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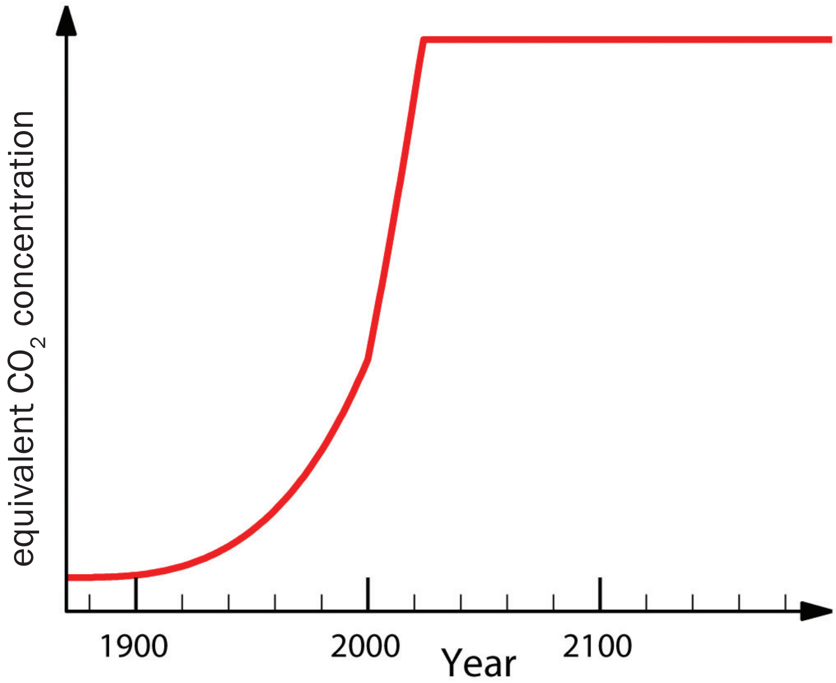

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Figure caption:
Idealized greenhouse gas concentration with time. After the increase from the pre-industrial period to the present-day level, a 1% per year further increase is assumed, until the amount, measured as CO_2-equivalents, reaches 450 ppmv, taking into account aerosol effects. In this case, this occurs at year 2023; in reality, this level might be reached either somewhat sooner, or later.

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Abstract/Sammandrag

The impact of a CO_2 stabilisation on the Swedish climate is investigated with the regional climate model RCA3 driven by boundary conditions obtained from a global coupled climate system model (CCSM3). The global model has been forced with observed greenhouse gas concentrations from pre-industrial conditions until today's, and with an idealised further increase until the stabilisation level is reached. After stabilisation the model integration continues for another 150+ years in order to follow the delayed response of the climate system over a period of time.
Results from the global and regional climate model are compared against observations and ECMWF re-analysis for 1961-1990. For this period, the global model is found to be too cold over Europe and with a zonal flow from the North Atlantic towards Europe that is too strong. The climate of the driving global model controls the climate of the regional model and the same deviations from one are thus inherited by the other. We therefore analyse the relative climate changes differences, or ratios, of climate variables between future's and today's climate.
Compared to pre-industrial conditions, the global mean temperature changes by about 1.5°C as a result of the stabilisation at 450 ppmv equivalent CO_2. Averaged over Europe, the temperature change is slightly larger, and it is even larger for Sweden and Northern Europe. Annual mean precipitation for Europe is unaffected, but Sweden receives more precipitation under higher CO_2 levels. The inter-annual and decadal variability of annual mean temperature and precipitation does not change with any significant degree.
The changes in temperature and precipitation are not evenly distributed with the season: the largest warming and increased precipitation in Northern Europe occurs during winter months while the summer climate remains more or less unchanged. The opposite is true for the Mediterranean region where the precipitation decreases mostly during summer. This also implies higher summer temperatures, but changes in winter are smaller. No substantial change in the wind climate over Europe is found.

Key words/sök-, nyckelord

Global climate modelling, CCSM3, regional climate modelling, RCA3, stabilisation scenario, climate change

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Contents

1 Introduction

Climate change, due to an enhanced greenhouse effect driven by anthropogenic emissions of carbon dioxide and other greenhouse gases, is one of the main components of global change, and one of the major obstacles for sustainable development. The contemporary climate change is often called global warming. The analogy is fitting since one of the global change primary aspects is a rise of temperature in the global mean, as well as regionally. Actually, many aspects of the climate such as sea level, glaciers, snow cover and permafrost, precipitation amounts and patterns, and soil moisture, also are affected. On the other hand, changes in other climatic features, such as severe storms, are, to a large part, less obvious. In addition to changes in the mean, the variability and extremes of different kind are also affected.

During the 20th Century, the global mean surface temperature rose by about 0.6°C. A significant part of this rise, and primarily during the second half of the past century, has been attributed to the increased concentration of greenhouse gases in the atmosphere. This attribution has not only been in the form of a temperature increase, but also in its global patterns and temporal evolution. These bear the signature that is expected from an enhanced greenhouse effect, modulated by the internal variability of the climate system, and, to a lesser degree, natural climatic-forcing factors. This warming is also apparent in cryospheric changes: melting of most glaciers and the reducing Arctic sea ice cover, biospheric changes: lengthening of the vegetation period in the Northern Hemisphere, and changes in the distribution and behaviour of some species.

Since the primary present-day climatic-forcing factor is the emission of greenhouse gases into the atmosphere, future climate will also depend on the future emissions. Today, the global emissions are still on the rise, despite the steps taken to reduce them. The main contributor to global emissions is the use of fossil fuels to produce energy for heating and cooling, for transportation, and in many other industries. Fossil fuels presently account for about 80% of the primary global energy demand and it is estimated to increase in the near future (IEA, 2004), Therefore, replacing fossil fuel with a cleaner and more efficient one will be challenging, to say the least (Bolin and Kheshgi, 2001). This needs to be addressed globally, and as such requires international co-operation. The main forum for intergovernmental efforts is the United Nations Framework Convention on Climate Change (UNFCCC, see www.unfccc.int) which was adopted in 1992, and has been in force since 1994. It recognises the fact that the climate affects everyone and, as well, the risks posed by anthropogenic emissions of greenhouse gases and aerosols. The UNFCCC allows countries to gather and share information, launch strategies to reduce the emissions and to promote adaptation, and to support developing nations' efforts.

The objective of the UNFCCC is defined in its Article 2:

> The ultimate objective of this Convention and any related legal instruments that the Conference of the Parties may adopt is to achieve, in accordance with the relevant provisions of the Convention, stabilisation of greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference with the climate system. Such a level should be achieved within a time-frame sufficient to allow ecosystems to adapt naturally to climate change, to ensure that

> food production is not threatened and to enable economic development to proceed in a sustainable manner.

In particular, Article 2 refers to a stabilisation of greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic influence. This level is not defined in the UNFCCC, nor does it lend itself to an easy definition. Indeed, what is regarded as dangerous is, at least to some degree, a value judgement (IPCC, 2004a). Scientific consideration of climate change outcomes and climate impacts can nevertheless provide policy-relevant information on the matter.

A global mean-temperature rise by 2°C, compared to the preindustrial era, has been quoted in climate policy contexts as a threshold of dangerous anthropogenic interference. This seems to trace back to the assessment of the German Advisory Council on Global Change (WBGU, 1995, 1997, 2004):

> The [intolerable] climate window is defined by two limits: Both a temperature change rate of more than 0.2°C per decade and a mean global temperature change of more than 2°C relative to pre-industrial levels are deemed unacceptable. (WBGU, 2004, p. 109)

Their definition of the intolerable climate window took into account both the warmest climate periods during the last several hundred thousand years and added a half degree centigrade to account for an adaptive capacity. However, the WBGU also noted that this dangerous interference could probably not be adequately described by a global mean measure. They also noted the risks that come with abrupt climate changes. It was also noted that these risks and their impact would become more probable with increased rate of emission and thus global warming as we approach the set limits. More recently, however, considerations have turned to regionally-varied measures and giving more room for weighing the risk for abrupt and other irreversible changes.

The 2°C upper limit of global mean warming has influenced the EU climate policy as early as 1996 (EU, 1996). It has also been reaffirmed more recently by the Council of the European Union in December 2004 and included in the European Council March 2005 Presidency Conclusions (7619/1/05) as:

> The European Council acknowledges that climate change is likely to have major negative global environmental, economic and social implications. It confirms that, with a view to achieving the ultimate objective of the UN Framework Convention on Climate Change, the global annual mean surface temperature increase should not exceed 2°C above pre-industrial levels.

Importantly, the December 2004 Council Meeting noted that the connection between some specific global mean temperature rise and a specific level of greenhouse gases is not precise:

...scientific uncertainties exist in translating a temperature increase of 2°C into greenhouse gas concentrations and emission paths; however, RECOGNISES that recent scientific research and work under the IPCC indicates that it is unlikely that stabilisation of greenhouse gas concentrations above 550 ppmv CO_2 equivalent would be consistent with meeting the 2°C long-term objective and that in order to have a reasonable chance to limit global warming to no more than 2°C, stabilisation of concentrations well below 550 ppmv CO_2 equivalent may be needed...

This is the same as the well-recognised scientific uncertainty on climate sensitivity that has been discussed in the First, Second and Third IPCC Assessment Reports (IPCC, 1990, 1996, 2001b), and more recently also by IPCC (2004b). Due to the complexity of the climate system, and in particular such feedback mechanisms as changes in cloudiness or reorganisation of ocean currents, it is not possible to arrive at exactly what the equivalent temperature increase would be for some specific atmospheric concentration of greenhouse gases. Considering the evidence of palaeoclimate periods together with historical climate events, observations made with modern meteorological instruments and information from climate models nevertheless points to a significant and increasing global warming for increasing amounts of greenhouse gases in the atmosphere. IPCC Assessment Reports so far state that a doubling of carbon dioxide in the atmosphere (which should be understood in terms of an equivalent carbon dioxide amount, i.e., accumulating the effect of all anthropogenic greenhouse gases) would raise the global mean temperature by an amount somewhere between 1.5 and 4.5°C.

To perhaps complicate the matters even further, it can be pointed out that there is a similar inexact relation between the actual emissions and the resulting atmospheric concentrations. Carbon dioxide, as well as some of the other anthropogenic greenhouse gases, participates in processes that circulate and store carbon in the atmosphere, the surface layer of the global ocean, and the biosphere. On much longer time scales, the deep ocean and geological processes also come into play. Presently, about half of the global emissions of carbon dioxide due to the use of fossil fuels and land use are captured in the ocean and the biosphere; the remainder is accumulated in the atmosphere. The oceanic and biospheric uptake is, however, affected by the climate and thus also by climate change.

This leads into uncertainties in defining emission reduction goals in climate policy as the effect goal needs to be propagated through, first, the impact on atmospheric concentrations and, thereafter, to the climate change, and possibly further into various climate impacts on the global mean and the regional distribution of the changes (see, e.g., Schneider and Lane, 2006).

The scientific uncertainties can, to some degree, be seen in policy-making. The 2°C target, albeit not quantitatively investigated for different impacts, different regions, and in terms of risks for abrupt and irreversible changes, provides a tangible goal. This goal might be attainable and even lead to the prevention of a range of, although not all, adverse effects. Nevertheless, accounting for the change in emissions to atmospheric concentrations to climate change, each transition lends itself to an incomplete range of values. The stabilisation of greenhouse gas concentrations in the atmosphere would need to be achieved at a rather low level to limit the risk of overshooting the temperature goal.

Climate policy goals are also formulated in terms of emission reductions and atmospheric concentrations at stabilisation. The formulation of emission pathways leading to a stabilisation can be done with various levels of complexity (e.g. Wigley et al., 1996; Azar and Rodhe, 1997; Cubasch et al., 2001; Swart et al., 2002; IPCC, 2001a; Elzen and Meinshausen, 2005). So far, the projection of such emission pathways into climate changes, by means of global climate models, and even more so regional climate models and climate impact models, has not been done to any great extent. Rather, inferences have been made on global mean temperature and sea level using simple climate models tuned to the sensitivity exhibited by the more complex atmospheric-oceanic global climate models (Cubasch et al., 2001, p. 557–559). Very few climate change projections under stabilisation scenarios are using the latter kind of advance global models (Mitchell et al., 2000; Cubasch et al., 2001). This is in contrast to the extensive application of such global models for transient (time-variant) climate change projections based on either idealised or rather explicit emission scenarios for the 21st Century. The investigation of the appropriateness of a global stabilisation target would, however, benefit from global climate model projections under stabilisation, especially if followed by regionalisation and analysis in terms of the regional human and environmental impacts.

It is against this background that this study should be comprehended. It represents a new kind of climate modelling exercise that attempts to provide a regional climate projection example based on advance regional and global climate modelling. The stabilisation pathway is described directly in terms of atmospheric carbon dioxide equivalent concentration that increases until the early 21st Century and then stays constant at the 450 ppmv level. This is an idealised approach and not a realistic assumption. Past greenhouse gas emissions have already increased their atmospheric concentration, evaluated as equivalent carbon dioxide, to about 440 ppmv. The emissions and concentration increases are also still on the rise. A 450 ppmv level does not have to be a totally unrealistic one, although it can reasonably only be reached with rather major emission reductions and only after a period of temporarily (i.e., possibly over several decades) higher atmospheric concentrations. Such overshooting, especially if it will be rapid, large, or last a long time, might give rise to significant impacts by pushing some systems or aspects into irreversible changes that might otherwise been avoided. The long term consequences, however, should more readily follow the amount and nature of changes corresponding to the ultimate stabilisation level.

In addition to the idealised treatment of stabilisation, a major caveat of the study is that only one global climate model and one regional climate model have been used, despite the fact that the models are state-of-the-art. The climate sensitivity of the global model in question is, on the low side, compared to other advance global climate models. Thus, the results are best considered as an illustrative example of the long term memory of the climate system of emissions and climate policy decisions in our own time (cf. Friedlingstein and Solomon, 2005). For more robust conclusions and recommendations, more extensive model projections should be pursued. These should perhaps be based on more realistic emission pathways and the results translated into relevant impacts that consider both the time-evolution and final climate state under stabilisation.

2 Global model CCSM3

2.1 Model description

The Community Climate System Model (CCSM3) is a state-of-the-art coupled global circulation model that has been developed under the auspices of the National Center of Atmospheric Research (NCAR) Boulder, USA. The modules for the atmosphere, land surface, sea ice, and ocean components are linked through a coupler that controls the exchange of energy and water between the components. The current version 3 of CCSM has been released in June 2004 and since then it has been widely used for climate studies. CCSM3 is extensively documented online (http://www.ccsm.ucar.edu) and in the literature (e.g. Collins et al., 2006, and references therein). Numerous multi-century controled runs with constant external forcing have been performed to guarantee that there is no significant drift in the model's climate. Variations in the external forcing have been applied to study the climate sensitivity for the forthcoming IPCC AR4.

Long climate simulations that are done at NCAR, or other laboratories, usually save data as monthly means. While monthly data are certainly sufficient to study many aspects of the global climate and its variability, they are not sufficient to drive a regional model in a downscaling experiment. For this purpose the model state needs to be saved several times every day. The archived results from NCAR were not suitable for the intended regionalisation and required a new CCSM3 climate simulation. Thus, the CCSM3 was run again and saved the necessary data for the regionalisation study report. Limitations in available computer time and memory[1] implied a lower horizontal resolution compared to NCAR's IPCC AR4 simulations. For this project the T31 setup was used (Fig. 1), that is, 3.75° horizontal resolution for the atmosphere and land component, and approximately 3.6° for the ocean and sea-ice model[2]. The atmosphere has 26 and the ocean 25 levels in the vertical. The coarse T31 resolution has implications on the quality of the simulation, e.g., the simulated climate will differ from the real climate as the distribution of land and sea, or the height of mountain ranges, are poorly represented. On the other hand, the large scale circulation should be reasonably well simulated which is a necessary prerequisite for successfully driving a regional climate model.

2.2 Climate sensitivity

Climate sensitivity typically refers to the response of the global mean temperature to a doubling of CO_2. The Coupled Model Intercomparison Project (CMIP2) compares the climate sensitivity of different coupled climate models when CO_2 increases by 1% annually (Covey et al., 2003). The average of 19 models at the time of CO_2 doubling (after 70 years) is found to be 1.8°C (IPCC, 2001b). CCSM3 is not included in this average but was later tested in the same type

[1]The new and powerful Tornado cluster at Nationellt Super Computer center (NSC) was not available at the time when the project started.

[2]The ocean and sea-ice models use a different grid with a displaced North Pole and stretched coordinates in the latitudinal direction with higher resolution at the equator.

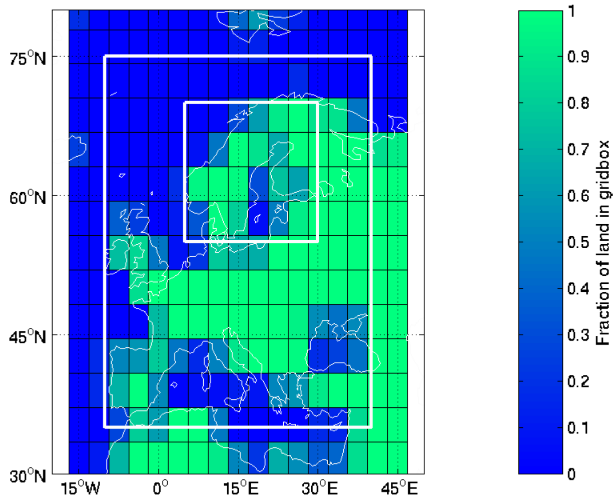

***Figure 1**: Distribution of land and sea in CCSM3 at T31 resolution. Averages of temperature and precipitation have been computed for Europe (EUR) and Scandinavia (SCAN) that are shown in white.*

of experiment. The transient climate response (TCR) of the fully coupled CCSM3 with T31 resolution is 1.43°C which is lower than the average CMIP2 response but still in the range of other climate models. There is a slight dependency on resolution in CCSM3 and the higher resolution T85 version shows a slightly larger TCR of 1.50°C (Kiehl et al., 2006).

2.3 External forcing

CCSM3 is a coupled global climate model. In climate change projections, it only requires the variation of the greenhouse gas concentrations and aerosols as the prescribed forcing. The relevant gases that are present in CCSM3 are the well-mixed CO_2, N_2O, CH_4, CFC-11 and -12. The concentration of O_3 varies in space and time. The sulphur cycle is explicitly represented in the model and requires the specification of sulphur emissions. For historical simulations between 1870 and 2000, variations in the solar constant and emissions from volcanoes are taken into account. As these forcing factors are not known in future simulations, they are kept constant after the year 2000.

Our simulation starts 1870 from an initial state that has been obtained from a 500 year long simulation with constant climate forcing that was done at NCAR. Coupled models require long spin-up times to make sure that all model components have come into balance and there is no significant long-term trend in the surface temperature. For our time-integrating simulation, we then use the reconstructed and observed climate forcing until 2000. The simulation was initially planned to continue with forcing according to the IPCC SRES B1 scenario (Nakićenović et al., 2000) until the set stabilisation level was reached. The forcing files for this scenario obtained from NCAR, however, were found to be corrupted and issues related to the consistency of the forcing data could not be resolved. Therefore, an alternative approach was

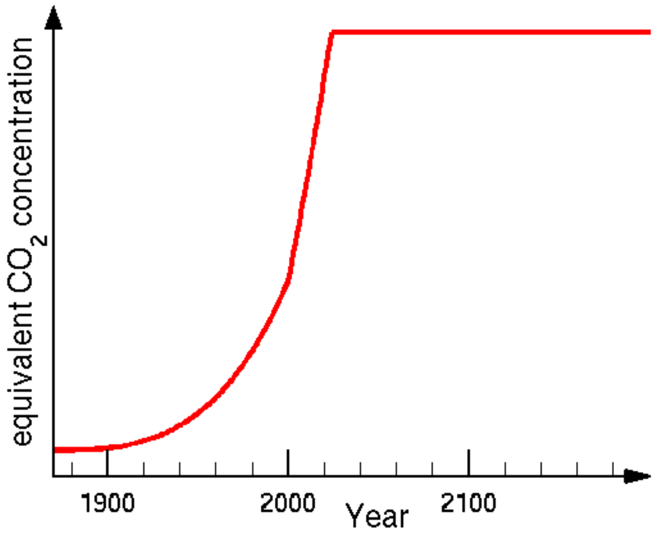

Figure 2: *Idealized greenhouse gas concentration with time. After the increase from the pre-industrial period to the present-day level, a 1% per year further increase is assumed until the amount, measured as CO_2-equivalents, reaches 450 ppmv. This assumption also takes into account aerosol effects. In this case, this occurs at year 2023; in reality, this level might be reached sooner or later.*

chosen to keep the concentration of all greenhouse gases and aerosols fixed at the 1950 levels. CO_2 was, however, allowed to grow, including even the effects of the other gases and aerosols. This method is known as the equivalent CO_2 approach as it treats the climate as if it were only forced by the variation of CO_2. For this study we let the CO_2 concentration increase from 1950 to 2000 following closely an inferred increase of the equivalent CO_2 concentration to present conditions. The CO_2 concentration increases thereafter by 1% annually until it reaches 450 ppm (equivalent) in year 2023 after which it is kept constant while the simulation continues until year 2200 (Fig. 2).

2.4 Reproducing today's climate

The results from the climate simulation for the 1961–1990 period are compared against data from ERA-40[3] (with 1.125 degrees horizontal resolution) and a similar climate simulation (20c3m) that was done at NCAR but with T85 resolution (1.41 degrees). We compare time averages for the Northern Hemisphere (hereafter referred to as NH), Europe (EUR) and Scandinavia (SCAN). The distribution of land and water as well as the size and location of EUR and SCAN is shown in Figure 1. It is obvious that the resolution of CCSM3 at T31 is too coarse to represent well the true distribution of land and sea. Therefore we include the comparison in

[3]ERA-40 is the re-analysis dataset of the European Centre for Medium Range Weather Forecasts that covers 1957-2002. It combines a wealth of meteorological observations with the first guess from a weather forecast model in an assimilation process to produce the best "snapshot" of the state of the atmosphere every 6 hours. For more information see Uppala et al. (2005).

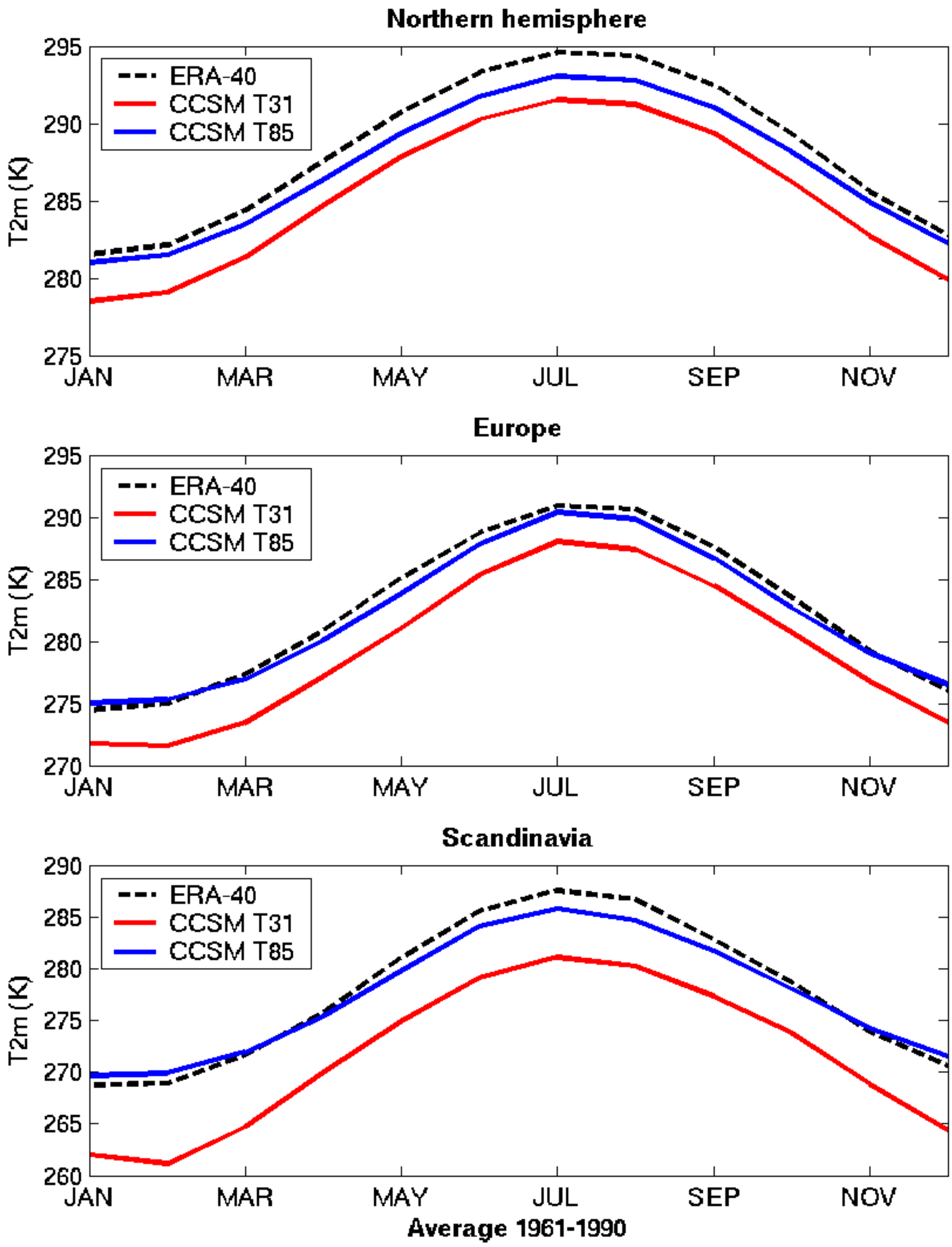

Figure 3*: Monthly averaged 2m temperature for the 1961–1990 period for CCSM3 with two different resolutions and ERA-40.*

the SCAN region mainly to illustrate how well or badly a T31 version is able to simulate the climate in a limited area. For larger areas, e.g., NH, we should expect a closer match between simulated and true climate.

The 2m temperature is found to be rather low in the T31 run (Fig. 3). For the NH the temperature is 3°C lower than in ERA-40. The difference increases for EUR (-3.2°C) and even more for SCAN (-6.2°C). The same model with T85 resolution shows a much better agreement with ERA-40. Despite the large bias of T31, we still find that the seasonal cycle is well reproduced, and, for example, the temperature difference between summer and winter is very close to that found in ERA-40.

Precipitation is found to be too low in CCSM3 compared to ERA-40 for the NH region, but generally too high for EUR and SCAN (Fig. 4). Spring and summer precipitations are too low in CCSM3 compared to ERA-40 in the NH average. For EUR, CCSM3 appears to be too wet except for summer. Unlike temperature where T85 and T31 are clearly distinct, precipitation is similar for the two resolutions in the NH and EUR regions. Small differences appear in SCAN but can easily be explained with the higher resolution of the topography in T85 that gives a better representation of small scale variability of the precipitation over mountainous terrain.

An analysis of the mean sea level pressure (MSLP) over the Atlantic Ocean illustrates the problems of CCSM3 to capture the observed pressure distribution during wintertime (Fig. 5). The dipole structure with the low pressure over Iceland and the high pressure over the Azores is present in the model. However, the pressure is too low in the centre of the Icelandic Low and too high in the Azores High, resulting in a too strong meridional pressure gradient. The problem with the pressure pattern over the Atlantic Ocean appears at both T31 and T85 resolution of CCSM, and it is actually more accentuated in T85. The consequence of too strong a pressure gradient is a zonal flow towards Europe that is too strong, implying a too large transport of mild and moist air from the Atlantic Ocean towards Central and Southern Europe. The analysis of long temperature records has revealed that the North Atlantic Oscillation (NAO) is the most important modulator of Europe's climate. NAO is usually expressed as an index based on the normalised wintertime pressure difference between Lisbon and Reykjavik. CCSM3 tends towards a more positive NAO which will result in mild and wet winters over Central Europe. Since the CCSM climate over Europe is still too cold, as shown before, we conclude that without this too-strong inflow of warm and moist air, the winters in Europe would be even colder, and the difference to ERA-40 even larger.

Apart from the differences in the meridional pressure gradient over the Atlantic Ocean, we also find a mismatch in the shape of the low pressure cell over Iceland between ERA-40 and CCSM3 with T31 resolution. The position of the pressure minimum in CCSM3 is shifted southward. Furthermore, the characteristic trough that stretches from the minimum towards the Barents Sea is not well reproduced in the model. This trough, also known as storm tracks, is the footprint of low pressure systems that sweep from the Atlantic Ocean towards Scandinavia. The weak storm track's signature of CCSM3 indicates that the low pressure systems don't occur frequently enough, or are too weak. Both effects result in a reduced transport of mild air towards Scandinavia and subsequently a climate over Northern Europe that is too cold. The problem with the location and strength of storm tracks is related to the resolution: the MSLP pattern of CCSM3 with T85 resolution agrees much better with ERA-40 (Fig. 5).

Why is there such a large difference, particularly in T2m, between T85 and T31 resolution? Some of the difference can be explained by differences in resolution, but there is also another aspect: initial conditions. The CCSM simulations, both with T31 and T85 resolution, start in

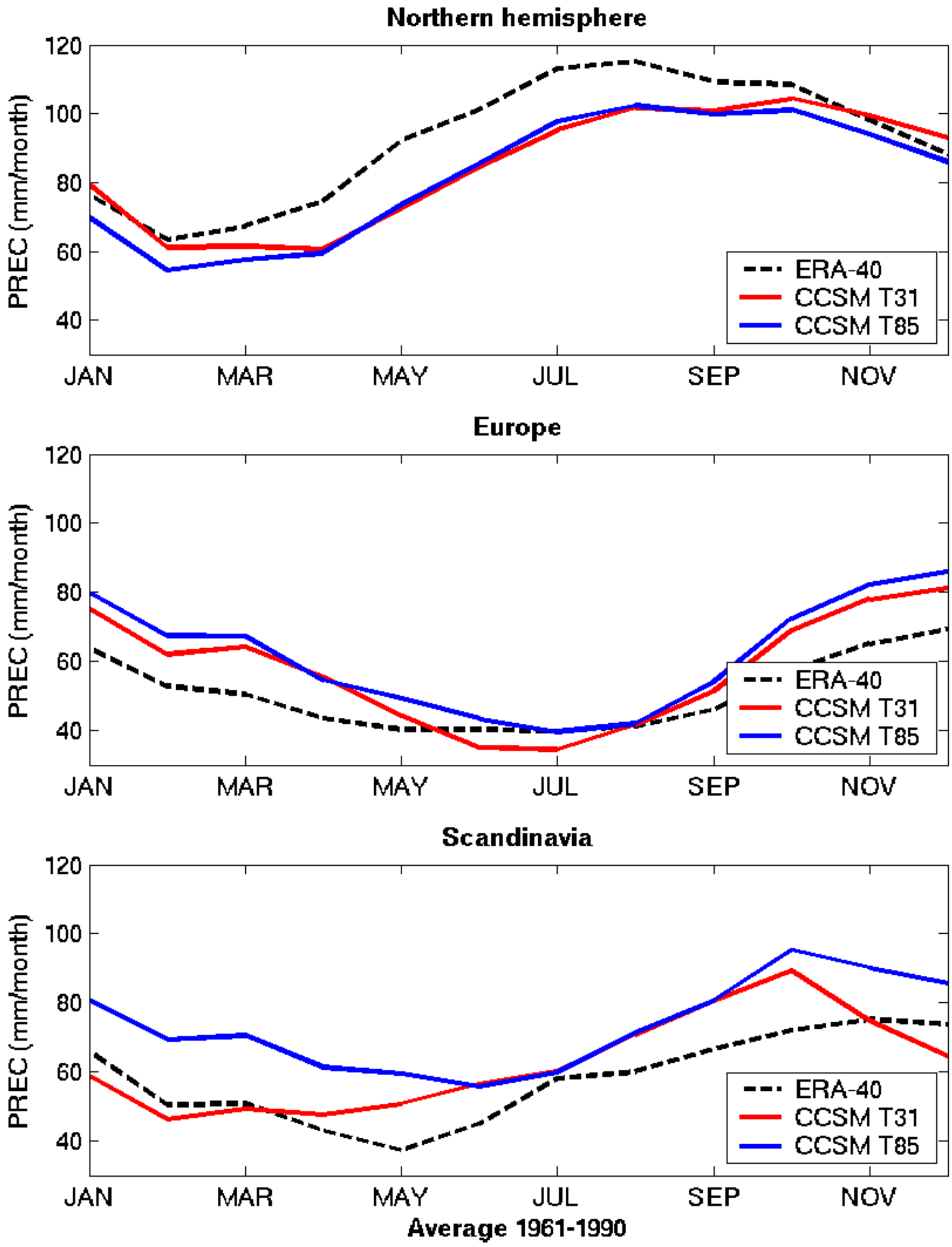

***Figure 4**: As Fig. 3 but for monthly accumulated precipitation.*

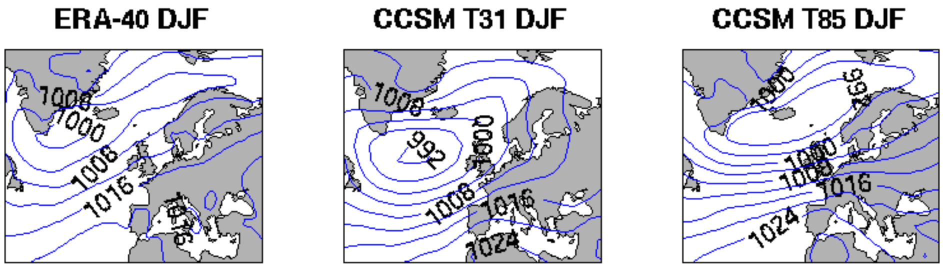

***Figure 5**: Winter (DJF) mean sea level pressure for the period 1961 to 1990 from ERA-40 and with two resolutions of CCSM3. Contour spacing is 4 hPa in all figures.*

1870, but they don't start from the "true" climate state. The initial state is produced by running the model for some hundred years with constant forcing that is representative for the year 1870. This long spin-up is required to bring the different components into balance. However, the resulting climatic state is not necessarily identical to the observed 1870 climate; it is just one realisation. Due to the natural climatic variability, the same forcing can yield different climatic states, and differences between the different realisations are more pronounced regionally or locally. A control experiment with perpetual 1990 forcing has been performed with T85 and T31 resolution, and in this case there was a considerably smaller bias between the two simulations. Thus, we conclude that the differences between our T31 run and the T85 runs from NCAR are mainly due to different initial states for the two setups.

To overcome the difficulties with the substantial bias between our T31 run, NCAR's T85 runs, and ERA-40, we will focus on relative differences between the future and the present climates. The downscaled results from any future time period are compared against the corresponding results from the period 1961–1990 and not against the observed climate. By doing so, we can largely eliminate the bias of the global model and clearly identify the impact of climate change from the imposed increase of the CO_2 concentration.

2.5 Global temperature evolution in a forced 1870–2200 simulation

The global mean temperature in our CCSM3 T31 simulation increases with time as the concentration of greenhouse gases increases (Fig. 6). The temperature increases by about 0.5°C from the early industrial period to present day climate, and then continues to rise sharply as long as

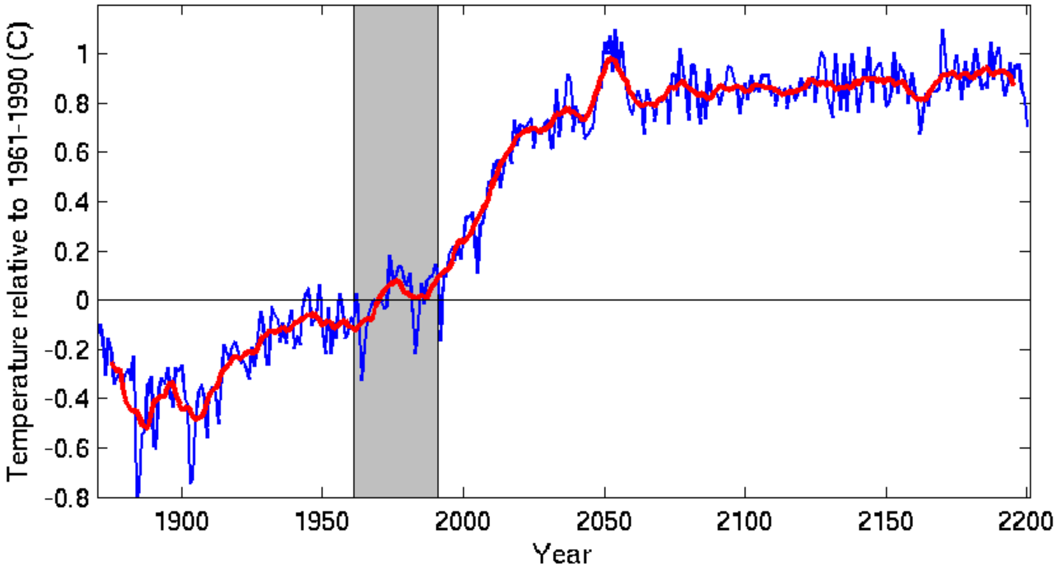

Figure 6: *Global mean annual temperature from the CCSM3 T31 simulation with external forcing as described in Sec. 2.3, with the 10-year running mean in red. The temperature is plotted relative to the average for the period 1961–1990 that is marked in grey.*

the forcing increases. Stabilisation is reached in 2023 after which greenhouse gas levels are kept constant. The temperature continues to rise, although at a slower rate. The temperature difference between 2020 and 2200 is about 0.1°C. There is no indication of an asymptotic levelling off in the temperature response even after 180 years of stabilisation. The temperature difference at the end of the simulation is about 0.9°C relative to present day, or about 1.5°C relative to pre-industrial conditions.

It is worth noting that the global temperature can undergo strong decadal and annual variations even under fixed CO_2 conditions. A relatively warm period occurs around year 2050, and a relatively cold period about 100 years later. However, this internal variability is considerably smaller than the response to the external forcing that is imposed in the beginning of the 21st century. The figure clearly shows that a new climate regime is entered when the CO_2 forcing stabilises at 450 ppm.

3 Dynamical downscaling of CCSM3 with RCA3

The Rossby Centre Atmospheric model is a comprehensive regional climate model that includes a description of the atmosphere and its interaction with the land surface. An overview of the most recent version, RCA3, is presented in Kjellström et al. (2005). Earlier versions of the RCA model are described by Rummukainen et al. (1998, 2001), Räisänen et al. (2003, 2004), and Jones et al. (2004).

The RCA3 model was used to dynamically down-scale the results from CCSM3 over Europe for the period 1961–2200. For this experiment we have set up RCA3 on a rotated latitude-longitude grid with a resolution of 0.44 degrees, corresponding to 49 km. The domain covers Europe with 102×111 grid points and 24 unequally spaced vertical levels. Thanks to the semi-Lagrangian advection scheme, RCA3 can be run with a comparatively long timestep of 30 minutes. Along the lateral boundaries, RCA3 is smoothly relaxed towards the CCSM3 forcing in an eight-point wide boundary zone. Sea surface temperature and sea-ice cover are also read in from CCSM3. The lateral and lower boundary forcing from CCSM3 is updated every 6 hours. In terms of greenhouse gas forcing, we have imposed an increase with time in equivalent CO_2, corresponding to that used for producing the CCSM3 data set.

4 RCA3 simulated climate for the 1961–1990 period

The annual and seasonal averages of mean sea level pressure, temperature, precipitation and winds from RCA3 are compared against the corresponding quantities from ERA-40 and CRU (Climate Research Unit, University of East Anglia, Mitchell and Jones, 2005) for the standard period 1961–1990. Seasonal averages are computed as 30 year means of June, July, August (JJA) or December, January, February (DJF) for summer and winter, respectively.

4.1 Mean sea level pressure

Not unexpected, the pressure distribution of the regionalisation follows closely the driving model (Fig. 7). In winter, the pressure in RCA3 is too low over the North Atlantic and too high over the Mediterranean Sea and Northern Africa, when compared to ERA-40. This stronger-than-observed meridional gradient gives rise to a too strong zonal flow and thus an increased transport of warm moist air from the Atlantic Ocean to Western Europe. Consequently, a warmer and wetter climate develops over Central and Southern Europe, and a colder climate over Northern Europe. In summer, there is less difference in the pressure pattern between RCA3 and ERA-40 than during winter. The low pressure over the North Atlantic is reasonably well captured by the model but the high pressure over the Mediterranean remains too high. The resulting flow is again too zonal, but the difference is less pronounced than in winter.

4.2 Temperature

Despite the enhanced maritime influence from the stronger zonal flow, the temperature in RCA3 is considerably lower than in ERA-40 or CRU, as illustrated in Fig. 8. The annual mean temperature is too low over almost all of Europe and the summer temperature is more than 2°C too low for most areas. In Scandinavia the temperature difference is largest in winter when Northern Scandinavia is 6–10°C too cold. On the other hand, Central and Southeastern Europe are found to be about 2°C too warm during winter. The too-low temperatures in RCA3 over much of Europe are partly inherited from the driving model CCSM3. The temperature at the

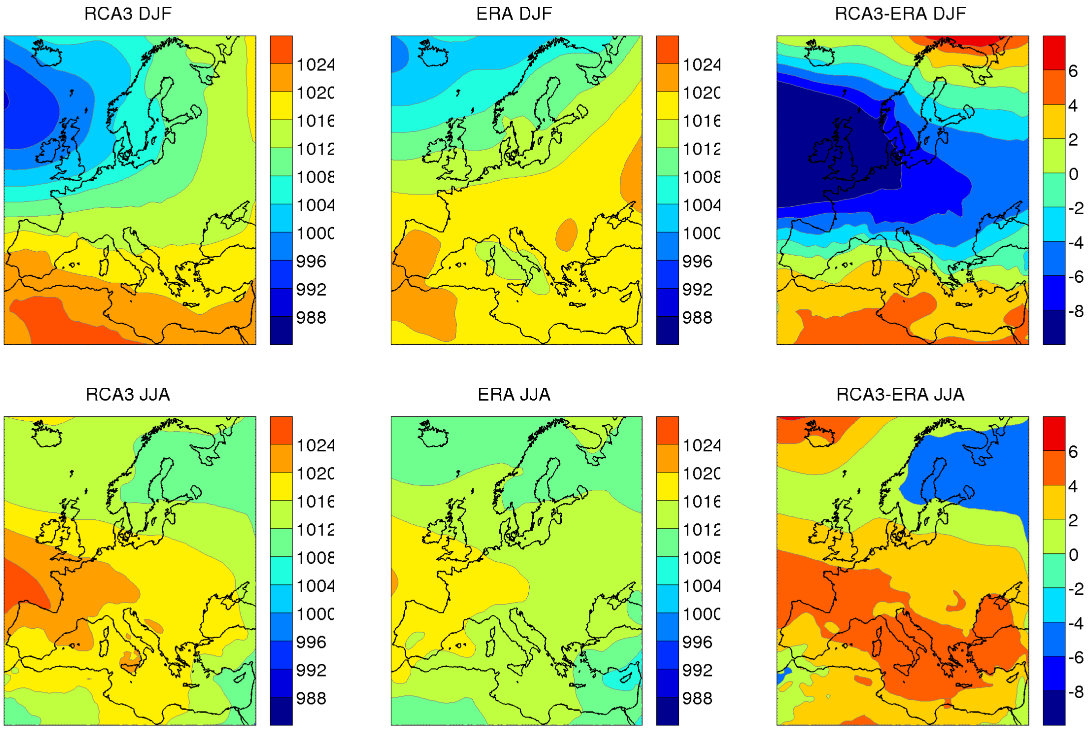

***Figure 7**: Winter and summer mean sea level pressure [hPa] from RCA3 (left), from ERA-40 (middle) and bias RCA3-ERA40 (right)*

lateral boundaries is too low, which keeps the temperatures low even in the regional climate model. Furthermore, SST and sea-ice distribution are taken directly from CCSM3 and the temperatures over the North Atlantic and the Baltic Sea are too low, even in RCA3. With the lateral boundaries and the sea surface temperatures being too cold, the regional model cannot diverge to a warmer temperature. Furthermore, the low temperature in RCA3 favours a climate with an extended snow season, and therefore, larger areas of Europe will be covered with snow. The snow increases the albedo which leads to a reduction of the absorbed shortwave radiation. The albedo feedback, in this case, is an additional mechanism promoting a too-cold modelled climate.

4.3 Precipitation

The regional climate model reproduces precipitation rather well (Fig. 9). Precipitation patterns are in fair agreement with ERA-40. The difference in annual precipitation is within ± 20 mm/month in most of Europe. The largest deviations from observed rainfall are found in spring when modelled precipitation is 20–40 mm/month too high. The largest differences occur in areas with abundant precipitation and the deviation from ERA-40 is less dramatic, if relative changes are considered. Large differences are found in Norway, northern Portugal, and the west coast of UK and Ireland. These areas lie along the western edge of Europe, and they all are under the strong influence of zonal inflow from the Atlantic. The humid air from the At-

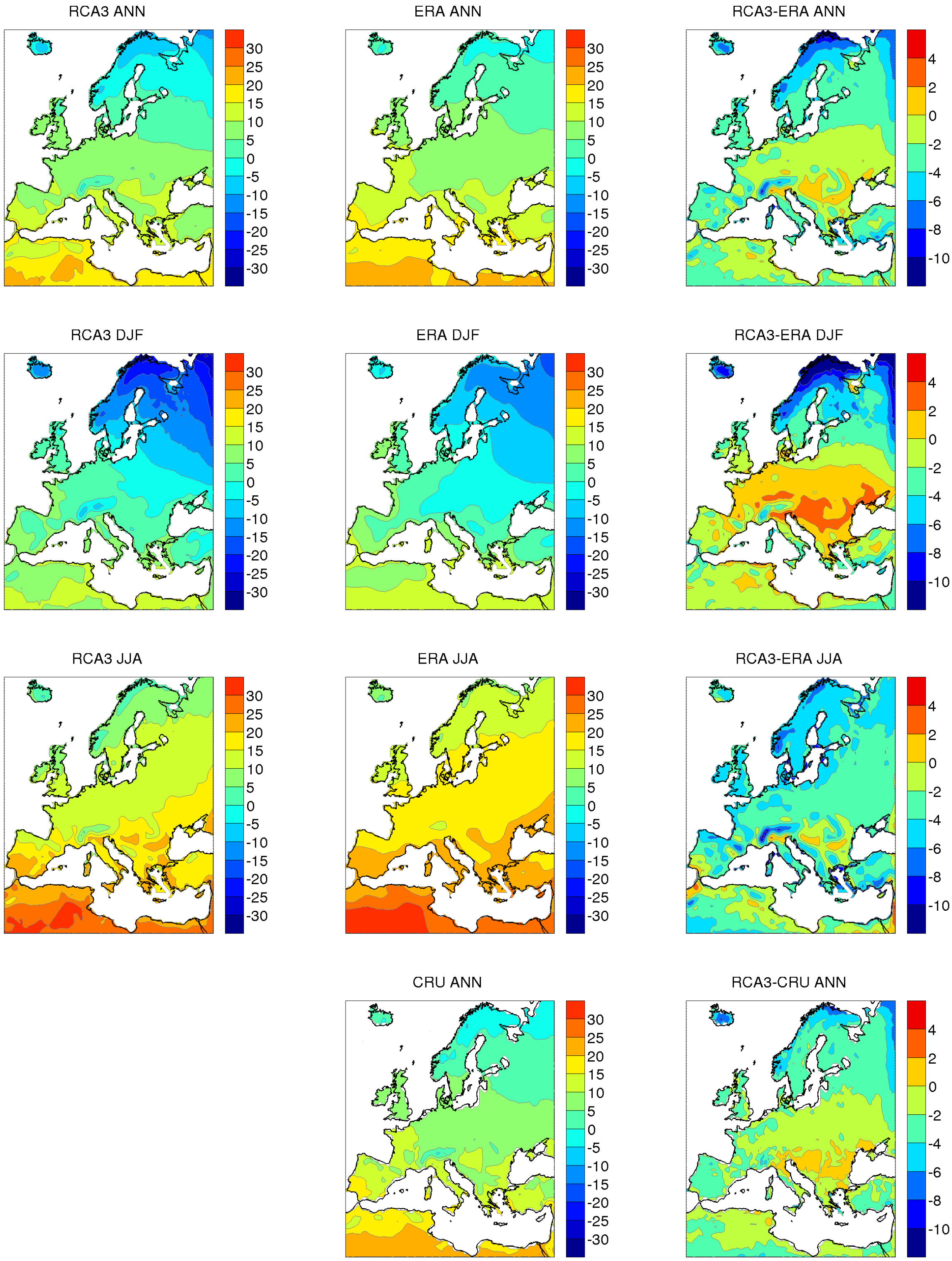

***Figure 8**: Temperature [°C] from RCA3 (left), from ERA-40 (middle) and bias RCA3-ERA40 (right). Bottom: CRU temperature [°C] (middle) and bias RCA3-CRU (right).*

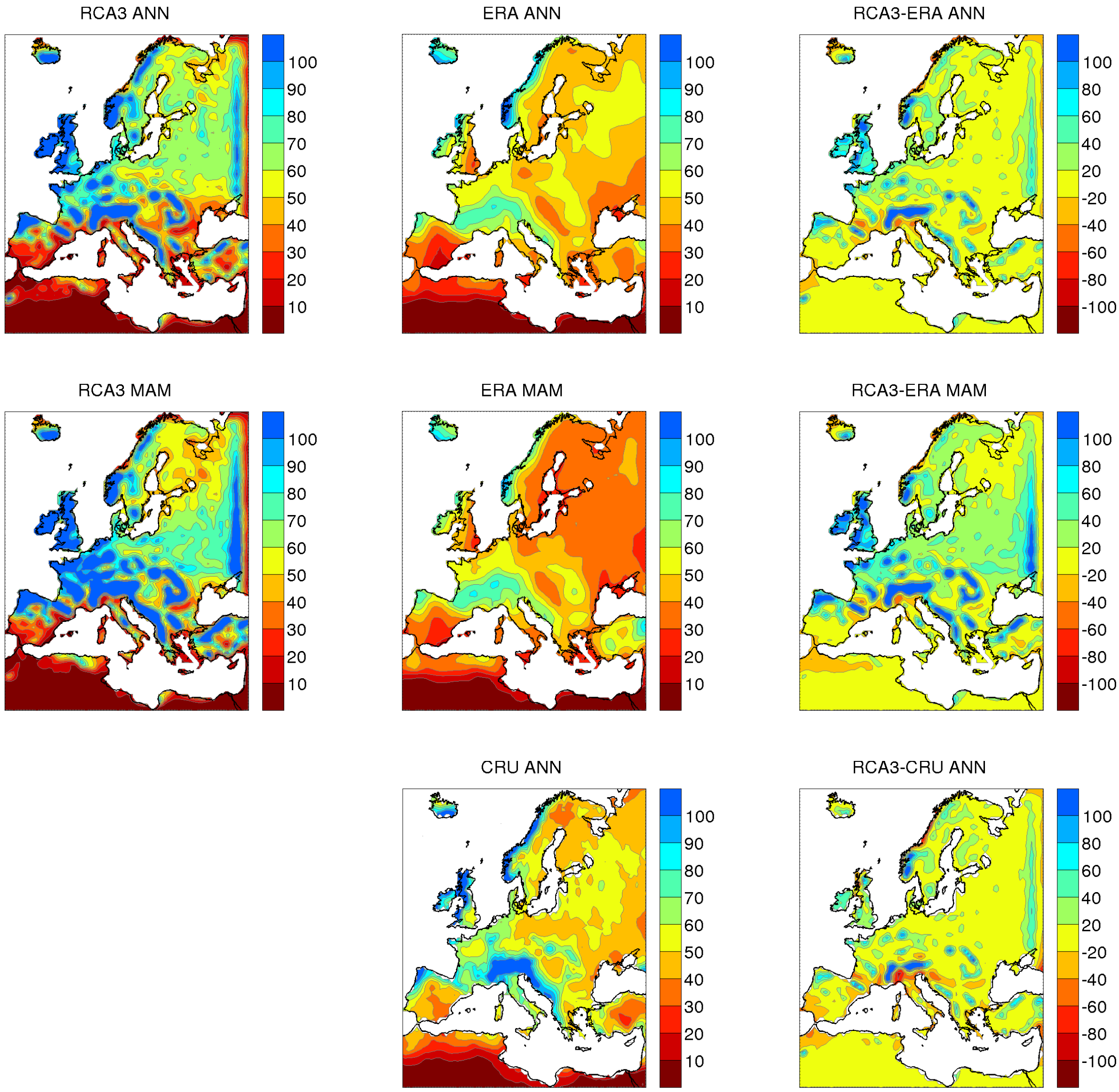

Figure 9*: Annual and spring precipitation [mm/month] from RCA3 (left), from ERA-40 (middle) and bias RCA3-ERA40 (right). Bottom: CRU annual precipitation [mm/month] (middle) and bias RCA3-CRU (right).*

lantic looses its water content when it reaches land and is forced to rise over coastal mountain ranges. The driving model and RCA3 both show a too-strong zonal flow compared to ERA-40. We can therefore expect increased amounts of precipitation along Europe's west coast.

The precipitation in RCA3 is also too high over the Alps and the Scandinavian mountain range. High amounts of rainfall are common for mountain ranges, and the excessive precipitation in the RCA3 simulation is not a matter of concern when relative changes are considered.

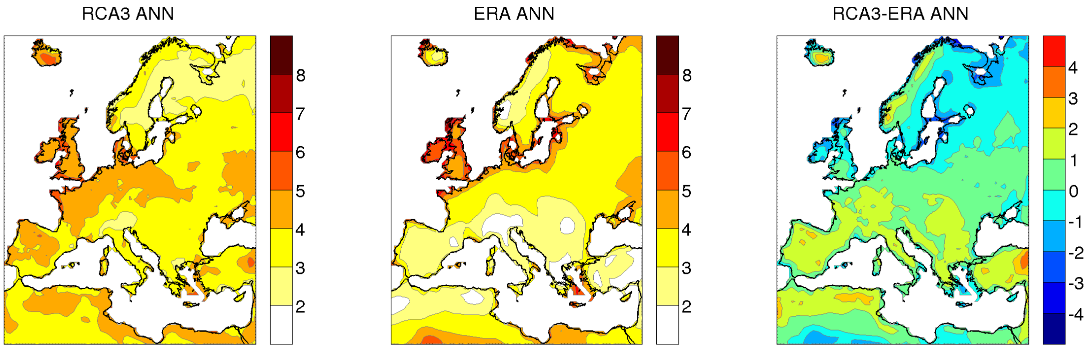

***Figure 10**: Average annual wind speed [m/s] from RCA3 (left), from ERA-40 (middle) and bias RCA3-ERA40 (right).*

4.4 Wind speed

The differences in wind speed between the model and ERA-40 are small (Fig. 10). The modelled wind speeds are up to 2 m/s too high in Western and Southern Europe, and up to 2 m/s too low in the north eastern part. No significant seasonal variation has been found in wind speed difference between RCA3 and ERA-40.

4.5 Summary

The RCA3 regionalisation driven by CCSM3 captures the observed mean climate reasonably well. However, the pressure patterns of RCA3 follow closely the driving model, and thus shows a too-strong zonal flow with an increased transport of humid air towards Central Europe. As a consequence, precipitation along the Atlantic coastline is on the high side due to the inflow of moist air.

The largest deviations between our results and the ERA-40 and CRU datasets are found for the temperature. This can be explained with the too-low temperature of the driving model, and also due to the above-mentioned bias in the circulation. Furthermore, the low temperature in RCA3 over Northern and Northeastern Europe favours an extended, both in space and time, snow cover that reduces the amount of absorbed solar radiation through the albedo feedback.

The large difference in the temperature of the control period makes it difficult to compare our results with other data in absolute terms. Therefore, the relative differences will be looked at and the mean temperature field from the 1961–1990 period will be subtracted from all results. What is then obtained is the change in temperature that is caused by the climate forcing from the increased greenhouse gas concentration. For precipitation and wind speed, the relative change as the ratio of the future value is expressed, and this is divided by the corresponding value of the control period.

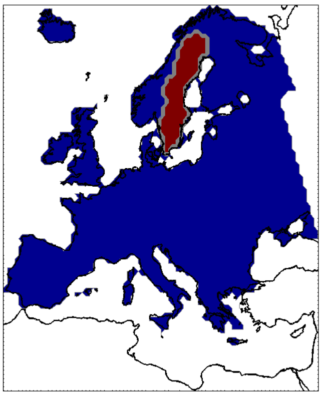

Figure 11: The regions for computing averages of RCA3 results. Sweden (SWE) is shown in red and Europe (EUR) as the red and blue areas combined.

5 Simulated climate change for the 2020–2200 period

5.1 Assessing the changes in a future climate

The analysis of the RCA3 simulation will focus on the key climate factors: temperature, precipitation, mean sea level, and wind speed. For all these quantities, the 30-year averages that represent the future climate of consecutive periods is computed. Seasonal averages of 30-year periods are analysed where appropriate. Changes are expressed relative to the period 1961–1990. For temperature and pressure, absolute differences in °C and hPa, respectively, have been computed, while relative differences (in %) are used for precipitation and wind speed. The results are presented as timeseries of maps that show the spatial pattern in the response of the climate to the forcing that has been imposed to the global climate model.

In addition to these maps, we have also computed timeseries of area-averaged temperature and precipitation. The two regions of interest are the land-covered grid points of RCA3 in Europe (EUR) and Sweden (SWE) (see Fig. 11). The change with time is expressed as the deviation from the period 1961–1990. Results have been filtered with a 10-year running mean to remove the inter-annual variability while still retaining the slow trend.

5.2 Variability after stabilisation

Stabilising the greenhouse gas concentrations in the atmosphere does not immediately imply a new stable climate. The climate response is slow and the change in greenhouse gases will affect the climate long after the stabilisation period has started (Figs. 12 and 19). The climate will also continue to vary from year to year similarly as it is the case now. It is clear from the figures that there is a considerable inter-annual as well as decadal variability. It follows that reaching a new stable climate state may take a considerable amount of time. There will therefore be the

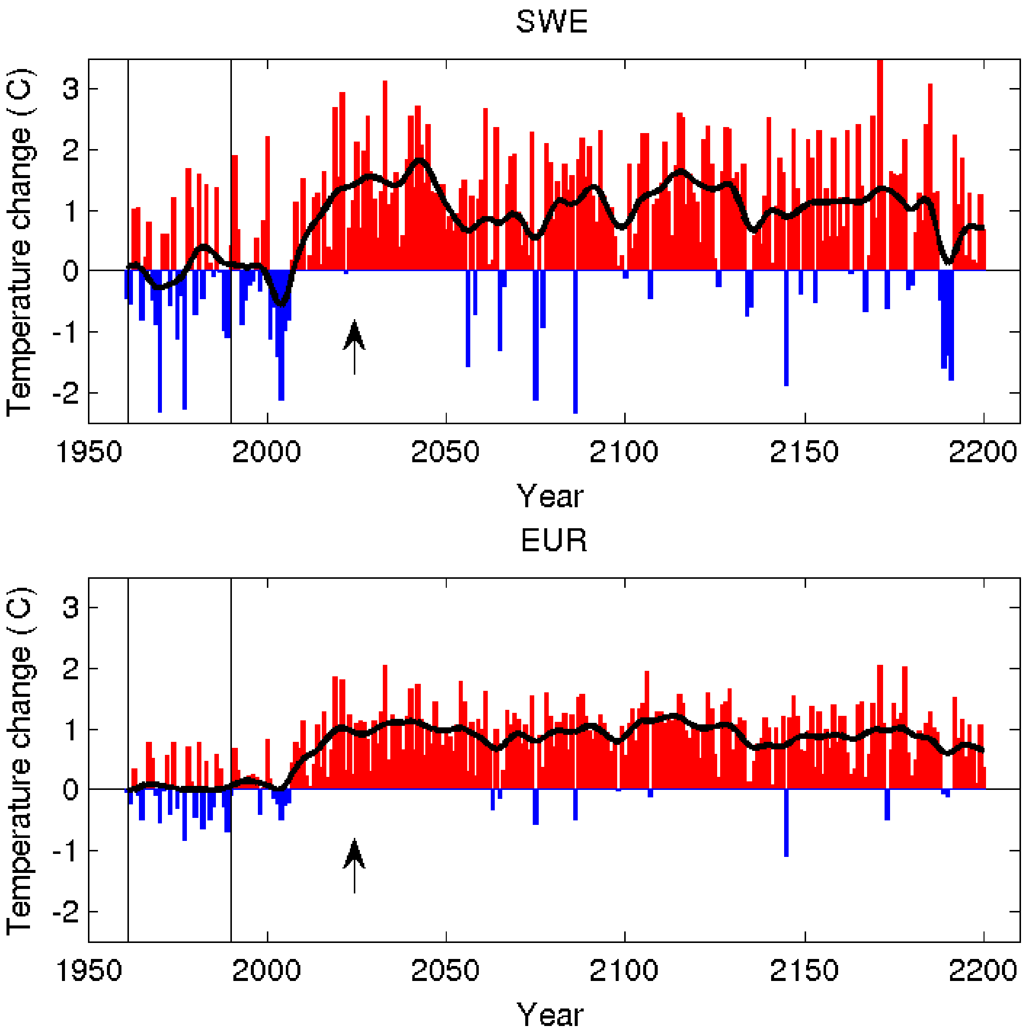

Figure 12: *Temperature anomalies for Sweden (top) and Europe (bottom). Bars represent deviation in annual average relative 1961–1990. Red bars indicate warm years and blue bars cold years. The black line shows the 10-year running average. Vertical lines mark beginning and end of the 1961–1991 period. The arrow indicates the begin of the stabilisation.*

need to study the climate evolution for a long time after the stabilisation in order to assess the impact of the stabilisation.

5.3 Temperature

Until stabilisation the average European surface temperature increases with approximately 1°C compared to the 1961–1990 period. In the following 180 years the temperature change varies from 0.65°C to 1.2°C, with an average of 0.9°C relative to 1961–1990. Over Sweden the variability is larger. The temperature peaks at a difference of 1.8°C in 2043 and varies thereafter between around 0.5 and 1.5°C, resulting in an average climate change of 1°C for the period after the stabilisation compared to 1961–1990. Figure 12 depicts temperature anomalies expressed

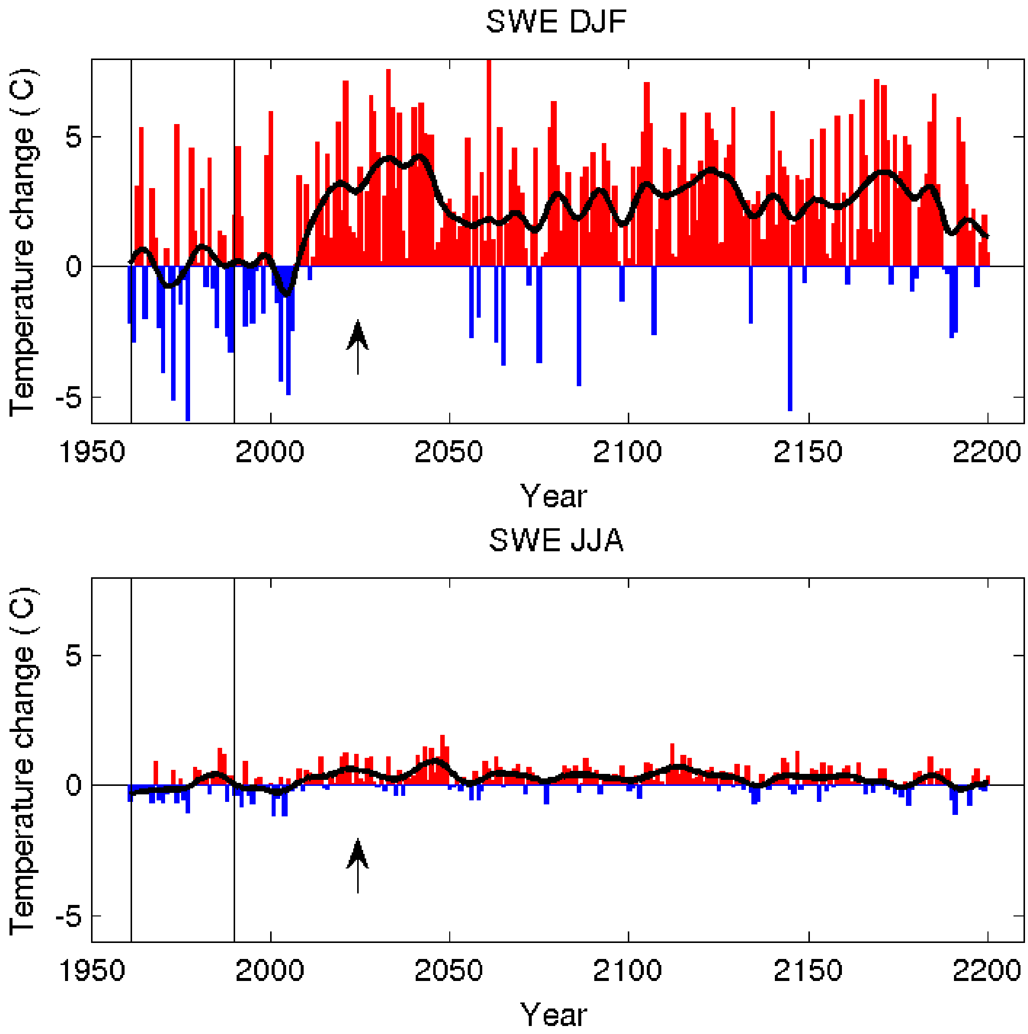

***Figure 13**: Change in mean winter (top) and summer (bottom) temperature over Sweden relative to 1961–1990. See Fig. 12 for explanation.*

as annual average and 10-year running average. Despite the large variability, a general warming is evident. Only few years after 2010 have annual temperatures below the average for the control period.

The change in temperature is not distributed equally among the seasons. The strongest warming is found during winter while summers will change rather insignificantly (Fig. 13). The results for spring and autumn lie between those for winter and summer.

When looking at the regional distribution of the temperature change, the largest warming is found over Northern and Eastern Europe. Here the annual average temperatures increase with up to 2°C (Fig. 14). The temperature increase is less pronounced, less than 1°C, in Central and Southern Europe, except in the southern part of the Iberian peninsula where temperature increases are slightly larger (1.5°C).

The largest warming is found over Scandinavia and North-East Europe during winter (Fig. 15).

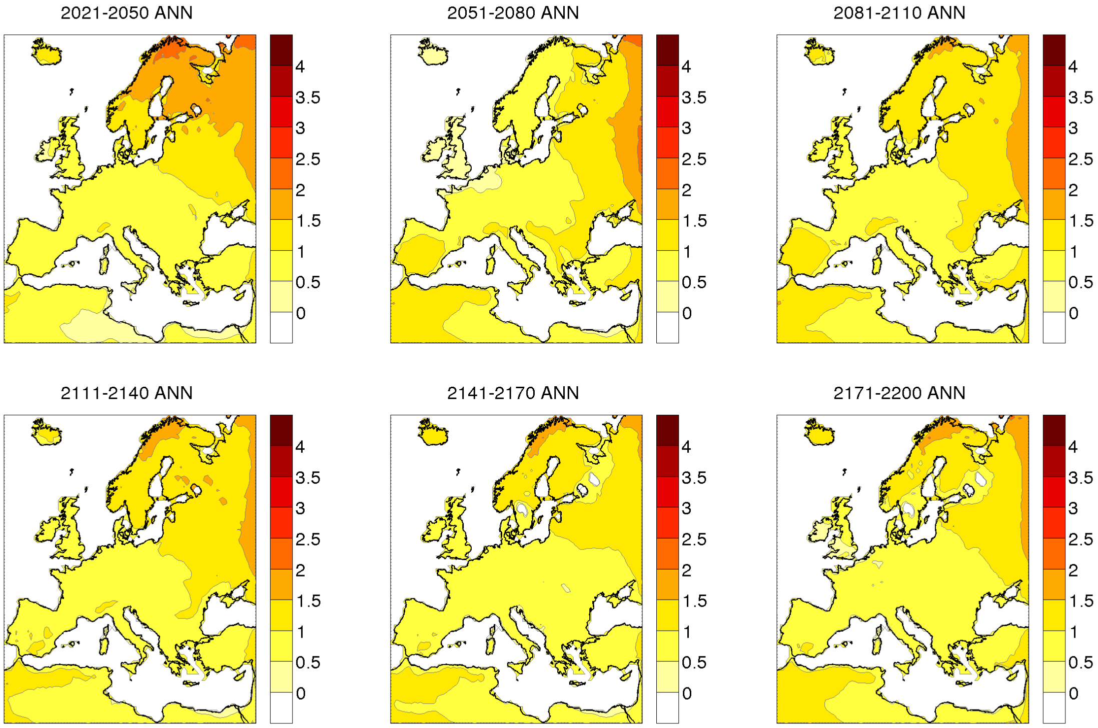

***Figure 14**: Change in annual mean temperature [°C] relative to 1961–1990*

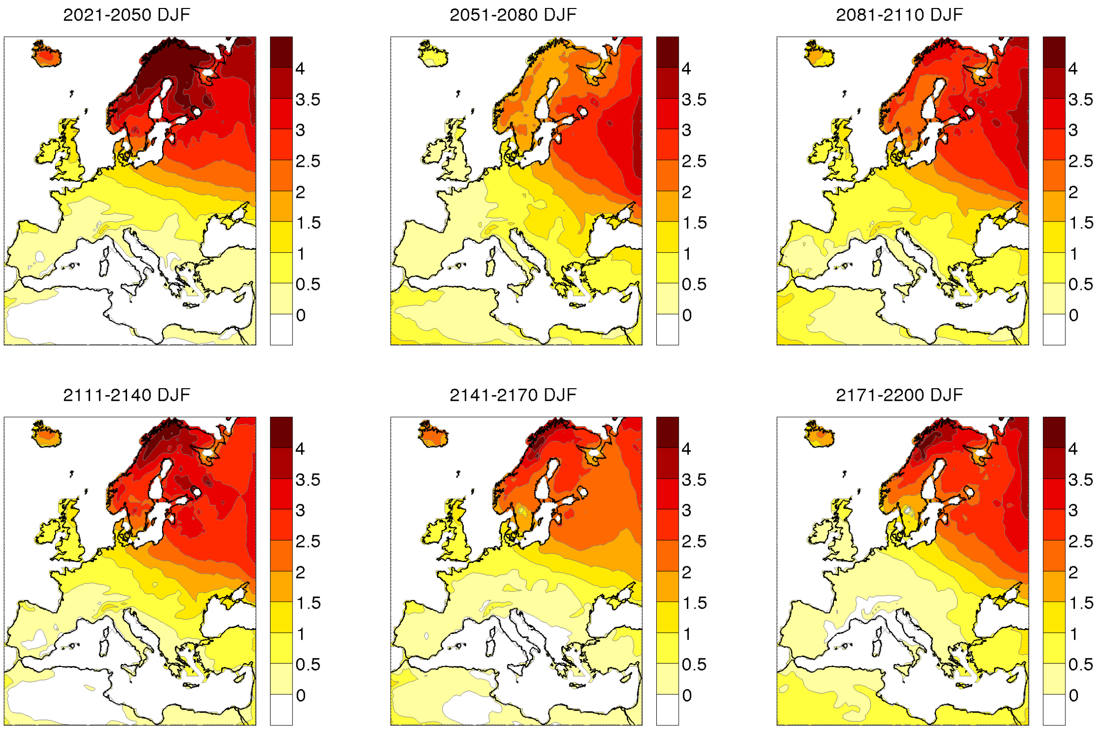

***Figure 15**: Change in mean winter temperature [°C] relative to 1961–1990*

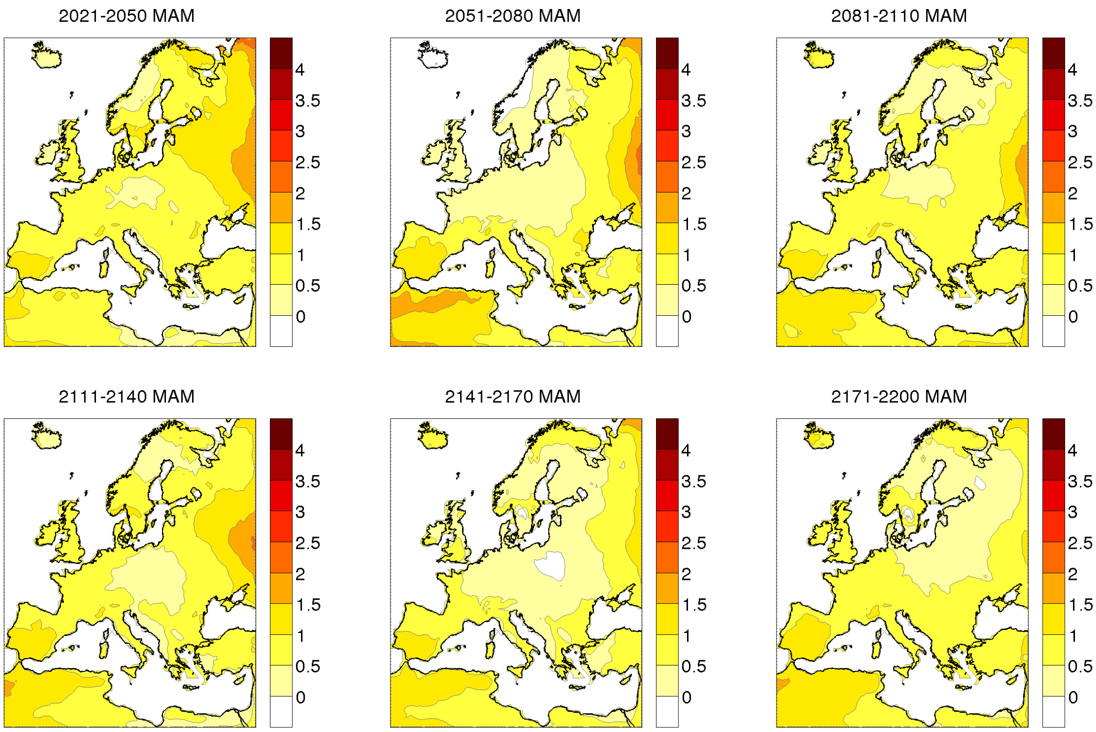

Figure 16: *Change in mean spring temperature [°C] relative to 1961–1990*

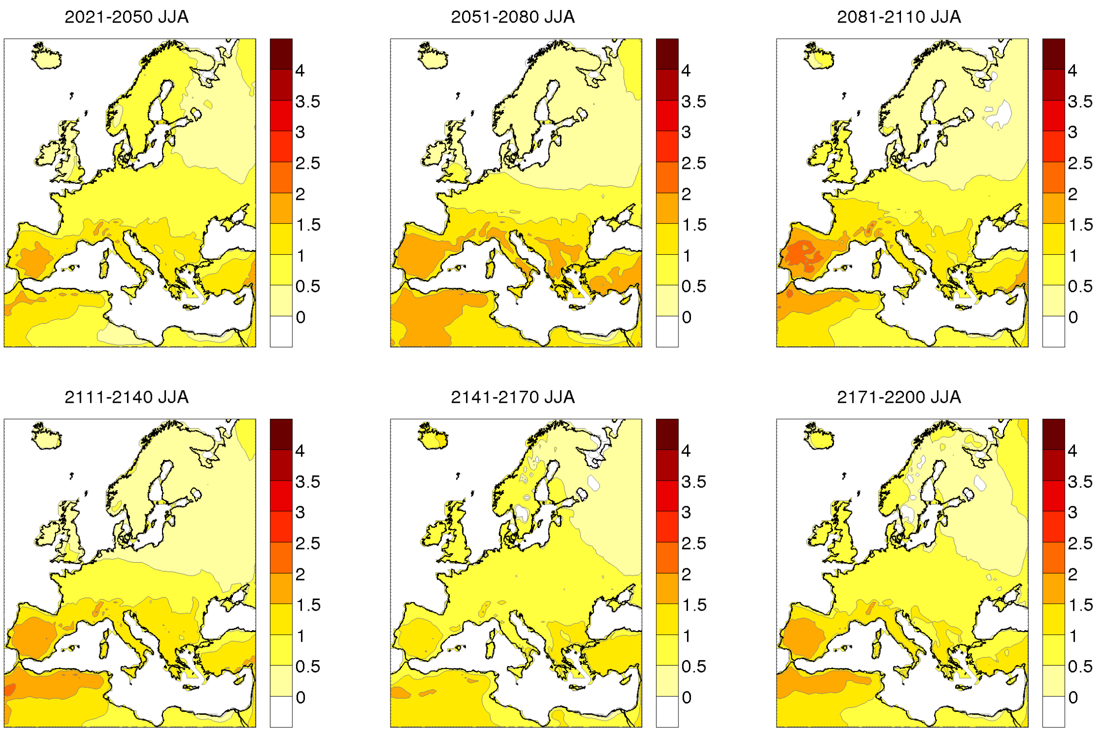

Figure 17: *Change in mean summer temperature [°C] relative to 1961–1990*

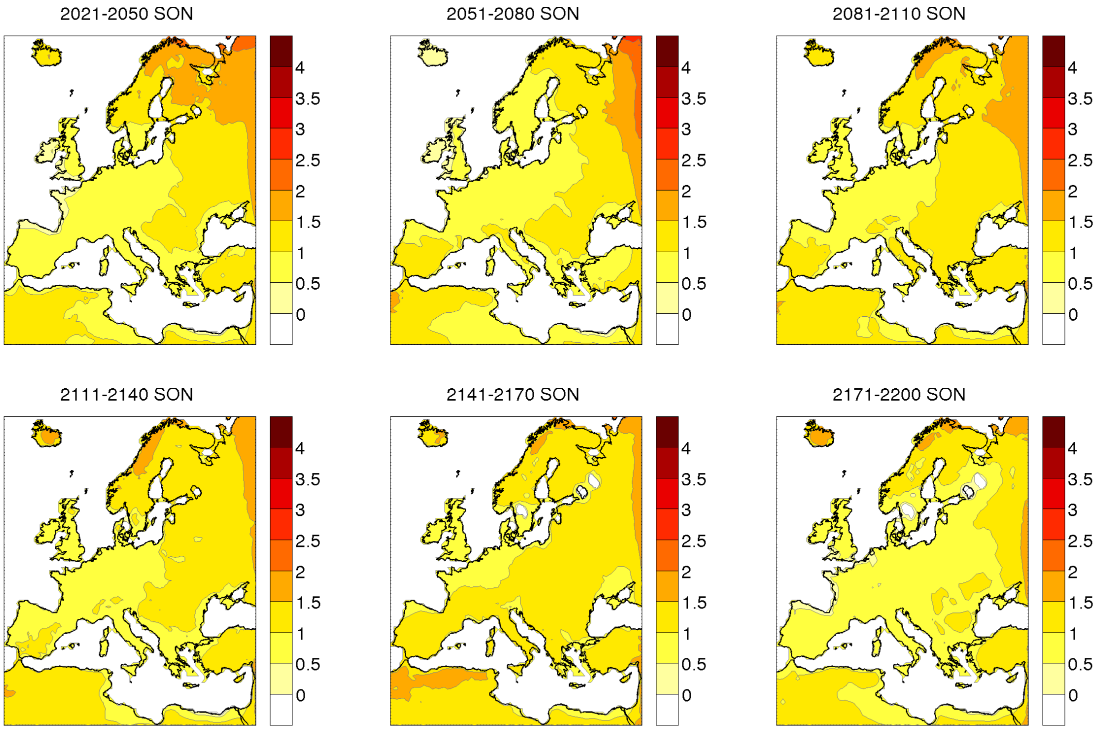

***Figure 18**: Change in mean autumn temperature [°C] relative to 1961–1990*

Temperatures increase by up to 2°C in Southern Sweden and more than 4°C in the most northern mountain areas. The variability over Scandinavia between the 30-year periods is large, from 2021–2050 the temperature change is about 4°C while it is less than 2°C for the 2051–2080 period. In Southern Europe, winter temperatures change less than or around 0.5°C. The strong warming during winter in the north-east of Europe, and the moderate warming over Southern Europe is a robust pattern of the whole experiment that appears in all 30-year periods.

Spring temperatures increase with approximately 0.5°C for most parts of Europe. Over Spain and Northern Africa, the warming can be slightly larger and exceed 1°C (Fig. 16).

The changes in summer temperatures show a pattern similar to the one for spring temperatures, but with an enhanced temperature increase around the western Mediterranean. The whole of the Mediterranean is, generally speaking, the area with the largest temperature increase during summer. Over Northern Europe and Sweden the temperature change is moderate with a warming that rarely exceeds 0.5°C (Fig. 17).

The distribution of changes in autumn temperatures resembles that of winter temperatures, when the largest warming is found over Scandinavia and Eastern Europe but the gradient across Europe is not as strong as during winter (Fig. 18).

5.4 Precipitation

Areas with small absolute amounts of precipitation under present day conditions are prone to produce large signals when analysing the relative change. Therefore, we introduce a precipitation threshold to avoid misleading interpretations of the results. The relative change (ΔP in %) of the precipitation is

$$\Delta P = 100 \frac{P - P_{ctrl}}{\max(P_{ctrl}, P_{thr})} \quad ,$$

where P and P_{ctrl} is the precipitation of the future and of the control period, respectively. P_{thr} is a lower threshold that is set to 5 mm/month in this study. The correction reduces the change

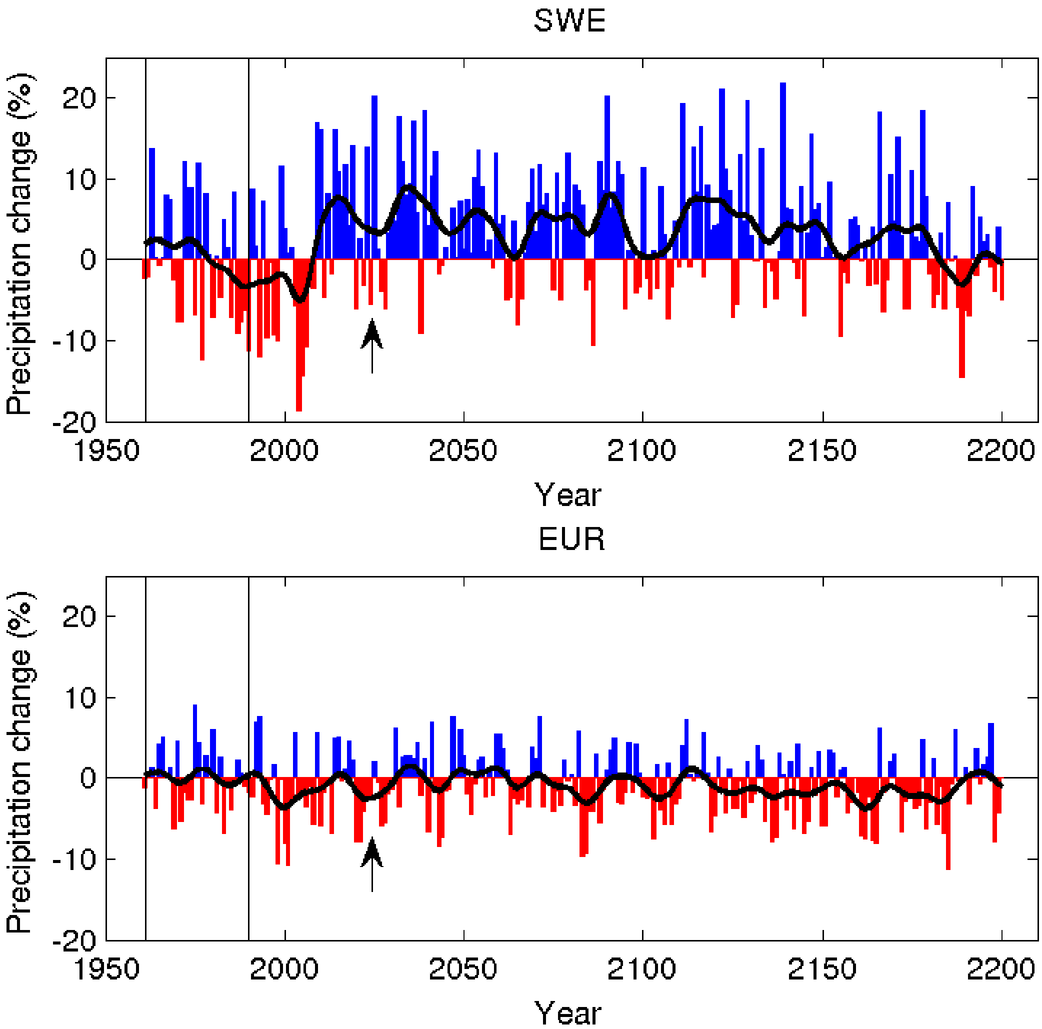

***Figure 19**: Precipitation change [%] for Sweden (top) and Europe (bottom). Bars represent deviation in annual average relative 1961–1990. Red bars indicate dry years and blue bars wet years. The black line shows the 10-year running average. Vertical lines mark beginning and end of the 1961–1991 period. The arrow indicates the begin of the stabilisation.*

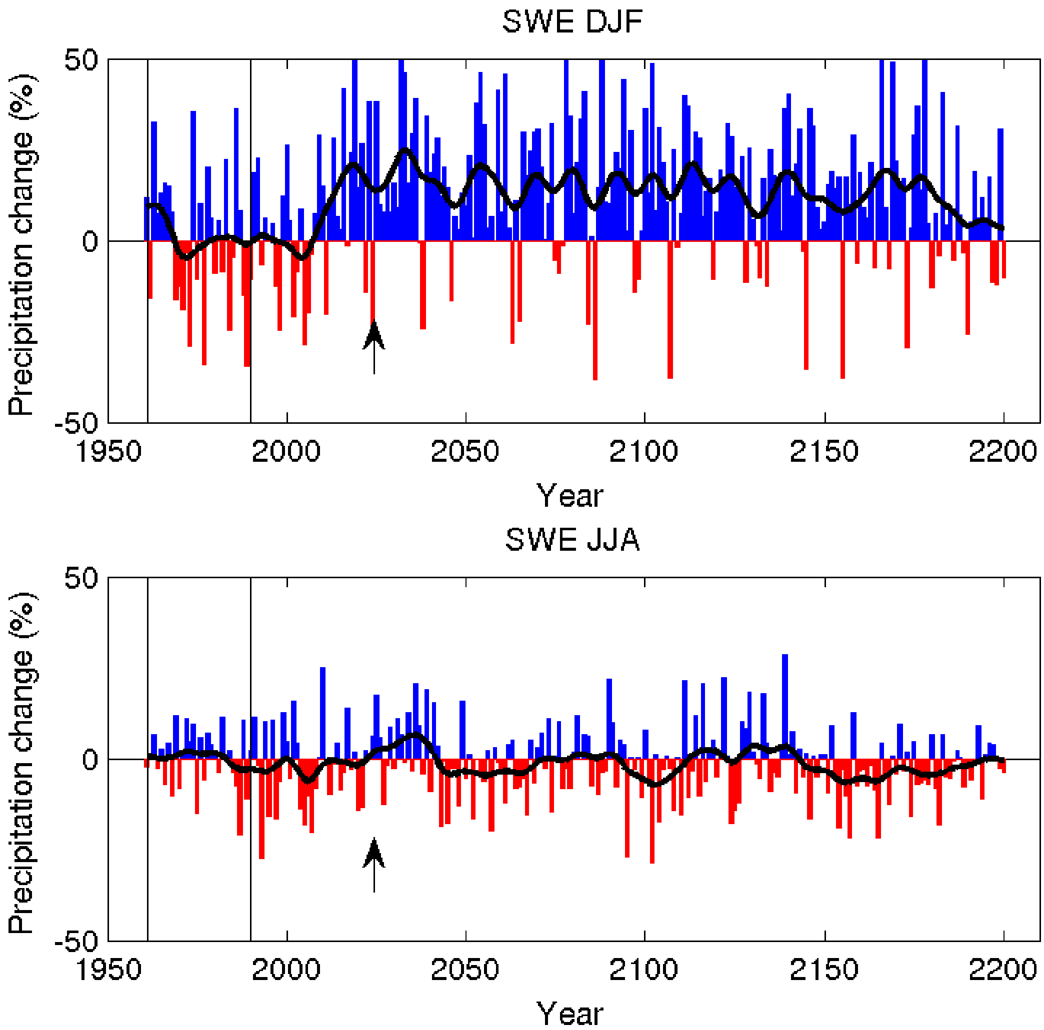

Figure 20: *Change in mean winter (top) and summer (bottom) precipitation over Sweden relative to 1961–1990. See Fig. 19 for explanation. The results for Europe are qualitatively identical with a smaller magnitude of the temperature change in winter (not shown).*

in precipitation in dry regions and removes spuriously large changes in arid regions where P_{ctrl} is close to 0. It has little impact in regions with moderate and high amounts of precipitation as we find it over much of Central and Northern Europe. The threshold P_{thr} has been chosen conservatively and we can still find a very strong signal over Northern Africa after applying the correction.

Figure 19 illustrates annually averaged precipitation change over Sweden and Europe. The precipitation in Europe remains relatively unchanged in the beginning of the stabilisation period, but during the last 50–100 years of the simulation there is a small but distinct change to a drier climate. Sweden on the other hand gets more precipitation after the stabilisation. Towards the end of the simulation, the precipitation change decreases and the precipitation approaches the same amounts as during the control period. Whether this has to do with climate change or variability cannot be determined.

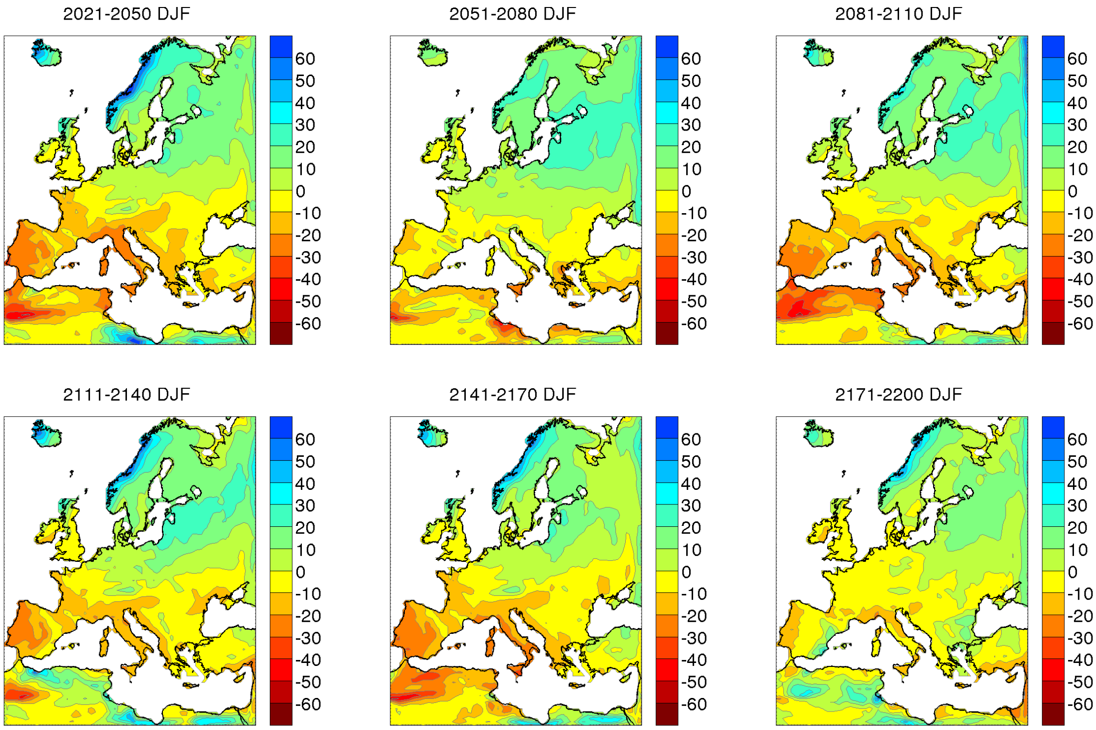

Figure 21: *Winter precipitation [%] relative to 1961–1990*

An analysis of the contributions from different seasons to the annual averaged change in precipitation reveals that most of the change is due to an increase in the winter precipitation (Fig. 20). While winters get clearly wetter, we also find a weak tendency towards drier summer months over Sweden. Compared to the change over Sweden, the changes over Europe are less pronounced in the winter precipitation, and the summers tend to be distinctly drier (not shown).

The regional distribution of precipitation changes during winter is shown in Figs. 21 and 22. The general pattern is up to 40% less precipitation over Southern Europe and, more or less, unchanged amounts of precipitation in Northern Europe. The large change around the Mediterranean Sea should be viewed with caution as this region is dry today and the relative change can be large despite the correction with P_0. However, the decrease in the amount of precipitation around the Mediterranean Sea is a robust signal. The variability over Central Europe is quite large when we compare the different 30-year periods. France, for example, can exhibit both drier (during the 2051–2080 period) and wetter summer climate (2141–2170).

Winter precipitation increases over Scandinavia and Northeastern Europe (Fig. 21). The largest changes, up to 50%, are found on the Atlantic side of the Norwegian Mountain range, while the average increase in precipitation over Northern Europe is between 10 and 30%. Precipitation over Southwestern Europe is reduced; but the variability is large between different 30-year periods. The precipitation over the Iberian Peninsula is reduced by 20–30% in 2021–2050, but it ‘bounces’ back to current levels for 2171–2200. Central and South-Eastern Europe show a weak tendency towards a drier climate, but the trend is not strong.

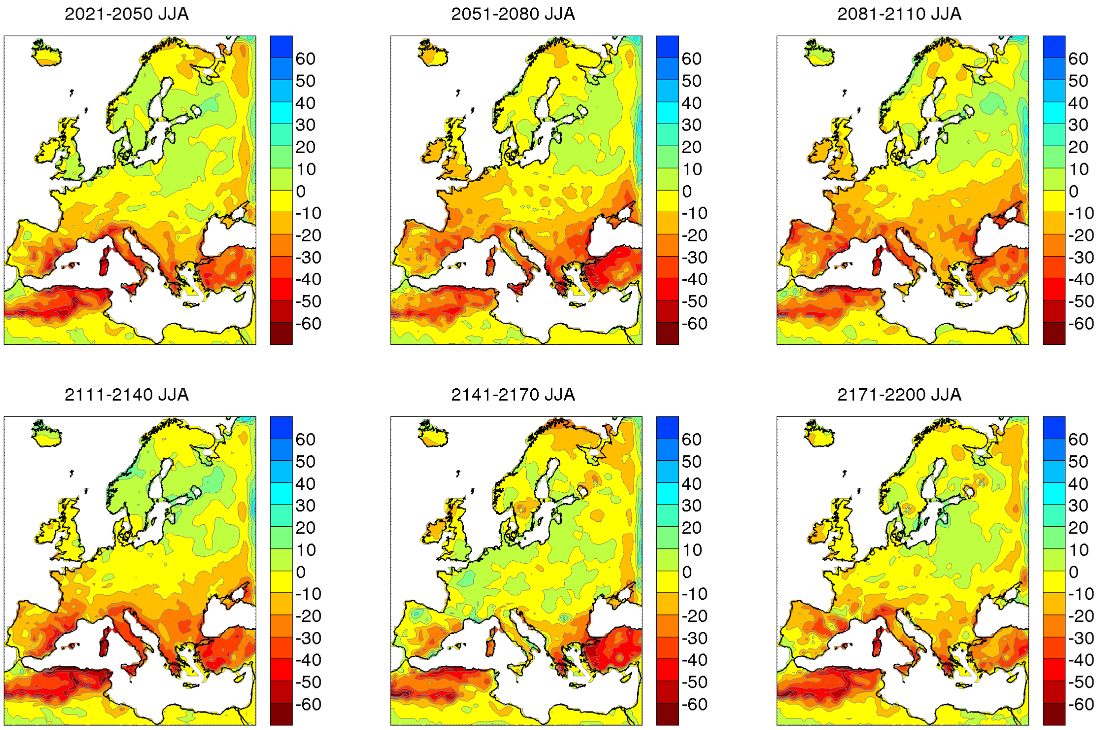

***Figure 22**: Summer precipitation [%] relative to 1961–1990*

5.5 Mean sea level pressure

The changes in mean sea level pressure are in general negligible. The most apparent changes appear during winter when the pressure decreases in northern Scandinavia and increases in central and south western Europe (Fig. 23). This implies a stronger gradient between South-western and Northern Europe and suggests an increase in the mean westerly geostrophic winds. There is also an indication of a slight northward shift of the cyclonic activity (stormtracks) during winter. The same pattern of change also appears in spring although it is less pronounced. Pressure changes in summer and autumn are very small.

5.6 Wind speed

The changes in annual wind speed are moderate, on the order of ±4% over the whole of Europe. However, the small annual change is a result of compensating trends: the differences are larger and of opposite sign for winter and summer (Figs. 24 and 25). During winter, the wind speed increases over Northern Europe and the British Isles in general, but it decreases during summer months. The winds over the Mediterranean region, on the other hand, become stronger in summer and weaker in winter. The seasonal changes are largest in winter which is also the period when we have found the largest changes in the pressure distribution.

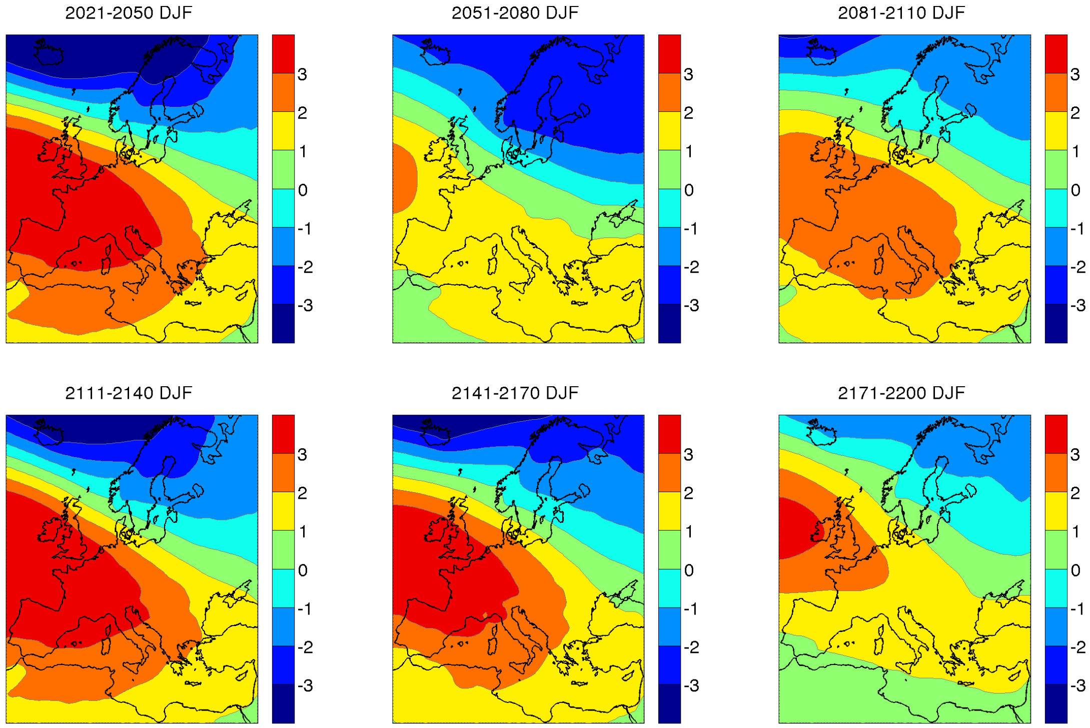

***Figure 23**: Change in winter mean sea level pressure [hPa] relative to 1961–1990*

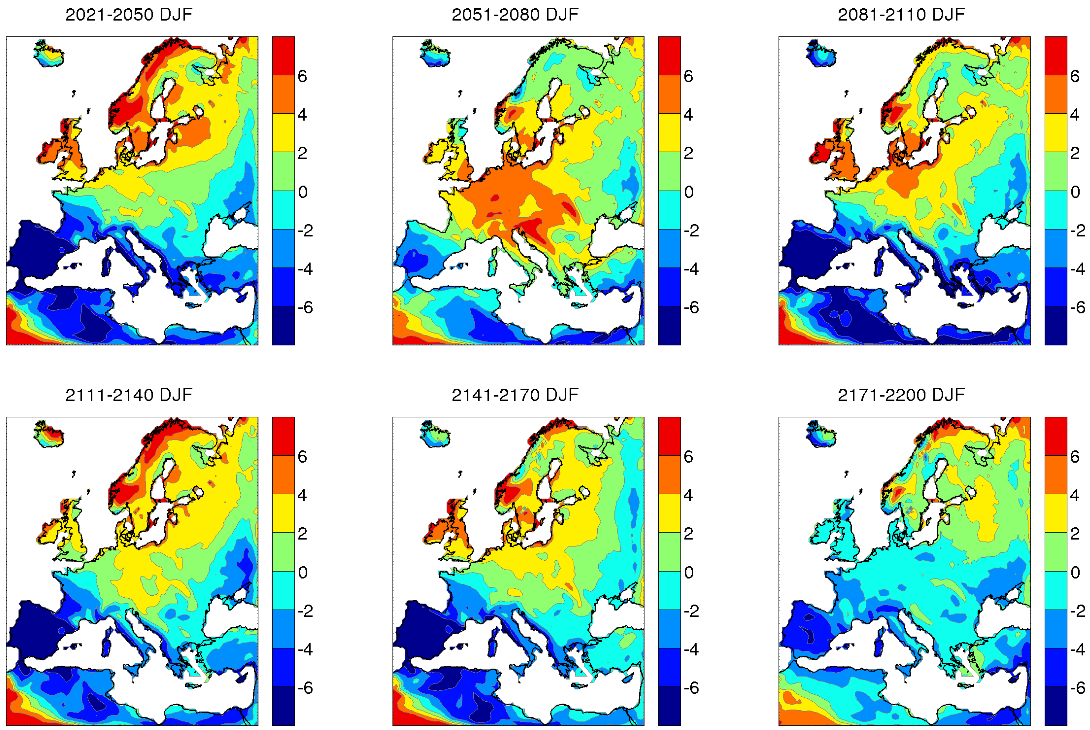

***Figure 24**: Change in winter wind speed [%] relative to 1961–1990*

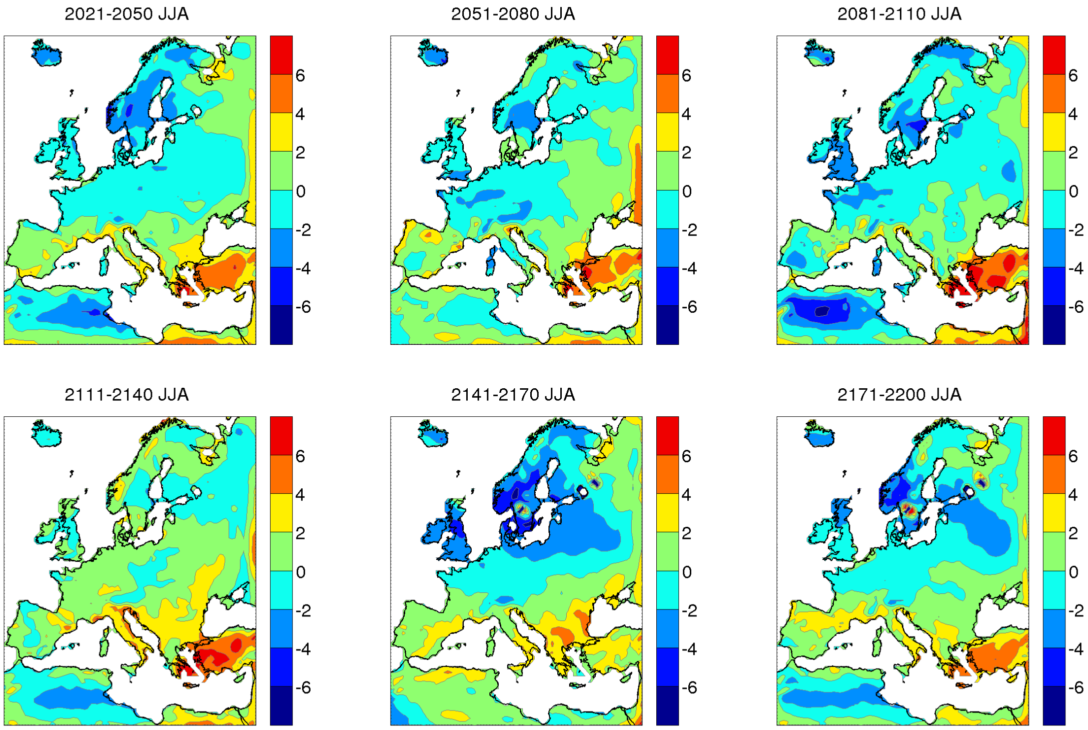

***Figure 25**: Change in summer wind speed [%] relative to 1961–1990*

6 Conclusions

In this study a state-of-the-art global coupled climate model was run from early industrial conditions to a level of 450 ppm equivalent CO_2. Thereafter. the experiment was continued until 2200 for the impact of the stabilisation to work its way through the climate system. The global simulation was then down-scaled over Europe with our regional climate model RCA3. The global climate model CCSM3 is able to reproduce the climate of today reasonably well, although with a substantial temperature bias, in particular, over Northern Europe. A possible reason for this discrepancy is the choice of a low horizontal resolution (T31) for this study. The restart files from the perpetual 1870 simulation that were used as initial conditions proved to be too cold compared to a similar simulation with T85. The T31 simulation was not able to overcome the cold bias from the initialisation during the transient simulation. The cold bias is most likely related to the coarse resolution of the ocean model in CCSM3. The SST distribution with T31 reveals large differences compared to T85, and even to ERA-40 for the period 1961–1990. A consequence of the cold North Atlantic water is a climate over Europe and Scandinavia that is too cold. Furthermore, we also find large differences to the re-analysis in the MSLP field over the North Atlantic that implies an overestimated transport of moist and mild air from the Atlantic ocean towards Europe in the model. Despite the shortcomings of the low resolution, CCSM3 is able to reproduce the climate and its variability reasonably well. The regional climate model RCA3 follows closely the driving model and remains in agreement with earlier studies with RCA3 (Räisänen et al., 2004; Rummukainen et al., 2004). The differences between today's observed climate and the model results will thus persist even after the regionalisation. However, this should not impact the forced changes between today's and the future climate.

The increased levels of CO_2 imply a shift towards a warmer climate both in the global and the regional simulation. After the stabilisation of CO_2 at 450 ppm, the temperature trend more or less levels off. The global mean temperature is about 0.9°C warmer than today, or about 1.5°C warmer than during pre-industrial conditions. The regional response over Sweden is somewhat larger than the global average. Our global simulation indicates that the climate has not stabilised even after 180 years of stable CO_2 concentrations; although the remaining trend is rather small towards the end of the simulation. The variability remains large, and there are extended time periods of warmer and colder years in the global simulation. The regionalisation shows an even larger variability than the global model especially when looking at the regional response over Sweden where the changes in annual mean temperature vary on decadal and longer timescales between 0.5 and 1.5°C compared to present day conditions, and with an inter-annual variability of ± 2°C. Precipitation amounts increase by about 5% over Sweden, and remain more or less unchanged over Europe as a whole. The variability in temperature and precipitation does not change in a future climate.

The regional distribution of the warming in the 450 ppm scenario shows a distinct south-west to north-east gradient. Although the gradient can be more or less pronounced in different 30-year periods of the future, the strongest annually averaged warming was always found over Northern Europe and Russia, and the weakest signal in the Mediterranean area. The strongest warming occurs during winter months that can become 3°C warmer over Northern Europe. During the summer the warming is largest in the Mediterranean region with a peak over the Iberian peninsula. The enhanced summertime warming over Southern Europe is related to the expected drier conditions. Rainfall is expected to decrease there in a future climate and, subsequently, periods with droughts and low ground water levels will become more common. The shortage of ground water reduces evaporation, and the reduced evaporative cooling implies a stronger heating of the surface. In other parts of Europe, summer rainfall changes less and the summertime warming is rather moderate.

Summing up for Sweden, it is found that the increasing levels of greenhouse gases and the stabilisation at 450 ppm equivalent CO_2 leads to a warmer and wetter climate. The largest changes relative to 1961–1990 occur during winter when mean temperatures can be 3°C warmer and precipitation increases by about 15%. Changes in temperature and precipitation are negligible for the summer months. The shift towards a warmer and wetter climate follows closely the evolution of the greenhouse gas concentration. The largest changes occurs before 2023 when the equivalent CO_2 concentration is increasing. After the stabilisation of CO_2, the regional climate also, more or less, stabilises at a warmer and wetter state, although it keeps fluctuating around this mean state. This year-to-year and decadal variability is not too different from today's climate. The changes over Sweden and Northern Europe are larger than in the rest of Europe, mainly due to the stronger wintertime response to the applied climate forcing.

The results presented here are in line with other climate change experiments done at the Rossby Centre. Kjellström et al. (2005) have used the same regional climate model, but driven with ECHAM4/OPYC3 at the boundaries and forced with IPCC SRES A2 and B2 scenarios (Nakićenović et al., 2000). The results of Kjellström et al. (2005) are in close agreement with our findings through the first 30 years of the 21st century, after which the results start to diverge. The forcing in our simulation is kept constant after 2023 while greenhouse gas concentrations keep increasing in Kjellström et al. (2005). Despite these differences in the experiment's setup, both studies agree qualitatively and come to the same conclusions: Sweden and Northern Europe can expect milder and wetter winters, and smaller changes during summer, while Southern Europe experiences the greatest change in summer, mainly as a consequence of the decreased precipitation.

Dynamical down-scaling with regional climate models strongly depends on the driving data imported from a global climate model (Räisänen et al., 2004). From this, it follows that the study presented here is limited by the use of a single realisation of one climate model with low resolution. Robustness and, thus, confidence in the results, would increase if multiple sets of driving data could be used in an ensemble experiment. This is, however, beyond the scope of the present study.

A number of question marks remain in the context of CO_2 stabilisation. It is difficult to predict the evolution of atmospheric greenhouse gas concentrations and there are many pathways to achieve a stabilisation and many conceivable stabilisation levels. This study uses a simple approach with a constant increase until a given stabilisation level is reached. It is, however, more likely that the CO_2 concentrations may overshoot a target set at a rather low stabilisation level, at first, and then decrease again to reach the target level from above. Since there is a memory in the climate system, the magnitude and timescale for the overshooting will also have an effect on the resulting climate state.

The global mean temperature in our simulations increases by less than 1°C compared with today's climate, or about 1.5°C compared to pre-industrial conditions. This is still well below the so-called 2°C limit. The results of the present study should thus be interpreted with caution and taken as a lower limit of climate changes under a stabilisation at 450 ppm equivalent CO_2. We suggest to extend the investigation with additional experiments possibly including multiple models, stabilisation pathways and stabilisation levels.

Acknowledgements

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