates that the loss of ice, mainly from melting and calving in summer, is larger than the accumulation from snowfall in winter.

Glaciers are an important freshwater resource and act as 'water towers' for lower-lying regions. In the coming decades, we can first expect more melt water from the glaciers running into rivers. As the glaciers diminish, however, the annual melt water, and therefore their contribution to river flow and sea-level rise, will decrease. This will have serious consequences for freshwater supply, river navigation, ecosystems fed by water from rivers, irrigation facilities, and power generation. Furthermore, the solute release from melting rock-glaciers may affect the water quality of high mountain lakes adversely by the intrusion of heavy metals (Thies *et al.*, 2007).

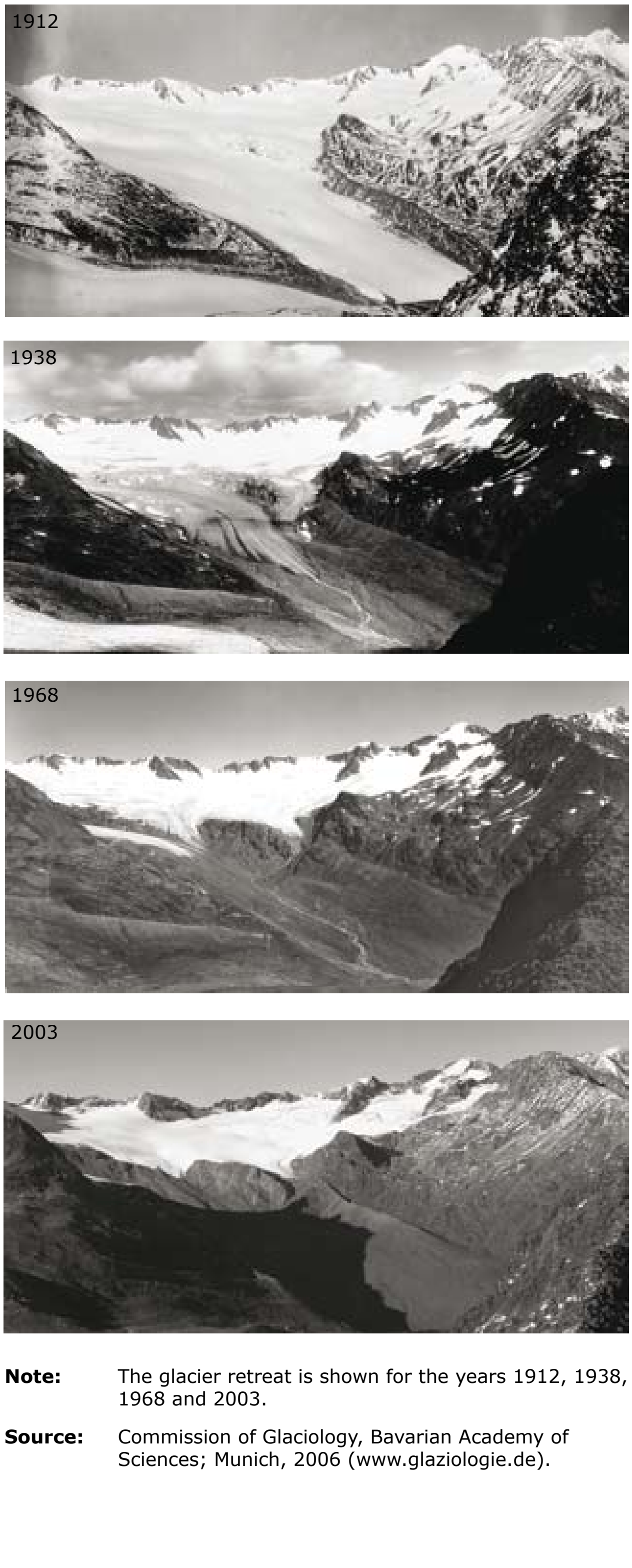
Strong retreat of glaciers can cause instabilities resulting in hazardous incidents such as glacier lake outbursts, rock-ice avalanches and landslides (Pralong and Funk, 2005; Huggel *et al.*, 2007). This may cause damage to infrastructure. Glacier retreat affects tourism and winter sports in the mountains (OECD, 2007) and changes the appearance of mountain landscapes.

Improved glacier monitoring, and adaptation options such as water management measures, draining of glacier-lakes and construction of protective walls can reduce some of the risks and negative consequences, but not all.

Past trends

According to the high-quality data records of the WGMS, a general loss of glacier mass has occurred in nearly all the European glacier regions (Figure 5.8). Glacier retreat in Europe started after the maximum glacier extent of the so-called 'Little Ice Age' in the middle of the nineteenth century. In the Alps, glaciers lost one third of their surface area and one half of their volume between 1850 and the end of the 1970s. Since 1985 an acceleration in glacial retreat has been observed, which led to a loss of 25 % of the remaining ice by 2000 (Zemp *et al.*, 2006). This was followed by a further loss of 5–10 % in the extraordinary hot and dry summer of 2003 (Zemp *et al.*, 2005), resulting in a total loss of about two thirds of the 1850 ice mass. This is illustrated by

Figure 5.9 Shrinking of the Vernagtferner glacier, Austria



Note: The glacier retreat is shown for the years 1912, 1938, 1968 and 2003.

Source: Commission of Glaciology, Bavarian Academy of Sciences; Munich, 2006 (www.glaziologie.de).

the shrinking of the Vernagtferner-glacier in Austria (Figure 5.9).

The Norwegian coastal glaciers, which were expanding and gaining mass due to increased snowfall in winter up to the end of the 1990s, are also now retreating, as a result of less winter precipitation and more summer melting (Nesje *et al.*, 2008; Andreassen *et al.*, 2005).

Glaciers in Svalbard are experiencing mass loss at lower elevations, and the fronts of nearly all glaciers there are retreating (Haeberli *et al.*, 2005, 2007; Nuth *et al.*, 2007). Some ice caps in north-eastern Svalbard seem to be increasing in thickness at higher elevations (Bamber *et al.*, 2004; Bevan *et al.*, 2007). However, estimates for Svalbard as a whole show that the total balance is negative (Hagen *et al.*, 2003), and there is a clear sign of accelerated melting, at least in western Svalbard (Kohler *et al.*, 2007).

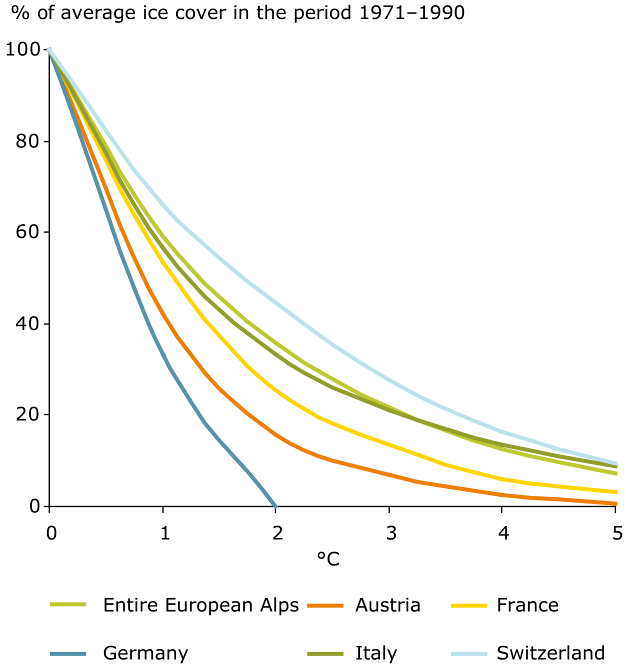
Very recent findings by the WGMS (UNEP, 2008) indicate a clearly increasing annual reduction of the global mean ice-thickness of glaciers since the turn of millennium (0.5 m) compared with the 1980–1999 period (0.3 m). Some of the most dramatic shrinking has been in Europe (Scandinavia, Alps, and Pyrenees).

The centennial retreat of European glaciers is attributed mainly to increased summer temperatures. However, changes in winter precipitation, the decreased glacier albedo due to the lack of summer snow-fall and various other feedback processes are altering the pattern on a regional and decadal scale. The recent strong warming has made disintegration and down-wasting increasingly dominant causes of glacier decline in the European Alps during the most recent past (Paul *et al.*, 2004).

Projections

According to a recently published sensitivity study (Zemp *et al.*, 2006), the European Alps could lose about 80 % of their average ice cover for the period 1971–1990 if summer air temperatures rose by 3 °C; a precipitation increase of 25 % for each 1 °C would be needed to offset the loss of cover. The modelled remains of Alpine glaciers as a consequence of warming are presented in Figure 5.10. Sugiyama *et al.* (2007) investigated the potential evolution of

Figure 5.10 Modelled remains of the glacier cover in the European Alps for an increase in average summer air temperature of 1 to 5 °C



Note: The total of 100 % refers to the average ice cover 1971–1990.

Source: Zemp *et al.*, 2006.

the Rhone Glacier, Switzerland, in the 21st century using a model which included more consideration of glacier flow dynamics. They found increasing mass loss as well as decreasing glacier cover, but at a gradually slower rate. However, neither modelling studies considered feedback processes such as the development of glacier lakes, which could accelerate glacier retreat dramatically.

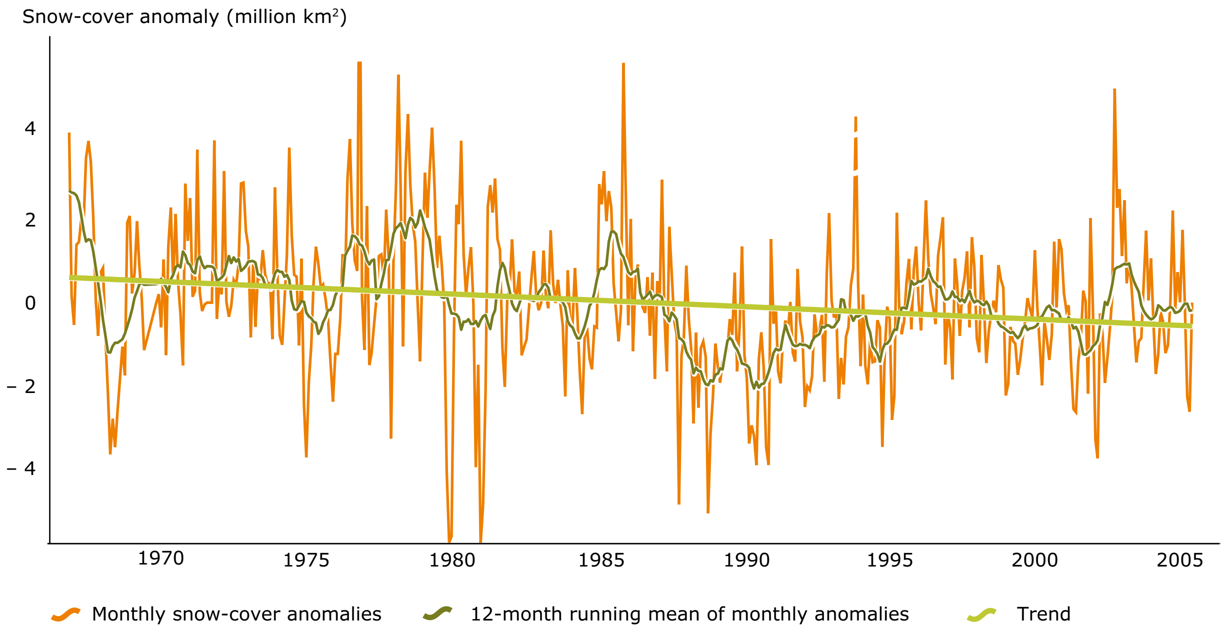
Recent climate scenarios for Norway, based on model calculations by the British Headley Centre and the German Max Planck Institute which follow the SRES B2 emission-scenario, indicate a rise in summer temperature of 2.3 °C and an increase in winter precipitation of 16 % in the period 2070–2100 compared with 1961–1990. As a result, nearly all the smaller Norwegian glaciers are likely to disappear and overall glacier area as well as volume may be reduced by about one third by 2100 (Nesje *et al.*, 2008).

5.3.3 *Snow cover*

Key messages

- Snow cover in the northern hemisphere has fallen by 1.3 % per decade during the past 40 years. The largest losses are during spring and summer.
- Model simulations project widespread reductions in the extent and duration of snow cover in Europe over the 21st century.
- Changes in snow cover affect the Earth's surface reflectivity, river discharge, vegetation, agriculture and animal husbandry, tourism, snow sports, transport and power generation.

Figure 5.11 Northern hemisphere snow-cover extent variation 1966–2005



Note: Snow-cover anomalies are expressed in deviations from monthly means.

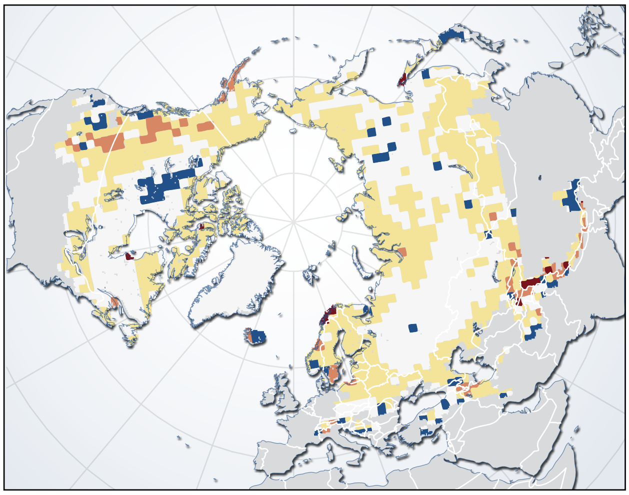
Source: Brodzik, 2006 (NOAA-data); UNEP, 2007. (http://maps.grida.no/go/graphic/northern-hemisphere-snow-cover-extent-anomalies-1966-2005).

Relevance

Snow covers more than 33 % of the land surface north of the equator from November to April. It reaches a maximum of about 45.2 million km² in January, and a minimum of about 1.9 million km² in August (Clark *et al.*, 1999).

Snow cover is an important feedback mechanism of the climate system. The extent of snow cover depends on the climate, for example on temperature and precipitation, and on solar radiation. But it also influences the climate and climate-related systems because of its high reflectivity, insulating properties, effects on water resources and ecosystems, and cooling of the atmosphere. Thus a decrease in snow cover reduces the reflection of solar radiation, contributing to accelerated climate change. Changes in the extent, duration, thickness and properties of snow cover can affect water availability for domestic use, navigation and power generation. Changes in snow cover affect human well-being through influences on agriculture, infrastructure, the livelihoods of indigenous Arctic people, environmental hazards and winter recreation. Snow-cover retreat can reduce problems of winter road and rail maintenance, affecting the exploitation and

Map 5.15 Observed change in spring snow-cover duration 1970–2004



Observed change in spring snow-cover duration 1970–2004 (days/year)

– 4 to – 2 | – 2 to – 1 | – 1 to – 0.25 | – 0.25 to 0.25 | 0.25 to 3

Note: Data are in days/year. Negative values (brown/yellow) indicate reduced duration of snow cover/positive values (blue) indicate extended duration of snow cover.

Source: R. Brown, Environment Canada; data from D. Robinson, Rutgers University. (http://maps.grida.no/go/graphic/trends-in-spring-snow-cover-duration-for-the-northern-hemisphere-1970-2004).

transport of oil and gas in cold regions (UNEP, 2007; ACIA, 2004).

Shallow snow cover at low elevations in temperate regions is the most sensitive to temperature fluctuations and hence most likely to decline with increasing temperature (IPCC, 2007a, b).

For several of these impacts, adaptation can reduce the negative effects of snow-cover change. Some adaptation options, such as artificial snowmaking in the Alps to maintain tourism as a main source of income, have to be balanced against their negative implications for mitigation, due to increased energy use and greenhouse gas emissions.

Past trends

Data from satellite monitoring (NESDIS-database at NOAA) from 1966 to 2005 show that monthly snow-cover extent in the northern hemisphere is decreasing by 1.3 % per decade (Figure 5.11), with the strongest retreat in spring and summer (UNEP, 2007). Snow cover fell in all months except November and December, with the most significant decrease during May to August (Brodzik *et al.*, 2006). This was accompanied by lower springtime water content, earlier disappearance of continuous snow cover in spring (Map 5.15) by almost two weeks in the 1972–2000 period (Dye, 2002), less frequent frost days (days with minimum temperature below 0 °C) and shorter frost seasons (period of consecutive frost days).

The trends in duration and depth of northern hemispheric snow cover at higher latitudes differ between regions. In contrast to a reduced duration of snow cover over North America, a long-term increase in depth and duration has been observed over most of northern Eurasia (Kitaev *et al.*, 2005).

Snow-cover trends in the mountain regions of Europe vary considerably with region and altitude. Recent declines in snow cover have been documented in the mountains of Switzerland (e.g. Scherrer *et al.*, 2004), Slovakia (Vojtek *et al.*, 2003), and in the Spanish ski-resorts in the Sierra Nevada and the Pyrenees (Rodriguez *et al.*, 2005), but no change was observed in Bulgaria over the period 1931–2000 (Petkova *et al.*, 2004). Declines, when observed, were largest at lower elevations, and Scherrer *et al.* (2004) statistically attributed the declines in the Swiss Alps to warming. Lowland areas of central Europe are characterised by recent reductions in annual snow-cover duration of about 1 day/year (Falarz, 2002). At Abisko in sub-Arctic Sweden, increases in snow depth have been recorded since 1913 (Kohler *et al.*, 2006), and trends towards greater maximum snow depth but shorter snow season have been noted in Finland (Hyvärinen, 2003).

Photo: © M. Zebisch, 2004

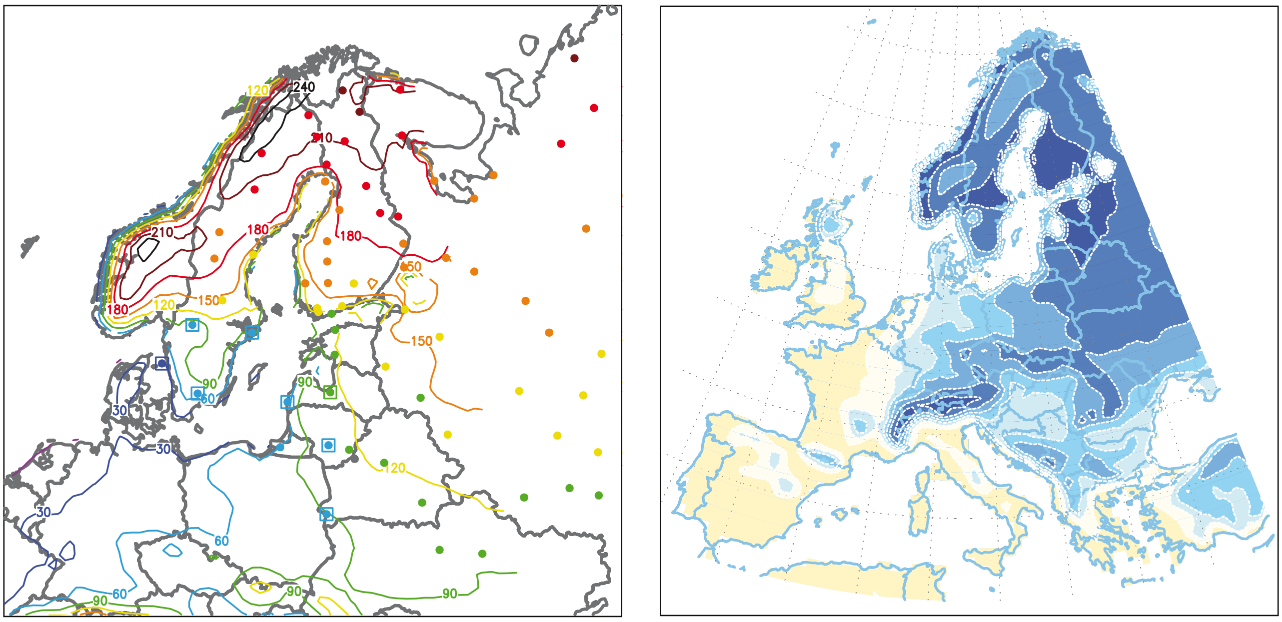
Projections

Model simulations project widespread reductions in snow cover over the 21st century (IPCC, 2007a). Decreases of between 9 and 17 % in the annual mean northern hemisphere snow cover by the end of 21st century are projected by individual models (ACIA, 2004). Although winter precipitation is projected to increase in northern and central Europe (Christensen and Christensen, 2007), less frequent frost occurrences associated with higher temperatures are projected to reduce the number of days with snow cover (Map 5.16). Decreases of more than 60 snow-cover days are projected to occur (for the period 2071–2100 compared with 1961–1990) around the northern Baltic Sea, on the west slopes of the Scandinavian mountains and in the Alps (Jylhä *et al.*, 2007). The beginning of the snow accumulation season is projected to be later and the end earlier, and snow coverage during the snow season is projected to decrease (Hosaka *et al.*, 2005).

For every 1 °C increase in average winter temperature, the snowline in the European Alps rises by about 150 metres (Beniston, 2003). Regional climate model runs, following the SRES emission scenarios A1B, B1 and A2, project milder winters with more precipitation in this region, increasingly falling as rain (Jacob *et al.*, 2007). A recently-published study on the sensitivity of the Alpine snow cover to temperature by Hantel and Hirtl-Wielke (2007) reported a distinctive and strong variation of snow-cover sensitivity to temperature change with altitude. The study estimated that a 1 °C increase in temperature over central Europe (5–25°E and 42.5–52.5°N) would result in a reduction of about 30 days in snow duration (snow cover of at least 5 cm) in winter at the height of maximum sensitivity (about 700 m).

Snowfall in lower mountain areas is likely to become increasingly unpredictable and unreliable over the coming decades (Elsasser and Bürki, 2002), with consequences for natural snow reliability and therefore difficulties in attracting tourists and winter sports enthusiasts (OECD, 2007).

Map 5.16 Annual number of days with snow cover over European land areas 1961–1990 and projected change for 2071–2100



Annual number of days with snow cover

Observed means in nothern Europe, 1961–1990 (coloured dots) and modelled means for Europe, 1961–1990

5 30 60 90 120 150 180 210 240

Projected change of annual number of days with snow cover

Projected multi-model mean changes for the period 2071–2100, relative to 1961–1990

– 60 – 50 – 40 – 30 – 20 –10

Note: Results are based on seven regional climate-model simulations.

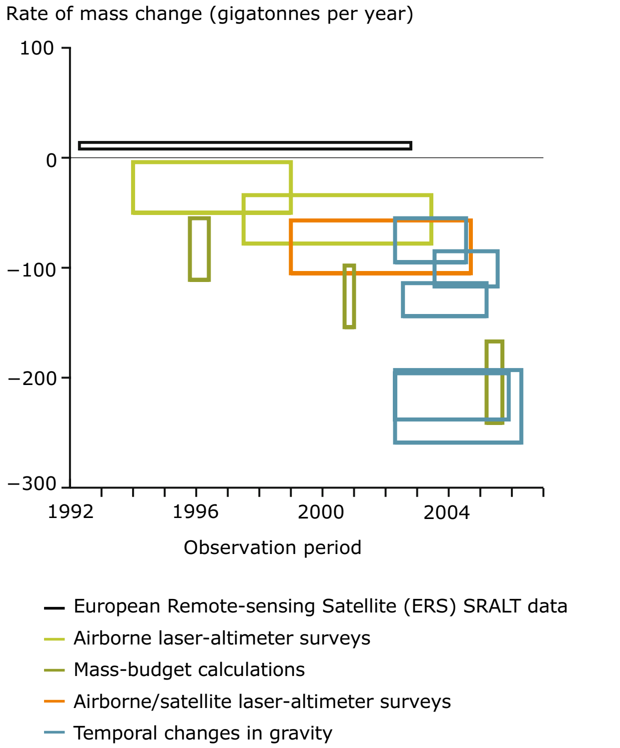
Source: Jylhä *et al.*, 2007.

5.3.4 *Greenland ice sheet*

Key messages

- The Greenland ice sheet changed in the 1990s from being in near mass balance to losing about 100 billion tonnes of ice per year. Ice losses may have doubled again by 2005. Accelerated flow of outlet glaciers to the sea accounts for more of the ice loss than melting.
- The contribution of ice loss from the Greenland ice sheet to global sea-level rise is estimated at 0.14–0.28 mm/year for the period 1993–2003 and has since increased. In the long term, melting ice sheets have the largest potential to increase sea level.
- No reliable predictions of the future of the ice sheets can yet be made; the processes causing the faster movement of the glaciers are poorly understood and there is a lack of long-term observations.

Figure 5.12 Estimated changes of the ice mass in Greenland 1992–2006



Note: The rectangles depict the time period of the observations (horizontal) and the upper and lower estimates of mass balance for that period (vertical), calculated by different techniques as marked with colour codes. The uncertainty in assessing the trend is largest in periods when the vertical parts of rectangles from different estimates do not overlap. The main factors determining whether the Greenland ice sheet gains or loses ice (mass balance) are: (1) Surface mass balance = the difference between net snow accumulation and loss from melting (meltwater runoff and evaporation) and (2) Dynamic ice loss from the movements of glaciers, leading to ice berg calving.

Source: Thomas *et al.*, 2008.

Relevance

The ice sheets of Greenland and Antarctica contain 98–99 % of the freshwater ice on earth's surface. To illustrate their sizes, the volumes of the Antarctic and Greenlandic ice sheets are equivalent to a 57 and 7 m layer, respectively, of water on top of the world's oceans. When setting their upper estimate of a projected 59 cm sea-level rise by the end of this century, the IPCC did not take into account increased discharges into the ocean from the moving outlet glaciers of the ice sheets. The uncertainty about their future is therefore a main reason for uncertainties in projections of sea-level rise. The Greenland ice sheet is the most susceptible to warming because of its closeness to the Atlantic Ocean and other continents. But the more isolated Antarctica now also seems to be experiencing a net loss of ice, which may be accelerating (UNEP, 2007) (See indicator on sea-level rise (Section 5.4.2)).

The speed of ice loss is important as well as its magnitude because a faster rise in sea level reduces the time available to take appropriate adaptation measures.

The melt water from Greenland will contribute to reducing the salinity of the surrounding ocean. An upper layer of fresher water may reduce the formation of dense deep water, one of the mechanisms driving global ocean circulation.

Past trends

The Greenland ice sheet is a huge inland glacier with several glacier tongues calving into the sea. It covers roughly 80 % of Greenland. The average ice thickness is 1 600 m, with the highest summit reaching

3 200 m above sea level. It has a volume of about 3 million km^3.

Until recent improvements in remote sensing, it was hard to measure whether the polar ice sheets were growing or shrinking. Most time-series remain short. There is however a general consensus from different approaches that ice loss from the Greenland ice sheet has accelerated. From a near balance in the early 1990s, about 100 billion tonnes were being lost annually at the end of the century. This may have doubled again by 2005 (UNEP, 2007). However, there is still considerable discrepancy between different estimates of ice loss rates (Figure 5.12).

The IPCC estimated the Greenland ice sheet contribution to sea-level rise during 1993–2003 to be 0.14–0.28 mm/year, based on an annual ice loss in that period of 50 to 100 billion tonnes (IPCC, 2007a). A study estimating an ice loss of 224 billion tonnes/year in 2005, found a corresponding contribution to sea-level rise of 0.57 mm/year (Rignot and Kanagaratnam, 2006). This is one of the upper estimates in Figure 5.12 and illustrates the effects that different rates of ice loss can have on sea level.

The ice in the interior of the Greenland ice sheet at high elevations has thickened since it has received more snowfall — on average about 4 cm/year since 2000 (UNEP, 2007). This gain has been more than offset by the loss in lower-lying regions from melting and increased calving of ice bergs. Air temperatures in summers have increased significantly along the coast since the early 1990s, whereas little change or slight cooling has been observed in the high interior (Steffen, 2007 unpubl.).

Ice loss can partly be caused by surface melting. On the low-lying edge of the ice sheet, the surface melts each summer causing evaporation and meltwater

Melt water forms rivers at the glacier surfaces

Photo: © John McConnico

runoff from the glacier. If this exceeds the net snow accumulation in winter, the glacier has a negative surface mass balance. Areas with melting can be measured from satellites, and from 1979 to 2007, the cumulative melt area increased by approximately 50 % (Figure 5.13). Melting has reached higher elevations, and the melting season is lasting longer. However, both snowfall and surface melting has increased. The resulting trend in surface mass balance for the whole Greenland ice sheet between 1958 and 2006 has been modelled to be insignificantly negative (Hanna *et al.*, 2007). Data from recent years extending to 2007 suggest a strong increase in the net loss of surface mass balance (Steffen, 2007 unpubl.).

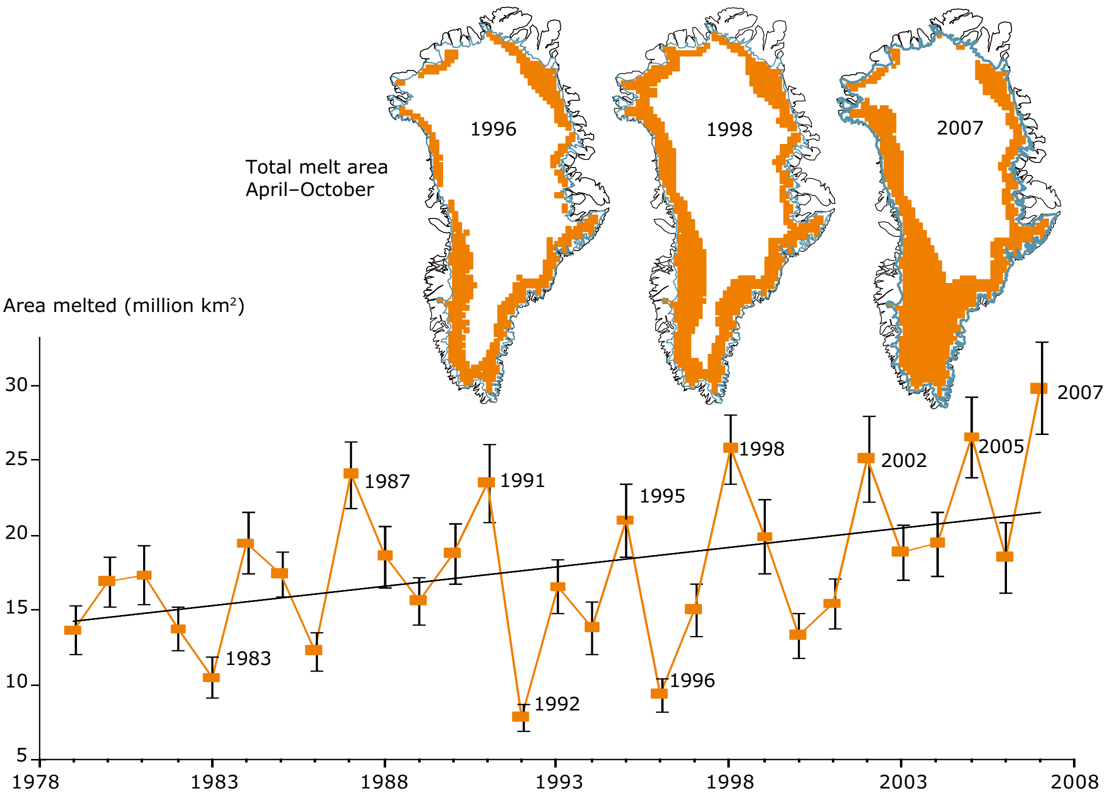
The other mechanism behind the ice loss since the 1990s is accelerated flow of outlet glaciers towards the sea. Large amounts of meltwater form rivers and melting ponds at the glacier surface and penetrate through crevasses to the bottom. This water probably lubricates the bedrock/ice interface, making the glaciers move faster. Another explanation is that the glacier fronts may be affected by increasing ocean temperatures, reducing their buttress effect. The outlet glaciers are also influenced by the topography of the fjords. Outlet glaciers act as 'bathtub drains' for the inland ice: ice is being transported into the melting zone, and calving into the ocean increases. The speed of the fastest-flowing glacier, Jakobshavn isbræ on the west coast, has nearly doubled, to about 14 km a year (UNEP, 2007). Some glaciers are however reported to be slowing down from the maximum speeds measured, possibly around a new equilibrium position. Acceleration is widespread mostly on the southeast coast and has moved northwards to about 70°N. It is associated with large retreats and thinning of the ice sheet. Near the outlets, glacier surface elevation can subside by tens of metres.

The ice losses caused by accelerated flow of the outlet glaciers (ice dynamics) have exceeded the losses from melting processes (negative surface mass balance) several times during the recent few warmest years. For 2005, it has been estimated that two thirds of the ice loss was caused by ice dynamics (Rignot and Kanagaratnam, 2006).

Projections

It is currently not possible to predict the future development of the Greenland ice sheet with confidence. Glacier models account mostly for accumulation of snow during winter and melting in summer (surface mass balance). The accelerated ice flow has been observed for a rather short period of time. Scientists are now trying to understand the

Figure 5.13 Area of Greenland ice sheet melting 1979–2007



Note: The area of the Greenland ice sheet where there is at least one day of surface melting in summer increased to a new record extent in 2007. Melting passed the 2 500 m elevation and probably led to a record ice loss that year.

Source: Updated from Steffen *et al.*, 2004; Witze, 2008.

processes driving this phenomenon. That should in turn allow better ice models to be developed. But for models to predict the future well, they must be validated by data from long-term measurements describing key processes. Until science comes closer to this, our ability to predict the sensitivity of the Greenland ice sheet to global warming will remain limited.

Further temperature increases can accelerate the ice loss because of positive feed-back mechanisms, like thinning of the ice sheet that exposes larger areas to melting. It is hard to say how strong these mechanisms are, how rapidly the ice sheet will react to them and whether ice loss will be irreversible.

Throughout the earth's history, the ice sheets have shrunk in response to warming and grown in response to cooling. Deep ice core drillings reveal past climates and can give some indications of how they have changed. The Eemian era was an interglacial period 120 000 years ago when temperatures over Greenland were about 5 °C warmer than today. But the Greenland ice sheet did not melt completely. Sea level rose to about 5 m above today's level, with melting Greenland ice contributing 1–2 m (Dahl Jensen, pers. com.). Because global warming is amplified near the poles, a future temperature rise of 5 °C in Greenland may be reached when global average temperature rises by around half of this, which is within the range of IPCC projections for this century.

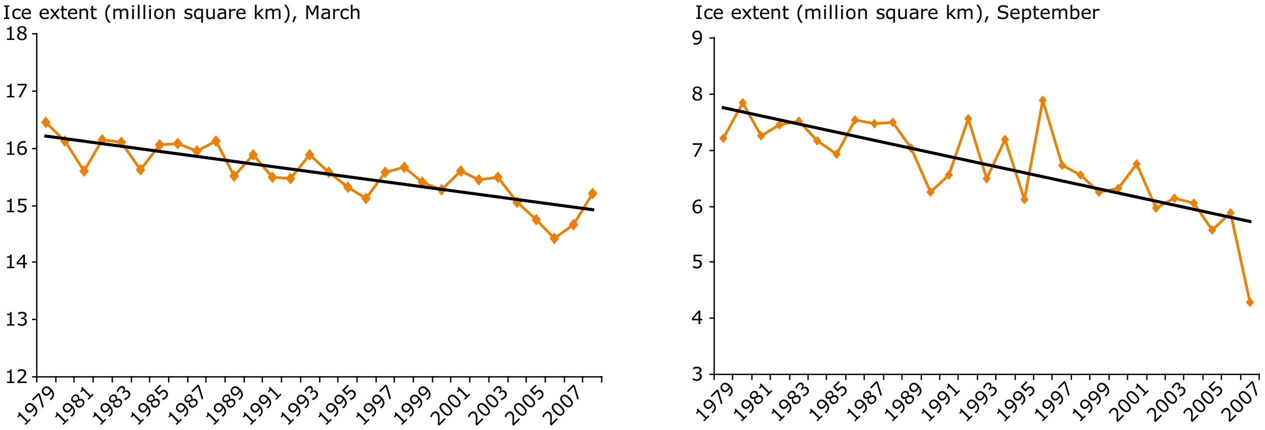
The ice sheets of Greenland and Antarctica have previously been associated with slow climate responses over thousands of years. But the acceleration of ice movement has caused a rethinking of how rapidly they respond to warming. Paleodata show periods of rapid melting of the large continental ice sheets after the last ice age, resulting in an average rise of sea level of 1 cm/year and peak rates up to 4 cm/year (UNEP, 2007). Shrinkage seems to be a faster process than growth, probably because accelerated ice flow plays an important role in retreat. A better understanding of these processes is of vital importance for assessing how much we can expect the flow of meltwater from the Greenland ice sheet to increase.

5.3.5 *Arctic sea ice*

Key messages

- The extent of the sea ice in the Arctic has declined at an accelerating rate, especially in summer. The record low ice cover in September 2007 was roughly half the size of the normal minimum extent in the 1950s.
- The summer ice is projected to continue to shrink and may even disappear at the height of the summer melt season in the coming decades. There will still be substantial ice in winter.
- Reduced polar ice will speed up global warming and is expected to affect ocean circulation and weather patterns. Species specialised for life in the ice are threatened.
- Less ice will ease access to the Arctic's resources. Oil and gas exploration, shipping, tourism and fisheries will offer new economic opportunities, but also increase pressures and risks to the Arctic environment.

Figure 5.14 Average extent of arctic sea ice in March and September 1979–2007



Note: Arctic sea ice grows to its greatest yearly size in March and melts to its lowest size in September. The figure shows the average ice extents for these two months after 1979. The linear trend for March indicates that the Arctic is losing an average of 44 000 km^2 of ice per year in winter. The corresponding value for September and summer is 72 000 km^2.

Source: National Snow and Ice Data Centre, Boulder (http://nsidc.org/data/seaice_index/).

Relevance

Reduction in Arctic sea ice has several feedbacks to the climate system. Snow-covered ice reflects 85 % of the sunlight (high albedo), whereas open water reflects only 7 % (low albedo). Less ice and snow will therefore accelerate both sea-ice decline and global warming. Reduced ice formation will also reduce the formation of dense deep water which contributes to driving ocean circulation. As the ice cover influences air temperature and the circulation of air masses, changes in weather patterns such as storm tracks and precipitation can be expected even at mid-latitudes (Serreze *et al.*, 2007). Warming over the Arctic Ocean can also penetrate into the surrounding continents, raising concern about thawing of the permafrost with release of additional greenhouse gasses (Lawrence *et al.*, 2008).

The sea ice is an ecosystem filled with life uniquely adapted to these conditions, from micro-organisms in channels and pores within the ice, rich algal communities underneath, to fish, seals, whales and polar bears. The diversity of life in the ice usually grows with the age of the ice floes. As the ice gets younger and smaller, the abundance of ice-associated species will be reduced, with a risk of extinction for some of them. Indigenous Arctic peoples adapted to fishing and hunting will face large economic, social and cultural changes.

Less summer ice will ease access to the Arctic Ocean's resources, though remaining ice will still

Map 5.17 The 2007 minimum sea-ice extent

Average sea-ice extent for September 2007

Sea ice

Median September monthly sea-ice extent 1979–2000

Note: The extent of the summer sea ice in September 2007 reached a historical minimum, 39 % below the climatic average for the first two decades of satellite observations (red line). The weather conditions that summer were dominated by clear skies. Continuous warm winds blew the ice towards the coast of Canada-Greenland and out into the north-east Atlantic, where it melted.

Source: National Snow and Ice Data Centre, Boulder (http://nsidc.org/news/press/2007_seaiceminimum/20070810_index.html).

pose a major challenge for operations most of the year. Expectations of large undiscovered oil and gas resources are already driving the focus of the petroleum industry and governments northwards. As marine species move northwards with warmer sea and less ice, so will the fishing fleet. It is however hard to tell whether the fisheries will become richer or not; fish species react differently to changes in marine climate, and it is hard to predict whether the timing of the annual plankton blooms will continue to match the growth of larvae and young fish. Shipping and tourism are likely to increase, although drift ice, short sailing seasons and lack of infrastructure will impede a rapid development of transcontinental shipping of goods; it is more likely that traffic linked to extraction of Arctic resources on the fringes of the Arctic sea routes will grow first. These activities offer new economic opportunities. At the same time they represent new pressures and risks to an ocean that has so far been closed to most economic activities by the ice. This should be met by better international regulations of these activities.

High interest in getting access to the resources in the Arctic may create tensions and security problems. However most borders in the Arctic Ocean have been drawn, thereby clearly defining who has the ownership to the resources and right to manage them. In the remaining unresolved issues of delimitation of the Exclusive Economic Zones and extended continental shelves, all the coastal states of the Arctic Ocean follow the procedures of the UN Convention of the Law of the Seas.

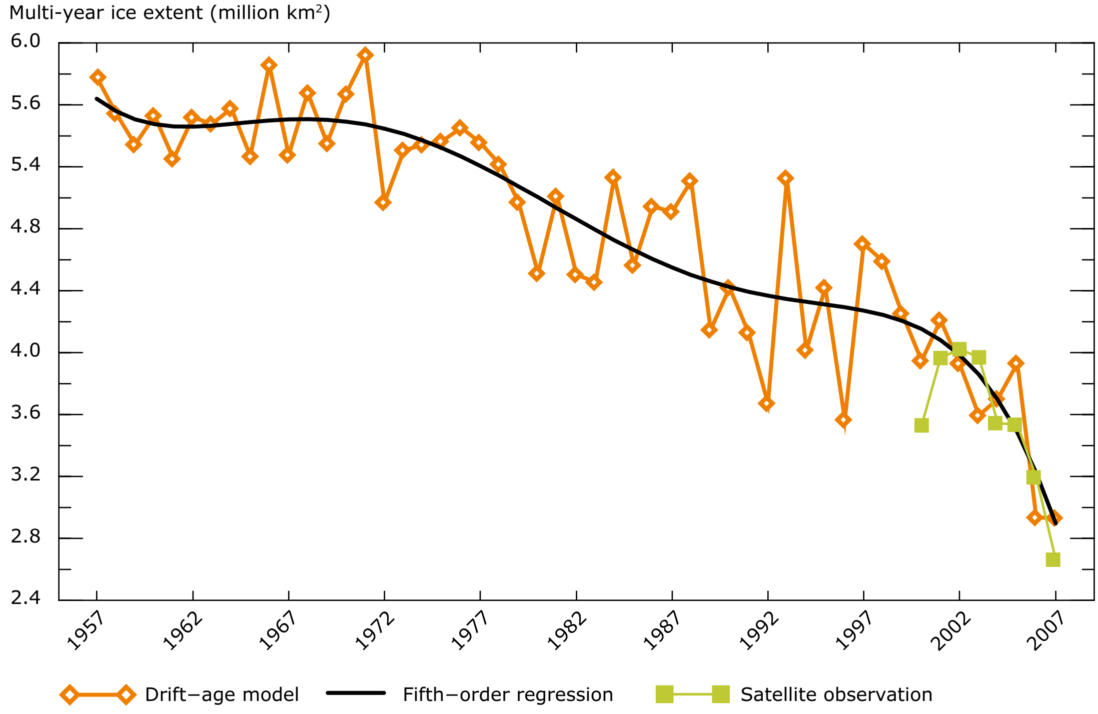
Past trends

The extent of the minimum ice cover at the end of the melt season in September 2007 broke all previously observed records. If older ship and aircraft observations are taken into account, sea ice coverage may have halved since the 1950s (NSIDC, 2007; Meier, 2007). Since the more reliable satellite observations started in 1979, summer ice has shrunk by 10.2 % per decade (Comiso *et al.*, 2008; NSIDC, 2007). This strong negative trend

The polar cod is a key species in the sea ice ecosystem; here in ice with ice algae at the surfaces

Photo: © Bjørn Gulliksen; www.UWPhoto.no

Figure 5.15 Area of multi-year Arctic sea ice in March 1957–2007



Note: The area of thick, multi-year sea ice in the Arctic Ocean is decreasing. The figure is based on a combination of modelling and satellite observations.

Source: Nghiem *et al.*, 2007.

was further reinforced when the second lowest minimum extent was recorded summer 2008. The reduction in maximum winter extent is smaller, with a decrease of 2.9 % per decade (Figure 5.14) (Stroeve *et al.*, 2007). Both summer and winter declines have accelerated (Comiso *et al.*, 2008).

The Arctic sea ice is also getting thinner and younger since less ice survives the summer to grow into thicker multi-year floes. There has been a remarkable shift in its composition towards less multi-year ice and larger areas covered with first-year ice (Figure 5.15). The first-year ice is weaker and melts easier in summer.

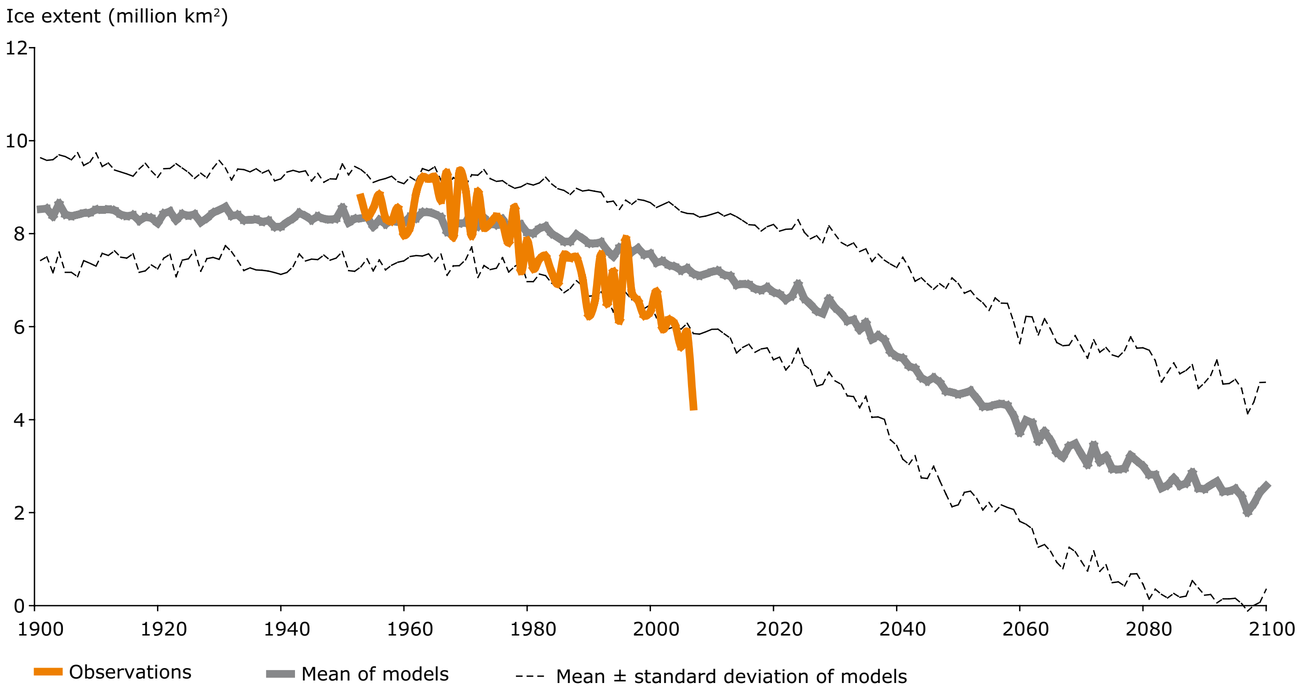
Observations of thickness are more scattered, and it is hard to calculate trends for the whole ice cover. Submarine data have been considered to be most representative and have demonstrated a decrease of 40 % from an average of 3.1 m in 1956–1978 to 1.8 m in the 1990s (UNEP 2007). British submarine data from 2007 show continued thinning (Wadham, pers. com.). German observations from the area around the North Pole and towards northeast Greenland indicate that ice thickness there decreased by 44 % from 2001 to 2007. This was due mainly to a fundamental regime shift from multi-year ice to first-year ice. But there has also been a general thinning of the ice (Nghiem *et al.*, 2007; Haas *et al., 2008*) These results are in stark contrast with observations between Ellesmere Island and 86°N, where ice thickness was still above 4 m in 2006 (Haas *et al.*, 2006).

Arctic sea ice reacts very sensitively to changes in air and ocean temperatures as well as winds, waves and ocean currents (both thermodynamic and dynamic forcing). There are strong imprints of natural variability in the observed changes, e.g. due to regular shifts in the circulation patterns of the polar atmosphere. However, the changes that can be attributed to increases in greenhouse gases seem to be increasing over time (Stroeve *et al.*, 2007).

Projections

The summer ice is very likely to continue to shrink in extent and thickness, leaving larger areas of open water for an extended period. It is also very likely

Figure 5.16 Observed and projected Arctic September sea-ice extent 1900–2100



Note: The retreat of the sea ice has been faster than predicted: Arctic September sea-ice extent from observations (thick orange line) together with the mean value (solid grey line) from 13 IPCC AR4 climate models and the variance (dotted black line) of models runs.

Source: Updated from Stroeve *et al.*, 2007.

that conditions for freezing in winter will persist so that winter sea ice will still cover large areas.

The speed of change is however uncertain. Several international assessments until recently concluded that mostly ice-free late-summers may occur by the end of this century (ACIA, 2004; IPCC, 2007a; UNEP, 2007). But the actual melting has been faster than the average trends simulated by the climate models used for these assessments (Figure 5.16). New model studies suggest that ice-free summers may occur in a much nearer future. (Winton, 2006; Holland *et al.*, 2006; Stroeve *et al.*, 2007). Exactly when is impossible to predict with confidence, due both to the limited understanding of the processes involved and the large variability of the system.

Photo: © John McConnico

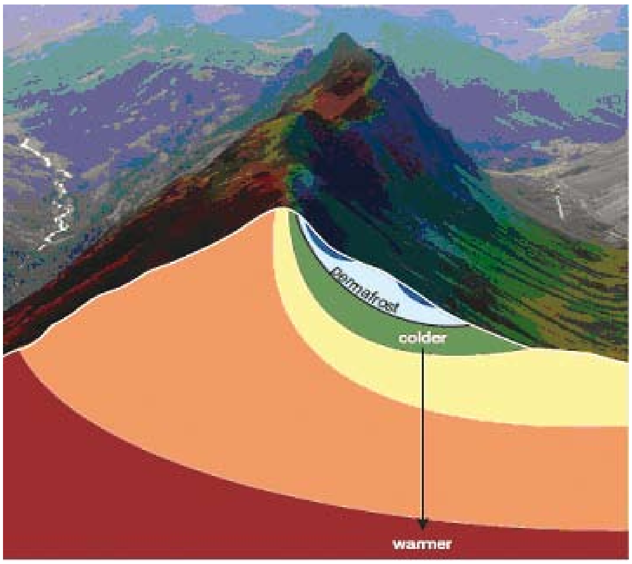
Most studies emphasise that it is very likely that thinner and more vulnerable ice will break up more easily so more heat from the sun will be absorbed in the open water. This can lead to abrupt melting and a high susceptibility to dynamic stress like strong winds when weather conditions are favourable, as in the summer of 2007. An increased influx of warm Atlantic water can also be an important mechanism for further weakening of the sea ice. Unless followed by consecutive years of cold winters, such events will produce a thinner, younger and even more vulnerable ice cover that can melt more easily the next summer, and be more easily transported out of the Polar Ocean.

5.3.6 *Mountain permafrost*

Key messages

- A warming of mountain permafrost in Europe of 0.5–1.0 °C was observed during the past 10–20 years.
- Present and projected atmospheric warming will likely lead to wide-spread thaw of mountain permafrost.
- Warming and melting of permafrost is expected to contribute to increasing the destabilisation of mountain rock-walls, the frequency of rock falls, debris flow activity and geotechnical and maintenance problems in high-mountain infrastructure.

Figure 5.17 Temperature distribution within a mountain range containing permafrost



Note: Permafrost is present in the blue area bordered by a black line.

Source: Gruber and Haeberli, 2007. http://maps.grida.no/go/graphic/mountain-permafrost-patterns-and-temperature-gradients.

Relevance

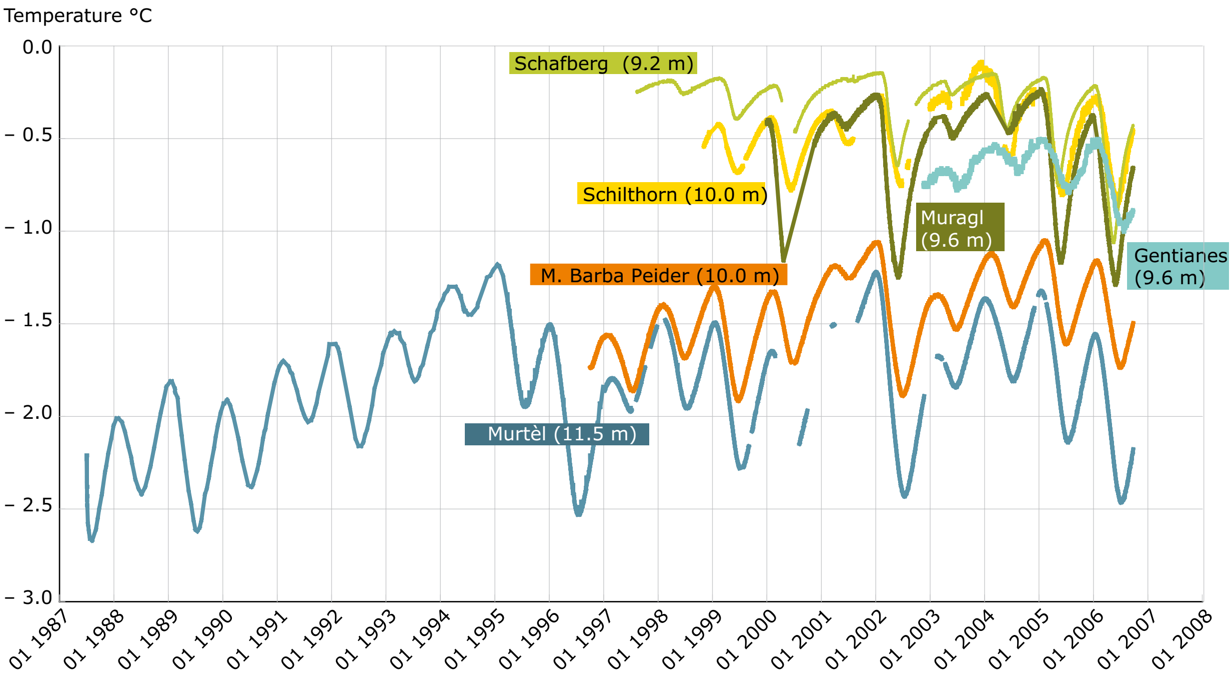
Permafrost is permanently frozen ground and consists of rock or soil that has remained at or below 0 °C continuously for more than two years. Mountain permafrost is the dominating permafrost in Europe, because Arctic permafrost is found in Europe only in the northernmost parts of Scandinavia. Permafrost is abundant at high elevations in mid-latitude mountains, where the annual mean temperature is below – 3 °C. It contains variable amounts of ice and exists in different forms: in steep bedrock, in rock glaciers, in debris deposited by glaciers and in vegetated soil. Because vegetation and circulating groundwater in mountain permafrost areas are mostly absent, the temperature in the deeper rock material is largely determined by the temperature history at its surface. Mountain permafrost therefore contains valuable information on climate change. Temperature profiles from alpine boreholes are difficult to interpret in terms of past trends due to the effects of the complex topography (Figure 5.17) and the availability of insulating snow-cover (Gruber *et al.*, 2004a). Nevertheless, monitoring of temperature change at depth provides valuable data on the thermal response of permafrost to climate change.

Permafrost influences the evolution of mountain landscapes and affects human infrastructure and safety. Permafrost warming or thaw affects the potential for natural hazards, such as rock falls (e.g. at the Matterhorn in summer 2003) and debris flows (Noetzli *et al.*, 2003; Gruber and Haeberli, 2007). At least four large events involving rock volumes of more than 1 million m^3 have occurred in the Alps during the past decade. Their effects on infrastructure have motivated the development of technical solutions to improve design lifetime and safety (Philips *et al.*, 2007).

Past trends

Data from a north-south transect of boreholes, 100 m or more deep, extending from Svalbard to the Alps (European PACE-project) indicate a long-term regional warming of permafrost of 0.5–1.0 °C during the recent decade (Harris *et al.*, 2003). In Scandinavia and Svalbard, monitoring over 5–7 years shows warming down to 60 m depth and current warming rates at the permafrost surface of 0.04–0.07 °C/year (Isaksen *et al.*, 2007). In Switzerland, a warming trend and increased active-layer depths were observed in 2003, but results varied strongly between borehole locations due to variations in snow cover and ground properties (PERMOS, 2007). At the Murtel-Corvatsch (rock-glacier) borehole in the Swiss Alps, the only long-term data record (20 years), permafrost temperatures

Figure 5.18 Temperature measured in different boreholes in mountain permafrost in Switzerland 1987–2007



Note: Measured at ca. 10 m depth in rock-glaciers and frozen rock-walls.

Source: PERMOS, 2007.

in 2001, 2003 and 2004 were only slightly below – 1 °C (Figure. 5.18) and were, apart from 1993 and 1994, the highest since measurements began in 1987 (Vonder Mühll *et al.*, 2007). Such data measured at rock-glaciers are difficult to interpret because the sub-surface thermodynamics in ice-rich frozen debris is rather complex. Complementary and clearer signals on thawing permafrost are expected from boreholes drilled directly into bedrock

Rock glacier Murtel-Corvatsch

Photo: © M. Phillips, SLF

(e.g. Schilthorn, M. Barba Peider; Figure 5.18). Corresponding monitoring programmes, such as PACE and PERMOS, however, only started less than a decade ago.

Projections

No specific projections on the behaviour of mountain permafrost are yet available, but changes in mountain permafrost are likely to continue in the near future and the majority of permafrost bodies will experience warming and/or melting. 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 is projected for the Alpine region (Jacob *et al.*, 2007). Further rises in temperature and melting permafrost could increasingly destabilise mountain walls and increase the frequency of rock falls, posing problems to mountain infrastructure and communities (Gruber *et al.*, 2004a). The warming and thaw of bedrock permafrost 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 joint systems will further accelerate destabilisation (Gruber and Haeberli, 2007).