Changing Temperature and Precipitation Extremes in Europe's Climate of the 20th Century

Albert Klein Tank

ISBN 90-369-2254-2

Design: Birgit van Diemen, Studio KNMI

Keywords: extremes, climate change, observations, daily dataset

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Veranderingen in Temperatuur- en Neerslagextremen
in het Europese Klimaat van de 20e Eeuw
(met een samenvatting in het Nederlands)

Proefschrift ter verkrijging van de graad van doctor
aan de Universiteit Utrecht op gezag van de Rector Magnificus,
Prof. Dr. W.H. Gispen, ingevolge het besluit van
het College voor Promoties in het openbaar te verdedigen
op maandag 4 oktober 2004 des middags te 12.45 uur door

Albertus Maria Gerardus Klein Tank

geboren op 15 mei 1965, te Groenlo

Promotor:

Prof. Dr. G.J. Komen,
Instituut voor Marien en Atmosferisch Onderzoek Utrecht (IMAU)

Copromotor:

Dr. G.P. Können,
Koninklijk Nederlands Meteorologisch Instituut (KNMI)

…aan mijn ouders

Contents

To the general public, news media and policy makers, 'climate change' is often synonymous with global warming. This is no surprise, because the reported increase in the mean temperature near the surface of the earth provides the most apparent signal to society that our climate is changing. According to the instrumental record, the global mean temperature increase over the period 1861–2000 was 0.61°C, with 90% confidence interval 0.45–0.77°C (Houghton *et al.*, 2001). Between 1901 and 2000 the observed warming was 0.57°C, with 90% confidence interval 0.40–0.74°C. For the period since 1976, the rate of change is roughly 3 times that for the past 100 years as a whole. Analyses of proxy data for the Northern Hemisphere indicate that the late 20th century warmth is unprecedented for at least the past millennium (Crowley and Lowery, 2000; Mann and Jones, 2003). Greenhouse sceptics who question the latter robust consensus view do this wrongly (Mann *et al.*, 2003).

Despite the focus on global warming in the climate change debate, scientists opt for a wider definition. They refer to climate change as a statistically significant variation in either the mean state of the climate or in its variability, persisting for an extended period of time, typically decades or longer (Houghton *et al.*, 2001). Thus climate change is not only characterized by changes in temperature, but also by changes in other variables of the climate system, for instance precipitation. Furthermore, climate change is not only characterized by changes in the mean, but also by changes in variability and extremes. Compared to the existing information on past changes in the mean climate, far less is known about past changes in extremes. For temperature extremes in the 20th century, Houghton *et al.* (2001) summarized as most noteworthy observations the lengthening of the freeze-free season in most mid and high latitude regions and the reduction in the frequency of extreme low monthly and seasonal average temperatures and smaller increase in the frequency of extreme high average temperatures. For precipitation extremes in the 20th century, the only findings were a 2 to 4% increase in the frequency of heavy events in mid and high latitudes of the Northern Hemisphere and the fact that in regions where total precipitation has increased even more pronounced increases were found in heavy precipitation events.

This thesis aims at increasing the knowledge on past changes in extremes through analysis of historical records of observations at meteorological stations. Knowledge on past changes in extremes is relevant in climate change research for various reasons. It is necessary information to establish baselines for adequate monitoring of climate change and setting the context for the interpretation of future trends and changes in variability (GCOS, 2003). Knowledge on past changes in extremes also supports the detection and attribution of anthropogenic influences on climate. Recent detection/attribution studies suggest that changes in extremes should be nearly as detectable (temperature) or even more detectable (precipitation) than changes in the mean (Hegerl *et al.*, 2004). Finally, knowledge on past changes in extremes serves as a reference against which climate models are validated when assessing their ability to simulate the climate. Only if climate models successfully simulate the changes in extremes observed in the past, confidence will exist in their projections of future changes in extremes.

This thesis adds to our knowledge on past changes in extremes by addressing the key question:

- **How did the extremes of daily surface air temperature and precipitation change in Europe's climate of the 20th century, and what can we learn from this?**

The contents is structured along the lines of four follow-up questions:

- **Are the available observational datasets adequate to analyse extremes?** Chapter 2

- **Which trends are observed for the daily extremes of surface air temperature and precipitation?** Chapter 3

- **Can the observed changes in temperature extremes in recent decades be regarded as a fingerprint of anthropogenic climate change?** Chapter 4

- **Do the observed changes guide the development of temperature scenarios for our future climate?** Chapter 5

Are the available observational datasets adequate to analyse extremes? Chapter 2

When assessing whether sufficient observational datasets exist to analyse extremes, one needs to define what is meant by 'extremes'. In disciplines of climate change research such as climate monitoring, climate change detection/attribution and climate modelling, extremes generally refer to rare events within the statistical reference distribution of particular weather elements at a particular place (Houghton *et al.*, 2001) and thus to the tails of the probability density function (PDF). This statistical definition of extremes is also adopted in this thesis. The words 'weather extremes' and 'climate extremes' are used interchangeably, following the paradigm that weather and climate are part of the same continuum. Climate means 'possible' or 'expected' weather and is usually defined as the statistical description in terms of the mean and variability of temperature, precipitation, etc. over a period of time ranging from months to thousands or millions of years (Houghton *et al.*, 2001). For practical reasons, climatologists use fixed 30 yr 'standard normal' periods (e.g. 1961–90) defined by the World Meteorological Organization (WMO) to characterize climate. Otherwise, climate descriptions for one country or application would be incompatible with those for other countries or applications. In climate change research the choice of a 30 yr study period is not obvious, since the climate system contains variability at all time scales. Clearly, the choice of the study period is of great importance for the analysis of extremes in the tails of the distribution.

Another definition of extremes is used in impact analysis. Here, extremes usually refer to the hazardous weather conditions that result in strong adverse effects on ecosystems or sectors of society, such as human safety and health, water management, agriculture, energy, insurance, tourism and transport. Examples of extreme impact events are the environmental disasters that are frequently reported in the media. Extreme impact events may be of short duration, but could also extend over several days, several months or perhaps even years (e.g. droughts). Extreme impact events do not necessarily match with extremes in the far tails of the distribution. The impacts that result from extreme values of weather elements are to a large degree dependent on the conditions of the system that is under investigation (including its vulnerability and capacities for adaptation and mitigation). For instance, the hydrology of an area determines whether or not it can cope with torrential rainfall without being flooded. Likewise, the design of buildings, availability of cooling gear and attitude of the inhabitants determines whether communities can withstand summer heat waves. Although the weather extremes in the tails of the distribution (as used in this thesis) are not directly related to environmental disasters, it is very likely that a systematic change in weather extremes will also be accompanied by a systematic change in extreme impact events.

From a scientific point of view, assessing changes in extreme impact events is problematic for several reasons. First, the inventories that are available to study extreme impact events (see Cornford, 2003) are biased towards recent years simply because of improved communication technology. Second, the methods to compare the reported events in an objective way are lacking. Third, even when corrected for improving communication technology, the growing number of reported events reflects global trends in population vulnerability more than an increased frequency of extreme events. In this respect it is no wonder that widespread claims that extreme events are more common nowadays than in the past also in Europe often come without scientific evidence.

Assessing changes in extremes in the tails of the distribution (the objective of this thesis) is not trivial as well. For statistical reasons, a valid analysis of extremes in the tails of the distribution requires long time series of observations to obtain reasonable samples. Also, series with at least a daily time resolution are needed to take into account the submonthly nature of many climate extremes. In various parts of the globe, there is a lack of homogeneous daily station records covering multiple decades that are part of integrated datasets. Global studies on extremes over land (like Frich *et al.*, 2002) therefore suffer from blank and data scarce regions. Even the recommendations of the Global Climate Observing System (GCOS) initiative of several UN organizations to establish global baseline networks such as the GCOS Surface Network (GSN) has not yet led to a comprehensive dataset of daily series held in a single international data centre (GCOS, 2003).

Until recently, Europe was one of the regions lacking a dataset of high-resolution observational series with sufficient density and quality that was readily available and accessible. This situation changed when a new dataset was developed within the European Climate Assessment (ECA) project (Klein Tank *et al.*, 2002). The ECA dataset consists of daily series of temperature and precipitation at a large number of meteorological stations throughout Europe. It is distributed via the Internet. Earlier initiatives in this direction failed, mainly because of the restricted data policy of National Meteorological and Hydrological Services in Europe. For a long time, the only datasets of temperature and precipitation observations covering entire Europe were the gridded global datasets with monthly resolution that are used for quantitative estimates of global temperature and precipitation change (Jones and Moberg, 2003; New *et al.*, 2000; Hulme *et al.*, 1998).

Chapter 2 *(published in Int. J. Climatol.)* describes the ECA dataset of daily resolution station series of temperature and precipitation. A comparison is made with the monthly gridded datasets and examples are given of the potential of the ECA dataset for climate change research.

☰ Which trends are observed for the daily extremes of surface air temperature and precipitation? Chapter 3

Substantial work has been completed in recent years in the area of climate trend research (Houghton *et al.*, 2001). But in the absence of daily datasets, Europe-wide trend studies on extremes needed to rely on combining the results of different analyses from different areas. Unfortunately, the exact definition of the measures that were used to characterize the extremes in the tails of the distribution differed from one study to the other. To gain a uniform perspective on observed changes in climate extremes, a core set of standardized indices was recently defined by the joint Working Group on Climate Change Detection of the World Meteorological Organization – Commission for Climatology (WMO–CCL) and the Research Programme on Climate Variability and Predictability (CLIVAR; Peterson *et al.*, 2001). Each index describes particular aspects of climate extremes. Examples are the number of days with minimum temperature below the long-term 10th percentile (baseline period 1961–90) or the maximum rainfall amount observed in any period of 5 consecutive days.

Using the standardized indices, the trends in the extremes of Europe's climate of the past can be analyzed unambiguously. But what is a 'trend' in climate? When defining a trend in climate, similar problems arise as described in the previous section with respect to the definition of 'extremes'. Scientists have different understanding of what a trend exactly means. Often, wide formulations are used like in the IPCC assessment reports, in which the word 'trend' is used to designate a generally progressive change in the level of a variable (Houghton *et al.*, 2001). Subsequent sentences like 'Chapter 2 describes the observed trends and inter-decadal variability in climate as recorded in the instrumental record' imply that trends refer to longer time scales than variability. But, since the climate system contains variability at all time scales, which time scale differences do distinguish trends from variability? And consequently: for which time span should trends be calculated? In another definition, trends and variability refer to the same time scale, but to different causes. Trends refer to the portion of climate change that is directly or indirectly attributable to human activities and variability to the portion of climate change at the same time scale that is attributable to natural causes. In this definition trends can only be analyzed in conjunction with formal detection/attribution of anthropogenic influences on climate.

The pragmatic approach followed throughout this thesis is that trends are calculated for any specified period, regardless of the time scale constraints with respect to variability and regardless of the causes. Trends are the simplest component of 'climate change' and only have a statistical meaning in the perspective of the time domain considered. In line with the definition in Houghton *et al.* (2001), trend detection in this thesis refers to the process of demonstrating that climate has changed in some defined statistical sense, without providing a reason for that change (no attribution). This implies that the physical mechanisms behind the detected trends remain unknown for the time being. The calculated trends represent changes that can be due to natural internal processes within the climate system and/or external forcing, which can either be natural (solar irradiance, volcanic aerosols, ozone, etc.) or anthropogenic (greenhouse gases, etc.).

Chapter 3 *(published in J. Climate)* describes the observed trends in the recently defined indices of temperature and precipitation extremes over entire Europe. Also, the inherent limitations are given of trend analysis of extremes in terms of the return periods that can still be handled in order to detect significant trends using the available daily series (~50 yr long).

≡ Can the observed changes in temperature extremes in recent decades be regarded as a fingerprint of anthropogenic climate change? Chapter 4

There is increasing concern that climate extremes may be changing in frequency and intensity as a result of human influences on climate. But natural variability masks anthropogenic trends, and hence creates uncertainty in detecting and attributing the causes of climate change. Extremes are, and always have been, part of natural climate variability. Single extreme events (such as the 2003 summer heat waves in Europe) cannot be simply and directly attributed to anthropogenic climate change if the event in question might have occurred naturally.

To date, detection/attribution of the influence on climate of increasing greenhouse gas emissions by mankind has required investigation of the global mean warming in the past decades or the evolution of temperature averaged over large continental areas (Zwiers and Zhang, 2003). But scientists are making progress in this discipline of climate change research and they soon can say (at some pre-specified confidence level) that past greenhouse gas emissions are likely to have increased the risk of a particular extreme event over its pre-industrial value by X % (Allen, 2003). An anomalously higher number of individual extremes in recent years may then provide a fingerprint of anthropogenic climate change. For this detection/attribution to be successful, a careful selection of parameters to be subject to formal detection/attribution studies is essential.

Chapter 4 *(in press with Int. J. Climatol.)* searches for signals of anthropogenic warming over Europe in the trend patterns of selected measures that describe the variance and skewness of the distribution of daily mean temperature. Comparisons are made between these patterns in the station records of the ECA dataset, the patterns associated with natural variability in the observations, and the patterns of future warming and natural variability as simulated by the National Center for Atmospheric Research Community Climate System Model in the Challenge ensemble experiment (Selten *et al.*, 2004). Candidates are identified for fingerprints of anthropogenic warming in formal climate change detection/attribution studies.

Do the observed changes guide the development of temperature scenarios for our future climate? Chapter 5

The rationale behind studying the changes in daily temperature and precipitation extremes in Europe's climate of the 20th century is the intention to learn from the past in order to predict the future. But extrapolating the observed trends to the future is not allowed as long as it is not known whether the physical mechanisms behind the trends continue to affect the future climate in the same way as in the past. General Circulation Models (GCMs) explicitly account for these mechanisms and within the scientific community GCMs are therefore seen as the only source for reliable predictions of the future climate. GCM predictions are inherently probabilistic, due to uncertainties in forecast initial conditions, representation of key processes within models, and climatic forcing factors. They also involve assumptions about future socio-economic and technological developments that are subject to substantial uncertainty. GCM predictions of future climate are therefore referred to as projections or scenarios.

Unfortunately, GCMs are yet unable to provide scenarios with sufficient detail at the regional and local scale for many applications. The coarse spatial resolution of GCMs (~250km) affects in particular the projections for changes in extremes, because extremes are often smaller in extent than the effective spatial resolution of the GCMs. For this reason, downscaling of GCM results using Regional Climate Models (RCMs) or statistical techniques has become popular nowadays. RCMs typically apply advanced physical models (although with numerous statistical parameterisations of subgrid scale processes) to simulate the climate of a limited area in high temporal and spatial resolution (~50km). Their boundaries are driven by the large-scale output of a GCM. Statistical downscaling techniques rely on statistical relations between local and large-scale weather elements. The projected changes in one element govern the changes in the others. Provided that the relations do not change in a changing climate, statistical downscaling techniques can be used to obtain plausible scenarios with more detail than a GCM or RCM can possibly provide.

Chapter 5 *(published in Climatic Change)* goes beyond the observations of the climate of the past as described in the earlier chapters. It speculates on future changes in extremes and presents a 'what-if' scenario with seasonal details for a Gulf Stream induced cooling of the temperature climate in Western Europe. The scenario is based on the statistical relationship between the wind direction and temperature in observational series from the past. As an example of statistical downscaling, the scenario describes the day-to-day response of surface air temperature in the Netherlands on changes in sea surface temperature of the Atlantic Ocean.

Finally, Chapter 6 summarizes the main findings of this thesis. In particular, it reviews what we have learned as an answer to the key question:

- **How did the extremes of daily surface air temperature and precipitation change in Europe's climate of the 20th century, and what can we learn from this?**

The chapter describes which aspects of our knowledge on past changes in extremes improved, for which purposes the new information is relevant, and which research needs remain.

Abstract – We present a dataset of daily resolution climatic time series that has been compiled for the European Climate Assessment (ECA). As of December 2001, this ECA dataset comprises 199 series of minimum, maximum and/or daily mean temperature and 195 series of daily precipitation amount observed at meteorological stations in Europe and the Middle East. Almost all series cover the standard normal period 1961-90, and about 50% extends back to at least 1925. Part of the dataset (90%) is made available for climate research on CDROM and through the Internet (at www.knmi.nl/samenw/eca).

A comparison of the ECA dataset with existing gridded datasets, having monthly resolution, shows that correlation coefficients between ECA stations and nearest land grid boxes between 1946 and 1999 are higher than 0.8 for 93% of the temperature series and for 51% of the precipitation series. The overall trends in the ECA dataset are of comparable magnitude to those in the gridded datasets.

The potential of the ECA dataset for climate studies is demonstrated in two examples. In the first example, it is shown that the winter (October-March) warming in Europe in the 1976-99 period is accompanied by a positive trend in the number of warm-spell days at most stations, but not by a negative trend in the number of cold-spell days. Instead, the number of cold-spell days increases over Europe. In the second example, it is shown for winter precipitation between 1946 and 1999 that positive trends in the mean amount per wet day prevail in areas that are getting drier and wetter.

Because of its daily resolution, the ECA dataset enables a variety of empirical climate studies, including detailed analyses of changes in the occurrence of extremes in relation to changes in mean temperature and total precipitation.

Daily series of temperature and precipitation observations

2.1 Introduction

There is a growing interest in extreme weather events, like food- or flood-producing rains, droughts, severe heat/cold spells, and gales (Easterling *et al.*, 2000; Houghton *et al.*, 2001). Identification of changes in the occurrence of these events requires accurate, complete and spatially consistent climatic time series with at least daily resolution (Jones *et al.*, 1999; Folland *et al.*, 2000), since only daily and higher-resolution series account for the submonthly time scale nature of extreme weather events. For the USA (Peterson *et al.*, 1997), Canada (Vincent *et al.*, 2000), the former Soviet Union (Razuvaev, personal communication) and Australia (Lavery *et al.*, 1997; Trewin, 1999), datasets of daily temperature and precipitation have been developed. For Europe, such a dataset of daily station observations is not available, so far. Existing climatic datasets with satisfactory spatial coverage of Europe consist of monthly (mean) values only.

Until now, studies on climate extremes that consider Europe have usually had a strong national signature (e.g. Herzog and Müller-Westermeier, 1997; Brunetti *et al.*, 2001), or have had to make use of either a dataset with daily series from a very sparse network of meteorological stations (e.g. eight stations in Moberg *et al.*, 2000) or standardized data analysis performed by different researchers in different countries along the lines of agreed methodologies (e.g. Brazdil *et al.*, 1996; Heino *et al.*, 1999). The only other studies based on networks that cross the national borders refer to the Nordic countries, where the National Meteorological Services developed a dataset that includes some submonthly information, like the absolute daily maximum temperature in a month and the maximum 1-day precipitation amount (Forland *et al.*, 1998; Tuomenvirta *et al.*, 2001). Owing to all these limitations, studies on European climate extremes so far suffer from inherent low spatial coverage, restricted information about daily extremes, and lack of standardization in the definition of extremes. It is obvious, therefore, that systematic analysis of changes in climate extremes requires a single European dataset with daily resolution.

Although Europe has a long history of routine meteorological observations, the compilation of a good quality dataset with daily resolution requires much effort. The main reasons are that archiving, maintaining and dissemination of daily climatic time series are the individual responsibility of more than 30 countries, each with its own storage system and data policy. Yet these multiple sources need to be opened up, as the alternative – collecting the daily data from synoptic observations transmitted on a routine basis over the Global Telecommunication System – has proven to result in a dataset of inferior quality (Folland *et al.*, 1999).

We compiled a comprehensive European dataset consisting of daily temperature (minimum, maximum and/or daily mean) and precipitation series to address a whole range of questions of the type:

- Is the increase in mean temperature, as observed in many areas over the last few decades, accompanied by a decreasing number of frost days and a systematic advance in the start of the thermal growing season?

- Is the change in the number of very wet days more pronounced than the change in total annual precipitation, as in Groisman *et al.* (1999), who hypothesizes that in many areas of the world only the scale of the daily precipitation distribution changes and not its shape, leading to an amplification of the changes in extreme precipitation relative to changes in the total amounts?

The objective of the current chapter is to present and evaluate the contents of the dataset. As outlined in Section 2.2 the dataset was developed in the framework of the European Climate Assessment (ECA) project. Section 2.3 describes the quality control procedure that was applied to the daily series in the dataset; and in Section 2.4 the ECA dataset is compared with existing (gridded) datasets with monthly resolution. Section 2.5 illustrates the potential of the ECA dataset in two examples: one on the relation between winter warming and trends in the number of days belonging to cold or warm spells, and the other on the relation between precipitation changes and wet-day trends.

2.2 Data collection

In the ECA project the temperature and precipitation climate is analyzed for WMO Region VI (Europe and Middle East: Lebanon, Syria, Jordan and Israel), putting particular emphasis on changes in daily extremes.

Figure 2.1 Participating countries in the ECA project that contributed daily climatic time series to the ECA dataset (status December 2001).

The ECA aims at applying uniform analysis methodologies to daily observational series from as many European meteorological stations as possible. So far, 34 countries participate in the ECA project (Figure 2.1; see also the authors' list), and the current version of the ECA dataset (December 2001) contains data from over 200 stations. Figure 2.2 shows the geographical distribution of stations for which daily time series have been collected. In the area covered by the participating countries, there are a total of 109 GCOS Surface Network (GSN) stations, assigned by the WMO for their importance for climate monitoring (Peterson *et al.*, 1997). However, the participants included only 45 of these GSN stations in the ECA dataset.

The ECA project focuses on the 20th century from 1901 to 1999. For this time interval, station series with daily resolution have been collected where possible. Figure 2.3 indicates that most of the series cover the standard normal period 1961–90, and about 50% extends back to at least 1925. The strong decline in recent years in Figure 2.3 is due to the delay in archiving and quality controlling meteorological data in the collaborating institutes and the time needed for collection in the ECA. There are 199 temperature series and 195 precipitation series; 172 stations have both temperature and precipitation series, 27 stations have temperature only, and 23 stations precipitation only. Of the stations having temperature series, 127 stations feature minimum, maximum and daily mean temperature, whereas 64 stations have minimum, maximum temperature only and eight stations daily mean temperature only.

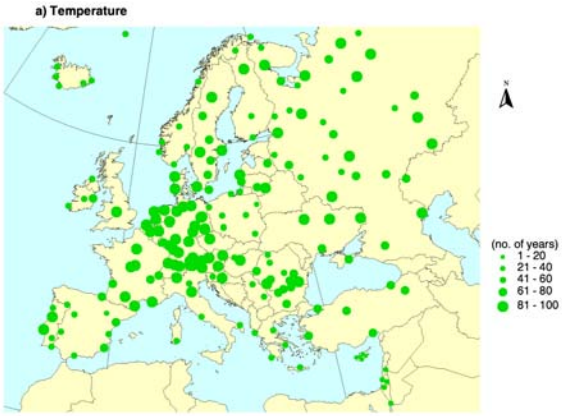


Figure 2.2 Stations with daily temperature (a) and daily precipitation (b) series in the ECA dataset. Station dots are scaled with the length of the time series. For details see the list of station series at www.knmi.nl/samenw/eca

Respecting the data policies of the participants, a selection of the daily series in the ECA dataset (90%) is made available to the public on CDROM and through the Internet (at www.knmi.nl/samenw/eca). The list of station series at the website, which gives the start and end dates of each series in the dataset, includes a column indicating whether the series is available online. At the moment (December 2001), this is true for about 60% of the data; that is, for 114 temperature series and 118 precipitation series from 20 countries. History metadata, like information on station surroundings, measuring instruments, observation times, and algorithms to calculate mean temperature, are not included in the December 2001 version. The release of an elaborated update of the ECA dataset is planned for 2002.

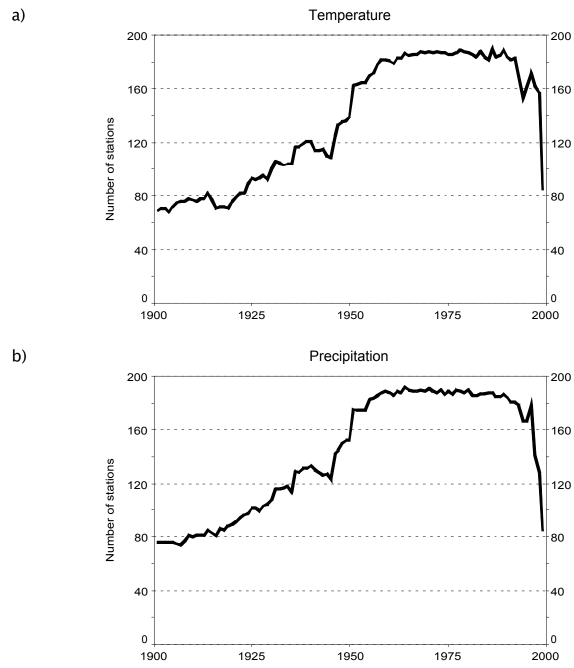


Figure 2.3 Number of stations with daily temperature (a) and daily precipitation (b) series in the ECA dataset.

2.3 Daily time series quality control

Climatic time series typically exhibit spurious (nonclimatic) jumps and/or gradual shifts due to changes in station location, environment, instrumentation or observing practices. In many daily resolution climatic time series, there is also a number of missing observation days. Because the degree of inhomogeneity and incompleteness of a daily resolution series determine the types of extremes analysis that can be undertaken, e.g. see Moberg *et al.* (2000) and Tuomenvirta *et al.* (2000), data quality control is an ongoing activity in the ECA project.

In the December 2001 version of the ECA dataset, the daily series were subjected to a basic quality control procedure only. Every time series is checked for the occurrence of miscoding, like: precipitation < 0 mm; minimum temperature > maximum temperature; nonexistent dates; and erroneous outliers. Although the series have usually undergone routine quality control procedures by the supplying institutes, our additional checks identified a number of days with non-correctable mistakes. Such days are assigned ‘missing values’ in the ECA dataset. Currently, statistical homogeneity tests are being applied to the ECA series. The results of these tests will be included in the updated version of the ECA dataset.

An evaluation of the quality of the ECA dataset was carried out by comparing the 1946–99 annual averages of the daily series with those of existing datasets of lower temporal resolution. Mean temperatures were compared with grid box values in the Jones dataset of land air temperature anomalies (Jones, 1994; Jones *et al.*, 2001); precipitation amounts were compared with grid box values in the Hulme land precipitation dataset (Hulme, 1992; Hulme *et al.*, 1998). These two gridded datasets have been derived from quality-controlled monthly series of stations that may also be on the ECA list. But the percentage of common stations is small: 56% of the ECA stations makes up only about 15% of the European stations in the gridded datasets (Jones, personal communication).

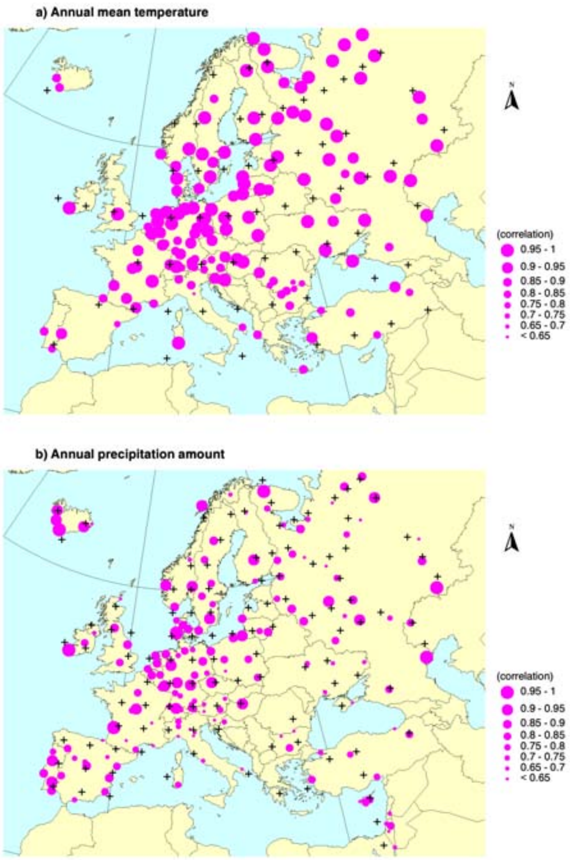


Figure 2.4 Correlation coefficients (1946–99) between annual mean temperature at stations in the ECA dataset and nearest land grid boxes of the Jones temperature dataset (a) and between annual precipitation amount in the ECA dataset and nearest land grid boxes of the Hulme precipitation dataset (b). The centres of the grid boxes used in the comparison with ECA are indicated.

Furthermore, sea surface temperature anomalies were merged into the grid box temperatures around coastal areas, whereas the ECA dataset contains station observations only.

For each ECA station, Figure 2.4a shows the correlation between the ECA temperature series and the temperature series of the nearest land grid box in the Jones dataset (1946–99). For the ECA stations where daily mean temperature is missing, but minimum and maximum temperatures are available, the average of minimum and maximum temperature was used. Time series with more than 20% of missing years were excluded from the analysis. In terms of the annual means, the daily ECA temperature series fit in well with the monthly series of the temperature grid. For 93% of the stations the correlation coefficient exceeds 0.8. A north-south gradient in correlation coefficients is apparent in Figure 2.4a. Precipitation correlation coefficients (Figure 2.4b) exceed 0.8 at only 51% of the stations, despite of the higher density of the precipitation grid (2.5°lat × 3.75°lon) compared with the temperature grid (5°lat × 5°lon). This lower correspondence results from the inherent lower spatial coherence of precipitation fields. A northwest-southeast gradient in correlation coefficients is apparent in Figure 2.4b.

Table 2.1 presents the European averages of temperature and precipitation trends in the ECA dataset and the gridded datasets. The 1946–99 period, its 30 yr subperiod 1946–75, and the remaining subperiod 1976–99 are considered separately.

Table 2.1 Comparison of average trends in annual mean temperature and annual precipitation amount for stations in the ECA dataset and corresponding land grid boxes of the Jones and Hulme datasets. The trends for ECA stations were calculated from the time series obtained by first averaging the ECA station series in each Jones/Hulme grid box, and then averaging the series over these grid boxes (see text). The numbers *n* in the table header denote the number of grid boxes used in the comparison. The 95% confidence intervals are shown in parentheses. For none of the (sub)periods did the trends for ECA stations and land grid boxes differ significantly at the 5% level (paired *t* test).

	Temperature trend (°C/decade); n = 41	
	ECA stations	Land grid boxes (Jones)
1946–99	0.04 (−0.04 – 0.08)	0.03 (−0.05 – 0.11)
1946–75	−0.04 (−0.24 – 0.16)	−0.03 (−0.21 – 0.15)
1976–99	0.42 (0.10 – 0.74)	0.38 (0.08 – 0.68)
	Precipitation trend (mm/decade); n = 97	
	ECA stations	Land grid boxes (Hulme)
1946–99	11.1 (5.5 – 16.8)	6.4 (1.3 – 11.4)
1946–75	16.1 (2.1 – 30.1)	4.7 (−8.4 – 17.8)
1976–99	−2.8 (−20.0 – 14.4)	−0.2 (−15.0 – 14.6)

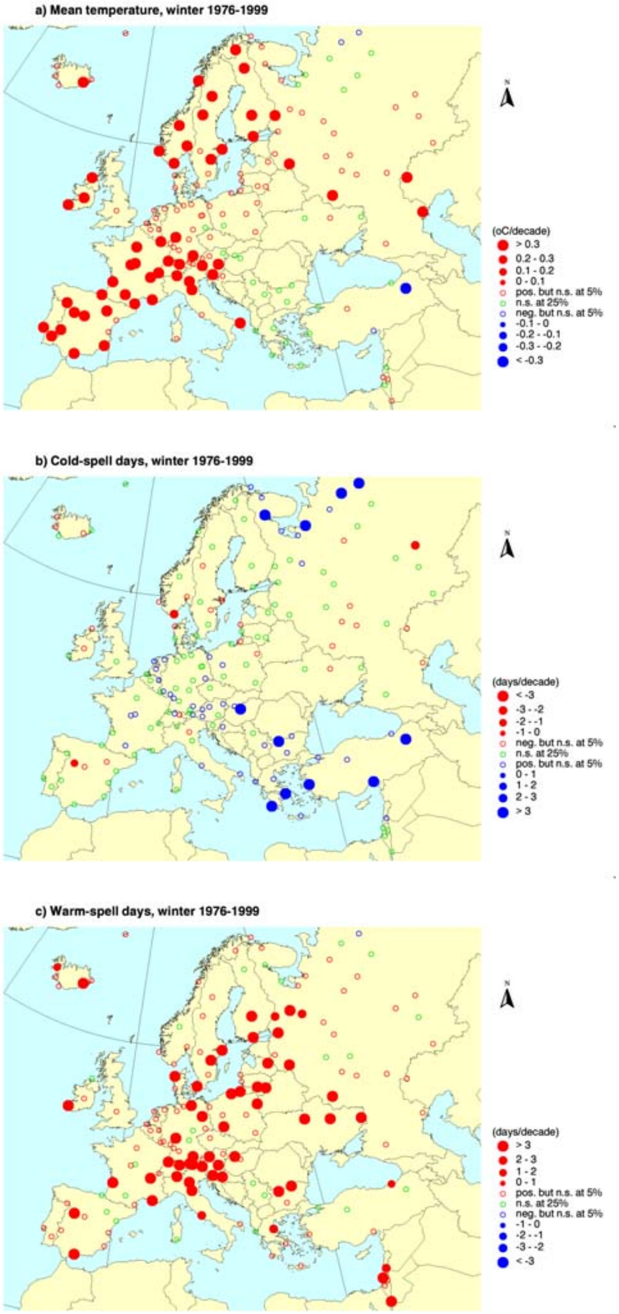


Figure 2.5 Trends in winter (October–March) mean temperature (a), number of cold-spell days (b) and number of warm-spell days (c) between 1976 and 1999. For trends significant at the 25% level, but not at the 5% level (Student's *t* test), only the sign of the trend is given and not its magnitude. Colour coding is applied: blue corresponds to cooling, red to warming. Green is used for trends that are not significant at the 25% level. No warming effect is seen in the number of cold-spell days.

European trends were calculated from grid box weighted average series, obtained as follows. For each Jones/Hulme grid box containing one or more ECA stations, the annual series of all stations in that grid box were arithmetically averaged. The resulting grid box series (41 for temperature and 97 for precipitation) were then averaged to derive the European ECA series. The corresponding European series for the gridded datasets were derived by arithmetically averaging the annual series of the same grid boxes as for ECA. Grid boxes without ECA stations, grid box series with more than 20% of missing years in a (sub)period, and years in grid box series for which either the ECA or gridded value were missing were omitted. Also, the years in the European series whose values were based on fewer than 80% of the total number of grid boxes considered were assigned missing values. Table 2.1 shows that an agreement exists between the average magnitude of trends at ECA stations and land grid boxes.

2.5 Two examples of dataset applications

The potential of the ECA dataset is demonstrated in two examples, highlighting the benefits of the daily resolution of the climatic time series.

In the first example, the effect of increasing winter (October–March) temperature on the number of days belonging to cold or warm spells is explored for the rapidly warming period 1976–99. Analogous to the cold/warm days definition of Jones *et al.* (1999) and Horton *et al.* (2001), we defined cold/warm spells at a given site as periods of at least six consecutive days with daily mean temperatures below/above the lower/upper 10th percentile of the temperature distribution for each calendar day in the 1961–90 standard normal period. These calendar-day specific percentiles were calculated from 5 day windows centred on each calendar day. This gives a total sample size of 30 yr × 5 days = 150 for each calendar day. The length of each spell is expressed in a single index that comprises the total number of days in that spell. For instance, a cold spell lasting 6 days yields 6; a cold spell of 7 days yields 7, etc., whereas a cold period of 5 days or less yields zero. The cold spell index of a winter is the sum of the indices of all spells in that winter, which equals the total number of days per winter that are member of any cold spell in that winter.

Trends in the number of cold/warm-spell days and in mean temperature were calculated for each individual ECA station and for Europe as a whole. Here, the European trends were calculated from European time series obtained by arithmetically averaging all 168 station series with no more than 20% of missing years in the 1976–99 period. In each year of the European series more than 80% of the total number of stations contributed to the European average.

The trends in winter mean temperature and cold/warm-spell days for stations in the ECA dataset between 1976 and 1999 are depicted in Figure 2.5. The figure shows that the winter warming is accompanied by an increase in the number of warm-spell days at nearly every ECA station. On the other hand, at less than 10% of the stations the winter warming is accompanied by a decrease in the number of cold-spell days. Averaged over all ECA stations, the increase in the number of warm-spell days of 3.0 days/decade is even accompanied by an increase in the number of cold-spell days of 0.2 days/decade, rather than a decrease in the number of cold-spell days (Table 2.2). A similar, though weaker, asymmetric behaviour of warm- and cold-spell days is found for summer warming, with a 5 times larger trend in the number of warm-spell days than cold-spell days. Given the small climatological-mean number of cold- and warm-spell days per season in the 1961–90 standard normal period, the trends are really large and show that changes in the outbreak of cold and warm spells are not trivially related with changes in mean temperature. This non-triviality was also noted by Walsh *et al.* (2001) in the relation between cold outbreaks and temperature in the National Centers for Environmental Prediction reanalysis data for 1948–99. Unraveling these kinds of relations is only possible with the aid of daily resolution observations, like those collated in the ECA dataset.

The second example deals with precipitation changes. Figure 2.6 shows that the station trends in the winter (October–March) total precipitation amount between 1946 and 1999 should not only be attributed to changes in the number of wet days ≥ 1 mm.

Table 2.2 Average change in summer (April–September) and winter (October–March) mean temperature, number of warm-spell days and number of cold-spell days for stations in the ECA dataset between 1976 and 1999. The changes refer to increases in temperature and number of warm-spell days, but to *decreases* in number of cold-spell days, so that positive numbers refer to warming throughout. The 95% confidence intervals are shown in parentheses. The climatological means (1961–90) for the number of warm/cold-spell days are shown in square brackets. All numbers were calculated from European time series obtained by arithmetically averaging over 168 ECA stations. The European trend in annual temperature according to this method is 0.45 (0.15 – 0.75); cf. Table 2.1.

	Temperature increase (°C/decade)	Warm-spell days increase (days/decade)	Cold-spell days *decrease* (days/decade)
Summer (April–September)	0.46 (0.24 – 0.68)	3.6 (2.5 – 4.6) [2.9]	0.7 (0.0 – 1.4) [1.9]
Winter (October–March)	0.47 (0.01 – 0.93)	3.0 (1.1 – 4.8) [2.4]	–0.2 (–1.9 – 1.4) [3.8]

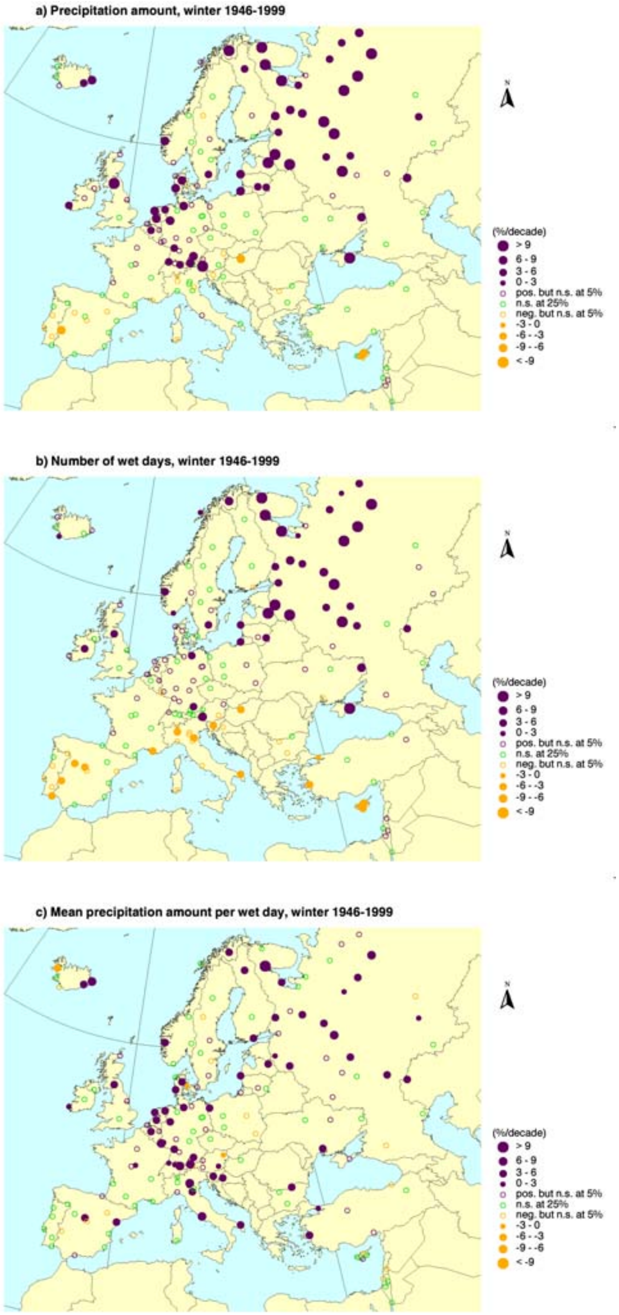


Figure 2.6 Trends in winter (October–March) precipitation amount (a), number of wet days ≥ 1 mm (b) and mean precipitation amount per wet day (c) between 1946 and 1999. Precipitation amount and number of wet days were calculated as percentage anomalies with respect to the 1961–90 means. Statistical significance is as in Figure 2.5. Colour coding is applied: yellow corresponds to drier conditions, violet to wetter conditions. Green is used for trends that are not significant at the 25% level. In contrast to the total winter precipitation amounts and the number of wet days, the mean precipitation amount per wet day increased throughout Europe.

For most stations the percentage changes in total amounts are different from the percentage changes in the number of wet days. This is caused by the accompanying changes in the mean precipitation amounts per wet day, that is, the average rain intensity per wet day. This rain intensity index predominantly increases over Europe, both at stations with positive trends and at stations with negative trends in total winter precipitation amount.

2.6 Conclusions and discussion

With a total of 199 station series of daily temperature and 195 station series of daily precipitation in the ECA dataset, we accomplished an important initial step towards the development of a high-resolution dataset for Europe and the Middle East that is suitable for the analysis of changes in climate extremes and allows for a variety of other climate studies. Although this daily dataset is made available for climate research on CDROM and through the Internet, it cannot yet be guaranteed that every temperature and precipitation series in the December 2001 version will be sufficiently homogeneous in terms of daily mean and variance for every application. It is clear that, at some stations, changes in instruments have led to discontinuities. It is also clear that the time series of some city stations include urban heating effects, as well as discontinuities resulting from station relocations to out-of-town airports. At the moment, there are no established methods to determine and adjust such inhomogeneities in daily resolution time series, although some techniques are under development, e.g. see Vincent *et al.* (2000) and Brandsma (2001). Within the ECA project, Wijngaard and Klein Tank (2001) explored the application to the ECA dataset of established statistical techniques that have been developed for homogeneity testing of lower time resolution series, e.g. see Szalai *et al.* (1999).

In this study, we have restricted ourselves to a first quality assessment of the ECA dataset by comparing the temperature and precipitation series to existing gridded datasets of monthly resolution. It may be argued that this comparison is biased, because the gridded datasets are based on monthly series from stations that partly overlap the list of ECA stations. However, the fact that the ECA stations make up only a small part of the gridded datasets and that the correlation coefficients between the annual values of the ECA series and the grid box series are generally higher than 0.8 strengthens our confidence in the ECA dataset. Besides, in terms of the average magnitude of temperature and precipitation trends, there is also a fairly good correspondence between the ECA dataset and the gridded datasets. Deviations between individual station trends and nearby grid box trends may correspond with suspect station series, but may also represent local climate anomalies that cannot be captured by the grid. However, it is not *a priori* clear that the gridded series can act as a homogeneous baseline. Differences might also originate from flaws in the gridded datasets (Moberg and Alexandersson, 1997).

The potential of the ECA dataset is demonstrated in two examples: (1) the observed counter-intuitive trends in the number of cold-spell days over Europe that accompany the warming trends in mean winter temperature between 1976 and 1999; and (2) the predominant increasing trends in mean precipitation amount per wet winter-day in Europe, inclusive of regions where the total precipitation amounts decrease (1946–99). These two examples illustrate that there need not be a trivial relation between the change in the mean (or total) of a climatic parameter and the change in other statistical characteristics, such as the extremes. The ECA dataset with daily time resolution facilitates the consistent analysis of such relationships for Europe and the Middle East.

The first study on climate extremes based on the ECA dataset (Frich *et al.*, 2002) proves that the dataset serves its initial purpose of climate change detection and analysis of submonthly climate extremes. In order to make the dataset fully applicable for systematic monitoring of changes in climate extremes, regular updates are required, together with concentrated and enhanced efforts on data quality control and daily time series homogenization.

Abstract – Trends in indices of climate extremes are studied on the basis of daily series of temperature and precipitation observations from more than 100 meteorological stations in Europe. The period is 1946-99, a warming episode. Averaged over all stations, the indices of temperature extremes indicate 'symmetric' warming of the cold and warm tails of the distributions of daily minimum and maximum temperature in this period. However, 'asymmetry' is found for the trends if the period is split into two subperiods. For the 1946-75 subperiod, an episode of slight cooling, the annual number of warm extremes decreases, but the annual number of cold extremes does not increase. This implies a reduction in temperature variability. For the 1976-99 subperiod, an episode of pronounced warming, the annual number of warm extremes increases 2 times faster than expected from the corresponding decrease in the number of cold extremes. This implies an increase in temperature variability, which is mainly due to stagnation in the warming of the cold extremes.

For precipitation, all Europe-average indices of wet extremes increase in the 1946-99 period, although the spatial coherence of the trends is low. At stations where the annual amount increases, the index that represents the fraction of the annual amount due to very wet days gives a signal of disproportionate large changes in the extremes. At stations with a decreasing amount, there is no such amplified response of the extremes.

The indices of temperature and precipitation extremes in this study were selected from the list of climate change indices recommended by the World Meteorological Organization - Commission for Climatology (WMO-CCL) and the Research Programme on Climate Variability and Predictability (CLIVAR). The selected indices are expressions of events with return periods of 5-60 days. This means that the annual number of events is sufficiently large to allow for meaningful trend analysis in ~50 yr time series. Although the selected indices refer to events that may be called 'soft' climate extremes, these indices have clear impact relevance.

Trends in daily temperature and precipitation extremes

3.1 Introduction

Surface air temperatures in most European regions have increased during the 20th century (Houghton *et al.*, 2001). In line with the characteristics of global temperature rise (Jones *et al.*, 1999a; Karl *et al.*, 2000), the European rate of change has been highest in the last quarter of the century (Chapter 2). The warming is projected to continue and is likely to be accompanied by changes in extreme weather and climate events (Houghton *et al.*, 2001). Yet, little is known quantitatively about the nature of these changes. In this context, it is relevant to learn how the past warming affected the occurrence of temperature extremes, or whether the past warming was accompanied by detectable changes in precipitation extremes. Studies on these issues are receiving increased attention in the last few years (Easterling *et al.*, 2000; Meehl *et al.*, 2000).

Although changes in extreme temperature and precipitation events have been analyzed for individual European countries and stations (see, e.g. Forland *et al.*, 1998; Tuomenvirta *et al.*, 2000; Moberg *et al.*, 2000; Brunetti *et al.*, 2000; Osborn *et al.*, 2000; Yan *et al.*, 2002), a coherent picture for Europe as a whole is lacking. The main reason is the limited spatial coverage of the high time-resolution European datasets used in such studies. The second reason is that until recently no accepted standardization existed in the definitions of climate extremes, which has made it difficult to compare the results of different studies. This situation has changed now. The objective of the present study is to investigate the trends in some of the recently defined (Peterson *et al.*, 2001) indices of temperature and precipitation extremes using the European Climate Assessment (ECA) daily dataset (Chapter 2).

The indices of temperature and precipitation extremes considered in the present study were selected from the list of indices for surface data recommended by the joint Working Group on Climate Change Detection of the World Meteorological Organization – Commission for Climatology (WMO–CCL) and the Research Programme on Climate Variability and

Predictability (CLIVAR; Peterson *et al.*, 2001). The selected indices refer to extremes with return period typically on the order of weeks, rather than to once-in-a-lifetime events. This choice ensures that the annual number of extremes is sufficiently high to allow for meaningful trend analysis in series of ~50 yr length and therefore in the series of the ECA dataset.

About half of the indices considered in the present study are expressions of anomalies relative to local climatology in the standard-normal period 1961–90, enabling comparisons between stations in different countries and regions. Using this type of indices, we compared trends in Europe-average cold temperature extremes with trends in Europe-average warm temperature extremes to investigate whether they suggest equal warming in both tails of the temperature distributions. For precipitation, we compared trends in extreme precipitation with trends in total amount to investigate whether extremes contributed disproportionately to overall wetting or drying. Some preliminary results of our study have been included in Frich *et al.* (2002; see also Houghton *et al.*, 2001), who provides trend estimates for temperature and precipitation extremes in about half the global land areas.

Section 3.2 describes the criteria for station selection from the daily ECA dataset and Section 3.3 the selection of indices from the Peterson *et al.* (2001) list. The procedures for estimating trend values for Europe-average indices, comparing cold to warm extremes, and comparing precipitation extremes to total amount are outlined in Section 3.4. Section 3.5 presents the observed trends for temperature and precipitation. In Section 3.6 the results are discussed. Section 3.7 summarizes the conclusions.

3.2 Analysis period and data selection

The ECA dataset (Chapter 2) comprises the period 1901–99, but the data coverage of Europe in the first part of the century is not adequate for our purpose. This restricts the period of analysis to 1946–99 (and its subperiods 1946–75 and 1976–99), for which 195 daily temperature series and 202 daily precipitation series are available in ECA (status February 2003). Not all these ECA series were used in the present study. First, the time series that received the lowest ranking in our homogeneity test (Wijngaard *et al.*, 2003) were excluded from the analysis, except for some series that were retained based on the metadata support and recommendations provided by the participants in the ECA project. Second, series with more than 20% missing or incomplete years in the analysis period or one of its subperiods were excluded from the analyses. The criterion applied for incompleteness is less than 361 observation days per year. Such a strict criterion is needed because some indices are critically dependent on the serial completeness of the data.

A total of 86 daily temperature series and 151 daily precipitation series from the ECA dataset survived all selection criteria. Only these station series (Figure 3.1) were used in the present study to calculate the indices.

Apart from trends for each individual ECA station, trends were also calculated for Europe as a whole. The European trends were obtained from Europe-average indices series calculated as the arithmetic average of the annual indices values at all 86 temperature stations or 151 precipitation stations. Annual values in the Europe-average indices series based on less than 75% of the stations were omitted when calculating the European trends.

Because of the nonuniform spatial distribution of ECA stations over Europe, areas with a higher density of stations are overrepresented in the Europe average.

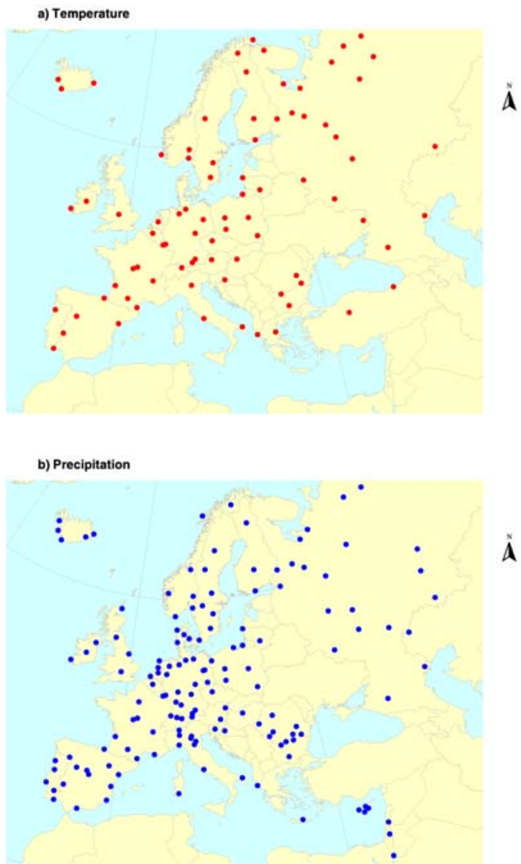


Figure 3.1 ECA stations with daily time series of temperature (a) and precipitation (b) used in this study.

Proper area weighing methods would include gridding of time series. Such methods are not applied in this study. Based on the comparison for mean temperature in Chapter 2, we estimate that the trends in our Europe-average indices series would agree within 10% with trends in area-weighed indices series.

3.3 Indices of climate extremes

From the internationally agreed WMO–CCL/CLIVAR list of over 50 climate change indices available online at www.knmi.nl/samenw/eca (Peterson *et al.*, 2001; see also Folland *et al.*, 1999; Nicholls and Murray, 1999; Jones *et al.* 1999b), we selected a set of 13 indices of climate extremes. Of this set, six indices refer to temperature and seven to precipitation (Table 3.1). Although the definitions allow for seasonal or monthly breakdown, subannual specification is not considered here. The indices selected in the present study are expressed in annual values Y_j, $j = 1,...,N$ with the index j referring to the year and N the length of the period covered by the station series. The temperature indices describe cold extremes as well as warm extremes. The precipitation indices describe wet extremes only. Most of the indices are defined in terms of counts of days crossing thresholds, either absolute (fixed) thresholds or percentile (variable) thresholds. These day-count indices refer to annual numbers of exceedences X_j, $j = 1,...,N$.

Annual day-count indices based on percentile thresholds are expressions of anomalies relative to the local climate. Consequently, the value of the thresholds is site specific. Such indices allow for spatial comparisons, because they sample the same part of the temperature and precipitation (probability density) distributions at each station. Annual day-count indices based on absolute thresholds are less suitable for spatial comparisons of extremes than those based on percentile thresholds. The reason is that, over an area as large as the European continent, annual day-count indices based on absolute thresholds may sample very different parts of the temperature and precipitation distributions. This implies that in another climate regime, the variability in such indices readily stems from another season. For instance, year-to-year variability in frost-day counts (days with minimum temperature < 0°C) relates to the variability in the spring and autumn temperatures for the northern part of Europe, whereas in the southern part of Europe annual variability in frost-day counts is determined by winter temperature variability (Heino *et al.*, 1999). Likewise, the threshold of 25°C in the definition of summer days (days with maximum temperature > 25°C) samples variations in summer temperatures in the north and variations in spring and autumn temperatures in the south.

Table 3.1 Definitions of the indices of cold and warm temperature extremes and the indices of precipitation extremes used in this study. The abbreviations and definitions follow the standardization of the CCL/CLIVAR Working Group on Climate Change Detection (Peterson *et al.*, 2001). See also online at www.knmi.nl/samenw/eca. In the present study only annually specified indices are considered; no separations are made to season or month.

		Extreme indices
Indices of cold temperature extremes:		
FD	Frost days (absolute threshold)	Number of days (per year/season/month) with minimum temperature below 0°C. Let $TN_{i,j}$ be the daily minimum temperature at day i in period j. Then counted are the days with $TN_{i,j} < 0°C$
TN10% TX10%	Cold nights (percentile threshold) Cold days (percentile threshold)	Number of days (per year/season/month) with temperature below a site- and calendar-day specific threshold value, calculated as the calendar-day 10th percentile of the daily temperature distribution in the 1961-90 baseline period. TN10% refers to low daily minimum temperature events (cold nights) and TX10% to low daily maximum temperature events (cold days). Let $TN_{i,j}$ be the daily minimum temperature at day i in period j and let TN_k10 be the 10th percentile for calendar day k calculated for a 5 day window centred on each calendar day in the 1961-90 period. Then counted are the days with $TN_{i,j} < TN_k10$. The cold days are calculated analogously from the daily maximum temperatures.
Indices of warm temperature extremes:		
SU	Summer days (absolute threshold)	Number of days (per year/season/month) with maximum temperature above 25°C. Let $TX_{i,j}$ be the daily maximum temperature at day i in period j. Then counted are the days with $TX_{i,j} > 25°C$
TN90% TX90%	Warm nights (percentile threshold) Warm days (percentile threshold)	Number of days (per year/season/month) with temperature above a site- and calendar-day specific threshold value, calculated as the calendar-day 90th percentile of the daily temperature distribution in the 1961-90 baseline period. TN90% refers to high daily minimum temperature events (warm nights) and TX90% to high daily maximum temperature events (warm days). Let $TN_{i,j}$ be the daily minimum temperature at day i of period j and let TN_k90 be the 90th percentile for calendar day k calculated for a 5 day window centred on each calendar day in the 1961-90 period. Then counted are the days with $TN_{i,j} > TN_k90$. The warm days are calculated analogously from the daily maximum temperatures.
Indices of precipitation extremes:		
RX1day RX5day	Highest 1-day and 5-day precip. amount (absolute extreme)	Maximum (annual/seasonal/monthly) precipitation sums for 1-day intervals (RX1day) and 5-day intervals (RX5day). Let $R_{i,j}$ be the daily precipitation amount for day i in period j. Then maximum 1-day values for period j are $RX1day_j = \max(R_{i,j})$. Let $R_{i,j}5$ be the precipitation amount for the consecutive five-day interval i, i-1, i-2, i-3, i-4 in period j. Then maximum 5-day values for period j are $RX5day_j = \max(R_{i,j}5)$.
R10mm R20mm	(Very) heavy precip. days (absolute threshold)	Number of days (per year/season/month) with precipitation amount ≥ 10 mm (R10mm) and ≥ 20 mm (R20mm). Let $R_{i,j}$ be the daily precipitation amount for day i in period j. Then counted are the days with $R_{i,j} \geq 10$ mm or $R_{i,j} \geq 20$ mm.
R75% R95%	Moderate and very wet days (percentile threshold)	Number of days (per year/season/month) with precipitation amount above a site specific threshold value for moderate and very wet days, calculated as the 75th (R75%) and 95th (R95%) percentile of the distribution of daily precipitation amounts at days with 1 mm or more precipitation in the 1961-90 baseline period. Let $R_{w,j}$ be the daily precipitation amount at wet day w (precipitation ≥ 1 mm) in period j and let R_n75 be the 75th percentile of precipitation at wet days in the 1961-90 baseline period. Then counted are the moderate wet days with $R_{w,j} > R_n75$. The very wet days are calculated likewise.
R95%tot	Precipitation fraction due to very wet days (percentile threshold)	Fraction of (annual/seasonal/monthly) precipitation amount due to very wet days (R95%-days). Let R_j be the sum of daily precipitation amount for period j and let $R_{w,j}$ be the daily precipitation amount at wet day w in period j and R_n95 the 95th percentile of precipitation at wet days in the 1961-90 baseline period. Then $R95\%tot_j$ is determined as the sum of $R_{w,j}$ at days with $R_{w,j} > R_n95$ divided by R_j.

In the present study, the values of the percentile thresholds were determined empirically from the observed station series in the climatological standard-normal period 1961–90. This was done as follows. For precipitation, the percentiles were calculated straightforward from the sample of all wet days in the series. For temperature, the percentiles were calculated from 5-day windows centred on each calendar day to account for the mean annual cycle. This yields a total sample size of 30 yr × 5 days = 150 for each calendar day. The procedure ensures that extreme temperature events, in terms of crossings of percentile thresholds, can occur with equal probability throughout the year. The same property holds for extreme precipitation events, because it was verified that the seasonal dependency in the chosen precipitation percentiles (Table 3.1) and in the occurrence of wet days can be neglected.

Percentile thresholds were also used by Jones *et al.* (1999b) and Horton *et al.* (2001) for determining the frequencies of temperature extremes. Contrary to our approach, their method accounts parametrically rather than empirically for the annual cycle of thresholds. As argued by Yan *et al.* (2002), the effect on the indices in Table 3.1 of using either empirical methods for percentile calculations or parametric methods relying on distributions is small.

The index R95%tot, that is, the fraction of annual precipitation amount due to very wet days, has been introduced in our study to explore the supposed amplified response of the extreme precipitation events relative to the change in total amount (Groisman *et al.* 1999; see also Houghton *et al.*, 2001). Indices like R95%tot are suitable to analyse such changes in the tail of the precipitation distribution, as they implicitly take into account the trends in the total amount. Similar to the indices used by Osborn *et al.* (2000), but unlike many other indices of precipitation extremes (Haylock and Nicholls, 2000), the R95%tot index is not sensitive to changes in the number of wet days.

Table 3.2 lists the climatological values of the 13 temperature and precipitation indices over the baseline period 1961–90 for five station series from different areas of Europe and for the Europe-average series. For the day-count indices, the corresponding mean return period T_{ret} is given by:

$$T_{ret} = 365/\overline{X} \tag{3.1}$$

if T_{ret} is expressed in days and the climatological value $\overline{X}$ of X_j in number of days per year. Table 3.2 shows that the 13 indices highlight extremes with T_{ret} roughly between 5 and 60 days. For the temperature indices based on the 10th- and 90th-percentile thresholds, a constant mean return period of 10 days is found throughout the table, as a direct consequence of the definitions. On the other hand, the values of the precipitation indices that are based on the 75th- and 95th-percentile thresholds are not constant in the table. The reason is that for precipitation the percentile thresholds were determined from the distribution of the daily precipitation amounts at wet days only, rather than all days. Accordingly, the mean return periods of R75% and R95% are 4 and 20 wet days, respectively. But since the number of wet days is always lower than 365 and is station dependent, mean return periods of 4 and 20 wet days translate into higher and station-dependent mean return periods if expressed in all (wet + dry) days.

The climatological values in Table 3.2 are valid for the standard-normal period 1961–90. Another baseline period would have resulted in different numbers for all indices, except for the temperature indices that are based on percentile thresholds. However, the trends in all indices are to a first approximation unaffected by the choice of the baseline period. This was verified by comparing the trend results obtained from the 1931–60 period with the results obtained from the 1961–90 period.

Table 3.2 Mean indices values over the 1961-90 baseline period for five stations from different areas of Europe and for the Europe-average series. For the day-count indices, the corresponding mean return period (in days) is also given. RX1day and RX5day are in mm and R95%tot is in % of total annual precipitation. The bottom rows give the averages of mean temperature TG, diurnal temperature range DTR, annual precipitation amount RR and annual number of wet days RR1.

	Reykjavik (Iceland)	Elatma (Russia)	De Bilt (Netherlands)	Salamanca (Spain)	Larissa (Greece)	Europe average
Indices of temperature extremes*:						
FD	123 (3)	162 (2)	66 (6)	71 (5)	42 (9)	67 (5)
TN10%	36 (10)	36 (10)	36 (10)	36 (10)	36 (10)	36 (10)
TX10%	36 (10)	36 (10)	36 (10)	36 (10)	36 (10)	36 (10)
SU	0 (–)	39 (12)	18 (30)	91 (4)	143 (3)	64 (6)
TN90%	36 (10)	36 (10	36 (10)	36 (10)	36 (10)	36 (10)
TX90%	36 (10)	36 (10)	36 (10)	36 (10)	36 (10)	36 (10)
Indices of precipitation extremes*:						
R10mm	19 (20)	14 (28)	22 (19)	10 (40)	11 (35)	21 (18)
R20mm	3 (171)	3 (177)	5 (102)	2 (218)	3 (141)	7 (55)
R75%	37 (10)	27 (14)	33 (12)	17 (23)	14 (28)	26 (14)
R95%	7 (56)	6 (83)	7 (66)	3 (134)	3 (176)	5 (71)
RX1day	31 mm	35 mm	33 mm	30 mm	53 mm	44 mm
RX5day	59 mm	57 mm	64 mm	52 mm	74 mm	77 mm
R95%tot	18%	20%	18%	20%	24%	20%
TG	4.4°C	4.3°C	9.4°C	11.7°C	15.7°C	7.6°C
DTR	5.1°C	8.5°C	8.0°C	12.4°C	12.7°C	7.7°C
RR	799mm	613mm	820mm	391mm	419mm	741mm
RR1	148	110	134	69	57	103

*The indices defined in terms of counts of days crossing thresholds are expressed in annual number of days.

Trends in an index of temperature and precipitation extremes Y_j were calculated by an ordinary least squares fit of the model:

$$Y_j = a + 0.1\,bj + e_j \qquad j = 1,\ldots,N \tag{3.2}$$

where Y_j is the value of the index at year j and e_j is a disturbance term (residual) with mean zero. The regression coefficient b gives the change per decade. Trend significance was tested using a Student's t test. As recommended by Nicholls (2001), not only the test results at the 5% significance level are presented but also the results at the 25% level. For the European trends, 95% confidence intervals were also calculated. The trend calculations were performed for two consecutive subperiods (1946–75 and 1976–99), as well as for the entire 54 yr period 1946–99. It should be noted that the simple weighed average of the trends for the two subperiods is not necessarily the same as the trend for the entire period. The reason is that the trend for the entire period is also affected by the average values of the indices in the two subperiods.

The probability of detecting trends in time series depends on the trend magnitude, the record length, and the statistical properties of the variable of interest, in particular the variance. For uninterrupted event-count records X_j with independence between successive events, and with the use of Eq. (3.1), a simple expression can be derived for the signal-to-noise ratio (see Appendix):

$$C \equiv \frac{b}{\sigma_{\hat{b}}} \approx \frac{1}{10\sqrt{12}}\,\frac{b}{\overline{X}}\left(T_{ret}/\,365\right)^{-1/2} N^{3/2} \tag{3.3}$$

Here the factor 10 arises because we choose to express $1/\overline{X}$ and the series length N in years, but b in an increase per decade (Eq. (2)); the factor 365 is because T_{ret} is in days (Eq. (1)).

The relative trend $b_q/\overline{X}$ that can be detected at a given significance level (e.g. 5%) with a probability of q%, follows from:

$$\frac{b_q}{\overline{X}} = C_q\,\frac{\sigma_{\hat{b}}}{\overline{X}} \approx 10\sqrt{12}\;C_q\left(T_{ret}/\,365\right)^{1/2} N^{-3/2} \tag{3.4}$$

with C_q the value of C that is required to find a significant trend in q% of the cases for the given level (Buishand *et al.* 1988). The dependence of C_q on the detection probability and significance level is shown in Figure 3.2a.

Figure 3.2b shows $b_q/\overline{X}$ for q = 80% as a function of N for various return periods T_{ret} (significance level 5%). As $b_q/\overline{X}$ is proportional to $T_{ret}^{1/2}$, equal detection probability at the given significance level for events with return period of 365 days instead of 10 days would require a trend that is 6 times larger. As the required trend value is proportional to $N^{-3/2}$, a 3 times larger trend is needed in a 24 yr record than in a 54 yr record to reach the same detection probability. Note that the trend values in Eq. (3.4) should be regarded as lower estimates, because of the presence of serial correlation in the event-count records. This leads to a higher standard deviation of the least squares trend estimate (fewer independent extreme events). The values from Eq. (3.4) and Figure 3.2a correspond well with those in the simulation experiment of Frei and Schär (2001), provided that the relative trend in the entire series is less than 50%.

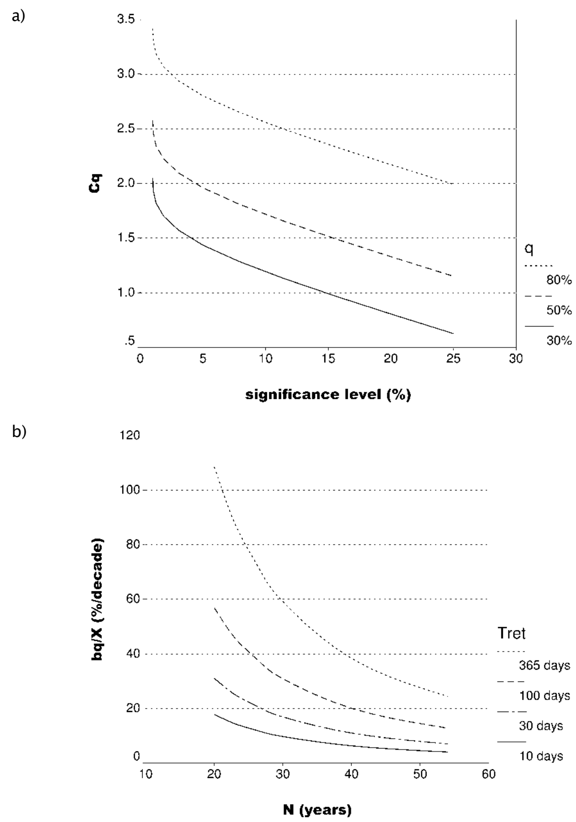


Figure 3.2 Dependence of signal-to-noise ratio C_q on the significance level for detection probability 30%, 50% and 80% (a); Relation between the relative trend $b_q/\overline{X}$ required for 80% detection probability (significance level 5%) and series length N , for extreme events with average return period T_{ret} =10, 30, 100, and 365 days, according to Eq. (3.4) (b). Here, $\overline{X}$ is the climatological value in a baseline period. As independence between successive extremes is assumed, the required trends represent lower estimates.

Taking the average over a number of station series S leads to improved trend detectability due to decreased variability in an average series. If there is no spatial correlation between the station series, Eq. (3.4) can be rewritten as:

$$\frac{b_q}{\overline{X}} \approx 10\sqrt{12}\; C_q \left(T_{ret}/\,365\right)^{1/2} N^{-3/2} S^{-1/2} \tag{3.5}$$

As the required trend value is proportional to $S^{-1/2}$, averaging over 86 (151) temperature (precipitation) series in Europe, as in the present study, would reduce the required trend value to reach the same detection probability by a factor of 9 (12). Note that for the temperature indices in particular, the actual reduction in the detectable trend is smaller than indicated by Eq. (3.5) because of spatial correlation.

In a second step of the analysis of temperature extremes, we investigated to what extent the trends in the percentile-based Europe-average temperature indices can be understood by a shift in the daily temperature distribution without a change in variance or skewness. The procedure illustrated in Figure 3.3 was followed.

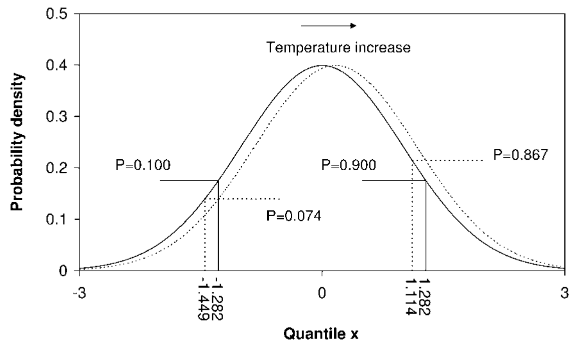


Figure 3.3 Effect of a change only in the mean and not in the variance of a Gaussian temperature distribution for the percentile-based temperature indices. The percentile-based indices TN10%, TX10%, TN90%, and TX90% have a value of 36 in a period with unchanging climate. Given the cumulative distribution function $P(x)$ of a Gaussian distribution (with mean 0 and standard deviation 1), a negative 1976-99 trend per decade in TN10% (or TX10%) of, for instance, 4 implies a decrease in probability from $P(x)=0.100$ in 1976 at the beginning of this 24 yr period to $P(x)=0.100-(2.4\times 2/365)=0.074$ in 1999 at the end of the period. Inverting $P(x)$ gives that x changes from -1.282 [because $P(x\leq -1.282)=0.100$] to -1.449 [because $P(x\leq -1.449)=0.074$]. Assuming a change only in the mean and not in the variance or skewness, the entire distribution must be shifted by $-1.449--1.282=-0.167$. The expected corresponding location change of x for TN90% (or TX90%) is then from $x=1.282$ [because $P(x\leq 1.282)=0.900$] to $x=1.282-0.167=1.114$. In terms of probability, this gives a change from $P(x)=0.900$ to $P(x\leq 1.114)=0.867$. Over the 24 yr period, the expected trend per decade in TN90% (or TX90%), which is denoted TN90%(expect) [or TX90%(expect)] in Table 3.3, is then $0.033\times 365/2.4=4.95$.

From the observed European trends in the cold extremes TN10% and TX10%, the expected values of the European trends in the corresponding warm extremes were calculated assuming a Gaussian distribution of the daily temperature anomalies. These expected trends, denoted TN90%-(expect) and TX90%(expect), were then compared with the observed values for TN90% and TX90%. If the expected and observed trends agree, then the cold and warm tail changed in a 'symmetric' way, whereas disagreement is indicative of 'asymmetric' change resulting from a change in variance or skewness. The comparison is sound, since it was verified that the distribution of temperature anomalies is near-Gaussian in the domains considered here (< 30th and > 70th percentile). If the comparison suggests asymmetry, it was investigated from which tail the effect mainly originates given the observed change in the median accompanying the observed trends in the cold and warm extremes.

In a second step of the analysis of precipitation extremes, we investigated to what extent the increase or decrease in the annual amount can be attributed to an increase or decrease in the number of very wet days R95% or the amount that falls at these days. The sign of the station trends for the annual amount were compared with the sign of the trends in the R95%tot index for the fraction of the annual amount due to very wet days. At stations where the annual amount increases, positive R95%tot trends are indicative of a disproportionate large contribution of the extremes to this wetting. On the other hand, at stations where the annual amount decreases, positive R95%tot trends indicate that the very wet days are less affected than the other wet days. Negative R95%tot trends indicate a smaller than proportional contribution of very wet days to wetting or drying.

3.5 Results

Trends in temperature extremes

Figure 3.4 shows that the 1946–99 trends in the indices frost days FD and summer days SU both indicate warming. For FD, the warming trend is significant at the 5% (25%) level for 39 (59) out of 86 stations; for SU, this is the case for 16 (42) stations. Cooling trends significant at the 5% (25%) level are not apparent (apparent for 6 stations) for FD, but for SU there are 4 (14) stations with significant cooling trends, concentrated in south-eastern Europe. The trend per decade in the Europe-average annual number of frost days is –1.7; for summer days this trend is 0.8. Consequently, averaged over Europe, there are 9.2 fewer frost days and 4.3 more summer days in the year 1999 than in the year 1946.

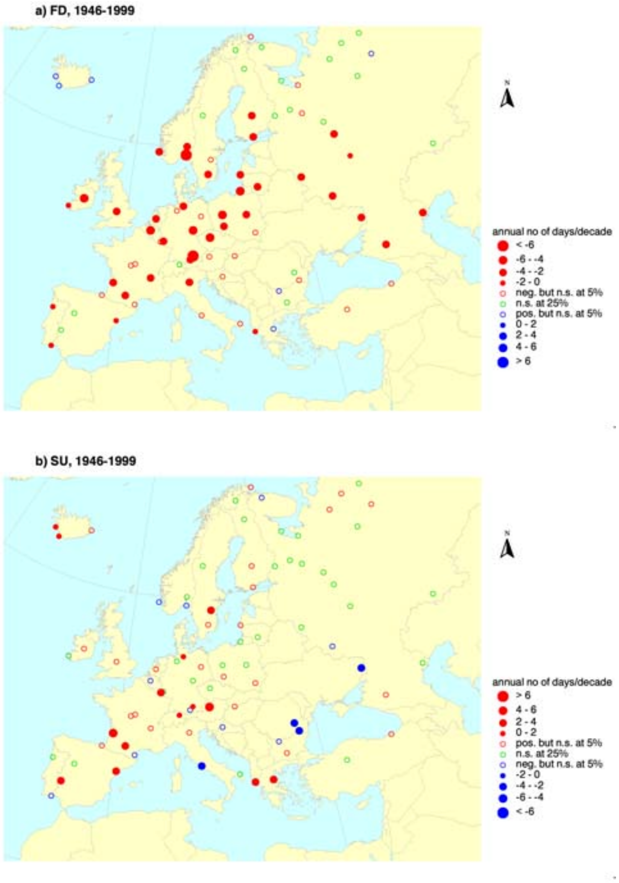


Figure 3.4 Trends per decade in the annual number of frost days FD (a) and summer days SU (b) for the period 1946–99. Dots are scaled according to the magnitude of the trend. For trends significant at the 25 % level, but not at the 5 % level (Student's *t* test), the sign of the trend is indicated (open circles). Colour coding is applied: red corresponds to warming trends (fewer frost days viz. more summer days), blue to cooling trends (more frost days viz. fewer summer days). Green is used for trends that are not significant at the 25 % level.

The 1946–99 station trends in the annual number of cold nights TN10% and warm nights TN90% also indicate warming (Figure 3.5a/b), as do the corresponding indices of maximum temperature: cold days TX10% and warm days TX90% (not shown). The spatial coherence in the trend patterns is higher for the two indices of cold extremes TN10% and TX10% than for the two indices of warm extremes TN90% and TX90%. Despite the dominating warming trend, local cooling trends are found for TN90%, particularly over Iceland and south-eastern Europe. These cooling trends mainly emerge from the summer months (April–September) and are also apparent in TX10% and TX90% (not shown). The high 1976–99 warming trend over Europe (0.43°C/decade compared with 0.11°C/decade for the entire 1946–99 period) results in larger trends in the indices of temperature extremes for that subperiod, see Figure 3.5c/d. The spatial coherence in the trend pattern of the number of warm nights TN90% is markedly enhanced, although the unspecified upper bound of the highest warming interval in the map overemphasizes the effect somewhat. In the 1976–99 period, no station saw a cooling trend in TN90%.

Table 3.3 gives the European trends in the four percentile-based indices of temperature extremes. For the full period 1946–99, all trends indicate warming, although the TX10% warming is not significant at the 5% level. For the subperiods, this significance level is harder to reach due to the shorter series; see Eq. (3.4). Nevertheless, the warming trends for the indices TN10%, TN90% and TX90% are significant at the 5% level in the 1976–99 subperiod. The trend per decade for TN90% of 11.3 implies almost a doubling of the annual number of warm nights from the climatological value 36 in 1976 (see Table 3.2) to 63 in 1999. This translates into a change in return period from 10 to 6 days. For the 1946–75 subperiod of average cooling (by 0.03°C/decade; not significant at the 5% level), the trends in the warm extremes also indicate cooling.

Table 3.3 European warming trends per decade (with 95 % confidence intervals in parentheses) in the percentile-based indices of cold and warm extremes for the periods 1946–99, 1946–75 and 1976–99. Values for observed trends significant at the 5 % level (*t* test) are set bold face. Expected trends in warm extremes TN90%(expect) and TX90%(expect) are calculated from the observed trends in the corresponding cold extremes, assuming Gaussian temperature distributions and no change in variance or skewness (see Figure 3.3). Expected trends in warm extremes that differ at the 5 % level (*t* test) from the observed values are marked with*. Note that the climatological values of the annual day-count indices are 36 (see Table 3.2). The observed European trends in mean temperature and diurnal temperature range are given in the headers.

Cold extremes decrease annual no. of days/decade		Warm extremes increase annual no. of days/decade			
European trends 1946–99 Increase per decade in mean temperature: **0.11** (0.01–0.22)°C Increase per decade in diurnal temperature range: **–0.04** (–0.07—0.01)°C					
TN10%	**2.1** (0.7–3.6)	TN90%	**2.5** (0.9–4.2)	TN90%(expect)	*2.8*
TX10%	1.3 (–0.6–3.1)	TX90%	**2.1** (0.3–3.9)	TX90%(expect)	*1.5*
European trends 1946–75 Increase per decade in mean temperature: –0.03 (–0.31–0.24)°C Increase per decade in diurnal temperature range: –0.03 (–0.11—0.05)°C					
TN10%	2.0 (–2.0–6.1)	TN90%	–2.3 (–5.4–0.7)	TN90%(expect)	*2.3**
TX10%	2.1 (–2.8–7.0)	TX90%	–2.1 (–5.8–1.6)	TX90%(expect)	*2.4**
European trends 1976–99 Increase per decade in mean temperature: **0.43** (0.09–0.77)°C Increase per decade in diurnal temperature range: 0.03 (–0.05—0.12)°C					
TN10%	**4.2** (0.2–8.2)	TN90%	**11.3** (6.6–16.1)	TN90%(expect)	*5.3**
TX10%	4.3 (–1.2–9.9)	TX90%	**10.9** (5.7–16.0)	TX90%(expect)	*5.5**

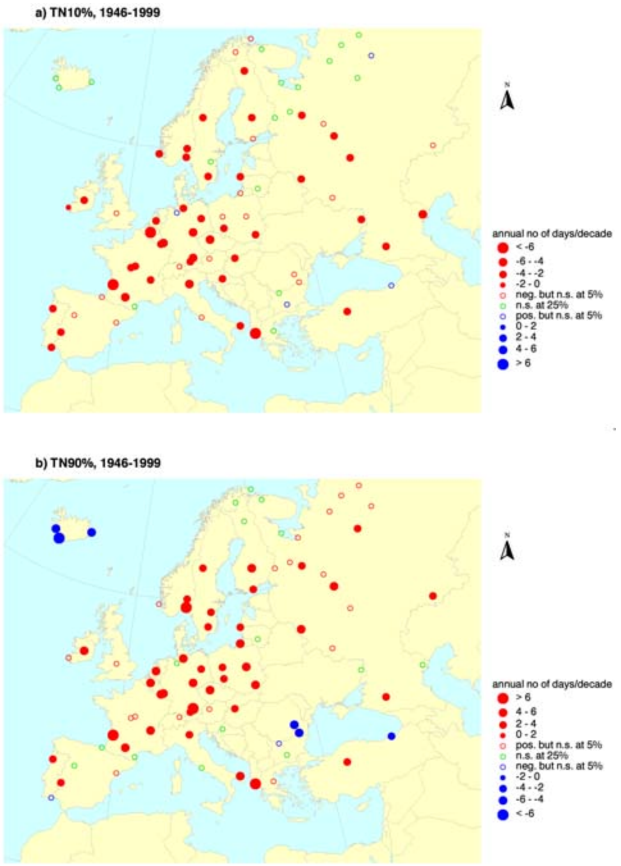


Figure 3.5 Trends per decade in the annual number of cold nights TN10% (a and c) and warm nights TN90% (b and d) for the periods 1946–99 (a and b) and 1976–99 (c and d). Red corresponds to warming trends (fewer cold nights viz. more warm nights), blue to cooling trends (more cold nights viz. fewer warm nights). The open circles are as described in Figure 3.4.

On the other hand, the number of cold extremes decreases, which is indicative of warming. This decrease is only significant at the 25% level and not at the 5% level.

For the 1946–99 warming period, Table 3.3 shows larger warming trends in the nighttime indices TN10% and TN90% than in the daytime indices TX10% and TX90%, which is in agreement with the observed negative trend in the mean diurnal temperature range (-0.04°C/decade). For the 1976–99 subperiod of pronounced warming, the almost equal warming trends in the nighttime indices and daytime indices suggest that there is no negative trend in the mean diurnal temperature range. The observed trend in the mean diurnal temperature range is indeed positive (0.03°C/decade), although not significant at the 5% level.

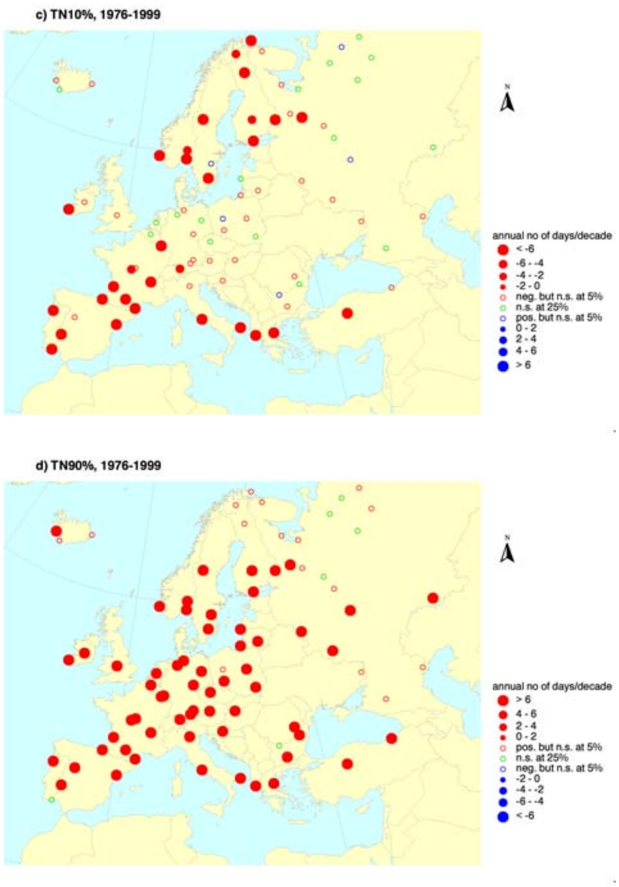


Figure 3.5 continued.

Comparison of trends in cold and warm extremes

The last column of Table 3.3 includes the expected European trends in the warm extremes TN90%(expect) and TX90%(expect), as calculated from the observed trends in the cold extremes TN10% and TX10% according to the scheme of Figure 3.3. For the 1946–99 period, these expected warming trends compare well with the observed trends TN90% and TX90%. However, for each of the two subperiods, the expected trends in the warm extremes differ significantly from the observed trends. For the 1946–75 subperiod, an episode of cooling, the cold extremes TN10% and TX10%, and therefore the expected warm extremes TN90%(expect) and TX90%(expect) show a warming trend. The sign of the expected trends disagrees with that of the observed trends in the warm extremes, which indicate cooling. For the 1976–99 subperiod, when Europe clearly warmed, the expected trends in the warm extremes are about half the observed values.

This result implies that for the two subperiods ‘asymmetric’ temperature change can be detected, whereas for the entire period asymmetry is undetectable. Asymmetry leads to a narrowing of the temperature distributions for the cooling subperiod and to a widening of the temperature distributions for the warming subperiod. Comparison with the trend in the median indicates that the asymmetry can mainly be attributed to the fact that the warming trends in TN10% and TX10% lag behind.

Trends in precipitation extremes

Figure 3.6 shows that a positive station trend in the indices of moderate wet days R75% and very wet days R95% dominates in the 1946–99 period.

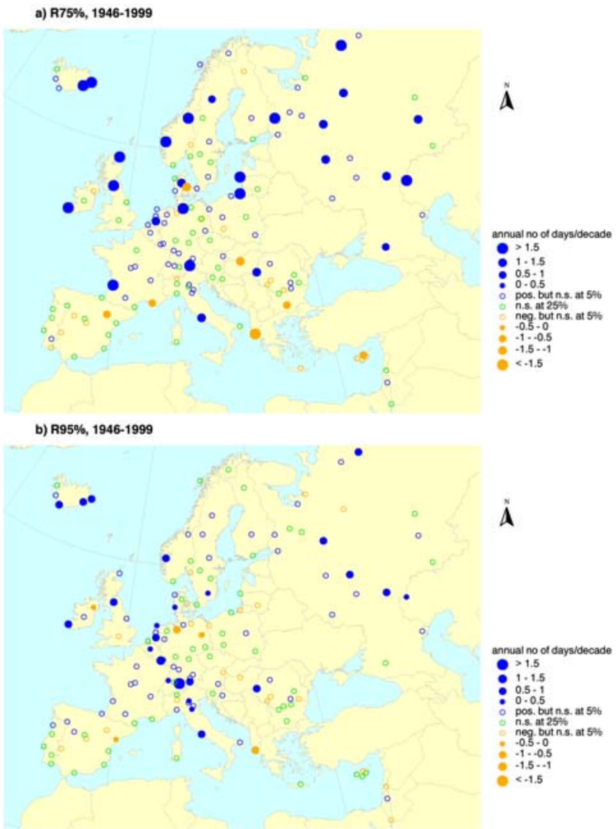


Figure 3.6 Trends per decade in the annual number of moderate wet days R75% (a) and very wet days R95% (b) for the period 1946–99. Blue corresponds to wetter conditions, yellow to drier conditions. The open circles are as described in Figure 3.4.

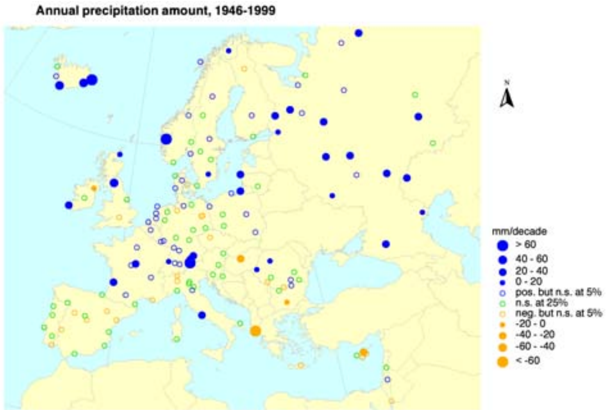


Figure 3.7 Trends per decade in the annual precipitation amount for the period 1946–99. Blue corresponds to wetter conditions, yellow to drier conditions. The open circles are as described in Figure 3.4.

A positive trend is seen mostly at stations where the annual amount increases (Figure 3.7). Vice versa, a negative trend does occur mostly at stations in drying areas.

For R75%, R95%, and the other indices of precipitation extremes, the spatial coherence is much lower than for the indices of temperature extremes. Even at short distances (~500 km), positive and negative trends in extreme precipitation are found at stations scattered all over Europe. The usually large local gradients in heavy precipitation events are among the possible explanations for the noisy spatial pattern. Table 3.4 shows that, averaged over Europe, six out of seven indices of precipitation extremes significantly increase between 1946 and 1999. Increases (though in most cases only significant at the 25% level) are also found in the 1946–75 and 1976–99 subperiods, with the exception of the highest 1-day precipitation amount RX1day that slightly decreases.

Figure 3.8 shows the 1946–99 station trends in the R95%tot index that defines extreme precipitation relative to the total amount. The trend pattern of R95%tot resembles closely the trend pattern of R95% in Figure 3.6.

Table 3.5 shows that from the 35 (73) stations for which the increase in the annual amount is significant at the 5% (25%) level, 11 (46) stations also have a significant increase in R95%tot. No station with a decrease in the annual amount significant at the 5% level shows a change in R95%tot significant at the 5% level. This means that a signal of a disproportionate large change in the extremes relative to the total amount is present, but this signal is only apparent in wetting areas. In drying areas, no signal is found.

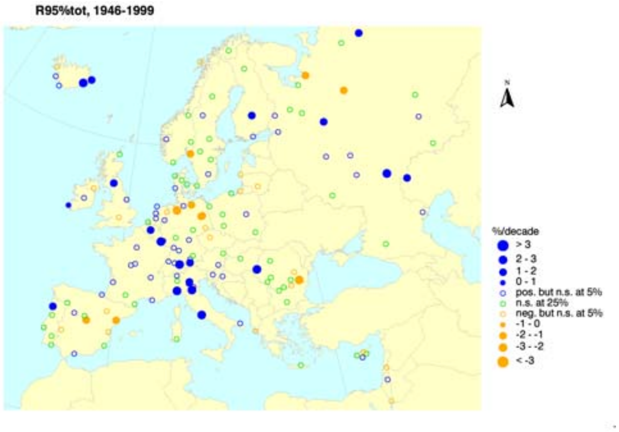


Figure 3.8 Trends per decade in the fraction of annual precipitation amount due to very wet days R95%tot for the period 1946–99. Blue corresponds to larger fractions, yellow to smaller fractions. The open circles are as described in Figure 3.4.

Table 3.4 European trends per decade (with 95 % confidence intervals in parentheses) in the indices of extreme precipitation for the periods 1946–99, 1946–75 and 1976–99. Values significant at the 5 % level (*t* test) are set bold face. The observed European trends in annual precipitation amount and annual number of wet days are given in the headers. As in Table 3.2, the day-count indices (R75%, R95%, R10mm, and R20mm) are expressed in annual number of days.

Percentile-threshold indices Increase per decade		Absolute-threshold indices Increase per decade	
European trends 1946–99 Increase per decade in annual precipitation amount: **7.6** (0.7–14.5) mm Increase per decade in annual number of wet days: 0.4 (–0.5–1.3)			
R75%	**0.4** (0.1–0.7)	R10mm	**0.3** (0.1–0.5)
R95%	**0.2** (0.1–0.2)	R20mm	**0.1** (0.0–0.2)
R95%tot	**0.3** (0.1–0.5)%	RX1day	0.2 (–0.1–0.5) mm
		RX5day	**0.6** (0.0–1.2) mm
European trends 1946–75 Increase per decade in annual precipitation amount: 14.0 (–5.0–32.9) mm Increase per decade in annual number of wet days: 1.3 (–1.1–3.8)			
R75%	0.5 (–0.3–1.4)	R10mm	0.5 (–0.2–1.1)
R95%	0.1 (–0.1–0.3)	R20mm	**0.2** (0.0–0.5)
R95%tot	0.1 (–0.4–0.6)%	RX1day	–0.1 (–1.0–0.7) mm
		RX5day	0.5 (–1.0–2.0) mm
European trends 1976–99 Increase per decade in annual precipitation amount: –1.0 (–19.9–18.0) mm Increase per decade in annual number of wet days: –1.6 (–4.2–1.0)			
R75%	0.4 (–0.4–1.3)	R10mm	0.3 (–0.3–0.9)
R95%	**0.3** (0.1–0.6)	R20mm	0.1 (–0.1–0.4)
R95%tot	**0.8** (0.1–1.5)%	RX1day	–0.2 (–1.2–0.9) mm
		RX5day	0.9 (–1.1–2.9) mm

Table 3.5 Contingency table showing the joint distribution of station trends in annual precipitation amount and station trends in fraction of annual precipitation amount due to very wet days R95%tot for the period 1946–99. Categories are: positive trend (+), negative trend (–) and not significant (n.s.) at the 5% level (and 25% level in parentheses).

	R95%tot +	R95%tot –	R95%tot n.s.
Annual precipitation amount +	11 (46)	0 (6)	24 (21)
Annual precipitation amount –	0 (4)	0 (13)	5 (10)
Annual precipitation amount n.s.	8 (17)	10 (10)	93 (24)

3.6 Discussion

Symmetric and asymmetric warming of extremes

The analysis of daily observations shows that the European 1946–99 trends in the annual number of cold and warm temperature extremes do not contradict the assumption of shifted temperature distributions with no change in the parameters other than the mean. This 'symmetric' warming of the cold and warm tails of the temperature distributions implies unchanged temperature variance. However, for the two subperiods, the temperature changes were accompanied by 'asymmetric' rather than 'symmetric' changes in temperature extremes. For the slightly cooling 1946–75 subperiod, the indices of warm extremes TN90% and TX90% show trends that are opposite to what is expected from the observed trends in the corresponding indices of cold extremes TN10% and TX10%. This asymmetry implies a narrowing of the distributions of both minimum and maximum temperature and therefore a lower temperature variance. For the warming 1976–99 subperiod, the indices of warm extremes TN90% and TX90% show considerably larger trends than expected from the observed trends in the corresponding indices of cold extremes TN10% and TX10%, implying a widening of the distributions of minimum and maximum temperature and therefore higher temperature variance.

Our results in Figure 3.5 indicate that the temperature rise in the Central England series in recent decades is basically associated with an increase in warm extremes, rather than with a reduction in cold extremes. This is in agreement with earlier studies of Yan *et al.* (2002), Horton *et al.* (2001) and Figure 3 of Jones *et al.* (1999b), although the latter authors drew the opposite conclusion from that figure. The Central England result matches the result of the majority of other stations in Europe, including the additional seven stations in Yan *et al.* (2002), as Table 3.3 shows that after 1976, the European trends in the indices of warm extremes TN90% and TX90% contribute much more strongly to the warming than the European trends in the indices of cold extremes TN10% and TX10%. Accordingly, in Chapter 2 we found that the recent warming over Europe in the winter half of the year was accompanied by an increase rather than a decrease in the number of cold days that are part of spells of at least five consecutive cold days.

The question remains why the recent warming is occurring 'asymmetrically' and why this asymmetry is not apparent in the long period. As cold/warm extremes are associated with specific atmospheric circulation patterns over Europe, the change in airflow may be the driving force. Cold extremes in winter are often associated with airflow from the snow-covered continent and in summer from the Atlantic Ocean. It is reasonable to assume that cold extremes are less sensitive to large-scale warming than warm extremes, because of the latent heat of snow and the thermal inertia of water. In such a warming scenario, small changes in the frequency of atmospheric circulation patterns may be capable of stabilising or increasing the number of cold extremes. The fact that the asymmetry of the subperiods is obscured in the long period may then be due to averaging of two opposite tendencies. Some of the possible changes in atmospheric circulation patterns over the second half of the 20th century that affect the temperature extremes are well documented, like the changes in the North Atlantic Oscillation (NAO). A systematic study of the relation between circulation changes and (seasonal) changes in cold/warm extremes is needed for a better understanding of the causes for asymmetric temperature change.

Amplified response of precipitation extremes

The positive European trends in the indices of wet extremes support the Houghton *et al.* (2001) statement that 'it is likely that there has been a statistically significant increase in the amount of heavy and extreme precipitation events when averaged across the mid and high latitudes'. According to the hypothesis of Groisman *et al.* (1999; see also Houghton *et al.*, 2001), there should be an amplified response of the extreme precipitation events relative to the change in total amount. The R95%tot trends as tabulated in Table 3.5 show that at about one- (two-) third(s) of the ECA stations with an increase in the annual amount significant at the 5% (25%) level, the trend in the amount falling on very wet days is significantly higher. Remarkably, at the ECA stations with a decrease in the annual amount, no negative trends in R95%tot are found. Although only 26% (66%) of the stations show a change in the annual amount significant at the 5% (25%) level, in wetting areas the observations support the notion of an amplified response of the extreme events compared with the annual amount. In drying areas, such an amplified response is not found. Future investigations of the relation between atmospheric circulation and R95%tot may clarify to what extent circulation changes (e.g. related to the NAO) contributed to the observed trends.

Relevance of selected indices for impact studies and climate change studies

Values of absolute extremes, like the highest 5-day precipitation amount RX5day in a year, can often be related with extreme events that affect human society and the natural environment. Indices based on the count of days crossing certain fixed thresholds (e.g. the 0°C threshold as used in the frost days index FD) can also be related to observed impacts, in particular if the thresholds refer to values of physical, hydrological or biological significance. Indices based on the count of days crossing percentile thresholds are less suitable for direct impact comparisons (Bonsal *et al.*, 2001), but they may provide useful indirect information relevant to impact studies. For instance, the same value for the index very wet days R95% often refers to larger amounts in wet climates than dry climates. The accompanying impacts are likely to differ accordingly. Yet, in every climate regime, nature and man have adapted to the local pattern of climate variability closely and local infrastructure is designed to withstand local extremes. Trends in the R95% index are thus relevant for comparing, for instance, the changes in demands on drainage and sewerage systems at different locations in Europe. Likewise, the trends in TX10% and TX90% are relevant for comparing changes in heating and cooling demands. Changes in percentile-based indices do not necessarily translate to changes in absolute extremes (Zhang *et al.*, 2001).

For climate change detection studies, indices based on percentile thresholds have a clear advantage, as they can be used to compare the changes in the same parts of the temperature and precipitation distributions Europe-wide. Our results suggest that the percentile thresholds in the selected day-count indices are adequate for detection of trends in extremes in the ~50 yr study period 1946–99. On average, more than half of the station trends is significant at the 5% level, and significant Europe-average trends could be detected for all but one of the temperature and precipitation indices. For the 24 yr subperiod 1976–99, significant Europe-average trends could be detected for three out of four temperature indices, even though this requires roughly a 3 times larger trend compared to the 54 yr period. The strong dependence of the detection probability of trends on the series length once again stresses the need for long-term climatic time series with daily resolution.

Due to the small (5–60 day) return periods of the events described by the selected day-count indices, they do not focus on situations that are rare enough to cause serious damage, like severe heat waves or large-scale flooding. In this respect the indices refer to 'soft' climate extremes. However, statistical analysis of trends in very extreme temperature and precipitation events ('hard' climate extremes with a return period of at least a decade) is not feasible, because of too few events in the short series. For instance, for extremes with return period of 365 days instead of 10 days, 6 times larger percentage trends are required to achieve equal detection probability. On the other hand, trend analysis of 'soft' climate extremes may be considered as empirical support for estimation of trends in 'hard' climate extremes.

Time series with a typical length in the order of ~50 yr are also generated by (regional) climate model simulations. This restricts the extremes that can be subject to trend analysis in these simulations to the same 'soft' climate extremes as in our observational series. Projections of changes in extremes by these models are then also limited to 'soft' climate extremes. Analysis of 'hard' climate extremes with return periods exceeding decades requires long-term ensemble simulations. Validation of the behaviour of the selected indices in any model against the results of our empirical study would support the credibility of the model projections.

Folland *et al.* (2000) proposed that WMO regularly distributes to all nations a consistently analyzed and internationally agreed set of environmental extremes indices. The WMO–CCL/CLIVAR-recommended temperature and precipitation indices used in this European study proved to be good candidates. The indices are applicable to a wide variety of climates and clearly demonstrate how trends in the frequency of cold temperature extremes differ from trends in the frequency of warm temperature extremes and how temperature and precipitation extremes relate to changes in mean climate.

- The selected indices of daily temperature and precipitation extremes in Europe show pronounced trends within the 1946–99 period.

- Europe-average trends in mean temperature are accompanied by 'asymmetric' rather than 'symmetric' changes in temperature extremes for two consecutive periods: 1946–75 and 1976–99.

- The pronounced warming between 1976 and 1999 is primarily associated with an increase in warm extremes rather than with a decrease in cold extremes.

- No such 'asymmetric' change in the cold and warm tails of the temperature distributions is seen from the indices trends for the entire 1946–99 warming period.

- At stations where the annual precipitation amount between 1946 and 1999 increases, the index for the fraction of the annual amount due to very wet days gives a signal of disproportionate large changes in the precipitation extremes.

Abstract – Signals of anthropogenic warming over Europe are searched for in the spatial trend patterns for the variance and skewness (expressed by the 10th and 90th percentiles) of the distribution of daily mean temperature. Comparisons are made between these patterns in the station records of the European Climate Assessment dataset for the 1976–99 period, the patterns associated with natural variability in the observations (which were empirically derived from the observations in the 1946–75 period), and the patterns of future warming and natural variability as simulated by the National Center for Atmospheric Research Community Climate System Model in the Challenge ensemble experiment.

The results indicate that, on the basis of the patterns for the variance, a distinction can be made between temperature change due to natural variability and temperature change due to changes in external forcing. The observed variance trend patterns for the spring (MAM) and summer (JJA) warming 1976–99 are clearly different from the patterns for the change in variance associated with a warming due to natural variability in the observations. This led us to conclude that a change in an external forcing has to be invoked to explain the observed spring and summer warming. From the evaluation of the greenhouse and natural variability patterns in the climate model simulations we infer that the observed spring and summer variance trend patterns contain imprints consistent with anthropogenic warming. The analysis of the variance trend patterns for the winter (DJF) season is inconclusive about identifying causes of the observed warming for that season. Unlike the other three seasons, the autumn (SON) is for Europe a period of cooling in recent decades. The observed variance trend pattern for this season closely resembles the estimated pattern for the change in variance associated with a cooling due to natural variability, indicating that the observed autumn cooling can be ascribed to random weather variations in the period under consideration.

Signals of anthropogenic influence on European warming

4.1 Introduction

Like for the world, higher future temperatures are expected for Europe as a result of anthropogenic climate change (Houghton *et al.*, 1990, 1996, 2001). This prospect of significant human-induced warming has led to increasing interest in detection of anthropogenic signals in observational records of the past. Whether such signals are actually seen in the temperature records of the last century depends on the spatial scale considered. At the global scale, Houghton *et al.* (2001) concludes that: 'most of the warming observed over the last 50 years is attributable to human activities'. At the continental scale, the attribution of the observed warming to anthropogenic influence is not (yet) firmly settled due to the larger natural variability in the records. However, positive results of such analyses are starting to be reported. Recently, Stott (2003) detected the warming effects of increasing greenhouse gas concentrations in six separate land areas of the Earth, including Europe. Zwiers and Zhang (2003) were able to identify the greenhouse gas and sulfate aerosol induced signal in the observed annual mean near-surface temperatures of the past 50 years over North America and Eurasia. Other regional detection studies are in progress (IDAG, 2004).

Anthropogenic signals have not yet been positively detected at all in series of extreme events (IDAG, 2004), despite their obvious relevance for society. The fact that extremes are by definition rare and observational series relatively short reduces the detectability of statistically significant trends in extremes, let alone trends that are induced by human activities. Other facts that hamper the detection of human-induced changes in extremes are the lack of high-resolution datasets, data inhomogeneities and uncertainties in climate model predictions of changes in extremes.

In our previous study of daily temperature extremes in Europe (Chapter 3), we were able to detect significant changes in moderate, also called 'soft', extremes with average return periods of 5–60 days. In that study, we used indices recommended by the joint Working Group on Climate

Change Detection of the World Meteorological Organization – Commission for Climatology (WMO–CCL) and the Research Programme on Climate Variability and Predictability (CLIVAR; Peterson *et al.*, 2001). For the 1976–99 episode of strong and persistent warming over Europe, we found differences in the warming of the cold and warm tails of the distributions of daily maximum temperature and daily minimum temperature. Studying some of the same indices in observations and climate models, Kiktev *et al.* (2003) provide evidence that human-induced forcing has recently played an important role in extreme climate. However, they did not consider indices for both the cold and the warm tail of the daily temperature distributions.

The objective of the present study is to investigate whether the differential warming of the tails of the daily temperature distributions contains a possible clue for the attribution of the changes to anthropogenic influence. This is done by investigating whether the characteristics of the European warming of the recent past are different from the characteristics associated with natural temperature variability as derived from previous decades and whether similar differences are seen in climate model simulations of current and future climate. Daily mean temperature is investigated. Although a series of daily mean temperature is roughly equal to the average of maximum and minimum temperature, this does not imply that the daily distributions of these quantities show similar behaviour. The differential warming of the cold and warm tails of the daily mean temperature distribution studied here may be different from the differential warming found in Chapter 3 for maximum and minimum temperature.

As before, we focus on 'soft' temperature extremes with daily resolution. Soft extremes are defined as extremes with return periods of 5–60 days, which means that the annual number of events is sufficiently large to allow for meaningful trend analysis in ~50yr time series; see Eq. (3.3). The soft extremes in the present study are the 10th and 90th temperature percentiles rather than the corresponding day count indices used in Chapter 3. This choice is motivated by the fact that percentiles are less sensitive to inhomogeneities in the series than day count indices and by the fact that the use of percentiles avoids potential problems with the lower (zero) bound in day count indices. The 10th and 90th temperature percentiles form the basis for two carefully defined distribution measures that summarize variance and skewness. For these two distribution measures and for the mean, the trend patterns in the observations for the 1976–99 period are compared with the patterns associated with a warming due to natural variability. The latter were empirically derived from the earlier 1946–75 period, an episode with little temperature change over Europe. Also for the variance, skewness and mean, the trend patterns in the climate model simulations of future warming are compared with the patterns associated with a warming due to natural variability in the climate model.

We argue that, if the characteristics of the strong warming of the recent past are anomalous compared to natural variability, the trend patterns in one or more of the distribution measures may contain an imprint of anthropogenic warming. If the same measures in the climate model also show significant differences between the trend patterns for future warming under enhanced atmospheric greenhouse gas concentrations and the patterns associated with natural variability, then the trend pattern for that particular measure is sensitive for anthropogenic warming indeed. Formal detection techniques, like pattern-correlation methods (e.g. Santer *et al.*, 1995) or optimal detection methods (e.g. Hasselman, 1979), require realistic climate model estimates of natural variability and model fingerprints of forced climate change signals. These techniques have meanwhile evolved in sophisticated space-time and space-frequency detection studies (e.g. Stott et al, 2001). Our approach is different in that we empirically separate regional climate characteristics that are likely associated with a warming due to natural variability from characteristics that are likely associated with anthropogenic change. We do this separation on the basis of the patterns of a few simple measures for the daily temperature distribution, rather than on the basis of model estimates of (combined) greenhouse gas/aerosol patterns.

In our analysis a seasonal breakdown of the trend patterns is advisable, because the specific atmospheric circulations that govern the temperatures (including extremes) over Europe have different effect in each season. In large parts of Europe, circulations with airflow from the continent (snow-covered in winter; hot in summer) lead to cold extremes in winter and warm extremes in summer. Likewise, airflow from the Atlantic Ocean is associated with warm extremes in winter and cold extremes in summer. Thus a change in the frequency of a particular atmospheric circulation pattern has different effects on the trend patterns of temperature extremes in each season (Luterbacher *et al.*, 2004; Schaeffer *et al.*, 2004). The selected measures for the daily temperature distribution are suitable to reveal such differences.

Section 4.2 describes the observational series from the ECA daily station dataset (Chapter 2) and the climate model data from the Challenge experiment (Selten *et al.*, 2004). Section 4.3 details the measures that are used to summarize the daily temperature distribution. The procedure for estimation of trend patterns in the observations and model simulations and the procedure for partitioning changes due to natural variability and trends due to changes in non-natural external forcing are outlined in Section 4.4. Section 4.5 presents the results, which are discussed in Section 4.6.

A selection of daily series of temperature observations at meteorological stations in Europe was taken from the ECA dataset (Chapter 2). As in Chapter 3, the selection criteria were based on series completeness, adequate data coverage of Europe and homogeneity ranking (Wijngaard *et al.*, 2003). The present study employs the ECA dataset version October 2003 for the period 1946–99, in which the selection yields a total of 185 station series (Figure 4.1a) for which the daily mean temperature series met the criteria.

Climate model data are from ensemble integrations with the Community Climate System Model (CSM) version 1.4 of the National Center for Atmospheric Research (NCAR), which is a fully-coupled global climate model.

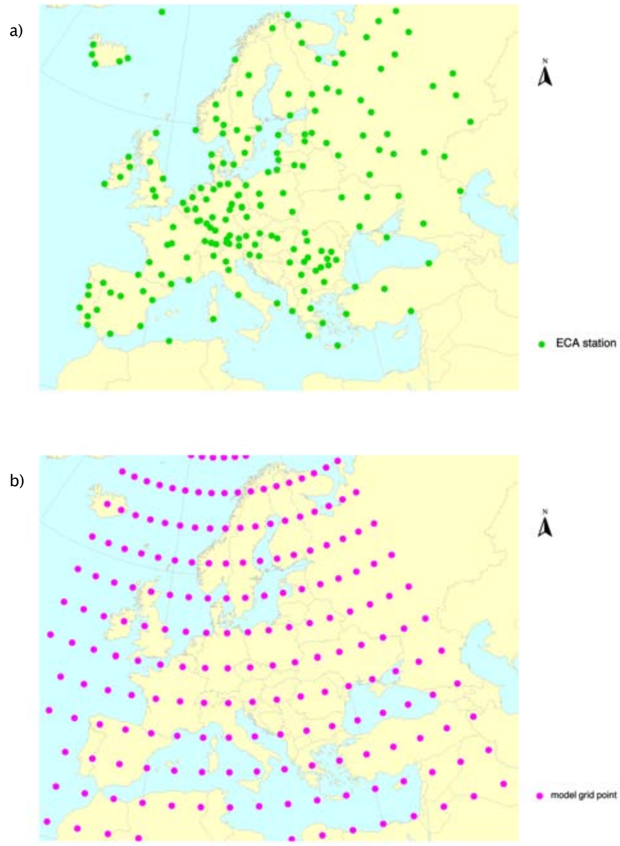


Figure 4.1 ECA stations (a) and Challenge model grid points (b) used in this study.

The ensembles were produced in the so-called Challenge experiment (Selten *et al.*, 2004), in which the CSM model has been used to produce 62 simulations of the earths' climate for the period 1940–2080, each starting from a slightly perturbed initial atmospheric state. The initial state of the other parts of the climate system was not perturbed. Between 1940–2000 the radiative forcings are according to historical estimates of atmospheric composition, solar irradiance and sulphate and volcanic aerosols. From 2000 until the year 2080 the radiative forcing change is calculated according to the BAU (Business As Usual) emission scenario of NCAR, which is similar to the SRES A1b scenario of IPCC (Houghton *et al.*, 2001). From the Challenge simulations, we use the daily mean near-surface air temperature data from all 62 members for grid points in a European window of the model (Figure 4.1b). Considered is the period 1946–99, which corresponds with the ECA observations, as well as the period 2046–75 for future warming.

4.3 Measures to summarize the distribution of daily mean temperature

The distribution of daily mean temperature is summarized by three descriptive measures: the mean MEA, and the variance VAR and skewness SKEW as expressed by the quantiles P10, P50 and P90, being the 10th percentile, the median and the 90th percentile, respectively. The measures VAR and SKEW give information about the shape of the distribution and primarily relate to the soft extremes in the tails of the distribution. The choice to express VAR and SKEW by the 10th and 90th percentiles rather than more extreme quantiles is motivated by the possibility to detect trends in extremes given the length of the record; see Eq. (3.3). The variance VAR and the skewness SKEW are defined as follows:

$$VAR = \frac{P90 - P10}{2} \tag{4.1}$$

$$SKEW = \frac{P90 - P50}{P50 - P10} \tag{4.2}$$

In this definition, VAR and SKEW are discrete expressions of the full statistical definitions of standard deviation and skewness. They are special cases of the sample L-scale and sample L-skewness parameters in Hosking (1990), which are known to be more robust than conventional moments to outliers in the data (Hosking, 1990). The unit for MEA and VAR is °C; SKEW is dimensionless. In the definition of SKEW in Eq. (4.2), the median P50 instead of the mean MEA was used. The reason is that, contrary to the mean, the median is not affected by a change occurring in one tail of the distribution only. For daily temperature distributions, the median is usually close to the mean. SKEW in Eq. (4.2) ranges from [0,∞], with SKEW = 1 corresponding to no skewness.

For the study periods, the annual and seasonal values of the three distribution measures MEA, VAR and SKEW were calculated for each observation station and model grid point as follows. First, the percentile values P10, P50 and P90 were calculated for five-day windows centred on each calendar day to account for the mean annual cycle. This yields a total sample size of 5*N* days, with *N* the length of the study period in years. The calendar day percentile values were then averaged over the year and the seasons: winter (December, January, February; DJF), spring (March, April, May; MAM), summer (June, July, August; JJA) and autumn (September, October, November; SON). No smoothing was applied (cf. Chapter 3). The 62 members of the climate model were treated as a single distribution, yielding sample sizes of 62×5*N* days for each calendar day. Jenkinson's empirical ranking formula (Folland and Anderson, 2002) was used to calculate the percentile values P10, P50 and P90. This formula gives satisfactory results for the (soft) extremes considered here (Folland and Anderson, 2002).

Table 4.1 shows the climatological values of MEA, VAR, SKEW, P10, P50 and P90 based on the standard normal period 1961–90 for 5 selected stations in different areas of Europe. The values for the grid points in the climate model nearest to the 5 stations are also given. The Europe-averages in Table 4.1 are calculated from the average values of MEA, P10, P50 and P90 over all ECA stations.

The values in Table 4.1 are illustrative for the climate conditions over Europe with regard to extremes. They show that Europe-average values of VAR are almost twice as high in winter than summer, with intermediate values in the transition seasons. This feature is in accordance with the fact that winter temperature variability is usually higher than summer temperature variability. The values of VAR vary greatly across Europe: the Russian station has the highest values of VAR (up to 9.4°C for winter), which is typical for a continental climate. In the Mediterranean region VAR is much smaller and its annual course is also small. The lowest value of VAR (1.9°C) is observed in the summer for the station in Iceland, whose (maritime) climate is very even compared to the other stations. The measure SKEW reveals other peculiarities. For the autumn and winter season, all stations have values smaller than 1, except Salamanca (Spain) in winter. SKEW values smaller than 1 imply that (P50-P10) > (P90-P50), that is, the cold tail (P50-P10) is heavier than the warm tail (P90-P50).

This indicates that cold outbreaks occur more frequent or are more severe than warm outbreaks in Europe during autumn and winter. For the spring season, both values smaller than 1 and values larger than 1 are found for SKEW. For the summer season, all stations show SKEW = 1 (no skewness), except De Bilt. The high value of 1.7 for SKEW at this station illustrates that relative severe warm summer spells occur in the otherwise mild summer climate near the west coast of Europe, where airflow from the nearby sea dominates.

Table 4.1 also shows that the climate model simulates the annual mean temperature at the five selected stations reasonably well, although the warm biases for MEA in winter and the cold biases for MEA in summer indicate that the annual cycle in the model is generally too weak. The exception here is the Russian station, for which the model simulates too low winter temperatures. The low winter temperatures are accompanied by values of P10 that are far too low, implying that very cold winter days are abundantly simulated by the model. Incorrect simulation of the amount of snow cover by the model may contribute to the temperature bias during winter. Daily temperature variability expressed by VAR is too high for the Icelandic and Russian station and too low for the Dutch and Spanish station. The values of SKEW are reasonably well reproduced by the model with the anomalous summer value of 1.7 at station De Bilt as marked exception.

4.4 Trend estimation and attribution methodology

Robust trend analysis was used to determine the trends in the three distribution measures MEA, VAR and SKEW. This method compares the values in the first half of a period with the values in the second half by means of a *t* test. The analysis was performed for all ECA stations and for all model grid points shown in Figure 4.1.

For the observations, the trend patterns in the three distribution measures were determined for the period 1976–99, an episode of pronounced warming in Europe (on average ~0.7°C/decade). To establish the patterns of change associated with a warming due to natural variability, the 1946–75 period, an episode of little temperature change, was split into a set of cold and a set of warm years. The cold set consists of all years with Europe-average temperature below the 1946–75 median and the warm set consists of all years with Europe-average temperature above the 1946–75 median. This splitting procedure is illustrated in Figure 4.2. The same procedure was applied for each of the four seasons. The differences in Europe-average temperature between the cold and warm sets are 0.6°C for the year, and 2.1, 1.1, 0.6 and 0.8°C for the winter, spring, summer

Table 4.1 Climatological values of the measures that summarize the distribution of daily mean temperature over the 1961–90 baseline period for 5 stations from different areas of Europe. The values for the nearest grid points in the NCAR climate model (Challenge ensemble experiment) are given in parenthesis. The Europe-average represents a mean over all ECA stations. MEA is the mean temperature; VAR and SKEW the variance and skewness as expressed by the temperature percentiles P10, P50 and P90. SKEW=1 implies no skewness. All units are °C, except for SKEW, which has no dimension.

	Reykjavik (Iceland)	Elatma (Russia)	De Bilt (Netherlands)
Annual:			
MEA	4.4 (4.9)	4.3 (2.1)	9.4 (8.8)
VAR	3.9 (4.6)	6.3 (8.3)	4.2 (3.3)
SKEW	0.8 (0.7)	0.8 (0.7)	1.0 (0.7)
P10	0.4 (−0.1)	−2.2 (−7.0)	5.2 (5.2)
P50	4.6 (5.3)	4.7 (3.0)	9.3 (9.2)
P90	8.1 (9.1)	10.3 (9.6)	13.7 (11.8)
Winter (DJF):			
MEA	−0.1 (1.1)	−9.5 (−13.4)	2.6 (4.4)
VAR	5.4 (6.5)	9.4 (14.1)	5.5 (4.6)
SKEW	0.8 (0.6)	0.7 (0.7)	0.7 (0.5)
P10	−5.8 (−6.1)	−19.5 (−28.8)	−3.3 (−0.8)
P50	0.4 (2.0)	−8.5 (−11.9)	3.1 (5.4)
P90	5.0 (6.9)	−0.7 (−0.7)	7.8 (8.3)
Spring (MAM):			
MEA	3.3 (2.7)	4.9 (1.2)	8.4 (5.9)
VAR	4.2 (5.2)	6.2 (8.0)	4.2 (3.6)
SKEW	0.8 (0.7)	0.9 (0.6)	1.1 (0.7)
P10	−1.2 (−2.9)	−1.4 (−7.7)	4.4 (2.1)
P50	3.6 (3.2)	5.3 (2.4)	8.3 (6.3)
P90	7.2 (7.5)	11.0 (8.2)	12.7 (9.3)
Summer (JJA):			
MEA	10.0 (9.3)	17.3 (15.6)	16.2 (14.4)
VAR	1.9 (2.8)	4.7 (4.4)	3.7 (2.0)
SKEW	1.0 (1.0)	1.0 (0.9)	1.7 (1.1)
P10	8.0 (6.5)	12.7 (11.1)	12.9 (12.4)
P50	10.0 (9.4)	17.3 (15.8)	15.7 (14.3)
P90	11.9 (12.1)	22.0 (19.8)	20.3 (16.4)
Autumn (SON):			
MEA	4.3 (6.3)	4.3 (4.7)	10.2 (10.3)
VAR	4.1 (3.9)	5.5 (6.9)	3.9 (3.2)
SKEW	0.9 (0.8)	0.8 (0.7)	0.9 (0.7)
P10	0.1 (2.3)	−1.4 (−2.7)	6.2 (6.8)
P50	4.4 (6.5)	4.7 (5.3)	10.2 (10.6)
P90	8.4 (10.1)	9.6 (11.1)	14.0 (13.2)

and autumn, respectively. Then, the patterns for the change in MEA, VAR and SKEW associated with a warming due to natural variability were calculated as the difference between the two sets for each station. They can be interpreted as resulting from a reordering of years that would have produced an artificial trend representative for an accumulation of overall warming due to natural variability. For instance, if a Europe-average warming due to natural variability of 0.4°C is found for a 20 yr period, it is caused by unequal distribution of warm and cold years in that episode.

	Salamanca (Spain)	Larissa (Greece)	Europe average
Annual:			
MEA	11.7 (14.4)	15.7 (14.7)	8.2
VAR	4.0 (3.1)	3.7 (3.5)	4.8
SKEW	1.0 (0.9)	0.9 (0.8)	0.9
P10	7.7 (11.3)	11.9 (11.1)	3.2
P50	11.6 (14.5)	15.8 (14.9)	8.3
P90	15.6 (17.4)	19.4 (18.0)	12.9
Winter (DJF):			
MEA	4.4 (10.3)	6.1 (7.2)	−1.0
VAR	4.2 (2.7)	4.4 (4.4)	6.5
SKEW	1.1 (0.7)	0.9 (0.7)	0.8
P10	0.3 (7.4)	1.7 (2.4)	−7.8
P50	4.3 (10.6)	6.2 (7.7)	−0.5
P90	8.7 (12.8)	10.4 (11.3)	5.1
Spring (MAM):			
MEA	10.1 (11.7)	14.4 (12.6)	7.4
VAR	4.3 (2.9)	3.9 (3.6)	4.9
SKEW	1.1 (1.1)	1.0 (1.0)	1.0
P10	5.9 (8.8)	10.5 (9.0)	2.4
P50	10.0 (11.6)	14.4 (12.6)	7.4
P90	14.5 (14.6)	18.2 (16.2)	12.3
Summer (JJA):			
MEA	19.7 (19.5)	26.0 (22.5)	17.3
VAR	3.6 (3.5)	3.3 (2.6)	3.9
SKEW	1.0 (1.1)	1.0 (1.1)	1.1
P10	16.0 (16.1)	22.7 (19.9)	13.5
P50	19.8 (19.5)	26.0 (22.4)	17.2
P90	23.3 (23.1)	29.3 (25.1)	21.3
Autumn (SON):			
MEA	12.3 (16.0)	16.3 (16.3)	8.9
VAR	4.1 (3.3)	3.9 (3.3)	4.6
SKEW	0.9 (0.9)	0.8 (0.7)	0.9
P10	8.1 (12.7)	12.2 (12.7)	4.2
P50	12.4 (16.2)	16.5 (16.7)	9.1
P90	16.2 (19.2)	20.0 (19.4)	13.4

In that case, the detected Europe-average trend is 0.2°C/decade, and the patterns of ΔMEA, ΔVAR and ΔSKEW should be the same as those obtained by the splitting procedure. Note that within the chosen framework, the splitting procedure introduces an upper bound to the Europe-average temperature rise by natural variability of 0.6°C for the year. The estimates for the patterns associated with a warming due to natural variability (being the difference between the cold and warm set of years) will not change when shorter (or longer) periods are considered.

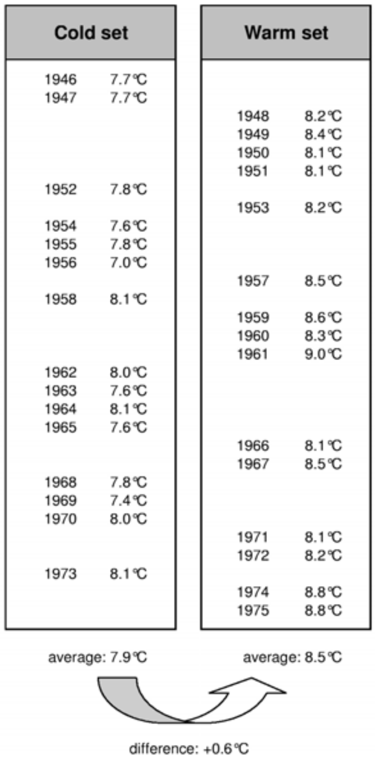


Figure 4.2 Cold and warm set of years distinguished on the basis of the Europe-average temperatures in the 1946–75 period. The patterns for the change in mean temperature ΔMEA, variance ΔVAR and skewness ΔSKEW associated with a warming due to natural variability are empirically determined by subtracting the patterns of MEA, VAR and SKEW for the cold set from the patterns of MEA, VAR and SKEW for the warm set. If a warming in a certain time interval arises from random weather variations, then the observed trend patterns of MEA, VAR and SKEW should correspond with the patterns of ΔMEA, ΔVAR and ΔSKEW estimated by the splitting procedure described above.

The robustness of the ΔVAR patterns associated with a warming due to natural variability as derived from the 1946–75 period was tested by calculating the patterns separately for the subperiods 1946–60 and 1961–75.

The splitting procedure assumes that natural variability primarily acts at short time scales (≤ 1 yr). The effect of natural variability at longer time scales was verified to be sufficiently small by considering a cold and warm set of temperature averages in 2 and 3 consecutive years instead of one year. This choice did not change the conclusions. Surface air pressure fields from NCEP-NCAR reanalysis data (Kalnay *et al.*, 1996) confirm that the splitting procedure is able to represent the small-scale spatial structures that are typically associated with natural temperature variability over Europe (not shown).

For the climate model, the trend patterns for anthropogenic climate change were calculated as the difference between the Challenge simulated values for the 1946–75 period (first half) and those for the 2046–75 period (second half). The Europe-average anthropogenic warming in the model is ~0.2°C/decade. Given the fact that the 62 ensemble members were treated as a single distribution, the characteristic signal of greenhouse forcing is well determined in this way, because the major part of natural variability cancels out. The patterns associated with a warming due to natural variability in the model were calculated from the 1946–75 simulations, following the same splitting procedure as with the observations and treating the 62 members of the climate model as a single distribution. The differences in Europe-average temperature between the cold set and the warm set in the model are 0.3°C for the year, and 0.6, 0.4, 0.1 and 0.3°C for the winter, spring, summer and autumn season, respectively. The patterns associated with a warming due to natural variability in the model obtained by the splitting procedure are in good agreement with the simulated warming patterns for the member exhibiting the largest Europe-average temperature increase in the 1946–75 period (correlation > 0.6). This supports the robustness of our splitting procedure.

Next, systematic comparisons were made for each of the three distribution measures between the trend patterns in the observations for the 1976–99 period and the estimated patterns associated with a warming due to natural variability in the observations. The same was done between the trend patterns in the climate model simulations of future warming and the estimated natural variability patterns in the climate model. The spatial resemblance of the patterns was examined by visual inspection and using fixed pattern-correlation techniques (Hegerl et al, 1996). To guide the interpretation of the comparisons the following conceptual relation is applied for the trend patterns in the 1976–99 period:

$$\Psi_s\left(observed\right) = a \cdot \Psi_s\left(forced\right) + b \cdot \Psi_s\left(natural\right) \qquad (4.3)$$

with Ψ_s the characteristic trend pattern for MEA and VAR in season s, a the trend portion due to changes in non-natural external forcing and b the trend portion due to warming/cooling by natural variability. Eq. (4.3) also holds in first order for SKEW, but since the SKEW trend pattern is generally more noisy than the VAR trend pattern, SKEW is only used to double check the results obtained with VAR.

If the patterns of trends and the patterns associated with natural variability for a distribution measure are different both in the observations and in the climate model, the observed 1976–99 patterns possibly contain an imprint of anthropogenic warming over Europe.

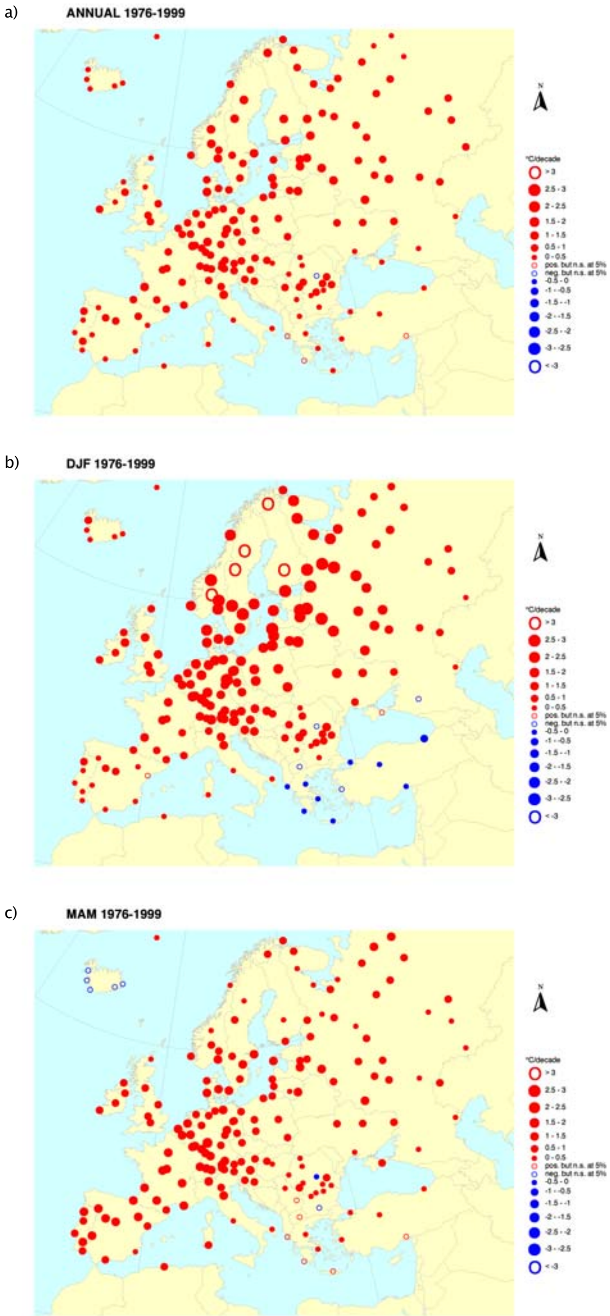


Figure 4.3 Trends per decade for mean temperature MEA observations in the 1976–99 period for the year (a), winter DJF (b), spring MAM (c), summer JJA (d) and autumn SON (e). The dots are scaled according to the magnitude of the trend. Colour coding is applied: red corresponds to warming trends, blue to cooling trends.

d) JJA 1976-1999

e) SON 1976-1999

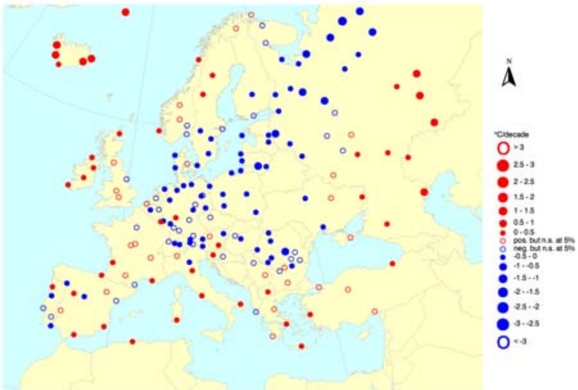


Figure 4.3 continued.

The following assumptions are made in the attribution method described above:

- The changes associated with a warming due to natural variability have a unique signature in the spatial patterns of ΔVAR and ΔSKEW.

- This unique signature of the changes associated with a warming due to natural variability can be obtained by the splitting procedure applied to a period of little overall temperature change, such as 1946–75.

These assumptions are partly validated with ECA and Challenge data.

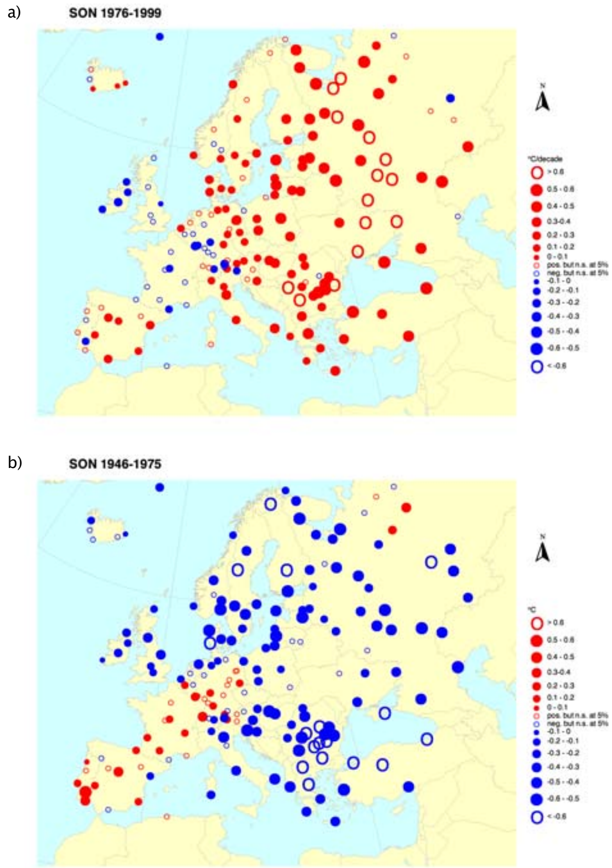


Figure 4.4 Autumn SON trends per decade for variance VAR in the observations for the 1976–99 period (a) and in the climate model simulations of future warming (c) together with the estimated ΔVAR patterns associated with a warming due to natural variability in the observations (b) and in the climate model (d). The latter were based on the 1946–75 period; for details see the text and Figure 4.2. The dots are as described in Figure 4.3.

4.5 Results

Mean temperature trends 1976–99

Figure 4.3 shows the station trends for mean temperature MEA observations in the 1976–99 period for the year and each of the 3-months seasons. Warming trends dominate with on average the strongest warming in winter. Apart from a north-south gradient, there is a weaker gradient from the ocean to the continent. As a result, the overall highest temperature rises are found over central and north eastern Europe and the lowest temperature rises over the Mediterranean and Iceland. The spatial gradients in winter (4.3b) are much stronger than those in summer (4.3d).

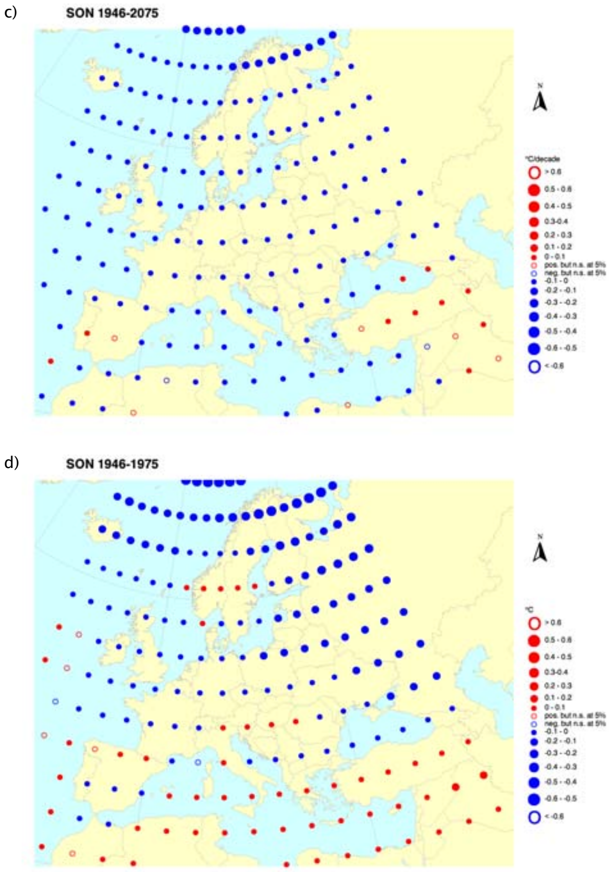


Figure 4.4 continued.

In winter, the strongest warming (up to about 3°C/decade) is observed over northern Europe, whereas southern Europe shows little warming or even cooling over Greece and Turkey. In summer, the warming pattern is much more flat with trends roughly between 0.5–1.5°C/decade throughout Europe. An intermediate trend pattern is found in spring (4.3c) with cooling rather than warming over Iceland.

The autumn season (4.3e) is a prominent exception to the overall temperature rise. It shows significant cooling up to about 1°C/decade over a large area of the continent. This result is in good agreement with the autumn cooling over parts of Europe found in the gridded temperature data-set from Jones *et al.* (2001; see Houghton *et al.*, 2001, Figure 2.10d).

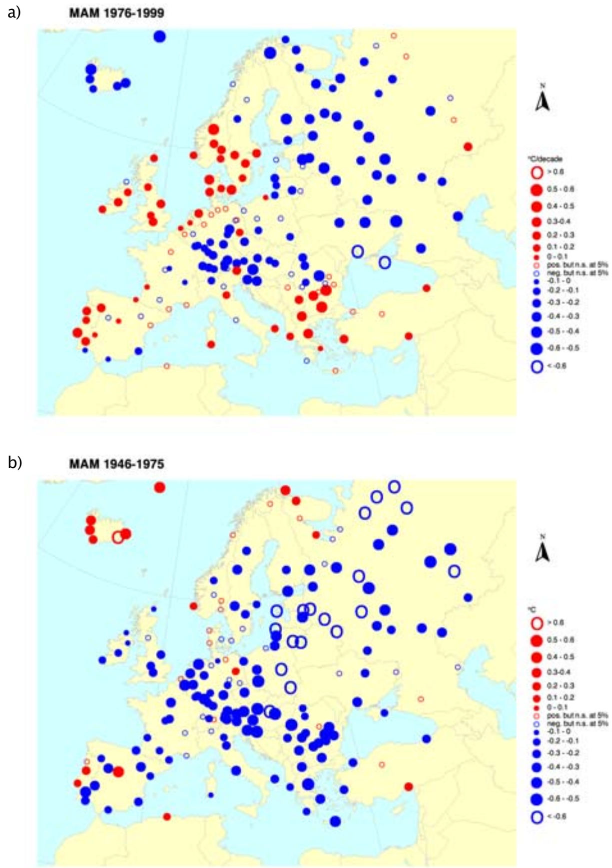
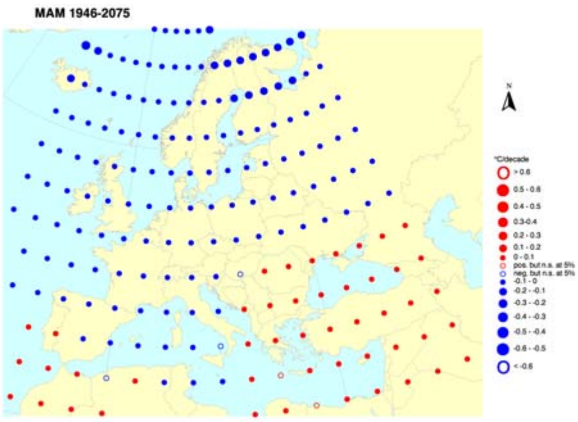


Figure 4.5 As Figure 4.4, but now for spring MAM.

Imprints of anthropogenic warming

The climate model simulations of future warming indicate that anthropogenic forcing leads to warming in every season (not shown). Clearly, this fact alone does not imply that the winter, spring and summer trend patterns of mean temperature MEA in Figure 4.3 contain imprints of anthropogenic warming. Our estimates of the patterns for the change in MEA associated with a warming due to natural variability (as derived from the observations in the 1946–75 period) indicate that, on the basis of natural variability alone, the possibility exists of a warming without changes in external forcing that has a similar pattern than the observed temperature rise between 1976–99. Table 4.2 shows that the spatial correlation coefficients between the observed trend patterns and the patterns associated with natural variability are significant at the 5% level for all seasons.

c)



d)

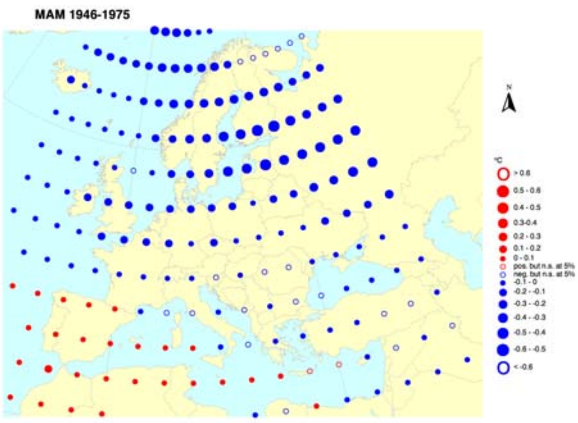


Figure 4.5 continued.

Table 4.2 Spatial correlation coefficients between the observed trend patterns 1976–99 and the estimated patterns associated with natural variability for the observations (top) and the same for the climate model simulations of future warming and natural variability in the model (bottom). Values significant at the 5% level (2-tailed) are set bold face.

	Winter (DJF)	Spring (MAM)	Summer (JJA)	Autumn (SON)
Observations:				
MEA	**0.71**	**0.23**	**0.55**	**−0.55**
VAR	**0.18**	0.10	0.04	**−0.45**
SKEW	**0.49**	0.07	−0.05	−0.03
Climate model:				
MEA	**0.78**	**0.67**	**0.38**	**0.76**
VAR	**−0.41**	**−0.16**	**0.21**	**0.15**
SKEW	**−0.18**	**0.14**	**0.30**	**0.46**

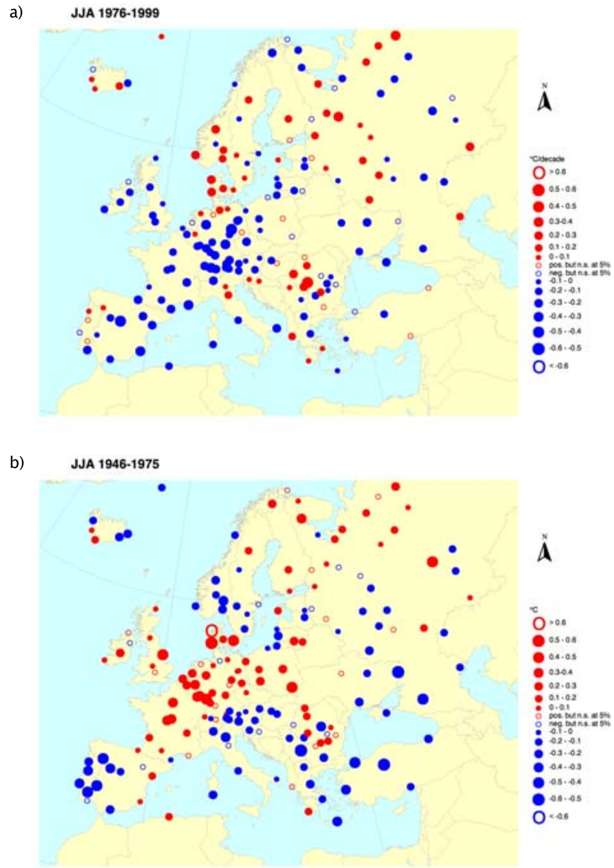


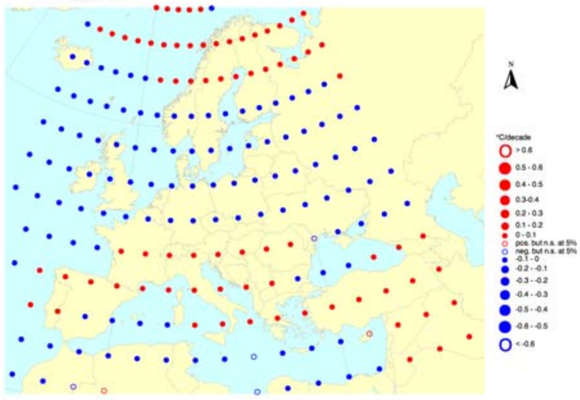
Figure 4.6 As Figure 4.4, but now for summer JJA.

The values range between 0.23 for spring and 0.71 for winter. Therefore, partitioning of the observed temperature change in either forced warming or natural variability on the basis of MEA is inconclusive, most so for winter.

The situation is different however for the measure of variation VAR and, to a lesser extent, the measure of skewness SKEW. A plausible distinction between the two causes of warming (forced and natural) can be made in several seasons on the basis of a comparison of the trend and natural variability patterns of these distribution measures.

For the autumn season, Figure 4.4 shows that the observed cooling is accompanied by an increase in temperature variance (4.4a). This VAR trend pattern is in good agreement with the ΔVAR pattern associated with natural variability (4.4b), if the signs are reversed. The spatial correlation coefficient between the observed pattern and that of natural variability is

c) **JJA 1946-2075**



d) **JJA 1946-1975**

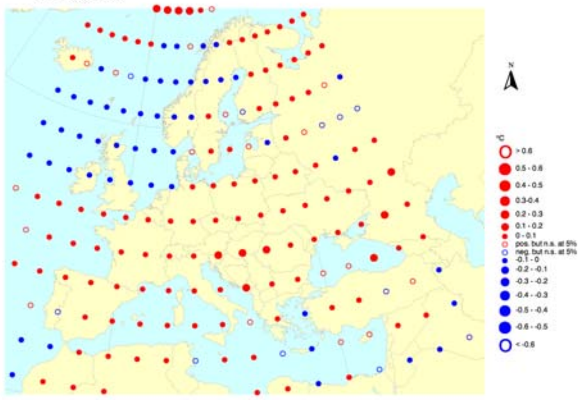


Figure 4.6 continued.

–0.45 (Table 4.2). Note that the negative sign is due to the fact that the pattern associated with natural variability was derived for a warming situation, rather than for a cooling situation; the same pattern but with reversed sign would have been derived for a cooling situation. In the climate model, the VAR trend pattern for the future warming scenario (4.4c) differs from the ΔVAR pattern associated with natural variability in the model (4.4d; correlation 0.15). This leads us to the conclusion that the observed autumn cooling in the 1976–99 period is primarily associated with natural variability.

For the spring season, the VAR trend pattern indicates that the observed warming cannot be explained by natural variability, because the pattern in the observations between 1976–99 in Figure 4.5a is clearly different from the pattern associated with a warming due to natural variability in 4.5b (correlation 0.10).

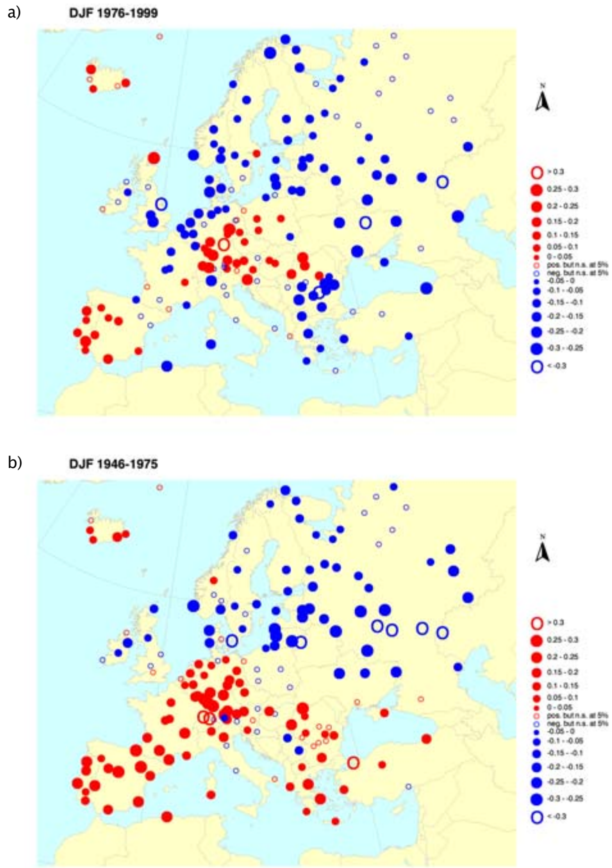


Figure 4.7 As Figure 4.4, but now for skewness SKEW in winter DJF.

The climate model also indicates differences between the simulations of future warming (4.5c) and natural variability in the model (4.5d; correlation –0.16). Therefore, it is likely that forced warming dominates over natural variability.

The situation for summer resembles that for spring. The observed summer VAR trend pattern in Figure 4.6a is different from the pattern associated with a warming due to natural variability in 4.6b (correlation 0.04) and a similar difference is also found for the climate model simulations (4.6c and 4.6d; correlation 0.21).

For the winter season, attempts to partition the observed temperature change in either forced warming or natural variability are less successful. The winter VAR trend patterns (not shown) give no conclusive answer at all about a dominant warming cause. In Figure 4.7 the corresponding trend patterns for skewness SKEW are presented.

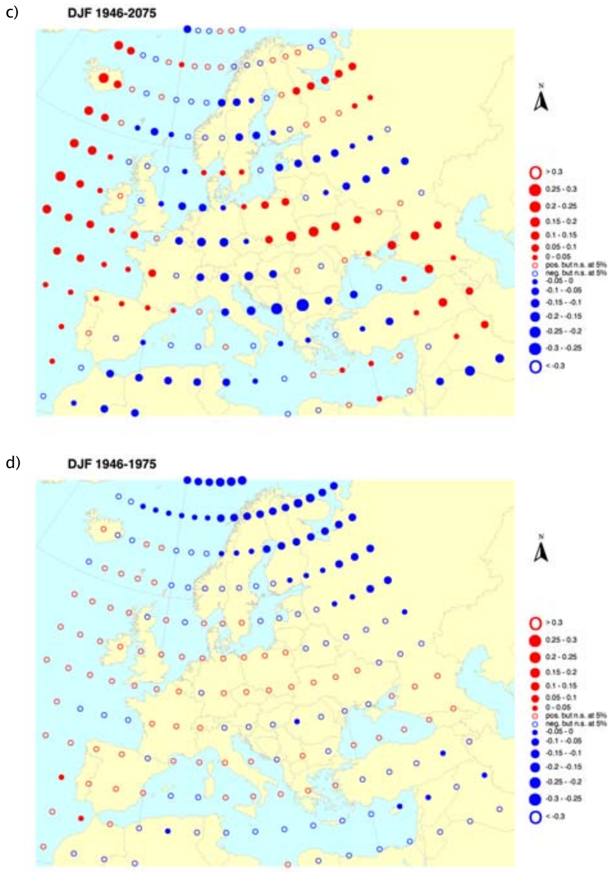


Figure 4.7 continued.

The figure and the correlation coefficients in Table 4.2 show that also the intrinsic noisier SKEW patterns give no indication of significant differences between the trend patterns and the patterns associated with natural variability.

Table 4.3 qualitatively summarizes the results of the comparison between the trend and variability patterns, identifying possible imprints of anthropogenic warming in the spring VAR trend pattern and the summer VAR trend pattern.

Finally, Figure 4.8 shows the patterns for ΔVAR associated with a warming due to natural variability as estimated from the subperiods 1946–60 (4.8a) and 1961–75 (4.8b) for the spring (MAM) season. A comparison with the corresponding pattern estimated from the full 1946–75 period in Figure 4.5b (repeated in Figure 4.8c for convenience) shows that the splitting procedure leads to robust estimates of the patterns associated with

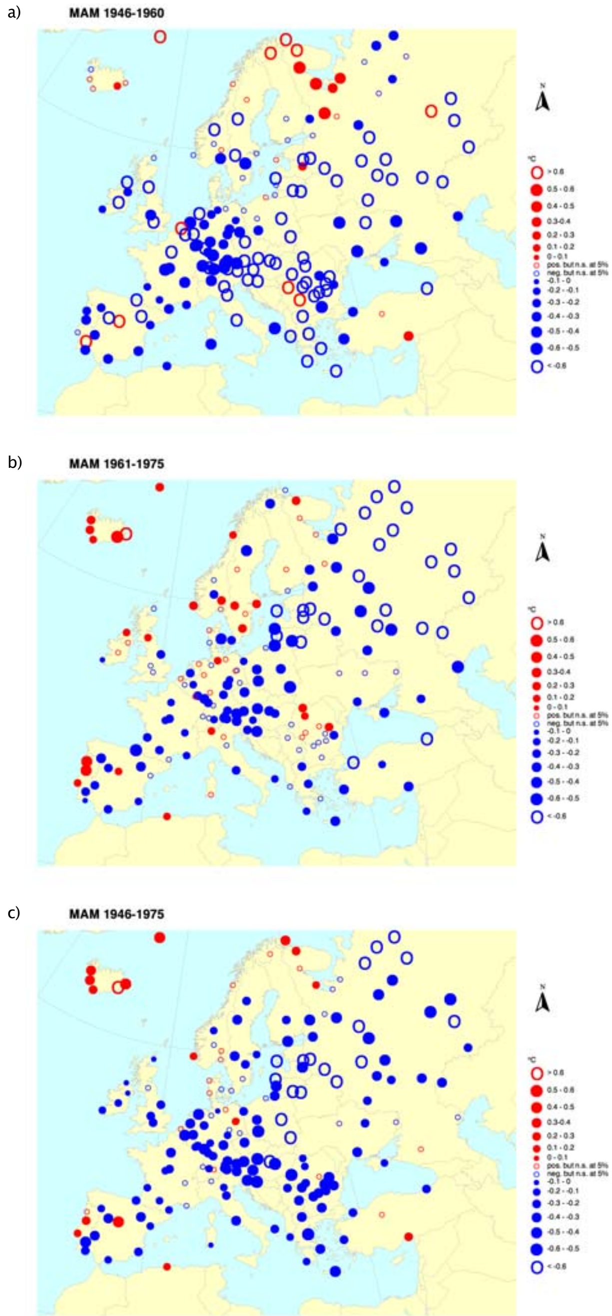


Figure 4.8 Spring MAM patterns for ΔVAR associated with a warming due to natural variability as derived from the observations in the 1946–60 period (a) and the 1961–75 period (b), compared with the corresponding pattern derived from the observations in the full 1946–75 period (c, repeated from Figure 4.5b). The good agreement between the patterns for the three different base periods is illustrative for the robustness of the splitting procedure (see Figure 4.2).

Table 4.3 Qualitative judgement of the results of the pattern comparison between observed trends and estimates of the changes associated with natural variability. Categories are: forced, indicating that an imprint consistent with anthropogenic warming is seen; natural variability, indicating that natural variability dominates; inconclusive, indicating that partitioning is not successful.

	Winter (DJF)	Spring (MAM)	Summer (JJA)	Autumn (SON)
MEA	inconclusive	inconclusive	inconclusive	inconclusive
VAR	inconclusive	forced	forced	natural variability
SKEW	inconclusive	forced	forced	natural variability

natural variability. The correlation coefficients between the 1946–60 and 1946–75 patterns and between the 1961–75 and 1946–75 patterns are 0.44 and 0.70, respectively. The correlation coefficients between these patterns for the other seasons (not shown) are ~0.6.

4.6 Discussion

On the basis of three measures for the distribution of daily mean temperature, we found that some of the seasonal patterns representing characteristics of European warming in the 1976–99 period are distinct from the estimated patterns that result from natural temperature variability. In particular, the observed spring (MAM) and summer (JJA) trend patterns for the measure of variance VAR are distinct from the estimated patterns for the change in VAR associated with a warming due to natural variability. Since the trend patterns in the climate model simulations of future warming for this measure are different from the estimated patterns associated with natural variability in the climate model as well, we conclude that the 1976–99 spring VAR and summer VAR trend patterns may contain imprints consistent with anthropogenic warming. For the warming in the winter season, no clear distinction could be made between forced warming and natural warming. The higher temperature variability in European winter is a possible explanation. For the autumn season, the results indicate that natural temperature variability is the most likely cause for the cooling observed over a large part of the continent, as the autumn pattern of the trends in VAR is very similar to the estimated pattern for the change in VAR associated with a cooling due to natural variability, whereas the climate model suggests clear differences between these patterns as a result of anthropogenic forcing. Consequently, for the autumn season the factor b in Eq. (4.3) for the trend portion due to natural variability is negative and dominant with respect to the factor a for the trend portion due to non-natural externally forced warming. In our view, this finding illustrates the power of our method.

A potential weakness of our method is that we obtained an estimate for the typical patterns associated with natural variability by artificially ranking the series of years (or seasons) in the 1946–75 period according to ascending Europe-average temperature. This simple method applied to a relatively short period can only partially account for the contribution from multi-decadal natural climate variability to the recent warming. We feel that this is acceptable, because the method is primarily intended to search for a possible imprint of anthropogenic warming in the measures that reflect the warming characteristics of the tails of the daily temperature distribution in the 1976–99 period. No formal attribution to anthropogenic influence is sought for. Our method relies on the existence of a unique signature of the changes associated with a warming due to natural variability in the spatial patterns of ΔVAR and ΔSKEW that can be estimated from a period with little overall temperature change, such as 1946–75. It is supported by the fact that the estimated patterns do not change markedly when they are based on the subperiods 1946–60 or 1961–75, instead of 1946–75. Our method is further supported by the fact that the estimated patterns associated with natural variability in the climate model resemble the trend patterns for the ensemble member in Challenge that has the greatest Europe-average warming in the 1946–75 period due to natural variability in the climate model simulations.

Formal climate change detection/attribution studies are required to determine whether the identified VAR trend patterns 1976–99 can indeed be attributed to anthropogenic influence. Such studies involve direct comparisons between the observations and the climate model data. A preliminary analysis of the climate model data from the Challenge experiment suggests that the model that was used is not suitable for such formal detection/attribution. The variance and skewness trend patterns observed in the ECA station data are not satisfactorily reproduced by the (older version of the) CSM model used in Challenge.

The advantage of the Challenge data for our study is that the trend patterns for the future warming resulting from anthropogenic forcing are well defined, because the major part of natural variability cancels out in the ensemble mean derived from 62 climate model simulations. The Challenge data could be further analyzed to find out in how many ensemble members imprints of anthropogenic warming can be detected using our method and how this number grows when the analysis period is extended beyond 1999 into the future. The 62 Challenge simulations underline that for an adequate representation of the characteristics of the daily temperature distribution in Europe by more state-of-the-art GCMs it is essential that these GCMs account for the natural variability that exists under the same external forcings (Selten *et al.*, 2004). Natural variability causes that the trend patterns for MEA, VAR and SKEW for a single simulation (such as an individual member in Challenge) do not need to be in close agreement with those in the ECA observations, even in a perfect model. Natural variability also makes it difficult to give an adequate interpretation of the differences in the climatological values of the distribution measures we found in Table 4.1 between the Challenge ensemble mean and the ECA data.

The measures we selected to represent the distribution of daily mean temperature proved to be able to separate in a plausible way the characteristics of forced warming from the characteristics of natural variability as derived from the 1946–75 period. Dominant causes for the temperature change between 1976 and 1999 could be identified for all seasons except winter. Although VAR and SKEW are based on the same series of daily mean temperatures, they are independent quantities with respect to MEA. Clearly, VAR is less noisy and therefore generally more suitable than SKEW. Among the possible reasons for the fact that the clearest signals of anthropogenic warming are found in the VAR trend patterns for spring are the amplifying effects of changes in the onset of spring and the amplifying effects of changes in snow cover. The fact that signals of anthropogenic warming are also found in the summer VAR trend patterns may be related to the soil moisture feedback associated with changes in summer precipitation (see Schär *et al.*, 2004).

Future climate change detection/attribution studies may profit from the results of our explorative analysis when choosing the measures that are prime candidates for the fingerprints of forced climate change signals. First studies in this direction (Hegerl et al, 2004) already suggest that changes in soft temperature extremes should be nearly as detectable as changes in mean temperature. The trend patterns of the distribution measures that we identified in the present study may be among the sensitive fingerprints for the early detection of anthropogenic change in climate extremes.

Abstract – Consequences of a Gulf Stream induced ocean surface cooling for the temperature climate of Western Europe were studied by means of a conditional perturbation of the observed daily temperature time series of the Netherlands. On days with advection of airmasses of maritime origin, the observed temperatures in the series were lowered with a fixed value, representing the influence of a cooler Atlantic Ocean. On the other days, the observed temperatures were left unchanged.

The perturbation results in a decrease in the mean temperature that is almost constant over the year, and in a change in the standard deviation of the daily temperatures that is seasonally dependent. Due to preferential cooling of warm winter days, the standard deviation decreases in the winter, whereas in the other seasons the standard deviation increases as a result of preferential cooling of days with low temperatures.

Although this ocean cooling scenario indicates an increase of the relative frequency of cold winters and cool summers, it is neither characterized by the occurrence of winters with unprecedented low temperatures nor by the disappearance of summer heatwaves.

Simple temperature scenario for a Gulf Stream induced climate change

5.1 Introduction

The Gulf Stream/North Atlantic Drift, which transports warm surface water from subtropical regions northwards, determines the mild climate at the west-side of the European Continent. An ocean cooling due to a sudden weakening of the northward water transport would have important consequences for the climate of Western Europe. To provide Dutch policy-makers with a temperature scenario for such a situation, a simple data perturbation study was performed. In this study, the observed time series of daily mean surface air temperature in the Netherlands for the period 1950–94 was perturbed, conditional on the origin of airmasses (continental or maritime) as determined using the observed daily atmospheric circulation patterns. The temperatures on days with airflow from the Atlantic Ocean were lowered with a fixed value, whereas the temperatures on the other days were left unchanged. The cooling in the Netherlands introduced this way is considered to be representative of the cooling of the airmasses overlying the ocean, which in turn is proportional to the assumed change in the sea surface temperature resulting from a weakened northward component of the Gulf Stream.

The methodology implicitly assumes that the temperatures of continental airmasses are not affected by an ocean cooling and that the atmospheric circulation remains unchanged. As such, the results are only first order estimates of possible regional climate changes. In fact, the scenario extrapolates the structure of the present climate, but by doing so it provides an insight in the nature of the temperature changes accompanying an ocean cooling with details that are not easily obtained from climate models or from reconstructions of the climates of the past.

5.2 Background

Various paleoclimatic indices (ice cores, ocean sediments, etc.) reveal that, both during the transition from the last ice age into the present interglacial (Johnsen *et al.*, 1992; Taylor *et al.*, 1993; Bond *et al.*, 1993) and during the last interglacial (GRIP members, 1993; Dansgaard *et al.*, 1993; Keigwin *et al.*, 1994; Field *et al.*, 1994), the climate in the North Atlantic region showed large and rapid changes. It is widely believed that these changes, that took place even within decades (Grootes *et al.*, 1993; Broecker, 1994), are related to changes in the ocean circulation. Similarly, the transitions in the 12th and 19th century, which mark the coming and going of the period with a relatively cold and variable climate known as the Little Ice Age, are brought in connection with changes in the ocean circulation (Meese *et al.*, 1994).

The ocean component that is most important for the present climate of Western Europe is the Gulf Stream/North Atlantic Drift. It consists of surface currents that are driven by westerly trade winds and modified by the earth's rotation and the presence of the continents. Along the European continent they transport warm water northward. This transport is thought to be associated with the Conveyor Belt; the thermohaline circulation system that links the worlds oceans, driven by deep convection at specific regions in the North Atlantic (see e.g. Gordon, 1986). A temporary perturbation in this perhaps delicately balanced ocean-atmosphere system might result in a transition into another ocean circulation mode, with shifted convective regions and a Gulf Stream that is altered in strength and path (Broecker *et al.*, 1985; Manabe and Stouffer, 1988; Held, 1993).

Given the observed changes in the past, natural variability in climate seems capable of inducing a transition between ocean circulation modes. According to ocean model studies (see e.g. Weaver and Hughes, 1994; Rahmstorf, 1994, 1995; Manabe and Stouffer, 1995), relatively small and temporary changes in the fresh water input into the North Atlantic Ocean (from ice melt or increased rainfall) are sufficient to trigger the transition. Both in the observations and in the ocean models, transitions occur on a time-scale of a few decades, so rapid changes in the regional Western European climate might occur (Houghton *et al.*, 1996).

It is possible that the global climate change due to the enhanced greenhouse effect alters the probability of a transition between ocean circulation modes. Several transient warming simulations with coupled ocean-atmosphere models show a weakening of the thermohaline circulation of the oceans (Cubash *et al.*, 1992; Manabe and Stouffer, 1994). In this perspective, the concern about a regional climate cooling in Western Europe induced by a change in the Gulf Stream is part of the global warming issue.

In the present climate of Western Europe, airmass advection from the Atlantic Ocean results in mild days in the winter and cool days in the summer, whereas airmasses originating from the Eurasian continent give rise to the coldest winter days and the warmest summer days. Figure 5.1 presents the annual cycle of the daily mean surface air temperatures in the Netherlands. The top and bottom lines connect the highest and lowest recorded temperatures of each calendar (month-day) date in the 1950–94 time series, whereas the middle line represents the temperature at each calendar date averaged over the 45 yr period. The time series on which the figure is based consists of the spatial averages of the 15 principal meteorological stations in the Netherlands.

From Figure 5.1 it is seen that the extreme temperatures fluctuate more during the continental influenced cold winters and warm summers than during the maritime influenced mild winters and cool summers. The temperature of airmasses overlying the land shows higher variability than the temperature of airmasses overlying the ocean, because of the lower heat capacity of the soil. The steep increase in the lower line at the end of February can be attributed to the snow-albedo feedback over the continent, where a rapid decrease of the snow cover leads to an accelerated warming in early spring.

A cooling of the Atlantic Ocean would result in lower temperatures of airmasses advected to Western Europe with the prevailing western circulations. What happens with airmasses originating from the continent is not *a priori* clear. Model results (Manabe and Stouffer 1988, Figure 22) indicate that the European continent would cool by about half the value of the ocean cooling. In the Gulf Stream induced ocean surface cooling scenario of the present study, it is assumed that the temperature of the continental airmasses are to first approximation not affected by the ocean cooling. This means that the average cooling over Western Europe originates from maritime days only. As is shown below, the mean cooling in our scenario is about half the ocean cooling, like in Manabe and Stouffers (1988) model experiments. This agreement suggests that the assumption that the temperature of continental airmasses is little affected by an ocean cooling is indeed justified.

A changed temperature contrast between the ocean and the continent will certainly affect the atmospheric circulation. However, there are no unambiguous indications from coupled ocean-atmosphere models (see e.g. Schiller *et al.*, 1996) of how atmospheric circulation will change and how, for instance, an ocean cooling will affect the frequency of blockings.

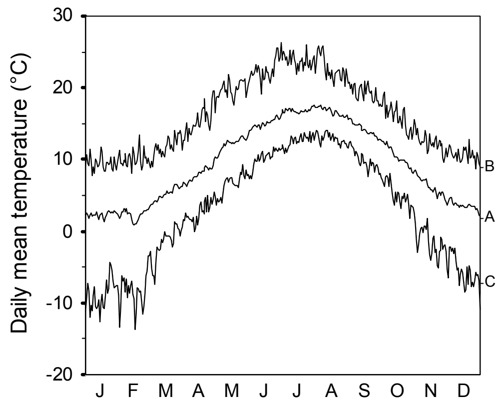


Figure 5.1 Annual cycle of the daily mean surface air temperatures in the Netherlands for the period 1950–94. The middle (A) line represents the averages over the 45 observation years for each calendar (month-day) date. The top (B) and bottom (C) lines represent the highest and lowest temperatures observed for each calendar (month-day) date in the 45 yr time series. J–D stands for January–December.

Figure 5.2 and Table 5.1 show the results of the simplest scenario for an ocean cooling. Here 4°C was subtracted from the observed temperatures on days in the 1950–94 time series with advection of airmasses of maritime origin and the temperatures on days with advection of continental airmasses was left unchanged. This scenario implicitly assumes an unchanged atmospheric circulation. The adopted value, 4°C, of the cooling at maritime days is within the range of ocean surface cooling observed in model experiments by Manabe and Stouffer (1988) or Rahmstorf (1995) and corresponds with the current temperature difference between the North Atlantic and the North Pacific, where no equivalent of the Gulf Stream is found (Levitus, 1994). The maritime days in the time series were selected with the use of Kruizinga's (1979) objective P30-classification of the daily atmospheric circulation at 500 hPa, assigning his classes 1–17, 19, 20 and 22 as maritime. We note that the results are very similar if the daily Grosswetterlagen (Gerstengarbe and Werner, 1993) or even a classification based on the direction of the surface winds in the Netherlands is used for the selection of maritime days.

According to Table 5.1, a 4°C ocean cooling would lead on average to a 2.7°C lower surface air temperature in the Netherlands. This decrease in the mean is almost independent of season, as can be seen from the hardly varying width of Figure 5.2's middle band. In contrast, the widths of the lower and the upper band do show a strong variation throughout the year. The annual courses of the widths of these bands are almost reversed, implying seasonal dependent changes in the standard deviation and in the

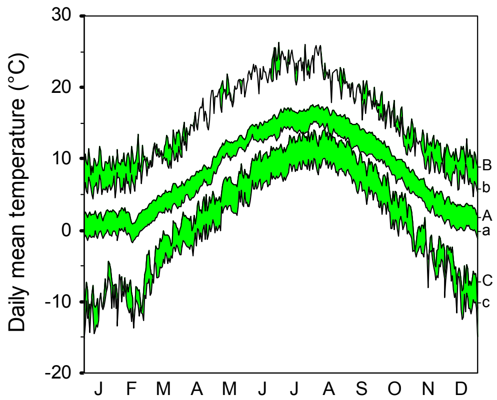


Figure 5.2 As Figure 5.1, but now the temperature decrease (shaded) resulting from the 4°C ocean cooling scenario is added. Lines A, B and C are identical to Figure 5.1. The lines a, b and c have the same meaning as A, B and C, but are obtained from the perturbed time series, in which 4°C was subtracted from the observed temperatures at days with advection of airmasses from the Atlantic Ocean. The middle band defined by the lines (A,a) represents the decrease in the 45 yr averages introduced by the perturbation; the upper (B,b) and lower (C,c) bands represent the decrease in the 45 yr extremes.

extremes of the daily temperatures. In the winter, the lower band is smaller than the upper band. Accordingly, the lower 5% quantile (the temperature below which 5% of the days fall) in Table 5.1 decreases less than the upper 5% quantile (the temperature above which 5% of the days fall). Hence, the standard deviation decreases in the winter. These results mean that an ocean cooling would lead to an increase of the relative frequency of winters with temperatures below the present normals, while at the same time the coldest winters will not be more severe than those of the present climate.

For the other seasons, Table 5.1 indicates that the lower 5% quantiles decrease with almost the full value of the assumed ocean cooling, whereas the upper 5% quantiles change little. Hence, the temperature in these seasons becomes more variable and cold days become colder, but at the same time warm days comparable to those in the present climate continue to occur.

Table 5.1 Changes in the mean, in the standard deviation (s.d.), in the lower 5% quantile and in the upper 5% quantile of the daily temperatures (°C) resulting from the 4°C ocean cooling scenario.

	Winter (DJF)	Spring (MAM)	Summer (JJA)	Autumn (SON)	Year
Δ mean	−2.9	−2.3	−2.6	−3.0	−2.7
Δ s.d.	−0.3	+0.6	+1.0	+0.4	+0.4
Δ lower 5% quantile	−2.2	−3.3	−3.9	−3.2	−3.1
Δ upper 5% quantile	−3.3	−1.2	−0.8	−1.9	−1.8

The values presented in Table 5.1 and Figure 5.2 should be viewed in the perspective of the adopted ocean cooling, whose real magnitude is unknown. But taken together, the results indicate distinct changes in the distribution of the daily temperatures, and the nature of these changes is the same for any adopted cooling.

5.4 Discussion

Our data perturbation study for a weakened northward component of the Gulf Stream indicates that a seasonally independent change in ocean temperature would result in seasonally dependent changes in the temperature climate of Western Europe. An analogue of the climate resulting from a weakened northward component of the Gulf Stream can be the Little Ice Age, which also showed an increased frequency of winters in the Netherlands with temperatures below the present normals. The occurrence of winters with unprecedented low temperatures was not characteristic of that period, neither is it seen in the scenario.

The temperature decrease on the coldest summer days in the scenario is probably not very realistic, because of the assumption of an equal cooling of airmasses over the Atlantic Ocean during the cold and warm part of the year. In fact, in the warm season the ocean is covered by a stable surface layer, hampering the cooling of the overlying atmosphere. Hence, the Gulf Stream determines the sea-surface temperature and thus the air temperature mainly during the cold season.

It should be emphasised that the present scenario represents only a first guess. Two assumptions are made, namely that of an unchanged temperature of continental airmasses and that of an unchanged atmospheric circulation. While the comparison with model results indicates that the former assumption is probably not too bad after all, such a justification lacks for the second assumption. Rather, it seems likely that a transition into a different mode of the Gulf Stream will be accompanied by a significant change in atmospheric circulation. However, simulations with coupled ocean-atmosphere models give no unambiguous indication as to what atmospheric circulation changes will occur in Western Europe in the situation of a weakened Gulf Stream, or how such changes would affect the ratio maritime/continental days. This uncertainty clearly limits the predictive value of this kind of a scenario.

The consequences of the ocean cooling scenario can be compared with that of a transient greenhouse warming, in which the ocean warming lags behind the warming of the continent. A scenario with analogue assumptions as before (unchanged atmospheric circulation and temperature changes of continental airmasses only) results in changes in the standard deviation of the daily temperatures which are similar to those in the ocean cooling scenario, but now the changes arise from cold winters and warm summers instead.

Contrary to the expected warming resulting from the enhanced greenhouse effect, of which the time dependence is predictable, it is impossible to determine whether or not a rapid change in ocean circulation will occur in the near future and whether such a sudden change can actually be realised in response to global warming (Houghton *et al.*, 1996). Little is known about the stability of the present ocean circulation mode and about the conditions favourable for triggering a transition into another mode. Nevertheless, the first-guess scenario presented in this study is answering some of the questions about the nature of a perhaps never occurring rapid temperature change caused by an Atlantic Ocean cooling, and in some sense the simple scenario has a surprising level of detail.

■ Our knowledge on past changes in extremes improved, *as…*

⚌ a new dataset of long daily temperature and precipitation series has become available Chapter 2

The dataset of climatic time series with daily resolution that is compiled for the European Climate Assessment comprises 215 series of minimum, maximum and/or daily mean temperature and 245 series of daily precipitation amount (status March 2004). The series stem from meteorological stations in Europe and the Mediterranean. Most series cover at least the period 1946–99. Almost all series are available for climate change research at www.knmi.nl/samenw/eca. In the near future the series at this website will be blended with daily data derived from synoptic observations that are transmitted on a routine basis over the Global Telecommunication System. In that way, each series is always updated until present.

Because of its daily resolution, the ECA dataset enables a variety of climate studies, including detailed analyses of changes in the occurrence of extremes over the past ~50 yr in relation to changes in mean temperature and total precipitation.

☰ statistically significant and non-trivial changes in extremes have been detected Chapter 3

The European trends (1946–99) in the indices of temperature extremes reflect the general warming: fewer cold extremes, more warm extremes. Averaged over all stations, the first decades of slight cooling saw narrowing of the distributions of daily minimum temperature and daily maximum temperature, whereas the last decades of strong warming saw widening of the distributions of daily minimum temperature and daily maximum temperature. For precipitation, all indices of wet extremes increased between 1946–99 when averaged over Europe. One index suggests that in wetting areas, the changes in the high intensity events were larger than expected on the basis of the changes in the average precipitation amounts, implying an amplified response of the wet extremes.

The indices results obtained in this thesis did already feed into the worldwide studies of changes in indices of extremes as performed for the Third Assessment report of the IPCC (Houghton *et al.*, 2001, Chapter 2, pages 156-159; see also Frich *et al.*, 2002). They will also feed into similar activities that are planned in the preparatory stage of the Fourth Assessment Report AR4.

Quantitative results for trends in indices of temperature and precipitation extremes are important not only from the point of view of climate monitoring, climate change detection/attribution or climate modelling, but also from the impact point of view. Ecosystems and society are generally more sensitive to changes in climate extremes than to changes in mean climate. The ~50 yr time series and moderate trend magnitudes limited the possibilities for trend analysis to extremes with average return period in the order of once each month. Considering extremes further in the tails of the distribution would not have provided an annual number of extremes that is sufficiently large to allow for meaningful trend analysis. Consequently, the extremes that were studied with the aid of descriptive indices generally translate into events of moderate or weak impacts. Nevertheless, decadal scale climate variations in extremes are captured that are important for engineering design practices and societal perception of extremes. For instance, the widening of the temperature distribution in recent decades indicates a coarsening of our climate. Some of the indices are a comprehensive representation of extremes in a form useful for decision makers. The European Environment Agency (EEA) relies on these indices for its European state of the environment reports, which are issued at regular intervals and aim to support sustainable development (EEA, 2004).

≡ candidate fingerprints have been identified for the early detection of anthropogenic change in climate extremes Chapter 4

Some of the characteristic patterns of the European warming between 1976 and 1999 provide a signal of anthropogenic influence, as they are distinct from the estimated patterns associated with natural temperature variability and similar differences are seen in the climate model simulations of the Challenge experiment, 1940–2080 (Selten *et al.*, 2004). Therefore, the identified patterns are good candidates for fingerprints in formal climate change detection/attribution studies.

Formal detection/attribution of anthropogenic climate change involves comparing the evolution of the most likely signal from anthropogenic change (the fingerprint) in GCM simulations of climate under changing greenhouse gas concentrations against the observations (IDAG, 2004). Such studies did form the basis for the strong conclusion in the third assessment report of the IPCC (Houghton *et al.*, 2001) that 'there is new and stronger evidence that most of the warming observed over the last 50 years is attributable to human activities'. Detection/attribution studies to date have focussed more on mean climate than climate extremes. Recent work (Zwiers *et al.*, 2003) shows that the potential for detection/attribution focussing on fingerprints of extremes is yet to be explored.

The fingerprint candidates derived in this thesis from the ECA dataset are based on carefully selected measures that describe the distribution characteristics of daily mean temperature (expressed by the 'soft' extremes: 10th percentile and 90th percentile). The clearest examples are the observed 1976–99 trend patterns for the spring and summer daily temperature variance. They are markedly different from the patterns for the change in variance associated with natural variability in the observations (which were empirically derived from the observations in the 1946–75 period). The potential for the variance trend patterns being sensitive fingerprints of anthropogenic warming follows from the differences that are seen for the same measure between the patterns in the greenhouse warming simulation and the control simulation of the Challenge ensemble experiment. The fact that the observed 1976–99 trend pattern for the autumn temperature variance is identified as a likely pattern associated with natural variability supports the discriminating power of the method that was used.

Although the carefully selected measures refer to extremes for which significant changes are in principal more difficult to detect than for the mean, the selected measures are based solely on the series of daily mean temperature, which makes them less sensitive for nonclimatic inhomogeneities than fingerprints that involve maximum and/or minimum temperature.

≣ an empirical temperature scenario has been developed for a shutdown of the thermohaline circulation Chapter 5

The scenario for the temperature climate of Western Europe resulting from a weakened Gulf Stream shows a decrease in mean temperature that is almost constant over the year, whereas the changes in the variance and extremes of daily temperature are seasonally dependent. Preferential cooling is induced of warm winter days and cool summer days, as only the temperature at days with circulation bringing airmasses from the ocean were perturbed. The extremes in the far tails of the overall distribution (coldest winter days and warmest summer days) are not affected by the perturbation. Their frequency remains unchanged, which implies that the daily variance increases in summer but decreases in winter.

The Gulf Stream scenario is plausible and internally consistent, while based on observational series of the past. It contains details on changes in temperature extremes that are valuable information for sensitivity and impact studies. Such details were up to now unavailable to decision makers. The changes in temperature extremes in the scenario show similar characteristics as in the winter warming example of Chapter 2. Both in the Gulf Stream scenario and in the observed warming of recent decades the change in the coldest winter extremes is different from the change in the mean or the warm winter extremes, since the coldest winter extremes are not affected by the cooling (Chapter 5) or warming (Chapter 2). As seen in Chapter 3, similar characteristics of differential warming of the cold and warm tails of the distribution were detected in the observations of the past decades when averaged over Europe and the year.

The Gulf Stream scenario is only valid under the assumption of unchanged atmospheric circulation, because it is derived on the basis of the empirical relation between wind flow and temperature. Changing atmospheric circulation would change the wind flow characteristics and thus the temperature climate for another reason than that of the changing Gulf Stream. Assuming unchanged circulation conditions means that the multidecadal and shorter time-scale variations that are reflected in the instrumental record are preserved. Because climate models (both GCMs and RCMs) do not yet provide reliable indications about systematic changes in atmospheric circulation at the regional and local scale (Houghton *et al.*, 2001), the assumption of unchanged atmospheric circulation is often made when developing scenarios for climate change studies. For instance Buishand and Klein Tank (1998) make use of the relation between daily temperature and precipitation amount in the series of station De Bilt (the Netherlands) to derive a precipitation scenario for a warming climate under unchanged circulation conditions. Both this precipitation scenario and the Gulf Stream temperature scenario is part of a suite of climate change scenarios that is issued by KNMI for the Dutch impact community.

◻ to keep its value for climate change research, the new daily dataset needs to be updated at a regular basis, as well as further quality controlled and elaborated with metadata information Chapter 2

In any long time series, changes in routine observation practices may have introduced inhomogeneities of nonclimatic origin that severely affect the extremes. Wijngaard *et al.* (2003) statistically tested the daily ECA series (1901–99) of surface air temperature and precipitation with respect to homogeneity. A two-step approach was followed. First, four homogeneity tests were applied to evaluate the daily series using the testing variables: (1) the annual mean of the diurnal temperature range (= maximum temperature – minimum temperature), (2) the annual mean of the absolute day-to-day differences of the diurnal temperature range and (3) the wet day count (threshold 1 mm). Second, the test results were condensed for each series into three classes: 'useful–doubtful–suspect'.

In the period 1901–99, 94% of the temperature series and 25% of the precipitation series were labelled 'doubtful' or 'suspect'. In the subperiod 1946–99, 61% of the temperature series and 13% of the precipitation series were assigned to these classes. The seemingly favourable scores for precipitation are due to the high standard deviation of the testing variable and hence to the inherent restricted possibilities for detecting inhomogeneities and trends. About 65% of the statistically detected inhomogeneities in the temperature series labelled 'doubtful' or 'suspect' in the period 1946–99 were traced back to observational changes that are documented in the metadata. For precipitation this percentage was 90%. The remainder of the detected inhomogeneities may still be real climate related changes. This illustrates that metadata are essential for climate observations. In their absence, exhaustive investigations are required to find, compile, and integrate information on how, where and when observations were taken in order to effectively interpret the data.

The high percentage of series labelled 'doubtful' or 'suspect' indicates that only part of the ECA dataset is adequate for reliable trend analysis. For this reason, only a subset of station series is used throughout this thesis. But even though data scrutiny has been careful, it cannot be excluded that results for individual stations may be affected by inhomogeneities in the underlying series that were not (yet) detected. Conclusions are therefore only drawn from tendencies over large areas and Europe-average trends.

In Wijngaard *et al.* (2003) statistical techniques for homogeneity testing were applied that were originally developed for testing of monthly or lower resolution data. They resulted in quality labels and subsequent subselections of data series, as described. Techniques for detection of discontinuities in monthly or lower resolution data are widely available (for a recent review see Ducré-Robitaille *et al.*, 2003). There is a clear need however for additional research on techniques for homogenization of daily data in order to create high quality daily datasets for the assessment of extremes without abandoning entire series or throwing out real extremes. This is of particular importance in areas where the density of stations with long daily data series is already low.

The ECA dataset proves that often the climate data do exist, but are of insufficient quality, unavailable to the scientific community or scattered over many data holders. Climate monitoring principles (like the GCOS principles under the UN Framework Convention on Climate Change) are intended to improve this situation. But adoption of such principles (for instance with the intention to establish the GCOS Surface Network) is insufficient without proper implementation. And current international implementation of climate monitoring principles is spotty at best (Karl and Trenberth, 2003). At present, scientists are struggling to address all aspects of data archaeology. This includes retrieving data inaccessible because the recording media are outdated, national data exchange is restricted, or resources are inadequate to make them easily accessible (GCOS, 2003). National Meteorological and Hydrological Services play a key role in providing necessary and adequate climate information for research and impact assessment. They should all adopt the principle that the more their data are used and scrutinized, the more valuable they become, both for the originator and all subsequent users (IPCC, 2002).

Ξ the physical causes of the observed trends in extremes are unresolved even though the trends are statistically significant Chapter 3

The trends in the indices of temperature and precipitation extremes were determined using ordinary least square regression with time as predictor. Estimates of trend significance from standard Student's *t* tests provided information on the probability that an observed trend is caused by chance even though it passed certain detection thresholds. More advanced statistical estimates of trends and their significance (e.g. using the nonparametric Mann-Kendall rank test) would have resulted in slightly more reliable results, but a comparison (not shown) proves that the conclusions would not have changed.

To analyse the changes in extremes, no use was made of extreme value theory from statistics. Instead, trends were calculated in descriptive indices that describe various characteristics of extremes, including frequency, amplitude and persistence. The indices are simple to understand and standardization makes that the results of all countries are comparable to each other and European results to other regions of the world. But the indices only represent a limited number of all possible statistical characteristics of extremes. Projecting the indices on more formal measures of extremes that are common in extreme value theory will provide insight in the possibilities for improved indices definitions. Such an activity is recently started under the umbrella of the new Expert Team on Climate Change Detection, Monitoring and Indices of WMO/CCL and CLIVAR (Zwiers *et al.*, 2003).

In the pragmatic approach followed throughout this thesis, trends are calculated for any specified observation period. However, an important constraint is that trend detectability restricts the choice of time scale and event extremity according to the fundamental relation between these quantities provided in Chapter 3 (Eq. 3.3). The indices of temperature and precipitation extremes were defined with this relation in mind, in order to facilitate detection of significant trends.

The physical mechanisms behind the detected trends in the indices of extremes are unknown as yet. Earlier studies (e.g. Giorgi, 2002; Walsh *et al.*, 2001) show that changes in climate at the regional and local spatial scale are largely the result of variability in large-scale atmospheric circulation. This means that station trends in temperature and precipitation can partly be understood by studying the correlation with accompanying changes in large-scale atmospheric circulation. According to Hurrell (1996) part of the warming over Europe and Asia and associated cooling over the Atlantic at the end of the 20th century can be 'explained' by the recent upward trend in the North Atlantic Oscillation (NAO) index. Van Oldenborgh and Van Ulden (2003) conclude that the observed local temperature increase in the Netherlands, situated on the edge of the European continent, is associated with a combination of uniform large-scale warming and changes in the frequency distribution of circulation patterns. They found that about half of the yearly averaged temperature increase (~0.8°C) for station De Bilt over the 20th century is explained by an increase in average temperature per wind direction, roughly a quarter by an increased frequency of south-westerly wind in February to April, and the remainder by other, random, weather variations.

This thesis only dealt in a rudimentary and qualitative way with the question how the trends in the indices of daily temperature and precipitation extremes are related to accompanying changes in large-scale atmospheric circulation patterns, like NAO, or how they are related to small-scale variability in circulation and wind. This question needs to be further addressed in future studies. Dependent on the answer, the observed trends in extremes are either caused by changing atmospheric pressure and wind patterns over Europe or by other changes in climate, such as anthropogenic warming. But by establishing the correlations with atmospheric circulation, attribution of climate change to anthropogenic causes will not be settled, since the changes in air pressure and wind patterns may be a result of anthropogenic greenhouse gas build up in the atmosphere as well.

≡ formal climate change detection/attribution studies require fingerprints that are well represented by GCMs Chapter 4

The choice in this thesis for a wide definition of a 'trend', causes that the attribution question remains: what portion of the recent trends in the indices for extremes can be attributed to the effect of anthropogenic warming and what portion must be regarded natural climate variability (either internal or due to other external forcing factors, such as solar irradiance)? In formal detection/attribution studies the fingerprints of changes that result from the influence of mankind are usually derived from simulations with GCMs and compared with observations. Best estimates are obtained using comprehensive global models of the full climate system. Such models are the fundamental tools for understanding and predicting natural climate variability and providing reliable estimates of anthropogenic climate change. They are able to deal with issues related to: a changing chemical environment, nonlinearity in the response to forcings both at the global and regional scale (e.g. the hydrological feedback between the soil and the atmosphere), quantitative estimates of uncertainty guided by observations, and nonlinear feedbacks between climate and the impacts of climate change (e.g. water resource management, changes in land use, energy needs).

In this thesis no direct comparisons were made between the fingerprints in the climate model data generated by the Challenge experiment and the imprints of anthropogenic warming that were seen in the observations. Although these imprints in the form of the trend patterns of the daily temperature variance provide promising new candidate fingerprints, a preliminary analysis of the climate model data suggests that the model is not suitable for formal detection/attribution. The variance trend patterns observed in the ECA station data are not satisfactory reproduced by the (older version of the) NCAR GCM used in Challenge.

The ensemble simulations of Challenge do clearly indicate that for an adequate representation of the daily temperature variance in Europe by more state-of-the-art GCMs it is essential that these GCMs account for the natural variability that exists under the same external forcings (Selten *et al.*, 2004). Only then valid comparisons can be made between the trend patterns in the climate model and the observed trend patterns. A complicating factor is that comparisons between area-based extreme events in climate models and extremes in station observations require some form of area averaging (gridding) of the point observations in order to adjust for the different scales involved (Osborn and Hulme, 1997). Gridding of the daily resolution ECA series will be part of the European Union Integrated Project ENSEMBLES, which will start in the autumn of 2004 and run for 5 years.

In the meantime, station monitoring of the measures describing the distribution characteristics of daily mean temperature as identified in Chapter 4 as well as station monitoring of the indices of temperature and precipitation extremes from Chapter 3 contributes to better understanding and appreciation of their potential future role in detection/attribution of anthropogenic climate change. Near real time station monitoring of extremes using ECA series that are updated with daily data from synoptic observations is part of the ECA project plans for the coming years. This monitoring effort will serve as show case in the framework of WMO's World Climate Data and Monitoring Programme.

≣ the Gulf Stream scenario describes one of many possible futures; we still don't know how extremes will change Chapter 5

Although the Gulf Stream scenario provides a plausible representation of the state of the future climate, it must be emphasised that the Gulf Stream scenario is only one of many possible future climate scenarios. Current GCMs and RCMs are more certain about other climate changes for Europe until the end of the 21st century, resulting from enhanced greenhouse gas concentrations in the atmosphere (Houghton *et al.*, 2001). Mean temperature is predicted to increase between 1990 and 2100 with somewhat higher values than the predicted increase in the global mean of 1.4–5.8°C, with low and high estimates covering model and emission scenario uncertainties. The predicted values vary per region and season. Highest temperature increases are predicted for northern European winter and southern European summer. Along the Atlantic coast the temperature rise will be similar to the increase in the global mean. In accordance with the average warming, the probability of summer heat waves will increase and the probability of winter frost days will decrease. Precipitation in the northern half of Europe is predicted to increase by 5–20% in winter, partly as a result of a higher number of extreme intensities. Precipitation in the southern half of Europe will decrease by 20% or more in summer, leading to a higher risk of summer drought. At present, no further predictions are available for the changes in the 'soft' extremes of temperature and precipitation that are the subject of this thesis.

In first order, the projections of the Gulf Stream scenario are additive to the changes indicated above. Only the exact timing of rapid climate change, such as the shutdown of the thermohaline circulation in the Gulf Stream scenario, is impossible to predict in the nonlinear climate system. At best, ensemble simulations with state-of-the-art GCMs provide probabilistic information on the question whether climate could make such a sudden switch. For this reason, the Gulf Stream scenario is not bound to a predictable moment in time, but can occur at any moment. Studies using paleodata indicate that an abrupt transition in the oceanic circulation is possible and could be completed typically in 10 years (Committee on Abrupt Climate Change, 2003).

Personal notes:

- *The risks associated with abrupt events involving changes in the Gulf Stream or surprises whose mechanisms are not (yet) understood, as well as the risks associated with changes in 'soft' extremes of temperature and precipitation warrant measures for reducing anthropogenic greenhouse gas emissions to the atmosphere. In the author's opinion these measures should be much more radical than currently negotiated, let alone implemented.*

- *Science continues to play a key role in underpinning adequate climate change policy by providing accurate and complete information on changes in extremes. This includes characterization and communication of the accompanying uncertainties and of the changes that can be ruled out as unlikely (see e.g. Allen and Ingram, 2002). This thesis contributes by providing new findings on changes in extremes of daily surface air temperature and precipitation as observed in Europe's climate of the 20th century. Better understanding of the causes of the observed trends needs to be given high priority on the (KNMI) research agenda, with the ultimate goal of better prediction of the occurrence of extremes in the future.*

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The variance of the estimated regression coefficient in event-count records

For an uninterrupted record, the variance of the least squares estimate $\hat{b}$ of the regression coefficient b in the model

$$Y_j = a + bj + e_j \qquad\qquad j = 1,\ldots,N \qquad\qquad \text{(A1)}$$

is given by

$$\sigma_{\hat{b}}^2 = \frac{\sum_{j=1}^{N}\left[\left(j-\bar{j}\right)^2 \operatorname{var} e_j\right]}{\left[\sum_{j=1}^{N}\left(j-\bar{j}\right)^2\right]^2} \qquad\qquad \text{(A2)}$$

where $\bar{j} = \frac{1}{2}(N+1)$ is the average of the year index j. This expression can easily be simplified in the case that var e_j is constant (e.g., Lettenmaier 1976; Kendall *et al.* 1983, Section 45.23):

$$\sigma_{\hat{b}}^2 = \frac{12 \operatorname{var} e_j}{N\left(N^2-1\right)} \approx \frac{12 \operatorname{var} e_j}{N^3} \qquad\qquad \text{(A3)}$$

For event-count records the error variance depends, however, on the underlying trend. If the successive events are independent, then the X_js have approximately a Poisson distribution with mean $a + bj$. Since the variance of a Poisson variable is equal to its mean, we have to substitute $a + bj$ for var e_j into Eq. (A2), giving:

$$\sigma_{\hat{b}}^2 = \frac{a + b\bar{j}}{\sum_{j=1}^{N}\left(j-\bar{j}\right)^2} + \frac{b\sum_{j=1}^{N}\left(j-\bar{j}\right)^3}{\left[\sum_{j=1}^{N}\left(j-\bar{j}\right)^2\right]^2} \tag{A4}$$

Because

$$\sum_{j=1}^{N}\left(j-\bar{j}\right)^2 = \frac{1}{12}N\left(N^2-1\right) \approx \frac{1}{12}N^3$$

$$\sum_{j=1}^{N}\left(j-\bar{j}\right)^3 = 0$$

and $\overline{X}$ is the least squares estimate of $a + b\bar{j}$, Eq. (A2) reduces to

$$\sigma_{\hat{b}}^2 \approx \frac{12\overline{X}}{N^3} \tag{A5}$$

Consequently, in the model

$$Y_j = a + 0.1\, bj + e_j \qquad\qquad j = 1,\ldots,N \tag{A6}$$

that we apply in Chapter 3; Eq. (3.2):

$$\sigma_{\hat{b}}^2 \approx \frac{1200\overline{X}}{N^3} \tag{A7}$$

Albert Klein Tank was born in Groenlo, Gelderland, in the eastern part of the Netherlands on 15 May 1965. He attended Marianum VWO school between 1977 and 1983 and then went to Utrecht University for a degree in Physical Geography. He took courses in geology, geomorphology, hydrology, statistics and environmental sciences. During a 9 months period in 1988, Albert visited Battelle Pacific Northwest Laboratories near Seattle (Washington, USA) to study trans-border acidification of Canadian lakes due to US sulfate and nitrate emissions. Returning to Utrecht, he concluded university in 1989 after completing his final work on the effect of acid rain on forests in Europe.

Albert fulfilled his (at that time mandatory) civil service at the Advisory Council for Research on Spatial Planning, Nature and the Environment (RMNO) in Rijswijk. For almost 2 years he stayed at the secretariat of this advisory council, preparing research policy advises for the national government. The advise on gamma aspects of climate change research brought him into contact with KNMI. In 1991, he started to work in KNMI's Climate Analysis Division on a climate scenario project funded by the National Research Programme on Climate Change (NRP). Empirical scenarios were developed that represent plausible future climate states for the Netherlands. The scenarios still serve as prime future climate information within the Dutch impact community.

In 1995, Albert continued his career at KNMI in a permanent position within the Climatological Services Division. Since then, he has initiated and led several R&D projects that support KNMI's routine day-to-day information and data services about the weather of the past to the general public and specific user groups. At present, he is leading the European Climate Assessment & Dataset project (www.knmi.nl/samenw/eca) that joins over 40 meteorological services and research centres in Europe and the Mediterranean. The scientific contacts resulting from this project formed the nucleus for the present PhD work. Albert has recently spent several months at WMO in Geneva in the position of scientific content coordinator of the 7th Global Climate System Review and the Statement on the Status of the Global Climate in 2003. He is a coordinator in the European Union Framework Programme 6 Integrated Project ENSEMBLES, which is planned to kick off in the autumn of 2004 and will run for 5 years. Finally, Albert is invited to serve as a Lead Author for the IPCC Working Group I Fourth Assessment Report, which is due in 2007.

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Samenvatting

Steeds vaker wordt beweerd dat ons klimaat extremer wordt door het broeikaseffect. In veel gevallen is de aanleiding een actuele gebeurtenis door gevaarlijk weer. Denk maar aan de hete zomer van 2003 met zijn vele hittedoden of aan de hevige regenval in de zomer van 2002 die de rivier de Elbe deed overstromen. Gewoonlijk blijft de wetenschappelijke onderbouwing van de bewering dat ons klimaat extremer wordt door het broeikaseffect achterwege. Er is immers nog maar weinig bekend over systematische veranderingen in de extremen van ons klimaat, laat staan over de precieze oorzaken of over de invloed van de mens. Terwijl die informatie in feite heel belangrijk is. Onze samenleving is immers veel kwetsbaarder voor veranderingen in extremen dan voor geleidelijke veranderingen in de normale situatie, zoals de langzame stijging van de gemiddelde temperatuur op aarde.

Dit proefschrift gaat over veranderingen in extremen en levert een bijdrage aan de kennis op dat terrein van de klimaatverandering. Het probeert een antwoord te vinden op de centrale vraag:

- **Hoe zijn de extremen van temperatuur en neerslag in Europa in de afgelopen eeuw veranderd, en wat kunnen we daaruit afleiden?**

Extremen worden omschreven als die waarden van temperatuur en neerslag die in het huidige klimaat uitzonderlijk zijn. Voorbeelden zijn een voorjaarsachtige temperatuur midden in de winter, een ijskoude en gure dag in het voorjaar, een wolkbreuk in de zomer of langdurige stortregen in de herfst. Dit betekent dat niet alleen extremen met catastrofale gevolgen zijn meegenomen, maar ook andere intense gebeurtenissen. Het aantal extremen dat zo wordt onderzocht is groter en dat leidt tot een grotere betrouwbaarheid van de uitkomsten. Het onderzoek is uitgevoerd op basis van lange reeksen waarnemingen die door de jaren heen zijn verricht op weerstations in Europa.

Achtereenvolgens komen in dit proefschrift de volgende vragen aan bod:

1. **Is voldoende gedetailleerde meetinformatie beschikbaar om de opgetreden veranderingen in extremen te kwantificeren?** Hoofdstuk 2

2. **Welke trends zijn te vinden in de meetreeksen van de dagelijkse temperatuur en neerslaghoeveelheid?** Hoofdstuk 3

3. **Kunnen de veranderingen in extremen toegeschreven worden aan de menselijke invloed op het klimaat?** Hoofdstuk 4

4. **Helpen de waargenomen veranderingen bij het opstellen van mogelijke toekomstscenario's?** Hoofdstuk 5

Op elk van deze vragen kon een antwoord worden gegeven. Die antwoorden worden nu kort besproken.

1. Nieuwe Europese dataset ontsluit gedetailleerde meetinformatie Hoofdstuk 2

Een kwalitatief hoogwaardige Europese dataset met lange reeksen van temperatuur- en neerslagmetingen bleek niet beschikbaar. Eerdere studies maakten voornamelijk gebruik van maandgegevens. Maar met reeksen op maandbasis kunnen extremen, die meestal van korte duur zijn, niet goed worden bestudeerd. Daarom is speciaal voor dit onderzoek een nieuwe Europese dataset met reeksen op dagbasis gemaakt. Dit lijkt eenvoudiger dan het is, omdat voor Europa de verantwoordelijkheid voor het archiveren en beheren van klimatologische reeksen vooral op nationaal niveau ligt. Veel (deels financiële) hobbels beletten een routinematige uitwisseling van gegevens. Maar uiteindelijk is het gelukt en de nieuwe dataset met een groot aantal dagreeksen van temperatuur en neerslag is nu voor iedere onderzoeker beschikbaar via www.knmi.nl/samenw/eca.

2. Opwarming aarde ook zichtbaar in extremen Hoofdstuk 3

Op basis van de nieuwe Europese dataset met dagreeksen van temperatuur en neerslag zijn trends in extremen vastgesteld. Om de extremen systematisch te onderzoeken is gebruik gemaakt van zogenaamde indicatoren. Meestal waren dat tellingen van dagen die aan een bepaald criterium voldoen, zoals het jaarlijks aantal dagen met vorst of het aantal dagen met meer dan 20 mm regen. In onderling overleg met wetenschappers in andere werelddelen zijn dezelfde indicatoren ook daar toegepast, zodat nu een wereldwijd beeld van de trends beschikbaar is.

Voor Europa laten de indicatoren zien dat er over de afgelopen 50 jaar steeds minder koude extremen en meer warme extremen voorkwamen. De opwarming van de aarde komt dus ook in de extremen tot uiting. Opmerkelijk is dat de gebruikte indicatoren voor de temperatuurextremen laten zien dat de grootte van de veranderingen in de extremen niet volledig in de pas loopt met de stijging van de gemiddelde temperatuur. Zo droeg de afgelopen decennia de toename van het aantal relatief warme dagen meer bij aan de opwarming dan de afname van het aantal relatief koude dagen. Die laatste stagneerde een beetje, waardoor in zijn geheel meer temperatuuruitschieters zijn ontstaan. Je zou kunnen zeggen dat het klimaat is verruwd.

Voor de neerslag laten de indicatoren voor de natte extremen gemiddeld over Europa bijna allemaal een toename zien. Dit betekent dat de neerslag intensiever en de buien heviger zijn geworden.

Aanwijzing voor menselijke invloed Hoofdstuk 4

De vervolgstudie leverde aanwijzingen op dat de vastgestelde veranderingen in Europese temperatuurextremen in direct verband kunnen worden gebracht met de menselijke invloed op het klimaat.

Bekend was al dat de mens in staat is het natuurlijke broeikaseffect te versterken door de grootschalige uitstoot van broeikasgassen. Tot voor kort kon die menselijke invloed op het klimaat uitsluitend worden aangetoond in de wereldwijde opwarming of in de gemiddelde temperatuurstijging over de grote continenten. Uit die studies bleek dat het merendeel van de wereldwijde opwarming sinds het midden van de 20^e^ eeuw is veroorzaakt door de menselijke invloed. Het resterende deel van de opwarming werd beinvloed door fluctuaties in de zonneactiviteit, vulkaanuitbarstingen en El Niño.

Bepaling van de menselijke invloed op de extremen van ons klimaat is ingewikkelder. Extremen zijn er immers per definitie maar weinig en de meeste extreme gebeurtenissen uit het recente verleden zijn in een verder verleden al eens eerder voorgekomen. Toch is de ongelijke verandering in relatief koude en warme dagtemperaturen, zoals die werd gevonden bij de trendstudie, mogelijk een vingerafdruk van de menselijke invloed op het klimaat. Die vingerafdruk wordt zichtbaar in de vorm van een karakteristiek geografisch patroon voor de trends over Europa. Dat karakteristieke patroon is anders dan het geografische patroon dat hoort bij een opwarming door natuurlijke oorzaken.

De belangrijkste reden om ons huidige klimaat en dat van het verleden te bestuderen is het maken van een voorspelling voor de toekomst. Het domweg doortrekken van de gevonden trends in extremen naar de toekomst is uit den boze, zolang we er niet zeker van zijn dat het fysisch mechanisme of de oorzaak die zorgt voor de gemeten veranderingen, ook in de toekomst dezelfde kant op blijft werken. Dit hoeft lang niet altijd het geval te zijn. Zo zou de bodem in toekomstige zomers door de hogere temperaturen wel eens volledig kunnen uitdrogen. Dan geeft de bodem geen vocht meer af aan de atmosfeer, waardoor de verwachte extreme zomerbuien boven centraal Europa grotendeels verdwijnen in plaats van heviger worden.

In het klimaat bestaan een groot aantal fysische mechanismen die ook nog eens allemaal op elkaar ingrijpen. Daarom werken klimaatonderzoekers met computermodellen waarmee ze het heden, het verleden en de toekomst nabootsen. Die modellen zijn nodig voor gefundeerde toekomstvoorspellingen. Maar ook zonder die modellen kan wel degelijk iets zinnigs worden gezegd over toekomstige scenario's, bijvoorbeeld op basis van onderlinge verbanden tussen weerkenmerken in de meetreeksen uit het verleden.

Zo'n alternatief scenario is hier gemaakt voor de temperatuur. Het beschrijft gedetailleerd wat er gebeurt met de temperatuur in West Europa als de Noord-Atlantische Golfstroom verandert. In het huidige klimaat transporteert de Golfstroom warm tropisch oceaanwater naar het noorden. De overheersende westenwinden op het Noordelijk Halfrond nemen die warmte mee naar het Europese continent. Dat levert een mild klimaat op. Een verzwakking van de warme Golfstroom zou in West Europa leiden tot een vrij abrupte en aanzienlijke verlaging van de temperatuur. Het zeer gedetailleerde scenario voor deze situatie laat zien dat dan vooral de relatief warme winterdagen en koele zomerdagen zullen afkoelen, terwijl de temperatuur op de koudste winterdagen en warmste zomerdagen ongewijzigd blijft.

Hoewel het onderzoek van dit proefschrift nieuwe kennis heeft opgeleverd over veranderingen in de extremen van ons klimaat, is het onderzoek nog verre van volledig.

Nieuwe Europese dataset vergt onderhoud Hoofdstuk 2

Zo moet de nieuwe Europese dataset steeds weer worden geactualiseerd en moet de kwaliteit van de dagreeksen van temperatuur- en neerslagmetingen regelmatig worden gecontroleerd. Alleen op die manier behoudt de dataset zijn waarde voor onderzoek.

Nader onderzoek nodig naar oorzaken trends Hoofdstuk 3

Voor een beter begrip van de vastgestelde veranderingen in extremen is een nadere analyse nodig van de achterliggende fysische mechanismen die de trends veroorzaken. Dat is ook noodzakelijk om te begrijpen waarom het klimaat op bepaalde plaatsen in Europa anders reageert op de gemiddelde opwarming dan elders. Bekend is dat lokale klimaatfluctuaties ten dele samenhangen met veranderingen in overheersende luchtstromingen. Onbekend is nog in hoeverre die ook de vastgestelde trends in extremen kunnen verklaren.

Modelstudies moeten menselijke invloed bevestigen Hoofdstuk 4

Willen we de veranderingen in extremen gebruiken om ondubbelzinnig de menselijke invloed op het klimaat vast te stellen, dan moeten de computermodellen in staat zijn de trends uit het verleden goed na te bootsen. In combinatie met de toekomstprojecties van dezelfde modellen zou dan over enige tijd het bewijs kunnen worden geleverd dat de menselijke invloed niet alleen zichtbaar is in de gemiddelde opwarming, maar ook in de veranderingen in de extremen van ons klimaat. Zover is het nog niet. Maar de gevonden vingerafdruk in de vorm van het karakteristieke geografisch trendpatroon is een belangrijke eerste stap in die richting.

Meer zekerheid over geleidelijke klimaatveranderingen Hoofdstuk 5

Tot slot is het scenario van de warme Golfstroom slechts één van de vele mogelijke scenario's voor toekomstige veranderingen in extremen. De kans dat zo'n situatie nog in deze eeuw optreedt wordt door wetenschappers als heel klein ingeschat. Als er geen onverwachte dingen gebeuren is voor Europa tot het jaar 2100 (t.o.v. 1990) een geleidelijke opwarming van het klimaat die iets groter is dan de wereldwijde temperatuurstijging van 1–6°C het meest plausibel. In samenhang daarmee zal de neerslag in het noorden van Europa met 5–20% toenemen en in het zuiden met 20% afnemen. De grote marges in deze cijfers zijn te wijten aan onzekerheden in de gebruikte computermodellen en de onmogelijkheid om nauwkeurig te voorspellen hoeveel broeikasgassen de mens in de toekomst zal gaan uitstoten. Ten opzichte van deze toekomstige veranderingen in het gemiddelde klimaat is nog lastiger in te schatten hoe in de komende honderd jaar de weerextremen in Europa precies gaan veranderen.

Dankwoord

Tegenwoordig schrijf je een proefschrift aan het begin van een wetenschappelijke carriere, meteen na het behalen van de universitaire graad. Dat het bij mij pas veel later zover is gekomen brengt nadelen met zich mee, maar ook voordelen. Zo heb ik de afgelopen jaren veel ervaring kunnen opdoen met projecten op het scheidingsvlak van operationele klimatologie en wetenschappelijk onderzoek. Kennis die ik goed heb kunnen gebruiken bij het onderzoek voor dit proefschrift. Maar omdat in een operationele omgeving de korte termijn over het algemeen prevaleert boven de lange termijn, schoof het proefschrift vaak door naar het tweede plan. Dit gebeurde bijna automatisch, ondanks de goede wil en inzet van veel vrije tijd. Daar is nu verandering in gekomen en dat is zeker niet alleen mijn eigen verdienste.

Ik bedank alle KNMI collega's voor hun hulp en voor de prettige samenwerking, maar vooral ook voor het plezier dat ik altijd heb gehad in mijn werk en hopelijk nog vele jaren mag hebben. In het bijzonder wil ik Günther Können bedanken voor zijn stimulerende rol als begeleider van dit onderzoek. De discussie-sessies en ideeënuitwisseling die hebben geleid tot onze gezamenlijke artikelen vormen de basis voor dit proefschrift. Günther was altijd gedetailleerd en kritisch in zijn commentaar, maar bovenal ook gedreven, creatief en 'to the point', met oog voor de grote lijn en de beste structuur. Zonder zijn stimulans zou de uiteindelijke afronding van dit proefschrift waarschijnlijk nog enkele jaren op zich hebben laten wachten. Mijn promotor Gerbrand Komen dank ik voor het bieden van de mogelijkheid om bij hem te promoveren. Ik waardeer het bijzonder dat hij ondanks alle drukte toch regelmatig de tijd vond om mij op het goede spoor te houden. Speciale dank gaat ook uit naar Aryan van Engelen. Hij heeft zich altijd ingezet voor een nationaal en internationaal geörienteerde R&D tak binnen de Klimatologische Dienst en hij heeft mij optimaal gestimuleerd om initiatieven in die richting te ontplooien en uit te werken in concrete projecten. Als afdelingshoofd gaf hij bovendien alle ruimte voor mijn promotie-onderzoek. Verder dank ik Adri Buishand voor de conscientieuze wijze waarop hij door de jaren heen zijn statistische kennis heeft proberen over te dragen en concreet voor de afleiding van vergelijking (3) en (4) in de Appendix. Harry Geurts dank ik voor de tekstsuggesties bij de Nederlandse samenvatting. De hardlopers hebben door de jaren heen gezorgd voor sportieve afleiding.

I am grateful to the scientists with whom I have had (and still have) the pleasure to work in international projects, expert teams and working groups. In particular the participants of the European Climate Assessment (ECA) project have contributed greatly to the success of this thesis. ECA was initiated by the European Climate Support Network (ECSN) and supported by the Network of European Meteorological Services (EUMETNET). Following the recommendations of the Asheville IPCC workshop organized by Tom Karl and the Abisko EUROCLIVAR workshop organized by Lennart Bengtsson and Gerbrand Komen in 1997, ECA aimed at developing a daily dataset of observations at European meteorological stations, analyzing daily extremes, and subsequently disseminating the data and the analysis results. With the intention to build on the success of the 1995 assessment report by Cor Schuurmans, the former KNMI director Harry Fijnaut was keen on playing a leading role in ECA. As a result, EUMETNET appointed KNMI as responsible member for this project in 1998. I thank KNMI for giving me the responsibility of coordinating ECA and providing ample opportunity to travel abroad and attend conferences and workshops. I enjoyed the privilege of giving presentations and meeting colleagues, which has resulted in a great many contacts worldwide. The long list of missions given below was crucial for successfully embedding the work of this thesis in an international context.

Norwich, United Kingdom, 11/ 2003: CCl/CLIVAR ETCCDMI meeting; Erfurt, Germany, 6/2003: WMO CLIPS workshop; Arnhem, Netherlands, 6/2003: CHR workshop; Nice, France, 4/2003: EGS-AGU-EUG Joint Assembly; Brussels, Belgium, 11/2002: Fourth ECAC; Geneva, Switzerland, 9/2002: ISDR/WMO/UNDP workshop; Beijing, China, 6/2002: IPCC workshop on Extremes; Geneva, Switzerland, 11/2001: Scientific lecture at WMO/CCl 13th Session; Funchal, Madeira, 10/2001: COST-719 meeting; Utrecht, Netherlands, 8/2001: Climate conference; Helsinki, Finland, 3/2001: NORDKLIM workshop; Kingston, Jamaica, 1/2001: NOAA workshop; Pisa, Italy, 10/2000: Third ECAC; Oslo, Norway, 6/2000: ECSN meeting; Oslo, Norway, 10/1999: ECSN workshop; Koblenz, Germany, 3/1999: Int. conference on hydrology; Asheville, USA, 3/1999: Update of the 6/1997 Asheville conference; Vienna, Austria, 10/1998: Second ECAC; Bracknell, United Kingdom, 9/1998: Joint CCL/CLIVAR task group on indices; Asheville, USA, 6/1997: Workshop on indices and indicators for climate extremes; New York, USA, 5/1997: 8th Global Warming Int. Conference; Nörkopping, Sweden, 5/1996: First ECAC; Volterra, Italy, 3/1996: European School of Climatology and Natural Hazards; Maastricht, Netherlands, 12/1994: Int. Conf. on Climate Change Research; Karlsruhe, Germany, 6/1994: IGBP-BAHC-Focus 4 workshop; Boulder, USA, 11/1993: 18th Annual Climate Diagnostics Workshop; Harpenden/Rothamsted, United Kingdom, 8/1993: Second Int. SPRUCE Conf.; Koblenz, Germany, 5/1993: CHR workshop

De meeste dank ben ik tot slot verschuldigd aan mijn vader, moeder, broers, familie en vrienden voor hun vertrouwen, gezelligheid, liefde en steun. De vraag ‘weet je al wanneer’ is nu eindelijk beantwoord.

Albert Klein Tank, Juli 2004