5 Climate change and Europe's mountains

Europe's mountains stretch from the Arctic through the temperate and into the subtropical climatic zone of the Northern hemisphere, as well as from maritime to continental environments. As such, they encompass a wide range of bioclimatic zones. Across these very diverse mountains, local climatic and other environmental controls vary enormously as their effects are superimposed upon macro-scale factors influencing mountain climates, such as continentality and latitude. Recognising the sensitivity of mountain environments and the potential vulnerabilities of these environments to climate change, the scientific community has increased research on global change in mountain regions including the possible impacts of anthropogenic climate change (Becker and Bugmann, 2001; Huber *et al.*, 2005; EEA, 2009). This chapter presents recent observed changes in the climate of Europe's mountains and likely changes during this century. The likely impacts of these changes on glacier, hydrological and ecological systems are presented in Box 6.2 and Sections 6.5, 6.6, and 8.3, respectively.

5.1 Changes in climate across Europe

The availability of climatic data across Europe's mountain regions is highly variable in both space and time, with particularly high spatial density and length of record in the Alps, and lower densities and lengths of record in other mountain regions (Price and Barry, 1997). Consequently — and also because the spatial resolution of Global Climate Models (GCMs) generally does not permit detailed prediction of climates of regions such as mountains, and relatively few studies using statistical downscaling methods or regional climate models have considered mountain areas — this introductory section mainly presents data for Europe as a whole, rather than mountains specifically, to provide a context for the following sections.

5.1.1 Observed changes in climate

Observations of increases in global average air and ocean temperatures, widespread melting of snow and ice, and rising sea level are unequivocal evidence of warming of the climate system globally. Direct observations and proxy records indicate that historical and recent changes in climate in many mountain regions are at least comparable with, and locally may be greater than, those observed in the adjacent lowlands. Global mean temperature has increased by 0.8 °C compared with pre-industrial times for land and oceans, and by 1.0 °C for land alone (EEA, 2008). Most of the observed increase in global average temperatures is very likely due to increases in anthropogenic greenhouse gas concentrations (Albritton *et al.*, 2001). During the 20th century, most of Europe experienced increases in average annual surface temperature (average increase 0.8 °C), with more warming in winter than in summer (IPPC, 2007). European warming has been greater than the global average, with more pronounced warming in the southwest, the northeast, and mountain areas. As the observed trend in western Europe over the past decade appears stronger than simulated by GCMs, climate change projections probably underestimate the effects of anthropogenic climate change (van Oldenborgh *et al.*, 2009).

5.1.2 Projected regional changes

Landmasses are expected to warm more than the oceans, and northern, middle and high latitudes more than the tropics (Giorgi, 2005, 2006; Stendel *et al.*, 2008; Kitoh and Mukano, 2009; Lean and Rind, 2009). Warming in the atmosphere is also expected to be more pronounced at progressively higher elevations in the troposphere, along a latitudinal gradient from the northern mid-latitudes to approximately 30 °S, with a maximum above the tropics and sub-tropics (Albritton *et al.*, 2001). Many European mountain regions are situated in these high-latitude zones of anticipated enhanced warming.

Projections from GCMs generally show increased precipitation at high latitudes (Frei *et al.*, 2003). With more precipitation falling as rain rather than snow in a warmed atmosphere, soil moisture in northern areas in winter would increase, while in summer, simulations suggest a general tendency towards mid-latitude soil drying (Christensen, 2001). Despite possible reductions in average summer precipitation over much of Europe, precipitation amounts exceeding the 95th percentile are very likely in many areas, thus episodes of severe flooding may become more frequent despite the general trend towards drier summer conditions (Christensen and Christensen, 2002; Christensen, 2004; Pal *et al.*, 2004; Frei *et al.*, 2006).

The details of outputs from different models vary, and so ensemble-based approaches have been used to bring together outputs from a range of models. In such an approach using outputs from 20 GCMs for three of the emission scenarios of the Inter-Governmental Panel on Climate Change (IPCC), the Mediterranean, northeast and northwest Europe are identified, in this order, as warming hot spots (Giorgi, 2006), albeit with regional and seasonal variations in the pattern and amplitude of warming (Giorgi and Lionello, 2008; Faggian and Giorgi, 2009; Brankovic *et al.*, 2010).

Most climate change studies for mountain areas rely on simulations of the future climate using statistical downscaling models (SDMs) or regional climate models (RCMs) forced by boundary data from GCMs. Table 5.1 lists RCM-based studies for different European regions, some of which evaluate model performance in mountainous areas. RCMs also project rising temperatures for Europe until the end of the 21st century, with an accelerated increase in the second half of the century. However, for many regions, there are substantial differences between the RCM surface temperature and precipitation simulations, depending on the driving GCM. There is no clear correlation of differences with regions, but the driving GCM has a dominant effect on temperature during spring, winter, and autumn, which seems to be larger than the effect of the specific RCM (Christensen and Christensen, 2007).

For precipitation, the driving model seems to be relatively most important in spring and summer (Christensen and Christensen, 2007; Déqué *et al.*, 2007). Despite the complex local character of simulated summertime change in RCMs, the larger-scale pattern shows a gradient from increases in Northern Scandinavia to decreases in the Mediterranean region (Frei *et al.*, 2006; Schmidli *et al.*, 2007). In contrast, increases in wintertime precipitation primarily north of 45 °N are a robust feature of RCM projections over Europe, with decreases over the Mediterranean (Frei *et al.*, 2006; Schmidli *et al.*, 2007; Haugen and Iversen, 2008). Overall, therefore, there is likely to be an increase in precipitation in the north and a decrease in the south, with all models agreeing in the north, and 12 out of 16 models agreeing in the south (van der Linden and Mitchell, 2009).

The previous paragraphs refer to changes in mean values. However, for both ecological and human systems, changes in extremes may be far more important (Box 5.1). With regard to temperatures, biases in maximum temperatures during summer, and minimum temperatures during winter, tend to be larger at the extremes than in the mean values (Beniston *et al.*, 2007; Hanson *et al.*, 2007). RCMs generally underestimate maximum temperatures during summer in northern Europe and overestimate them in eastern Europe (Frei *et al.*, 2006). In winter, minimum temperatures are overestimated over most of Europe. The spread between the models is generally also larger at the tails of the probability distributions (Frei *et al.*, 2006). With regard to precipitation, simulated change in extremes from various RCMs shows a seasonally-distinct pattern (Frei *et al.*, 2006; Jacob *et al.*, 2007; Koffi and Koffi, 2008). In winter, land north of about 45 °N would experience an increase in multi-year return values, and the Mediterranean region would experience small changes, with a general tendency towards decreases (Hanson *et al.*,

Table 5.1 Recent literature, RCM projections and evaluations for European mountain regions

Year	Author	Literature type	Type of study	Region addressed
2003	Frei *et al.*	Journal paper	RCM evaluation	European Alps
2006	Schmidli *et al.*	Journal paper	Downscaling methods comparison	European Alps
2007	Coll	Unpublished PhD thesis	RCM evaluation	Scottish Highlands
2007	Schmidli *et al.*	Journal paper	Downscaling methods comparison	European Alps
2008	Lopez-Moreno *et al.*	Journal paper	RCM inter-comparison	Pyrenees
2008	Noguez-Bravo *et al.*	Journal paper	GCM projections	Mediterranean mountains
2009	Smiatek *et al.*	Journal paper	RCM inter-comparison	European Alps

2007). The increase in wintertime precipitation extremes is a robust feature in RCM projections over Europe, whereas the character of change for summer is more complex (Beniston *et al.*, 2007; Christensen and Christensen, 2007; Déqué *et al.*, 2007; Schmidli *et al.*, 2007; Lopez-Moreno *et al.*, 2008). The larger-scale pattern shows a gradient from increases in northern Scandinavia to decreases in the Mediterranean region which is fairly similar between models. Addressing uncertainty in scenarios of summer precipitation extremes is a research priority (Frei *et al.*, 2006).

Box 5.1 Climate change and extreme events in the mountains of northern Sweden

Climate warming in the Swedish sub-Arctic since 2000 has reached a level where the current warming has exceeded that of the late 1930s and early 1940s and, significantly, has crossed the 0 °C mean annual temperature threshold that causes many cryospheric and ecological impacts. The accelerating trend of temperature increase has driven trends in snow thickness, loss of lake ice, increases in active layer thickness, and changes in tree line location and plant community structure. Changes in the climate are associated with reduced temperature variability at the seasonal scale, particularly a loss of cold winters and cool summers, and an increase in extreme precipitation events that decrease the stability of mountain slopes and cause infrastructure failure. Both mean annual precipitation and extreme precipitation events have increased, especially the number of days with more than 20 mm precipitation.

Even more important from a landscape change perspective, the 'extremes of the extremes' have also increased. Except from one extreme precipitation event in the 1920s, these extremes have reached higher and higher levels, with increasing daily maxima up to 60 mm. Several of the geomorphological and hydrological impacts of these extreme events are well known in the Abisko area, where both a railroad and a road pass close to mountain slopes. The extreme precipitation events have caused disturbances for traffic; the latest extreme precipitation event, on 20 July 2004, triggered a number of debris flows and landslides and, for the first time in this area, badly damaged a road bridge. Parts of the road-bank were eroded and transported away by the running water, and it was only because of an attentive driver that severe car accidents were avoided. The trajectory of increasing extremes of extremes over time renders the planning, building and meteorological concept of 'return frequency' of extreme events obsolete, as each new extreme has not been experienced earlier in the instrumental record. Planning adaptation to climate change therefore requires the formulation of new concepts and building guidelines.

Not only *precipitation* affects and causes changes in these landscapes; extreme *temperature* events are also occurring more frequently in winter. Experimental studies and findings from observations following natural events show that short winter warming events can cause major damage to plant communities even at the landscape scale. In such an event in December 2007, the temperature rose to 7 °C within a few days, resulting in more or less complete loss of the snow cover and hence exposure of the vegetation when low temperatures returned. After a short period of no or little snow cover, the temperature fell and, a few days later, the vegetation was again covered by snow. This single warming event, about 10 days long, caused substantial impacts to the vegetation cover. In the following summer, satellite-derived Normalised Differential Vegetation Index (NDVI) showed damage of dwarf shrubs over almost 15 000 km². Field studies in the affected areas showed that the frequency of dead roots of the dominant shrub, *Emperum hermaphroditum*, increased up to 16-fold, resulting in almost 90 % less summer growth compared with undamaged areas. Similarly, field experiments using infra-red heating lamps and soil warming cables to simulate extreme temperature events have shown that single-day snow-free conditions followed by freezing result in c. 20 times greater frequency of dead roots and almost 50 % less shoot growth of *E. hermaphroditum* and near complete absence of berry production in *Vaccinium myrtillus*.

These events are of major concern both for conservation — as animals such as lemmings that depend on continuous snow cover decline, resulting in loss of predators such as the snowy owl and arctic fox — and for the reindeer-herding Sami, as damaged vegetation needs to be replaced by alternative pastures or expensive supplementary food pellets.

Source: Christer Jonasson and Terry Callaghan (Abisko Scientific Research Station, Sweden).

5.2 Changes in climate in European mountains

5.2.1 Long-term trends in climatic variables

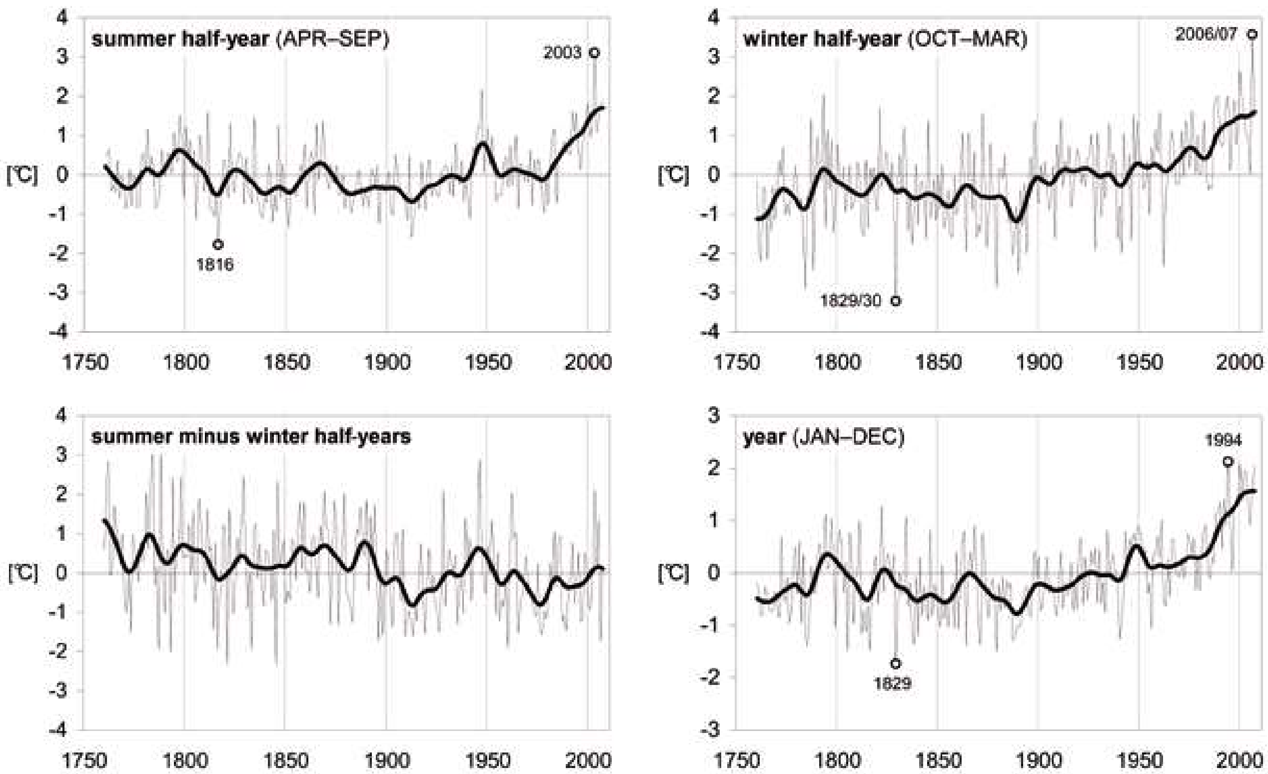
Evidence of recent climate change comes from observations at high-altitude sites across the globe, with observed changes including increased winter rainfall and rainfall intensity (Groisman *et al.*, 2005; Malby *et al.*, 2007) and temperatures increasing more rapidly than at lowland sites, particularly through increases in minimum (nocturnal) temperatures (Bradley *et al.*, 2006). However, evidence of altitude-based differences in warming is not equivocal (Pepin and Seidel, 2005). Actual and potential responses in cryospheric variables include: a rise in the snowline; a shorter duration of snow cover (Martin and Etchevers, 2005); changes in avalanche frequency and characteristics; glacier recession (Haeberli, 2005; Box 6.2); break-out of ice-dammed lakes; warming of perennially-frozen ground; and, thawing of ground ice (Barry, 2002; Harris *et al.*, 2003; Harris, 2005).

As noted above, the availability of climate data is greatest for the Alps (EEA, 2009). A compilation of 87 temperature records, with documentary and narrative reports and gridded reconstructions, some dating back to 1500, shows that 1994, 2001, 2002 and 2003 were the warmest years in the record (Casty *et al.*, 2005). Over the past 250 years, in the Greater Alpine Region (GAR):

- there has been an overall annual temperature increase of ~ 2.0 °C from the late 19th to early 21st century;
- following a decrease in temperature from 1790 to 1890, 20th century warming was more pronounced in summer than in winter;
- during the past 25 years, winters and summers have warmed at comparable rates, leading to an annual mean temperature increase of 1.2 °C, an increase unprecedented in the instrumental record (Zebisch *et al.*, 2008).

While temperature changes have followed similar patterns across the Alps (Figure 5.1), trends at the

Figure 5.1 Change in temperature for the Greater Alpine Region, 1760–2007: Single years and 20-year smoothed mean series



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

Source: ZAMG-HISTALP database (version 2008, including the recent Early Instrumental (EI) period correction (Böhm *et al.*, 2009).

sub-regional scale are different for precipitation (EEA, 2009; Figure 5.2). Over the past two centuries, there has been a trend of increasing precipitation in the north-west Alps (eastern France, northern Switzerland, southern Germany, western Austria) and a decreasing precipitation in the south-east (Slovenia, Croatia, Hungary, south-east Austria, Bosnia and Herzegovina) (Auer *et al.*, 2005).

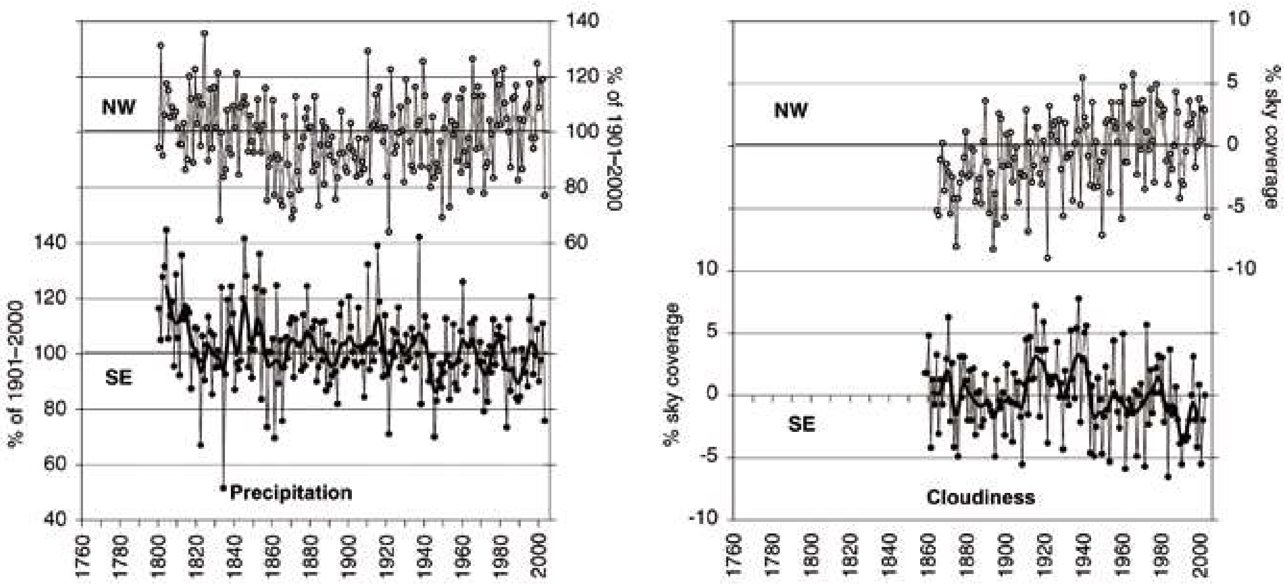
The frequency of temperatures exceeding the freezing point during the winter season in eastern Switzerland has more than doubled during periods of high North Atlantic Oscillation (NAO) index, compared to periods with low index values, thereby increasing the chances of early snowmelt. Despite strong inter-annual variability, overall trends in snow cover have not changed much, as the rate of warming during the 20th century is modest in relation to future projections (Beniston, 2006). However, the upper tens of metres of permafrost warmed by 0.5 °C to 0.8 °C during the 20th century (Gruber *et al.*, 2004), especially at higher altitudes, with accompanying thickening of the seasonal active layer (Harris *et al.*, 2009).

After the Alps, the longest records and most dense networks are in parts of the Carpathians, the mountains of the British Isles, and the mountains of Scandinavia (Price and Barry, 1997). Changes have also been observed for areas of the more maritime UK uplands, including evidence of more rapid warming (Holden and Adamson, 2002) and marked precipitation changes (Barnett *et al.*, 2006; Fowler and Kilsby, 2007; Maraun *et al.*, 2008). In the Carpathians, annual temperature variability increased from 1962 to 2000 (e.g. from 0.3 °C to 0.5 °C in the Bucegi Mountains; from 0.5 °C to 0.7 °C in the Semenic Mountains; and, from 0.8 °C to 0.9 °C in the southern Carpathians and Apuseni Mountains (Ionita and Boroneant, 2005; Micu, 2009)). At other Carpathian locations, winter temperature increases of ~ 3 °C characterised the end of the 1961–2003 period compared to the long-term average (Micu and Micu, 2008; Micu, 2009).

Central European station data for 1901–1990 and 1951–1990 indicate that mountain stations show only small changes of the diurnal temperature range from 1901 to 1990, while low-lying stations in the western Alps show a significant decrease in the diurnal temperature range, caused by a strong increase in the minimum temperature. For 1951–1990, the diurnal temperature range decreased at the western low-lying stations, mainly in spring, but remained roughly constant at the mountain stations (Weber *et al.*, 1997). Proxy measures elsewhere in European mountain regions also offer evidence of recent changes. For example:

- Borehole monitoring of permafrost temperatures showed that relief and aspect led to greater variability between Swiss and Italian Alpine

Figure 5.2 Annual precipitation series (left graph) and annual cloudiness series (right graph) for the northwest (NW) and southeast (SE) Alps



Note: All values relative to 1901–2000 averages. Single years (thin lines) and 10-year smoothed means (bold lines).

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

boreholes than between those in Scandinavia and Svalbard. However, 15 years of thermal data from the 58 m-deep Murtèl–Corvatsch permafrost borehole in Switzerland, drilled in ice-rich rock debris, showed an overall warming trend, with high-amplitude inter-annual fluctuations reflecting early winter snow cover fluctuations more strongly than air temperatures (Harris *et al.*, 2003).

- In upland lakes, spring temperature trends were highest in Finland; summer trends were weak everywhere; autumn trends were strongest in the west, in the Pyrenees and western Alps; while winter trends varied markedly, being high in the Pyrenees and Alps, low in Scotland and Norway and negative in Finland (Thompson *et al.*, 2009).

5.2.2 *Climate change scenarios*

A number of studies (Giorgi *et al.*, 1994; Beniston and Rebetez, 1996; Fyfe and Flato, 1999) suggest that the highest mountainous areas are expected to experience the most intense increases in temperature. If this occurs, the impact of climate warming could be enhanced due to the high dependence of surrounding regions on the water resources provided by the mountains (Beniston, 2003, 2006); this could be particularly important in river basins where snow and glaciers play a major part in regulating seasonal hydrological cycles (Barnett *et al.*, 2006); this is discussed further in Chapter 6.

Figure 5.3 presents predicted seasonal changes in precipitation and temperature in the Alps up to the end of the 21st century. By 2071–2100, summers in Europe's southern mountains are projected to warm by 5–6 °C (Räisänen *et al.*, 2004; Christensen and Christensen, 2007), in the Alps by up to 5 °C (Smiatek *et al.*, 2009; van der Linden and Mitchell, 2009; Box 5.2) and in the north by 3–5 °C. A similar latitudinal contrast is projected for 21st century precipitation, with northern mountains experiencing increases of 20–50 %, and decreases of ~ 25–50 % in southern ranges, associated with a north-eastward extension of the summer mean Atlantic subtropical high pressure system. In summer, most RCMs simulate a strong decrease in mean precipitation for the Alps (Frei *et al.*, 2003, 2006; Schmidli *et al.*, 2007; Smiatek *et al.*, 2009), a pattern also found for the Pyrenees (Lopez-Moreno *et al.*, 2008). One significant outcome may be an increased frequency of lightning fires (Box 5.3). Mean net shortwave length radiation is projected to increase by around 10 watts per square metre (W/m^2) over much of Europe during the summer (Lenderink *et al.*, 2007). Another climatic element strongly affected by circulation change is wind speed. In general, summer wind speeds are projected to decrease in southern Europe but to increase in the north (Räisänen *et al.*, 2004), as the Atlantic storm track shifts polewards (Bengtsson *et al.*, 2006).

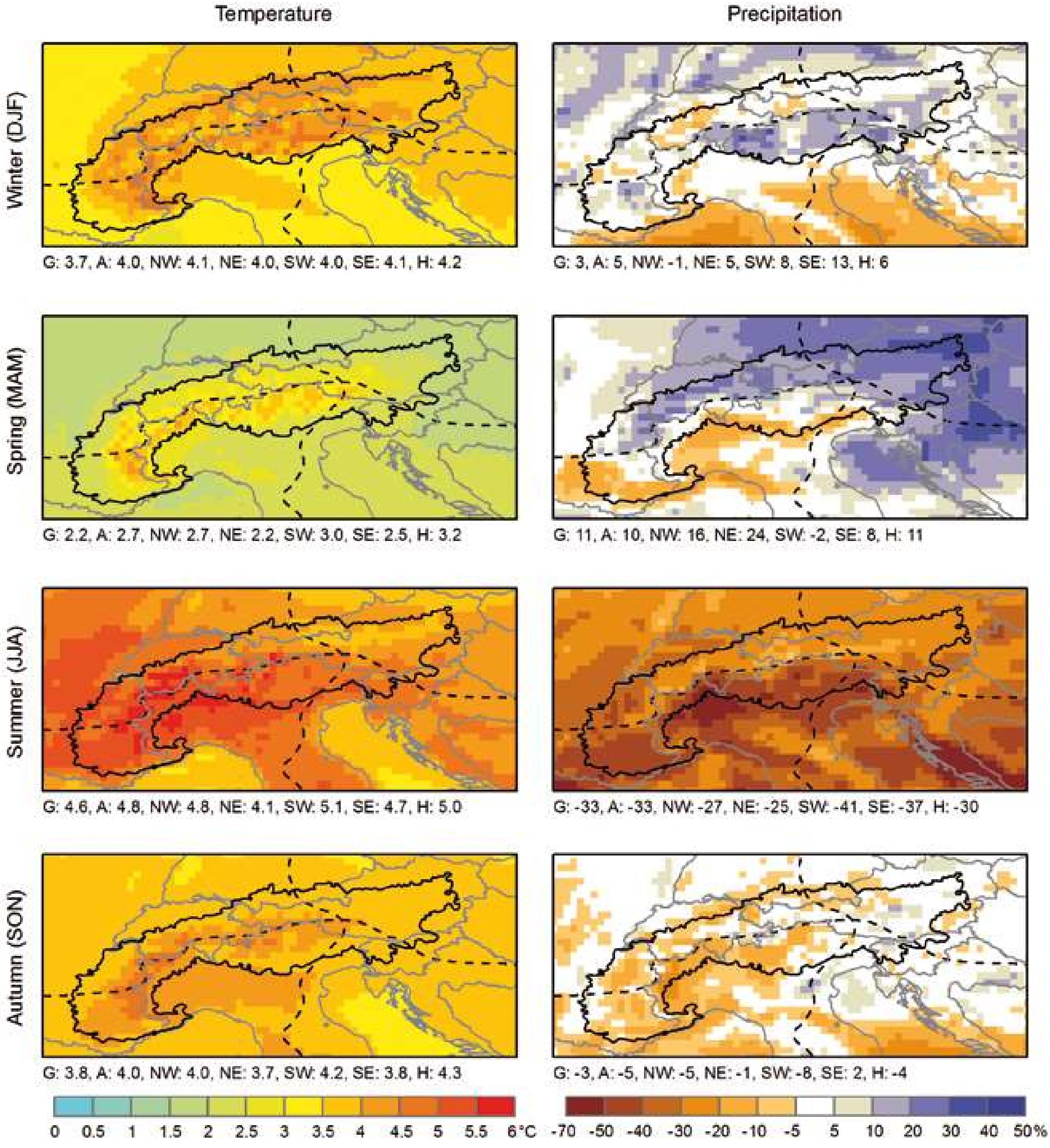
Winters are also projected to warm, with a geographically consistent pattern of 4–5 °C increases in mean winter temperature in Europe's eastern mountains, but increases of 1–3 °C in western, more maritime, settings (Christensen and Christensen, 2007; Räisänen *et al.*, 2004). All scenarios agree on a general increase in winter precipitation in northern and central Europe, and a decrease to the south of the Alps. However, large local changes in precipitation are projected for parts of Norway and the Alps, where the pronounced topography makes any change in precipitation pattern very sensitive to wind direction. A number of scenarios indicate a distinct wintertime increase in storm track density over the British Isles and across into Western Europe, but a decrease in the Mediterranean (Bengtsson *et al.*, 2006). However, while the basic dynamics governing shifts in the strength and path of the mid-latitude storm track are well understood, the ability of models to reproduce these is limited. As it is unclear which, if any, climate model is capable of satisfactory projections, there is considerable uncertainty about the future behaviour of storm tracks in the north-east Atlantic (Woolf and Coll, 2007).

5.2.3 *Changes in snow cover and permafrost*

Both temperature and precipitation increases to date have impacted mountain snowpacks simultaneously on a global scale. However, the nature of the impact is strongly dependent on geographic location, latitude, and elevation, among other factors (Stewart, 2009). In general, snow cover throughout the Alps decreased throughout the 20th century, in particular since the 1980s and during the latter part of the century (Stewart, 2009), and continues to do so (EEA, 2009).

Climate models suggest that future snowfall in the Alps could be reduced by 3 % in the winter, with altitudes above 1 500 m experiencing a loss of approximately 20 % up to the late 21st century (EEA, 2009); other results suggest that snow below 500 m could almost disappear completely (Jacob *et al.*, 2007). The duration of snow cover is expected to decrease by several weeks for each projected °C of temperature increase in the Alps, with the greatest sensitivity in the middle altitude bands (575–1 373 m) in winter and spring (Hantel *et al.*,

Figure 5.3 Seasonal changes in precipitation and temperature until the end of the 21st century, according to CLM Scenario A1B



Source: EEA, 2009.

2000; Wielke *et al.*, 2004; Martin and Etchevers, 2005). Keller *et al.* (2005) report an average decrease of a month in the modelled snowmelt for Alpine rock and sward habitats in response to a 4 °C increase in mean temperature. According to model projections following different greenhouse gas emission scenarios, the thickness and duration of snowpack in the Pyrenees will decrease dramatically over the next century, especially in the central and eastern areas of the Spanish Pyrenees (Lopez-Moreno *et al.*, 2008). The magnitude of these impacts will follow a marked altitudinal gradient. The maximum

Box 5.2 Future climate in the Greater Alpine Area

Over the past century, the mean temperature in the Alps increased by 1.1 °C. GCMs indicate that, by 2100, the temperature of the Alpine region, relative to the period 1980–1999, may increase by up to 5 °C (IPCC, 2007), and that summer precipitation will decrease significantly. Analysis of monthly mean values from six GCMs using the A1B emission scenario for the Greater Alpine Area for 2071–2100 showed increases in temperature of 3.4 °C in winter and 4.3 °C in summer relative to 1961–1990. On average, these models show that precipitation will increase by 10 % in winter and decrease by 30 % in summer (Smiatek *et al.*, 2009).

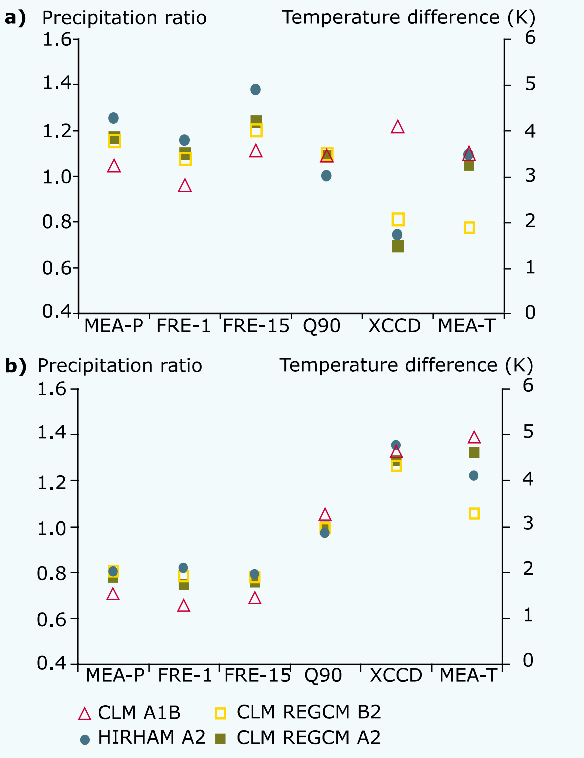
Several statistical and dynamic downscaling approaches have been applied to derive highly resolved climate change information for the Alpine region. While the regional models reproduce spatial precipitation patterns and the annual cycle in complex terrain, there are still large biases in precipitation when compared with observations. In the PRUDENCE project (Christensen and Christensen, 2007), an ensemble of 25 RCMs, mostly run with a horizontal resolution of 0.5 °C in a time slice experiment using the A2 scenario, showed a mean increase in the seasonal mean temperature in the Alps of 3.53 °C in winter and 5.04 °C in summer, compared to the 1961–1990 mean. The relative seasonal mean precipitation change was + 20 % in winter and – 26 % in summer. Schmidli *et al.* (2007) evaluated six statistical and three dynamical downscaling models, and found a strong decrease in mean precipitation for the entire Alpine region in summer for 2071–2100; a substantial reduction in the frequency of wet days in summer resulted in a large increase (50–100 %) in the maximum length of dry spells. Most models also simulate an increase in precipitation intensity on wet days in summer and in the 90 % quantile of precipitation on wet days in winter, compared to 1961–1990. Some models indicate increased precipitation intensity in summer, despite the strong decrease in mean precipitation.

Figure 5.4 shows the simulated changes in temperature and various precipitation statistics as simulated by two RCMs — HIRHAM (Christensen and Christensen, 2007) and RegCM (Gao *et al.*, 2006) — driven with boundary forcings from the HadAM3 GCM, and also the transient CCLM (Rockel *et al.*, 2008) RCM, driven with boundary data from the ECHAM5 GCM as evaluated by Smiatek *et al.* (2009). For the Alpine region, the RCM models simulate a winter temperature increase for 2071–2100 of 2 °C to over 3 °C and, in summer, of almost 5 °C compared to 1960–1990. Summer precipitation decreased up to 29 %, with a substantial increase in the maximum length of dry spells. For winter, all models indicate a precipitation increase, with more wet days and strong precipitation events. In particular regions, however, the RCMs simulate much greater differences: an increase of more than 30 % in winter and a decrease of almost 40 % in summer.

The analysis of the regional climate simulations shows that results based on different regional models, different driving global models, and different emission scenarios show similar trends — but that these differ in the magnitude of the expected climate change signal. Nevertheless, there are still large biases in the reproduction of the current climate, and therefore substantial uncertainties in the magnitude of expected climate change.

Source: Gerhard Smiatek and Harald Kunstmann (Institute for Meteorology and Climate Research, Karlsruhe Institute of Technology, Germany).

Figure 5.4 Simulated change in precipitation (2071–2100 to 1961–1990) and temperature (2071–2100 to 1961–1990) statistics in the Greater Alpine Area in (a) winter and (b) summer for four Regional Climate Models



Note: Statistics: MEAP: mean climatological precipitation, FRE-1: frequency (ratio) of wet-days with precipitation > 1 mm, FRE-15: frequency (ratio) of days with precipitation > 15 mm, Q90: 90 % quantile of the distribution function on wet days, XCCD: maximum number of consecutive dry days, MEA-T: mean climatological temperature.

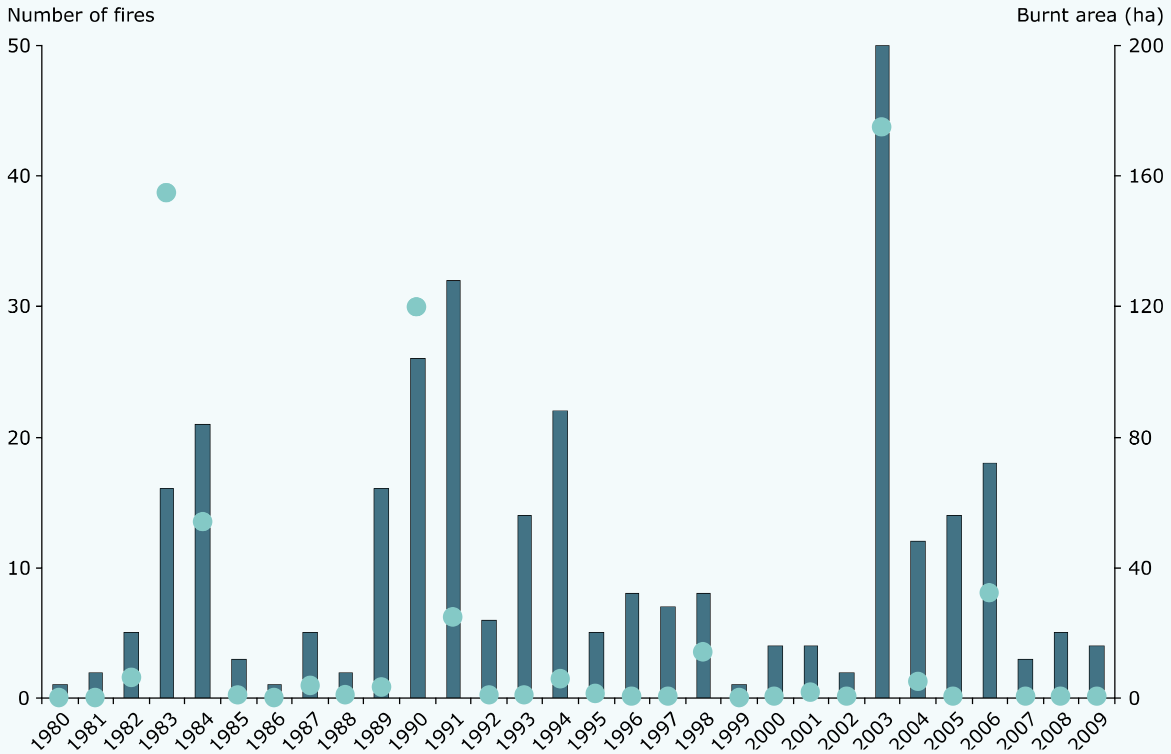
Box 5.3 Lightning-induced fires in the Alpine region

In most forest ecosystems, lightning is the only natural source of ignition (Pyne *et al.*, 1996). As well as factors such as fuel (type, moisture, density and depth) and topography, the frequency and distribution of lightning-caused forest fires greatly depend on weather (drought or lack of precipitation, frequency and type of the thunderstorms and of the associated lightning discharges, and ventilation). This makes lightning-fires of particular relevance for assessing the possible impact of climate change (Street, 1989; Flannigan and van Wagner, 1991; Balling *et al.*, 1992; Weber and Stocks, 1998).

In Europe, most lightning-induced forest fires take place in the southern boreal forests of Fennoscandia (Granström, 1993; Larjavaara *et al.*, 2005) and in the mountain regions from the Iberian Peninsula (Vasquez and Moreno, 1998; Galán *et al.*, 2002) to the Western and Central Alps (Conedera *et al.*, 2006). Lightning-caused forest fires may occur between May and October, but most events (90 % or more) take place during the warm summer months of June to August, with some differences due to the different elevation, expositions and start of the warm season (Granström, 1993; Wotton and Martell, 2005; Conedera *et al.*, 2006). In general, lightning causes fires in coniferous forests located on steep slopes at high elevations. Such fires are often started by an underground ignition that may keep smouldering locally for days and weeks resulting in small-size burned areas (Conedera *et al.*, 2006).

Given their natural origin, the frequency and extent of lightning-ignited fires depend strongly on seasonal weather conditions; data for the southern slope of the Swiss Alps show an increase with drought indices. In the Swiss Alps, the inter-annual variability in fire frequency and burnt area is high, with no clear increasing trend (Figure 5.5).

Figure 5.5 Annual variability in lightning-induced fire frequency (dots) and burnt area (bars) in the Swiss Alps



Source: Swissfire database.

Box 5.3 Lightning-induced fires in the Alpine region (cont.)

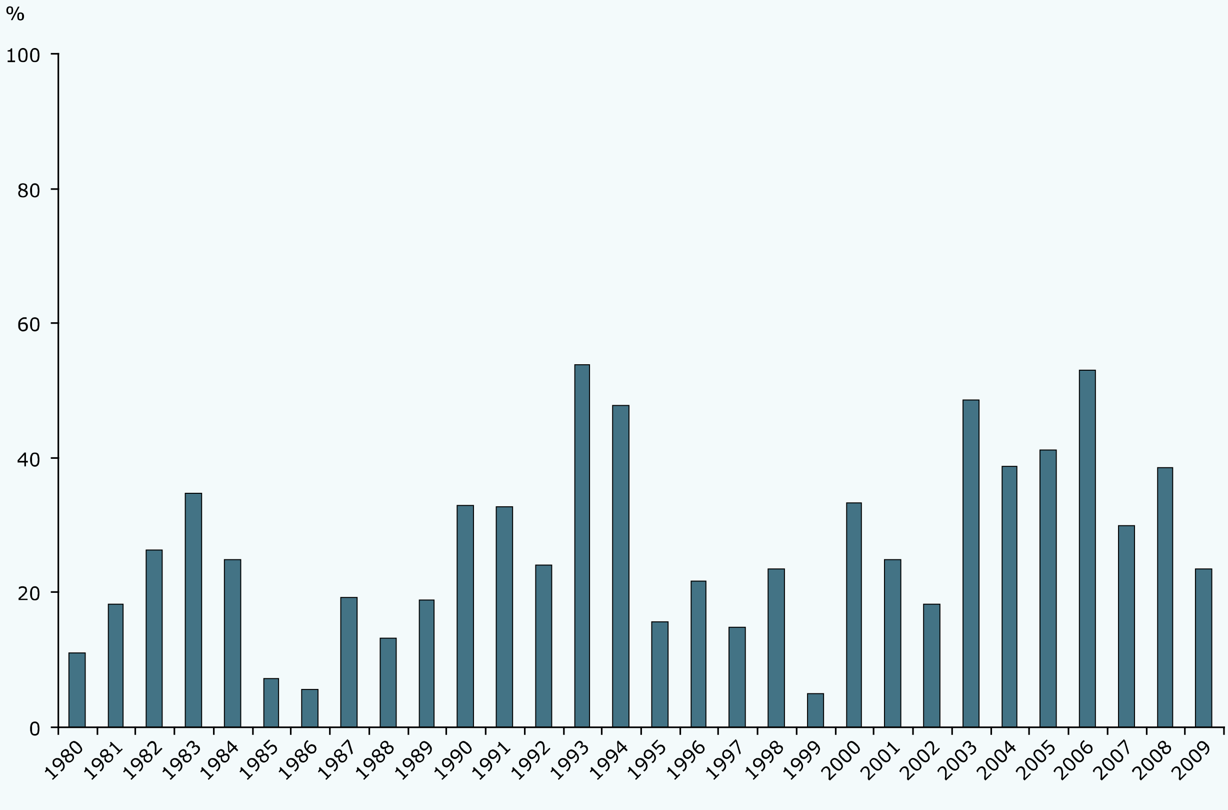
The relative importance of lightning-caused fires, however, increased in recent decades (Figure 5.6). In the period from May to October, the proportion of lightning fires changed from an average of 20.3 % in the 1980s to 29.1 % in the 1990s, and 41.1 % in the 2000s (Figure 5.6), highlighting the difficulty of preventing the ignition of fires of natural origin. In addition, in drought-summer years such as 1983–1984, 1990 and 2003, lightning fires are more likely to turn from underground into surface or crown fires, causing a significant increase in the burned area (Figure 5.5).

From a management point of view, lightning-induced fires occur mostly in remote locations and burn underground (Conedera *et al.*, 2006), making detection and suppression activities more difficult. When intense lightning activity occurs following a drought, lightning-ignited fires aggregate in both time and space, which may put a strain on the initial attack by the fire brigades and thus lead to longer and more difficult fire fighting campaigns (Podur *et al.*, 2003; Wotton and Martell, 2005).

As climate change may lead to an increased frequency of hot and dry summers (Schär *et al.*, 2004), these results suggest that, in the future, lightning-induced fires may assume a significant ecological role and have a higher economic impact in the Alps, as suggested by Schumacher (2004).

Source: Marco Conedera and Gianni Boris Pezzatti (Swiss Federal Research Institute, Switzerland).

Figure 5.6 Yearly relative frequency of lightning-induced fires with respect to total number of fires in the summer period (June to September) in the Swiss Alps



Source: Swissfire database.

accumulated snow water equivalent may decrease by up to 78 %, and the season with snow cover may be reduced by up to 70 % at 1 500 m (Lopez-Moreno *et al.*, 2009). However, the magnitude of the impacts decreases rapidly with increasing altitude, with snowpack characteristics projected to remain largely similar in the highest sectors (Lopez-Moreno *et al.*, 2009). Stewart (2009) summarises work examining observed and projected changes in snow cover and snowmelt-derived streamflow for the European Alps and European mid-elevation mountain ranges.

The lower elevation of permafrost is likely to rise by several hundred metres. Rising temperatures and melting permafrost will destabilise mountain walls and increase the frequency of rock falls, threatening mountain valleys (Gruber *et al.*, 2004; Harris *et al.*, 2009; Keiler *et al.*, 2010). In northern Europe, lowland permafrost will eventually disappear (Haeberli and Burns, 2002). Changes in snowpack and glacial extent (Box 6.2) may also alter the likelihood of snow and ice avalanches, depending on the complex interactions of surface geometry, precipitation and temperature (Martin *et al.*, 2001; Haeberli and Burns, 2002).

5.3 Research needs

5.3.1 Instrumental data and monitoring networks

Although some climatic information for mountain regions can be obtained from radiosonde measurements, significant differences between radiosonde and mountain surface data have been observed (e.g. Seidel and Free, 2003). This emphasises the need for paired station monitoring networks at lowland and mountain locations (Barry, 2008) and, while there have been encouraging developments in expanding the instrumental data provision for the Alps, an expanded monitoring network across Europe's mountain regions is needed (Schär and Frei, 2005; Bjornsen Gurung *et al.*, 2009; Smiatek *et al.*, 2009). This scarcity of instrumental data in many mountainous regions also hampers the performance assessment of outputs from this and subsequent generations of RCMs; measures to address these data gaps could include the incorporation of more mountain areas in the integrated monitoring and observation system mooted for Europe (EEA, 2008).

5.3.2 Sources of uncertainty in climate change projections

Projections of climate change are subject to a high degree of uncertainty (Jones, 2000), as a consequence of both aleatory ('unknowable' knowledge) and epistemic ('incomplete' knowledge) uncertainty (Hulme and Carter, 1999; Oberkampf *et al.*, 2002; Foley, 2010); at least some of which relates to knowledge gaps in the understanding of the climate system (Albritton *et al.*, 2001; EEA, 2008). Adding to these, the accuracy of GCM performance in areas of complex terrain and the subsequent cascade through RCMs introduces a further tier of uncertainty.

5.3.3 Climate modelling challenges

Even with the evolution of ever more complex and sophisticated GCMs, issues remain concerning their robustness (Chase *et al.*, 2004), and their reproduction of the detail of regional climates remains limited (Zorita and von Storch, 1999; Gonzalez-Rouco *et al.*, 2000; Jones and Reid, 2001; Bonsal and Prowse, 2006; Connolley and Bracegirdle, 2007; Perkins and Pitman, 2009). For regions of heterogeneous terrain, such as mountains, RCMs provide more credible information on changes in climates than GCMs. However, since each RCM is constrained by the boundary conditions of the GCM used to drive it, uncertainties in GCM predictions are effectively cascaded (Carter and Hulme, 1999; Frei *et al.*, 2003; Jenkins and Lowe, 2003; Saelthun and Barkved, 2003; Déqué *et al.*, 2007; Jacob *et al.*, 2007).

An additional limitation of using RCM outputs in mountain regions relates to the fact that the true roughness of mountain terrain is represented by a smoothed surface in models. Consequently, the elevation of specific sites is poorly represented and the observed climate is not accurately reproduced (Coll *et al.*, 2005; Engen-Skaugen, 2007; Beldring *et al.*, 2008). Overall therefore, local controls on climate in mountain regions are not adequately captured by current GCMs and RCMs, and the best resolution of 50 x 50 km remains inadequate for impact assessment (EEA, 2008), particularly in mountainous areas. Finally, for both GCMs and RCMs, even if models improve in performance in simulating current climate, this may not be a reliable indicator of their performance for predicting future climate.