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RELIABLE ESTIMATION OF CLIMATIC VARIATIONS IN FINLAND

Heikki Tuomenvirta

ACADEMIC DISSERTATION

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Projektin nimi

Nimeke
Suomen ilmaston vaihteluiden luotettava arvioiminen

Tiivistelmä

Tutkimuksessa käsiteltiin Suomen ilmastoa kuvaavia järjestelmällisiä meteorologisia havaintoja. Erityisesti tarkasteltiin havaintoaineistojen luotettavuutta ja yhtenäisyyttä. Suomalaisissa keskilämpötilan ja sademäärän aikasarjoissa on koko maan kattavia, samanaikaisia häiriöitä, jotka aiheuttavat epäjatkuvuuksia sarjojen yhtenäisyyteen. Nämä epäjatkuvuudet on oikaistava ennen kuin aikasarjoja voidaan käyttää ilmaston muutosten tutkimukseen.

Tilastollista menetelmää (Standard Normal Homogeneity Test) käytettiin ilmastollisten aikasarjojen testaamiseen. Tutkimuksessa kehitettiin aikasarjojen testaus- ja oikaisumenetelmä, jossa hyödynnetään sekä tilastollista testiä että tietoja havaintojen historiasta. Työssä osoitettiin, että suuresta joukosta asemia voidaan laskea harhattomia aluekeskiarvoja myös ilman homogeenisuuden testausta. Muodostamalla alueellinen aikasarja tarkasteltavan suureen vuosien välisistä erotuksista voidaan käytettävissä olevien asemien lukumäärä maksimoida ja laskea harhaton aluekeskiarvo.

Suunnilleen viimeisten 150 vuoden aikana tapahtunut vuosikeskilämpötilan ja kevään (maalis-, huhti- ja toukokuu) keskilämpötilan kohoaminen on tilastollisesti merkitsevää. Keväisin lämpeneminen on ollut lähes lineaarista. Keväiden lämpeneminen on tilastollisesti merkitsevämpää kuin vuoden keskilämpötilan kohoaminen. Lämpötilat ovat nousseet nopeasti 1970-luvulta alkaen erityisesti talvisin. Lämpeneminen liittyy Pohjois-Atlantin värähtelyn vaiheeseen, jolloin länsivirtaukset ovat vallitsevia.

Fennoskandian alueen lämpötilan vuorokausiamplitudia kuvaamaan kehitettiin regressiomalli, jossa geostrofisen tuulen komponentit, ilmanpaineen poikkeamat ja pilvisyyden poikkeamat olivat selittäjinä. Jaksolla 1910-95 regressiomalli selittää huonoiten talvien (53%) ja parhaiten kesien (80%) havaittuja lämpötilan vuorokausiamplitudin vaihteluja. Malli tuottaa lämpötilan vuorokausiamplitudin vuosikeskiarvossa havaitun tilastollisesti merkitsevän pienenemisen. Pilvisyys on tärkein selittäjä, mutta ilmakehän virtauksien huomioiminen tarkentaa mallia merkittävästi. Lämpötilan vuorokausiamplitudin kaventuminen Fennoskandiassa johtunee pilvisyyden lisäyksestä ja ilmakehän kosteita ilmamassoja tuovien länsivirtausten voimistumisesta.

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Abstract

A study was made of the climate of Finland during the period since regular organised meteorological observations were recorded. Special attention was given to the reliability and homogeneity of the data. In the Finnish mean temperature and precipitation series there are synchronized, nation-wide homogeneity breaks that bias the original series. These discontinuities must be adjusted before performing studies of climatic changes.

The Standard Normal Homogeneity Test (SNHT) was used for homogeneity testing of climatic time series. A homogeneity testing and adjustment methodology was developed, in which SNHT and metadata (information on data) are used complementarily to produce homogenous series. In addition, homogeneity of area-averages based on a large number of stations was examined. It turned out that the use of the First Difference Method enables one to maximise the number of stations and to calculate unbiased area-averages without relative homogeneity testing and adjustment.

Statistical tests show that there has been a significant increase in the Finnish annual and spring (March-April-May) mean temperatures during the last 150 years or so. The spring temperature increase has been quite linear and it is more significant than the annual mean temperature increase. From the 1970s onwards there has been a rapid increase in temperature, especially during wintertime. A strong time-mean westerly wind, related to a positive North Atlantic Oscillation Index, was observed in connection with recent warm winters.

Atmospheric circulation indices defined by zonal and meridional sea level pressure differences, along with sea level pressure and cloud cover anomalies were used to build a multiple linear regression model for the Fennoscandian Diurnal Temperature Range (DTR). During the period 1910-95 the model explains from 53% (winter) to 80% (summer) of the variation in DTR, and reproduces the statistically significant decreasing trend at an annual level. Cloud cover is the dominant predictor, while circulation provides substantial improvement in explanation. The decrease of DTR can be explained primarily by cloud cover increase and a strengthening of the westerly flow bringing more humid marine air masses into Fennoscandia.

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PREFACE

It was my summertime job at the Finnish Meteorological Institute (FMI) in 1983 under the guidance of Dr. Raino Heino that got me interested in studying meteorology. Nearly a decade later I joined FMI, again under Raino's supervision. Meanwhile I had spent some inspiring years studying at the Department of Meteorology in the University of Helsinki, where Professors Eero Holopainen and Hannu Savijärvi guided my studies. I also had the opportunity of working as a research assistant at the University before some "internationalising" experience at the World Meteorological Organisation. I am grateful for the rich opportunities for learning that I was fortunate to have under these supervisors over those years.

Several people have helped and advised me during the years that it took to prepare my thesis. At FMI, I would like to thank Mr. Achim Drebs for his assistance and discussions related to data and metadata. I thank Dr. Reijo Solantie for guidance in applying the correction method for measured precipitation, Drs. Ari Venäläinen and Kirsti Jylhä for their help with various research projects, and Dr. Kimmo Ruosteenoja for processing climate model data of millennial control simulations and for his encouragement. Dr. Mikko Alestalo is acknowledged for providing stimulating working conditions. Mr. Tuomo Sankola and Mr. Juho-Pekka Kaukoranta kindly delivered station history information. The support for the research work provided by the staff of FMI is gratefully recognised; Mr. Jaakko Forsius is especially thanked for his help with the maps and pdf-files.

I would also like to thank my Nordic colleagues. Dr. Hans Alexandersson from SMHI, in particular, has given guidance on SNHT and constructive criticism. Eirik Førland, Per Øyvind Nordli, Inger Hanssen-Bauer and Ole-Einar Tveito from DNMI, Trausti Jónsson from VI, and Povl Frich while working at DMI have all supported me with data, advice and through stimulating discussions. Dr. Tim Carter made the linguistic revision of several of my papers and provided many helpful comments, as well as guided me in projects related to climate scenarios. I thank Mr. Robin King for carefully reviewing and correcting the language of this thesis.

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I wish to express my sincere gratitude to my dear wife Leena and our lovely children Katja, Sonja, Linda and Linus. Finally, I would like to thank my parents, Lauri (in memorium) and Elvi Tuomenvirta for their love and support.

CONTENTS

1. INTRODUCTION

Climate is known to vary on all time scales. Traditionally *natural* climate variability has been divided into two components. Firstly, there is variability caused by natural *external* forcing such as changes in the amount and distribution of solar energy available to the Earth or changes in the radiation balance of the atmosphere caused by debris from volcanic eruptions. Secondly, there is the unperturbed *internal* variability of the climate system. Even without variations in the external forcing the climate varies naturally. This is a consequence of couplings (physical, chemical, biological) between components of the climate system (the atmosphere, the hydrosphere, the cryosphere, the land surface and the biosphere). However, during the last 100-150 years mankind, through the combustion of fossil fuels, changes in land use and the release of new, man-made chemicals, has been changing the composition of the atmosphere. This has caused an external *anthropogenic* forcing of the climate system at a global level, which has begun and will continue to change climate (IPCC 1990, IPCC 1996, IPCC 2001).

Due to the possible anthropogenic effect on climate, during the last fifteen years or so special attention has been paid to the development and analysis of long surface climatological time series both globally (Jones 1988; Hansen and Lebedeff 1988; Vinnikov et al. 1990; Vose et al. 1992; Jones 1994; Easterling et al. 1996; Jones et al. 1999; Doherty et al. 1999; Jones and Moberg 2003) and in Northern Europe (Frich et al. 1996, Schmith et al. 1997, Førland et al. 1998, Tuomenvirta et al. 2001). These data are fundamental for monitoring the climate and documenting trends and variability. Data sets can be used to develop understanding of climate processes and to validate climate models. In studies investigating the possible attribution of part of the observed chnages to human influence, it is not trivial, even at the global level, to separate the anthropogenic signal from the signal caused by partly unknown natural external forcing and "noise" caused by internal variability (e.g. Hegerl et al. 1996 and Mitchell et al. 2001). This is because the different factors causing variability in the observations are mixed. In smaller regions like Northern Europe, although statistically significant trends can be found, the signal to noise ratio is even smaller than at a global level.

Climate is one of the physical factors affecting the biosphere, including humankind. The impacts of climate change and variability on natural systems and human activities are usually experienced at the regional or local scale. Therefore, creation of reliable data sets and determination of trends and fluctuations of regional and local climate are of practical importance. Climatic data are also needed in the development and application of various impact models.

In Finland there are several research areas that utilise climatic data. Table 1.1 gives some recent examples from the field of natural sciences. Climate data is used both as input into various impact models as well as for more general analysis of biological and physical processes in, *inter alia*, forest ecology, forestry, agriculture, hydrology. Many impact models have been used to study the possible effects of anticipated climate change. Climatic data have also been used in reconstruction of past climates. Furthermore, long-term time series can be important in planning. The 30-year normal period of climatological observations (usually 1961-1990 or 1971-2000) may not be sufficient, for example, for the estimation of the magnitude and frequency of extreme climatic events. For all of these applications, it is important that the climatological time series employed are as homogeneous as possible.

Table 1.1. Examples of recent research where climatic observations are used in Finland.

Field of research	**Some references**
Growth variations of trees (Forest ecology)	Beuker et al. (1996), Mäkinen et al. (2000)
Decomposition of cellulose in soils	Kurka (2000)
Plant phenology	Lappalainen (1994), Häkkinen (1999), Linkosalo (2000)
Dendrochronology (Reconstructions of temperature and precipitation)	Lindholm et al. (1996), Helama and Lindholm (2003)
Crop potential of spring wheat in changing climate	Saarikko (1999)
Oceanography	Haapala (2000)
Palaolimnology	Sorvari et al. (2002)
Soil frost	Venäläinen et al. (2001a), Venäläinen et al. (2001b)

The use of long-term climatic data is far from straightforward. Many types of disturbances can cause apparent changes in a record, complicating and sometimes even hiding the true climatic signal in the original time series. The adjectives "long" and "homogeneous" can seldom be used at the same time to characterise climatic time series. Peterson et al. (1998) give a review of the homogeneity testing and adjusting approaches used with surface climatic data. The homogeneity problem is not limited to surface observations, but also affects, for example, satellite data (Christy et al. 1995, Wentz and Schabel 1998, Christy et al. 2000) and numerical weather prediction data assimilation products (Kalnay et al. 1996).

This paper describes the construction of reliable long-term data sets based on meteorological observations. Special attention is accorded to the methods used to test and adjust the temporal homogeneity of time series. The constructed climate data sets are analysed, trends and fluctuations in Finland and the Nordic region are examined, and the physical linkages between different climatic parameters are explored.

1.1 Aims of the study

The objectives of this study are:

- To develop the methodology of homogeneity testing and adjustment. The aim is to use both statistical tests and "metadata" (i.e. information on observation methods, instruments and data processing). Although statistical tests are objective, subjective choices are still required in the application of tests and in the use of metadata.

- To produce reliable climatological data for studies of climatic variations and change as well as of impacts and adaptation. The source of data is meteorological observations carried out by the Finnish Meteorological Institute and its predecessors. The main emphasis is on improving the homogeneity of time series. The final data form the basis of climate-related studies in Finland and the Nordic region, as well as contributing to global data sets.

- To evaluate the uncertainty in the Finnish long-term, climatic time series, and to assess the value of homogeneity testing and adjusting.

- To compare the Finnish annual, national mean values of temperature and precipitation produced in this study with widely-used global data sets.

- To analyse trends and fluctuations of climate in Finland and the Nordic region. Long time series of temperature, precipitation, cloud cover and air pressure are used to characterise variability on the scale of a decade, and to determine trends and linkages between climatic elements.

1.2 List of original papers

This thesis is based on the following four original articles, referred to in the text by Roman numerals:

I Tuomenvirta, H. and Heino, R., 1996: Climatic changes in Finland - recent findings. *Geophysica*, **32**(1-2), 61-75.

II Tuomenvirta, H., Alexandersson, H., Drebs, A., Frich, P., and Nordli, P.O., 2000: Trends in Nordic and Arctic temperature extremes and ranges. *Journal of Climate*, **13**, 977-990.

III Tuomenvirta, H., 2001: Homogeneity adjustments of temperature and precipitation series – Finnish and Nordic data. *International Journal of Climatology*, **21**, 495-506.

IV Tuomenvirta, H., 2002: Homogeneity testing and adjustment of climatic time series in Finland. *Geophysica*, **38**(1-2), 15-41.

Papers I - IV are reprinted at the end of this thesis. Papers are reproduced by kind permission of the following: the Geophysical Society of Finland (I and IV), the American Meteorological Society (II), the Royal Meteorological Society (III).

The author of this thesis bore the main responsibility for writing papers I and II. He performed the homogeneity testing of Finnish data. Dr. R. Heino determined the adjustments of precipitation gauge type changes and changes in the averaging method in the calculation of daily mean temperature. The homogenised data from other Nordic countries were produced within the REWARD project (Førland et al. 1998). The analysis and calculations presented in paper I were carried out by the author together with Dr. R. Heino. The author performed all of the analysis in paper II, with the exception of the regression model, which was developed together with Dr. H. Alexandersson. In paper II, Mr. A. Drebs acted as a data manager, Mr. P. Frich introduced the use of extreme temperature range and Mr. P.Ø. Nordli determined the adjustments to Norwegian minimum temperatures. The author of this thesis is the sole author in papers III and IV. The Standard Normal Homogeneity Test widely used in paper IV and described in the appendix to that paper was developed by Dr. H. Alexandersson (Alexandersson 1986, Alexandersson and Moberg 1997).

Paper I also presents an analysis of snow cover, reporting a decrease in the number of snow cover days in southern Finland during the period 1938-95. Moberg et al. (2004) also report a decrease in the snow-covered area in Fennoscandia during the period 1967-2000. Recently, Solantie (2000) has analysed a comprehensive data set (1909-1998) of snow depths in Finland. Over this extended period there are no persistent linear trends in the snow cover. Snow cover observations are not discussed further in the present study.

2. DATA AND HOMOGENEITY

In order for a meteorological or climatological observational time series to be regarded as perfectly homogeneous, it should record variations that are attributable to weather and climate fluctuations alone (Conrad and Pollack 1950). This would require that observations be performed at the same site within an unchanged environment using the same calibrated instrument according to the same method. In reality, these requirements are rarely fulfilled in long time series, and their "absolute homogeneity" is always questionable. Instead, climatologists must make do with series that are "relatively homogeneous", where the differences or ratios between the candidate station series (i.e., that of the station being tested) and synchronous series at neighbouring (relatively) homogeneous stations are statistically random series. The use of non-homogeneous climatological time series (i.e., containing variations unrelated to climate) can lead to inconsistent conclusions. Therefore, besides routine quality control, the homogeneity of data should be evaluated before performing studies of climatic changes.

2.1 Building of data set for climatic studies

Fig. 2.1 gives a schematic description of the process of building a reliable data set from long-term climate observations. The aim is to detect suspicious data and to improve the relative homogeneity of the time series. Five general steps can be identified.

Firstly, sources of observations must be identified. In Finland, observations from 1959 onwards are in the digital database at the Finnish Meteorological Institute (FMI), but older data are in published meteorological yearbooks and on unpublished observation forms held in the archives of FMI (Heino 1994). Much of the older data of the 20th and late 19th centuries has been digitised at a monthly level. However, only part of the meteorological observations before about the 1880s is in digital form. For sub-daily records, only data since 1959 is in digital form.

The second step is to select the required part of the raw data for further processing. For example, monthly mean daily maximum and minimum temperatures were extracted for processing in paper II. The data were subjected to quality control where outliers were manually corrected or deleted.

Step 3 concerns the correction of known homogeneity breaks. Heino (1994) has documented the nation-wide methodological and instrumental changes that cause systematic biases in the original data. Adjustment factors that are needed for the mean temperature series, due to changes in the observation times and averaging methods, and for the precipitation series, due to changes in instrument type, are applied in correction routines for a large number of stations. These kinds of homogeneity breaks can affect large area averages. For example, the nationally-averaged annual mean temperature and precipitation presented in paper I would have been biased without adjustments.

In the fourth step, the homogeneity of individual time series is tested using statistical methods (Standard Normal Homogeneity Test, Alexandersson and Moberg 1997) and the test results are used to adjust the time series. The issue of homogeneity testing is only discussed briefly in papers I-III, but is considered in more detail in paper IV.

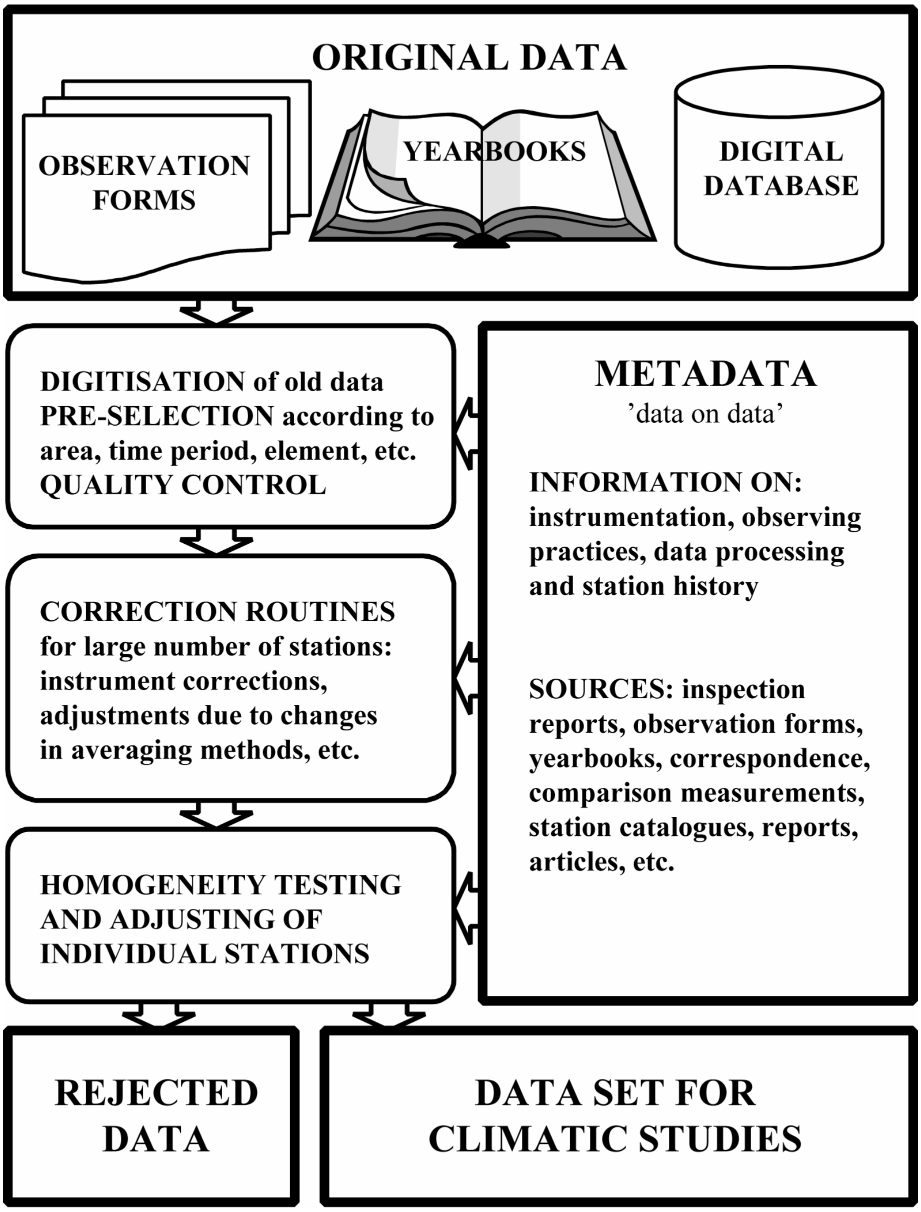

Fig. 2.1. The construction of observational climatological data sets (schematic).

In all of the first four steps, metadata (= information on data) are required to guide data processing. Some of the metadata are related to the observing network, e.g. changes in observing times. The largest volume of metadata, however, relates to individual stations and is found, among other sources, in station inspection reports. However, metadata information is usually incomplete, because it has not been collected systematically. In Finland, some comparison measurements and studies have been arranged to investigate the effect of changes in instrument type (Solantie and Junila 1995, Heino 1994, Tammelin 1984, Korhonen 1913), but there have seldom been comparison measurements in connection to changes at individual stations. Therefore, long time series can contain many unknown homogeneity discontinuities.

At the final step of data set construction, part of the data is labelled reliable according to the test results and can be passed on to other users of data. The remaining part of the data usually has to be rejected from the data set.

As part of a programme of Nordic co-operation, a number of data sets have been produced containing contributions from Finland: the North Atlantic Climatological Dataset (NACD) in Frich et al. (1996), North Atlantic - European pressure observations (WASA dataset) in Schmith et al. (1997), REWARD (Relating Extreme Weather to Atmospheric circulation using a Regionalised Dataset) in Førland et al. (1998) and the NORDKLIM data set in Tuomenvirta et al. (2001) that contains the updated NACD and REWARD data sets. In the new version of the CRU (Climatic Research Unit) global, land-based temperature database, Jones and Moberg (2003) have used the NACD and NORDKLIM data sets.

Table 2.1 contains a list of symbols that are used throughout this paper. Time series of annual and seasonal anomalies are quite often presented as absolute deviations from the mean values of the normal period 1961-90. Annual and seasonal means are calculated from monthly values. The seasons (winter, spring, summer, autumn) are defined conventionally as the three-month periods: December-February (DJF), March-May (MAM), June-August (JJA) and September-November (SON).

Table 2.1. List of symbols used in this study.

Symbol	**Description**
T	Mean temperature
Tx	Mean daily maximum temperature
Tn	Mean daily minimum temperature
Th	Highest maximum temperature
Tl	Lowest minimum temperature
DTR	Diurnal temperature range, Tx-Tn
ETR	Extreme temperature range, Th-Tl
R	Precipitation sum

2.2 On the homogeneity of Finnish climatic observations

The Finnish data in digital form at FMI start in 1829 with air temperature records from Helsinki. By the end of the 19th century monthly data for 12 climatic elements from 48 stations are available in digital form. In total there are about 1000 stations and 40 elements. Of the 1000 stations about two-thirds measure only precipitation. Heino (1994) describes the development of the climatological network in Finland, which can be monitored on an annual basis in the Meteorological Yearbook of Finland.

It is common for a large number of stations to participate in the homogenisation procedure, but for only part of them to be selected to make up the final data set. For example, nearly 300 precipitation series were tested to evaluate the homogeneity of 125 Finnish long (≥60 years) series in paper I (Tuomenvirta and Drebs 1994). Similarly, about 90 Finnish stations with monthly mean daily maximum and minimum temperature data were tested in order to produce the 19 time series for

REWARD (Førland et al. 1998) analysed in paper II. Chapter 5 in this summary contains an update to analyses presented in paper I. For the new analysis, nearly 300 Finnish mean temperature series were homogeneity-tested. About 20 stations from neighbouring countries were also used to carry out the homogeneity testing of mean temperatures. Aside from temperature and precipitation series, 42 long-term series of atmospheric pressure in Finland have also been tested.

A large part of the research was dedicated to the detection and adjustment of homogeneity breaks in the climatic time series (step four in the previous section). Paper IV describes in detail the testing procedure and gives examples. The Standard Normal Homogeneity Test (SNHT) developed by Alexandersson (1986) and Alexandersson and Moberg (1997) is a technique for indentifying an inhomogeneity without knowing *a priori* the time of the break point, and it can also estimate the statistical significance and magnitude of the identified break. SNHT together with available metadata were used in the process of creating reliable time series from meteorological observations made in Finland. The use of metadata enhances the testing and adjusting procedure, but the search through the relevant metadata can also be slow and laborious. One of the most notable findings of paper IV is that the risk of drawing wrong conclusions on climate changes due to flawed data can be much diminished by performing homogeneity control.

Essentially, paper III answers the question: is homogeneity testing and adjusting necessary? By analysing the adjustments needed to produce homogenous data sets, paper III justifies the efforts put into homogeneity testing and adjusting. The magnitude of homogeneity breaks can be substantial at individual stations. Typically there are only very few, if any, homogenous long-term time series. In addition, there are systematic errors in both long-term temperature and precipitation series that can bias large-scale area-averages. The biases are of the same order of magnitude as the observed trends over the 20th century. In general, the nation-wide changes in the formulas used for the calculation of mean temperatures and the simultaneous changes in the precipitation gauge type are the most detrimental breaks in the Finnish data. On the other hand, the remaining adjustments for the entire Finnish temperature and precipitation observation network appear to be random, and thus do not bias averages based on a large number of stations.

The results of papers III and IV provoke the question: Could Finnish averages of temperature and precipitation be calculated from the original data? The logic behind this approach would be to use the well-known adjustments for the nation-wide simultaneous changes in the formulas used for the calculation of mean temperatures and the changes in the precipitation gauge type, and assume that the effects of other homogeneity breaks cancel out in averaging. To ensure that the assumption of randomness holds, data from tens, preferably hundreds, of stations should be used.

By creating national-average time series from differences between successive years at individual stations it is possible to use all the data at hand. Peterson et al. (1998) call this method the First Difference Method (FDM). In chapter 5 of this summary, the construction of national-average series with FDM is presented and validated against other estimates, e.g., those presented in paper I. In the following section the data available for such an exercise are described.

2.3 Mean temperature and precipitation data used in an alternative approach to calculate national averages

During the year 1893 precipitation measurements were initiated at about ten stations; the following year was therefore chosen as the first year for the national average time series. Data from 32 stations are available for calculating the differences between annual precipitation sums between the years 1894 and 1895, i.e., the first value of the curve labelled "stations" in Fig. 2.2a (left axis). During 1908/09, the precipitation measurement network was made denser (Korhonen 1913), and this can be seen as an upward step in Fig. 2.2a. 1970 is a year with an increase from about 440 to 624 stations in the data base; apparently all pre-1969 precipitation data has not yet been digitised. In the case of temperature, the amount of digitised data increases more smoothly than that of precipitation. Somewhat arbitrarily, the year 1888 was chosen as the first year of the national average temperature series. At that time there were 25 stations for calculating the annual mean temperature change from 1888 to 1889 (Fig. 2.2b). The number of available temperature stations increases smoothly, peaking in 1977.

In Figs. 2.3 and 2.4 is shown the LPNN grid used in area-averaging. The LPNN grid was first introduced in the Meteorological Yearbook of Finland, Volume 70 Part 2 in 1970. It is a modified, mostly 1ºx1º latitude-longitude grid covering Finland with 80 grid boxes of varying size and shape. It is currently in use at FMI for the identifying system of stations.

In Figs. 2.2a and b are shown the percentage of the total Finnish land area covered by LPNN grid boxes with at least one (thin line) and two (thick grey line) stations. Until 1910 the precipitation network covered much less than 50% of Finland, after which a significant improvement was achieved. Since 1960, the precipitation stations have become so numerous and evenly distributed that the coverage of at least one station per grid box stays above 95%. The decrease in the number of stations, starting in the 1980s, is somewhat reflected by the drop in the coverage of at least two stations per grid box. All three curves in Fig. 2.2b show two facts quite consistently: firstly, there has been an increase in the number and coverage of temperature data. Secondly, a drop from the 1980s onwards has been accomplished by thinning out the station network. The network can be much sparser for temperature than for precipitation measurements and yet give sufficiently detailed information. This is due to the fact that the variations in temperature typically have a larger spatial scale than the variations in precipitation.

Fig. 2.3 shows the spatial distribution of precipitation stations available for calculation of the annual precipitation sum change from 1894 to 1895. After the improvement of the precipitation measurement network in 1909/1910, southern and central Finland are fairly evenly covered, but in northern Finland there are still large areas with no data. Fig. 2.3 also shows the present status of the precipitation measurement network (2000/2001). Fig. 2.4 shows similarly the available Finnish temperature data in 2000/2001. There are about 14 times the number of temperature stations in 2000/2001 than at the beginning of series in 1888/1889 (Fig. 2.4). Fig. 2.4 also shows the stations used in the calculation in paper I of the national average starting in 1901 (8 stations) and the four stations that can be used since 1847.

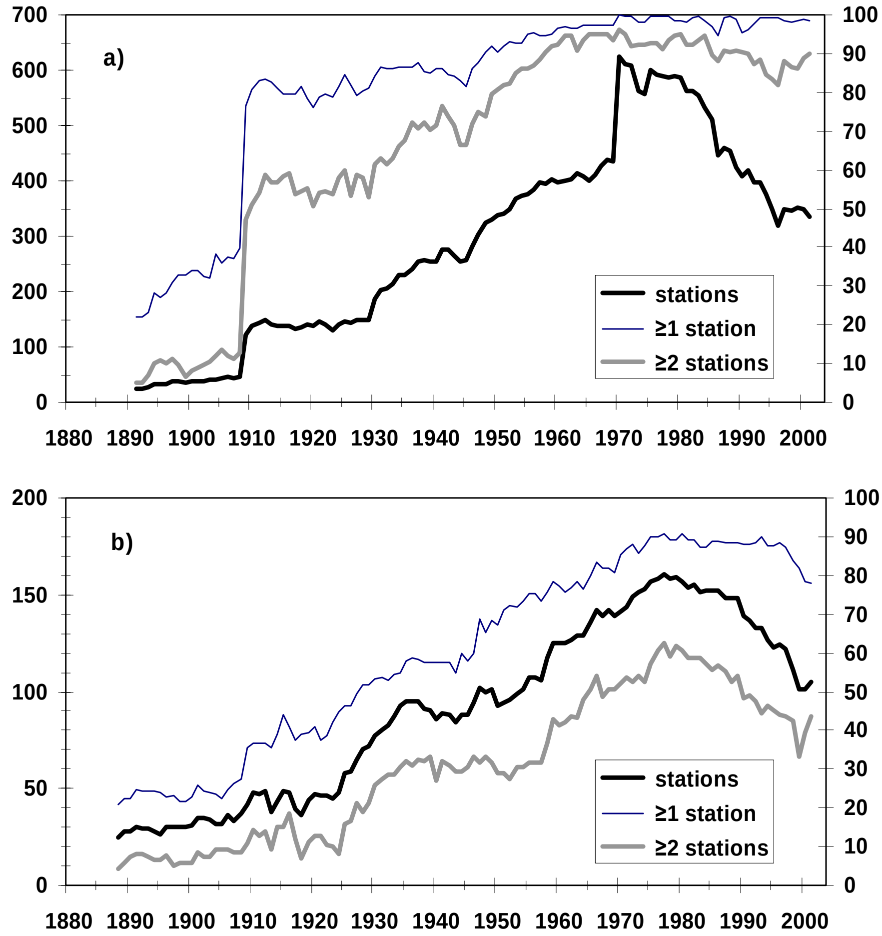

Fig. 2.2. The number of stations used in the calculation of national averages in sections 5 and 6 of this summary (thick black line, left axis), percentage of Finland covered by LPNN grid boxes with at least one station (thin line, right axis) and at least two stations (thick grey line, right axis): a) precipitation total, 1894-2001 and b) mean temperature, 1888-2001.

Besides data from Finland, stations close to the Finnish borders in Sweden, Norway and Russia have been used in testing and calculations. For example, data from Karasjok (LPNN box 81) and Karesuando (LPNN box 94) are crucial for the calculations of the early parts of the time series (Figs. 2.3 and 2.4). The Swedish temperature series were homogenised by Moberg and Alexandersson (1997). The Swedish precipitation series and the Norwegian data are also of good quality and homogeneity. They are from the NORDKLIM data set (Tuomenvirta et al. 2001) or are data used by Heino (1994). Data from Russia consist of old series from stations operated by the predecessor of FMI until the year 1940. Depending on the source, the Swedish and Norwegian data end in the 1990s.

Fig. 2.3. Precipitation stations available for the calculation of the change in the annual precipitation sum from 1894 to 1895 (upper left panel), from 1909 to 1910 (upper right panel) and from 2000 to 2001 (lower right panel). The LPNN grid used in the calculations of the area-averages is delineated with thick lines; the LP labels are shown on the right and on the bottom of the maps. Latitude values are marked on the left and longitude at the top of the maps. (Figure on following page)

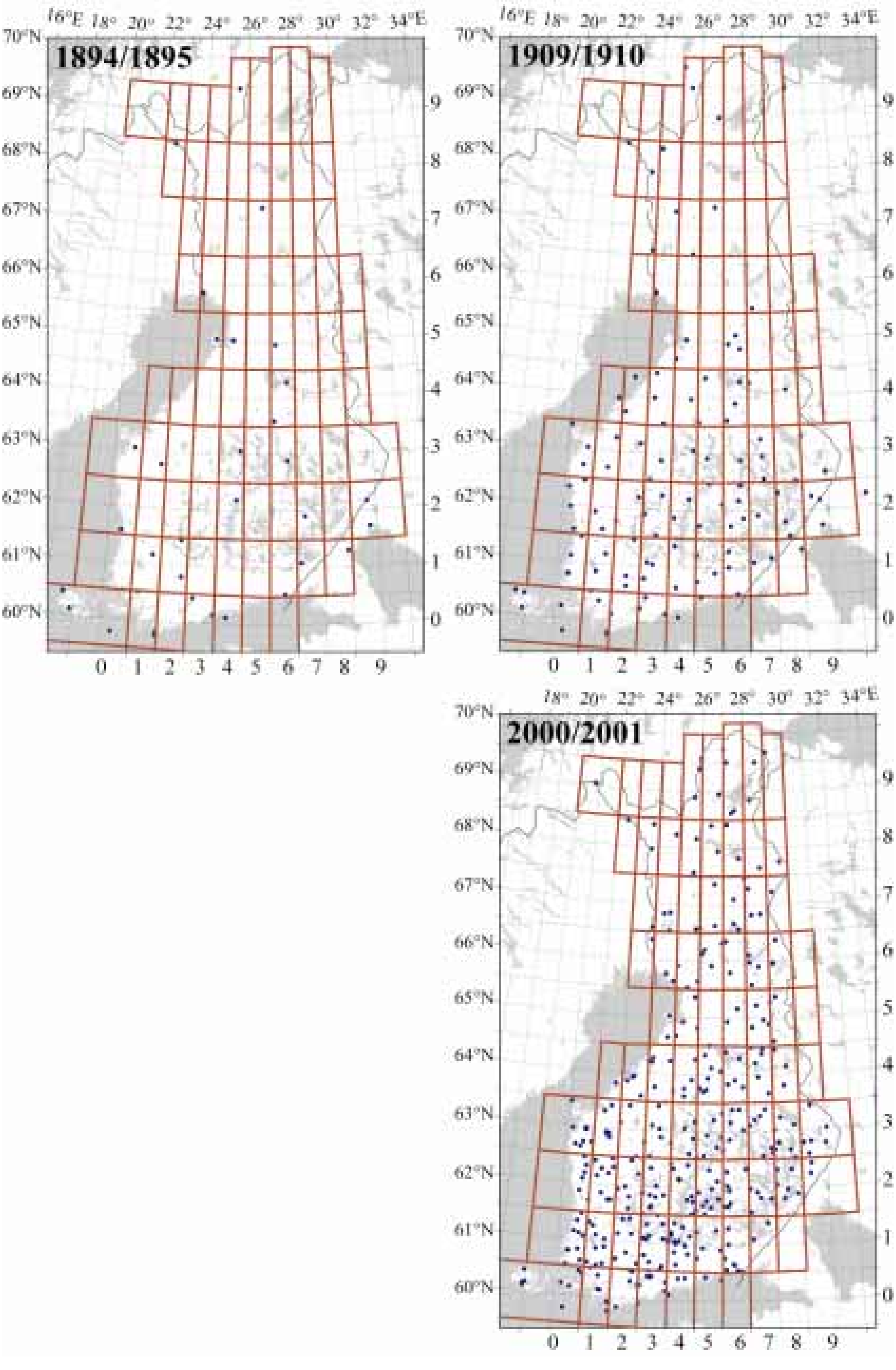

Fig. 2.3. (Caption on previous page)

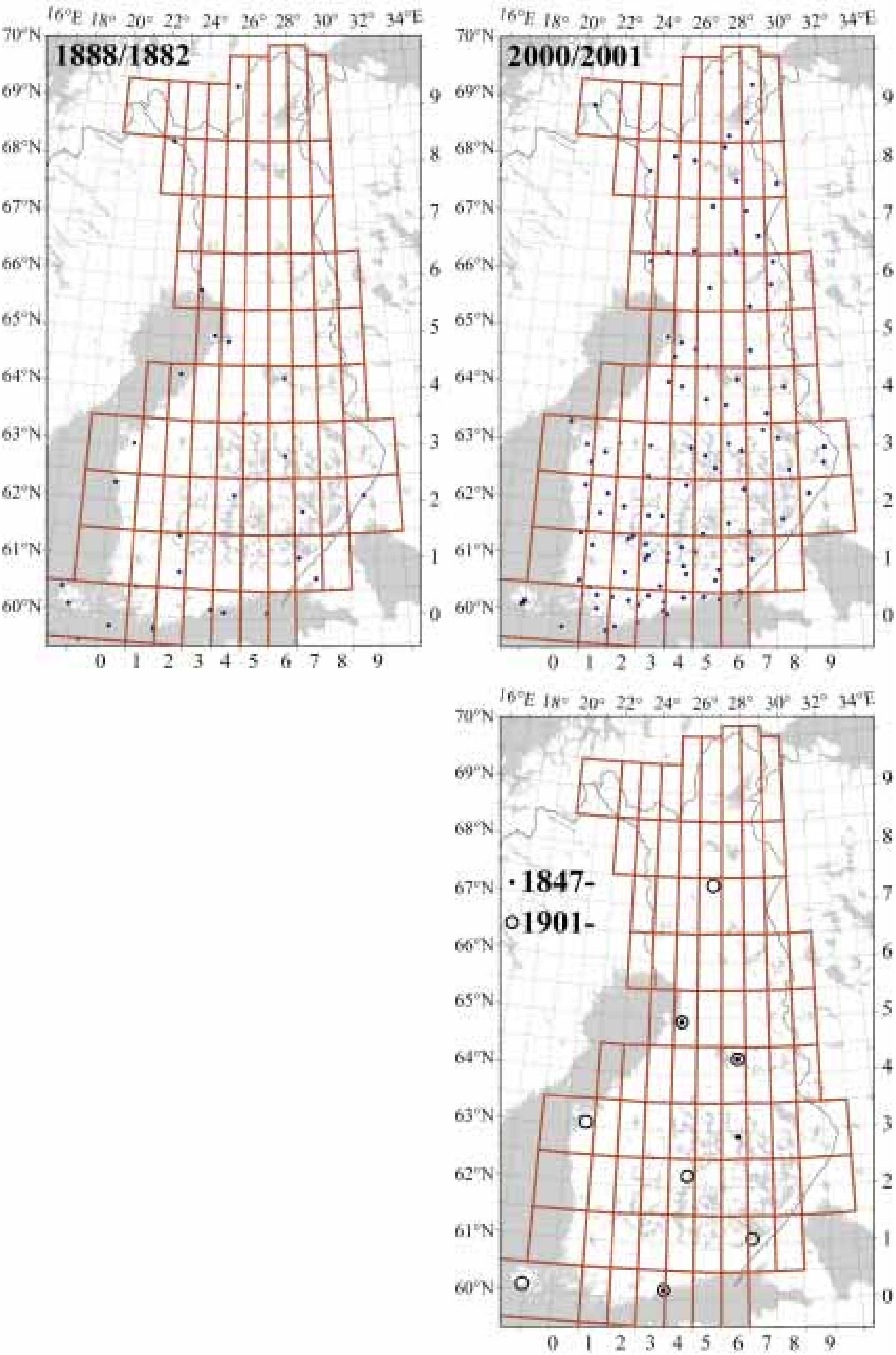

Fig. 2.4. (Caption on following page)

Fig. 2.4. Temperature stations available for the calculation of the change in the annual mean temperature from 1888 to 1889 (upper left panel) and from 2000 to 2001 (upper right panel). The lower right panel shows the stations used in the calculation of the national average starting in 1901 in paper I (8 stations) and the four stations that can be used since 1847. The LPNN grid is the same as that shown in Fig. 2.3. (Figure on previous page)

2.4 Results from testing Finnish mean temperature data

In papers I - IV an approach has mostly been used in which only part of the original data is homogeneity-tested. Section 5 of this summary describes another approach, in which all data are used without homogeneity testing. In order to be able to appropriately compare the effects of homogeneity processing on temperature data, all Finnish mean temperature data in digital form for the period 1847-2002 (described in the previous section) were tested and adjusted. In addition, the stations used in paper I were tested again, which resulted in some new adjustments and refinements to older adjustments for these stations.

For this summary, nearly 300 seasonal mean temperature series were tested. SNHT was not applied to series shorter than ten years in length. However, this is not a serious problem, because these short series are probably more homogenous than the longer ones. Due to the large number of stations, metadata were consulted only occasionally. For this reason the most common statistical level for defining a homogeneity break was 95%, and the low limit (90%) supported by some physical evidence from the station history (paper IV), was not much used. As a result, a little more than 200 homogeneity break points were detected from slightly over 100 mean temperature series, not including breaks due to formula changes.

The test results reported here are for the period 1888-2002 used in the calculation of the national mean temperature in chapter 5. About one quarter of the stations have records shorter than ten years (Fig. 2.5). The amount of data is fairly evenly distributed across records of all lengths, except that the longest records (115 years) contribute about 10%; this is seen as step-like increase in the cumulative frequency (black curve) in Fig. 2.5. Temperature series shorter than ten years were not tested, and forty years seem to be the length of record after which the probability of homogeneity discontinuity increases markedly. The longest record without adjustments is nearly a hundred years in length. The stations having full-length series (115 years) contain about 20% of the breaks (Fig. 2.5, grey curve). This is a result of the fact that it is practically impossible to maintain unchanged conditions for measurements over such long periods. The large number of breaks in the longest series is also the result of a somewhat artificial combining of records from nearby stations in order to create long-term series for data sets.

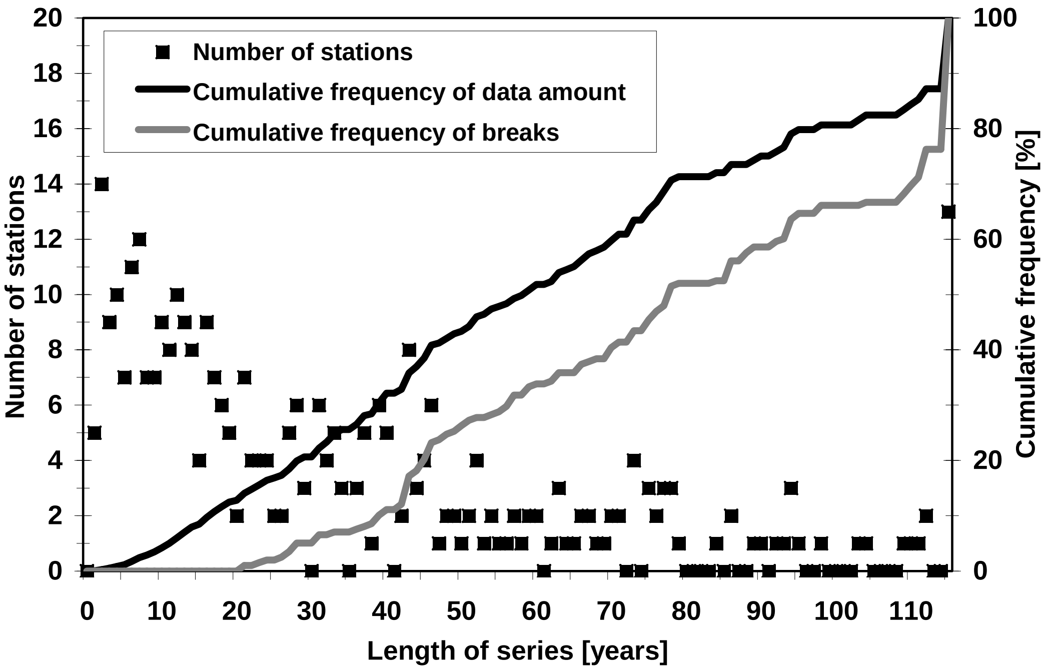

Fig. 2.5. The number of temperature stations as a function of record length (left axis). The curves show cumulative frequencies of the amount of data (i.e. number of stations multiplied by record length, in black) and of homogeneity breaks (right axis). Data are for the period 1888-2002.

The average value of the probability of finding a break at a certain year was about 0.02, while during some years no breaks were detected; the maximum number of breaks in a single year was ten (Fig. 2.6). The majority of breaks were found during years when there were many stations available. From the theoretical point of view, it should be difficult to detect small breaks in a sparse observing network with low correlation between the stations. In accordance, the mean of the intrinsic values of annual adjustments is somewhat larger in the period 1888-1940 than during the period 1941-2000.

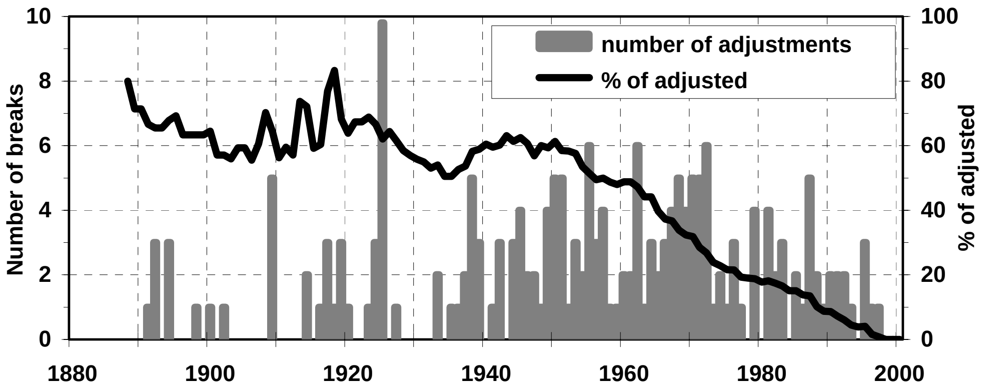

Fig. 2.6. The number of detected homogeneity breaks in annual mean temperature series per year (columns, scale on left axis). The curve (right axis) shows the percentage of adjusted series of the whole data. Data is for the period 1888-2000.

The largest instantaneous breaks are about ±1.5ºC at the monthly level. The annual or seasonal means of all adjustments did not statistically significantly differ from zero (Student's *t*-test, see paper III for description). In paper III, a similar conclusion regarding Finnish data was reached with the 1961-1990 mean temperatures as well as with the long-term daily mean maximum and minimum temperature series. Although the detected breaks for the whole study period appear to be random, it is possible that there is a time-varying bias in the data.

The mean of the temperature adjustments is close to zero during the period 1960-2000 (Fig. 2.7). However, going further back in time, there is a tendency for the mean adjustments to become negative, indicating that the old, original values were too warm, especially during winter. Similar results are reported in paper III for Finnish monthly mean maximum and minimum temperatures and by Böhm et al. (2001) for the Alpine temperature data set. The percentage of adjusted series rises from about 20% in 1970 to about 60% in 1950 (Fig. 2.6). This means that simple averaging of all data would have produced an unreliable series because on average the old data have a warm bias.

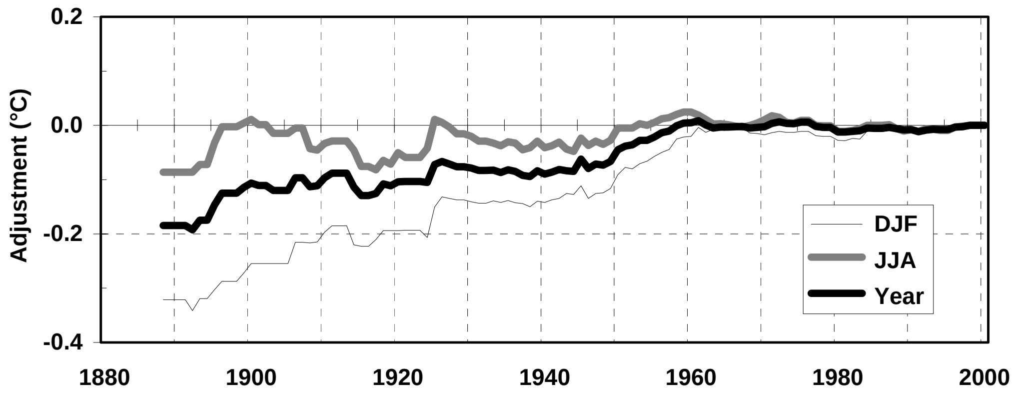

Fig. 2.7. The mean annual, winter (DJF) and summer (JJA) adjustments to Finnish mean temperatures, 1888-2000.

It is well-known that urbanisation may cause an apparent local warming trend in temperatures. The station in Helsinki has been located since 1844 in Kaisaniemi park, inside a growing town. Using neighbouring island and rural stations, Heino (1994) determined time-varying adjustments for the mean temperatures measured at the Helsinki Kaisaniemi station. These adjustments were used in this study, too. In other large Finnish towns the extent of urbanisation has been much more modest than in Helsinki, and the largest discontinuities have usually been caused by the station relocations (paper III). In this study, apart from the case of Helsinki just mentioned, any possible apparent trends due to urbanisation have been adjusted with homogeneous rural temperature series in connection with detected homogeneity breaks.

As the reasons for the homogeneity breaks were not systematically investigated, it is not known what causes adjustments to be negative compared to present-day measuring conditions (Fig. 2.7). According to paper III, several reasons may have contributed in Finland. During the 1940s and 1950s there were several station moves from town centres to open areas, e.g. airports, that are sensitive to radiative cooling.

During the 1910s and 1920s a transition to modern screens took place, perhaps improving the screening of thermometers and increasing ventilation. Nordli et al. (1997) have found that in Nordic countries, in general, the old screens were warmer than the modern ones. In addition, some of the old measurements were made at a height well above two metres above the ground, and were perhaps less strongly affected by shallow surface inversions.

3. TREND ANALYSIS AND THE SHAPE OF THE DISTRIBUTION

This section presents a selection of the statistical methods used in the study.

3.1 Gaussian filtering

A low-pass filter including Gaussian weighting coefficients was used to smooth out inter-annual variability and to display long-term trends. The filtered value in year j, G_j, is given by,

$$G_j = \frac{\sum_{i=1}^{n} w_{ij} \cdot x_i}{\sum_{i=1}^{n} w_{ij}} \tag{3.1}$$

where, the weighting coefficients w_{ij} are

$$w_{ij} = e^{\frac{-(i-j)^2}{2\sigma^2}} \tag{3.2}$$

and x_i is the original time series consisting of n years, and σ is the standard deviation in the Gaussian distribution. In the discussion, "year" is used instead of "time-step", although the time-step is not restricted to one year, e.g. in Fig. 4.1 time-step is one day.

The values of 3 and 9 were used as σ in two low-pass filters referred to in papers I and II as G3 and G9, which approximately correspond to 10- and 30-year moving averages, respectively. The first (last) few values in the filtered series are mainly determined by the original data following (preceding) the year in question. The filtered values near the both ends of the time series must therefore be interpreted with some caution. The shape of the curves can change when new values are added.

The filtering decomposes the series into low-frequency and high-frequency components. The high-frequency part of the series consists of the differences between the original and the low-pass filtered series (Moberg et al. 2003). Fig. 3.1 displays an example of the original, the low-frequency component produced with G3 and the high-frequency component calculated by subtracting the low-frequency part from the original series. In this particular example the high-frequency components of R_{ALL}(ori+) and R_{HYD}(ori) are very similar, although the low-frequency parts behave differently.

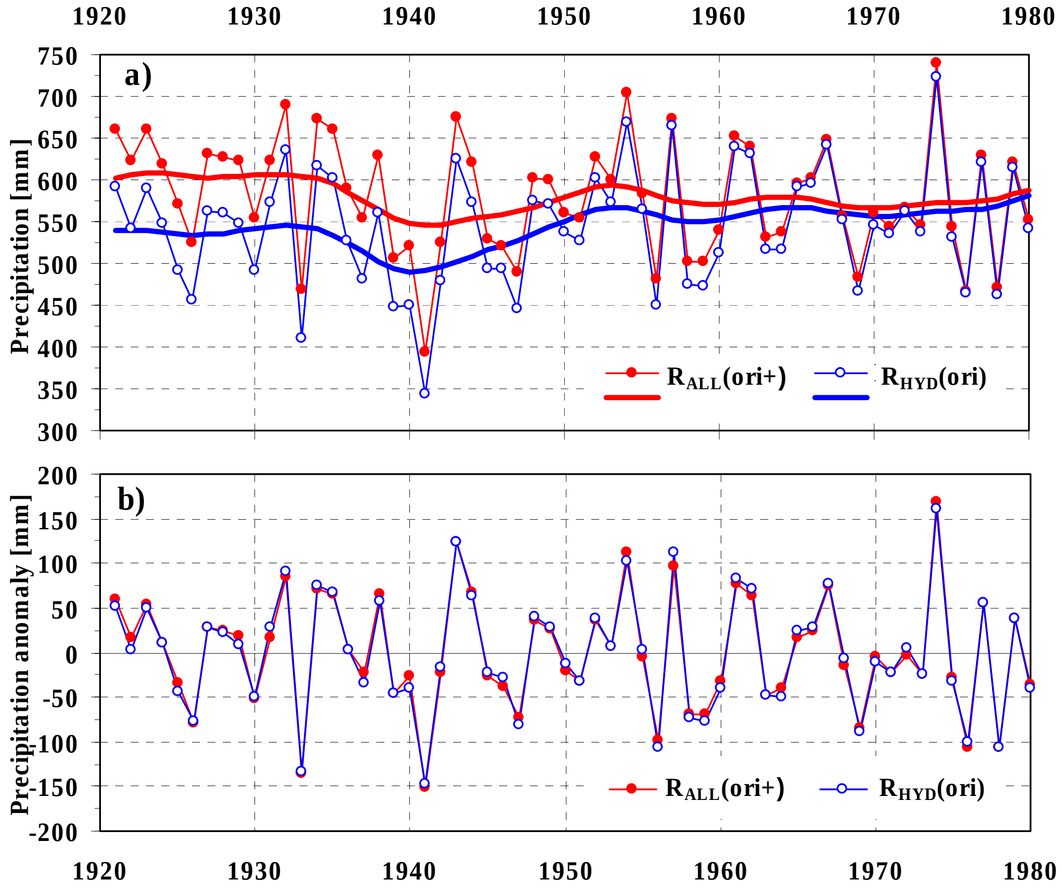

Fig. 3.1. G3 filtered precipitation series of R_{ALL}(ori+) and R_{HYD}(ori), 1921-1980 (see section 6.1 for details). a) Annual precipitation (R_{ALL}(ori+) with red closed symbols and R_{HYD}(ori) with blue open symbols) and the corresponding G3 filtered series with thick curves. b) High-frequency component of the original series (symbols as in a).

3.2 Mann-Kendall trend test and Sen's trend estimator

The non-parametric Mann-Kendall test was chosen for testing the significance of trends in paper II, as it can be used without knowing the exact distribution of the time series (Sneyers 1990). The test statistic, t, is defined by the equation

$$t = \sum_{i=1}^{n} n_i \tag{3.3}$$

where n is the number of elements and n_i is the number of smaller elements preceding element x_i (i= 1,2,.. n) that is being tested. Providing the data are independent and the number of elements in the series is more than 10, the test statistic, t, is nearly normally distributed under the hypothesis of randomness (the null hypothesis). Its expectation value, E(t), and variance, D^2(t), are given by the equations

$$E(t) = \frac{n(n-1)}{4} \tag{3.4}$$

$$D^2(t) = \frac{n(n-1)(2n+5)}{72} \qquad (3.5)$$

The normalised distribution of the test statistic, u(t), is then

$$u(t) = \frac{t - E(t)}{\sqrt{D^2(t)}} \qquad (3.6)$$

The cumulative distribution function for the standard normal distribution function may be used to decide whether the null hypothesis should be rejected or not.

To illustrate an application of the Mann-Kendall test, the significance of the linear trend in the annual, area-averaged Fennoscandian Tx, Tn and DTR during the period 1910-95 (see Fig. 6 in paper II) are tested. In paper II, no estimate of the size of the linear trend was given.

The time series are successively tested by starting from the first year of the series adding one year after another (u(t); forward testing). The test can be repeated by starting from the last year and moving backward in time (u'(t); backward testing). It is used widely, e.g., by Demarée (1990), Sneyers et al. (1998), Aksoy (1999) and Böhm et al. (2001).

Fig. 3.2 shows in graphical form the evolution of the standardised test statistics. The first and last ten years are also plotted although the 1%- and 5%-significance levels marked are not valid for those years. Both Tx and Tn show warming from the beginning of the series until the 1950s and 1960s, exceeding the 1%-level for a short period, but the trend over the whole time series is not significant, i.e. the last points of forward testing are positive but below the significance levels. The backward testing implies that the cooling since the warm 1930s is nearly significant (Tx is slightly above and Tn below the 5%-level). The only significant trend over the whole period 1910-95 at the 1%-level is the decrease of DTR.

The sizes of linear trends are calculated with the least squares method in papers I and II. More robust trend estimates can be calculated with Sen's non-parametric method (Gilbert 1987), where N' slope estimates, Q, are computed as

$$Q = \frac{x_{i'} - x_i}{i' - i} \qquad (3.7)$$

where $x_{i'}$ and x_i are data values at times i' and i, respectively, and where i' > i. N' is the number of data pairs for which i' > i. If there is only one datum in each time period, then N' = n(n-1)/2, where n is the number of time periods. The median of these N' values of Q is Sen's estimator of slope. It is not sensitive to outliers or gross errors and allows for gaps in the data. Both qualities are potentially useful in the analysis of extremes.

The previous example (Fig. 3.2) is continued in Table 3.1. It shows two estimates of linear trends in the Fennoscandian annual area-averaged Tx, Tn and DTR during the period 1910-95: the least squares estimate and the Sen estimate. Both estimates give quite similar trends, as was also the case in Heino et al. (1999). Estimates of trend significance based on the Mann-Kendall test and the standard *t*-test (e.g. Vining 1998) are also generally consistent with each other. The only exception is DTR in autumn, which shows a trend just above the 5%-significance level using the *t*-test but just below the same level using the Mann-Kendall test. Although there are some differences in the results depending on the test method used in paper II, the conclusions seem to be quite robust and do not depend on the statistical method chosen.

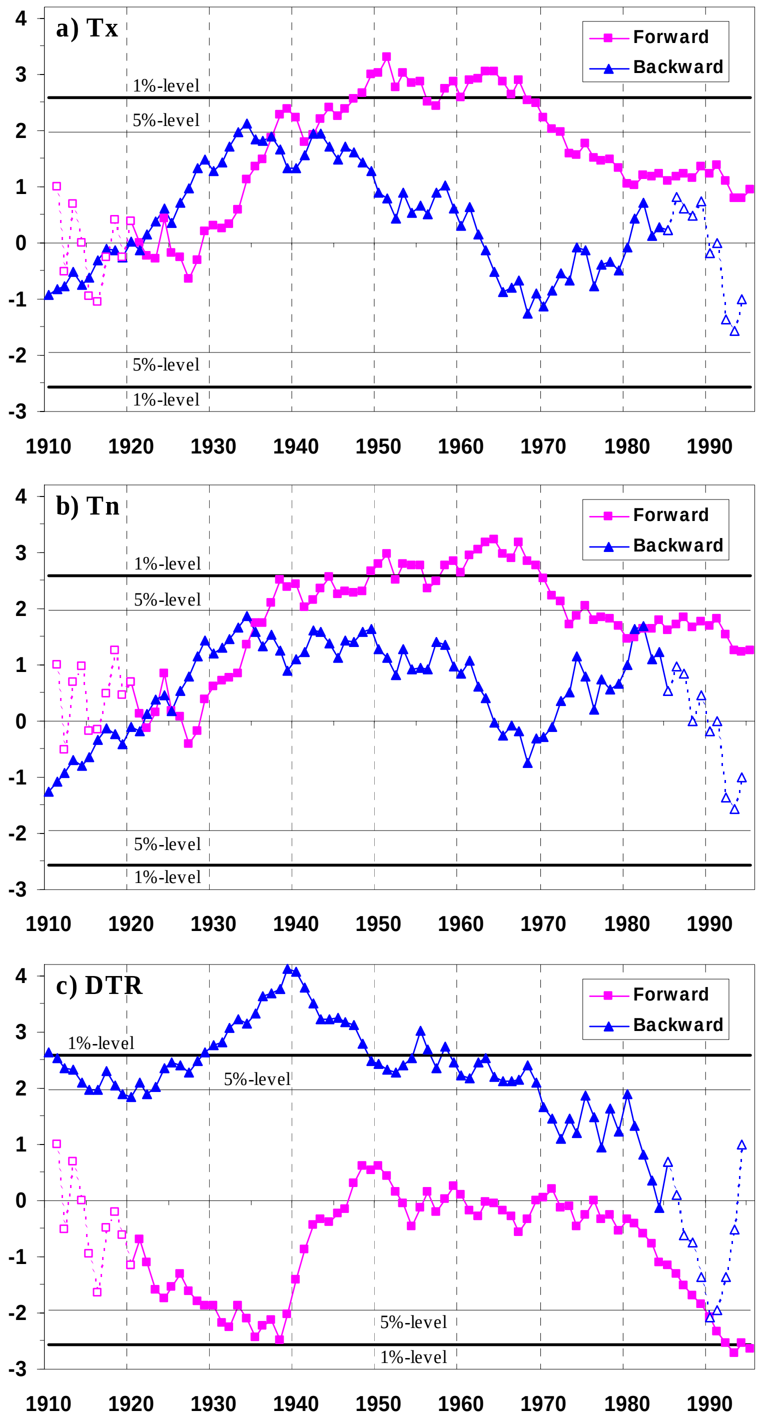

Fig. 3.2. The standardised Mann-Kendall test statistics for u(t) forward (from 1910) and u'(t) backward from (1995) testing of annual mean area-averaged Tx, Tn and DTR in Fennoscandia. 1%- and 5%-significance levels are also marked.

Table 3.1. Linear trends determined using least squares (LS) and Sen's method for Fennoscandian area-averages of Tx, Tn and DTR, 1910-95. Statistically significant trends (t-test for LS and Mann-Kendall test for Sen, both at the 5%-level) are indicated in bold type.

°C (10 yr)$^{-1}$	**Tx**		**Tn**		**DTR**	
	LS	**Sen**	**LS**	**Sen**	**LS**	**Sen**
DJF	0.006	0.015	0.007	0.012	-0.011	-0.014
MAM	0.096	0.098	**0.137**	**0.141**	-0.046	-0.042
JJA	0.022	0.020	0.056	0.061	-0.038	-0.043
SON	0.039	0.041	0.059	0.063	**-0.031**	-0.029
Year	0.037	0.041	0.059	0.063	**-0.030**	**-0.033**

3.3 The shape of the distribution

Of all the distributions in climatology, the normal distribution (or Gaussian distribution) is the single most important and widely used. Nevertheless, many observed climatic elements do not follow a normal distribution. Here two simple statistical parameters are introduced that describe non-gaussianity in the shape of a distribution.

The degree of asymmetry of the distribution is described with the coefficient of skewness, C_s,

$$C_s = \frac{\sum_{i=1}^{N}(x_i - \bar{x})^3}{(N-1)s^3} \qquad (3.8)$$

where x is the arithmetic mean, s the standard deviation and N is the number of observations (Kendall and Stuart 1958). A negatively-skewed distribution curve rises slowly, reaches its maximum and then falls rapidly. In other words, the "longer tail", as well as the mean and the median, are on the left-hand side of the mode. The opposite applies for positive skewness.

The degree of peakedness of the distribution is expressed with the coefficient of kurtosis, C_k,

$$C_k = \frac{\sum_{i=1}^{N}(x_i - \bar{x})^4}{(N-1)s^4} - 3 \qquad (3.9)$$

In statistics, kurtosis is the degree of flatness or peakedness in the region of the mode of a frequency curve. It is measured relative to the peakedness of the normal curve (the fourth-moment statistics for a normal distribution ~3, i.e. C_k ~0). C_k measures the extent to which a distribution is more peaked or flat-topped than the normal curve. For a normal distribution consisting of independent data the standard errors of C_s and C_k, i.e., E_s and E_k, respectively, are approximately

$$E_s = \sqrt{\frac{6}{N}} \qquad\qquad E_k = \sqrt{\frac{24}{N}} \qquad\qquad (3.10)$$

In the case of N=40, as in Table 5.8, E_s is 0.39 and E_k is 0.77. The 5% confidence levels for C_s and C_k are approximately twice the values of E_s and E_k.

4. ANALYSIS OF DIURNAL TEMPERATURE RANGE (DTR) AND RELATED TIME SERIES

An interesting recent finding based on observations is the worldwide decrease of DTR during the last fifty years or so (e.g. Easterling et al. 1997). It is possible that it is a signal of anthropogenic influence (cloud cover changes due to emissions of aerosols, land use changes, emissions of greenhouse gases) on the climate system (Folland et al. 2001, Nicholls et al. 1996). Observational studies of the possible mechanisms determining DTR variations have been performed by Karl et al. (1993), Dai et al. (1997a, 1999), Leathers et al. (1998), and Durre and Wallace (2001a,b). Detailed studies have focussed mainly on data from the USA. Tveito et al. (1998) present a mapped climatology of the annual mean DTR over Fennoscandia. Solantie and Drebs (2000) and Solantie (2003) have studied diurnal temperature variations in Finland. Geerts (2003) considered the main factors affecting DTR in all land-areas of the Earth. This section presents some recent analysis of DTR over the Fennoscandian region. Because no comprehensive DTR studies have been done in Finland, some aspects of DTR climatology are examined and area-averaged Fennoscandian DTR dependencies are studied.

4.1 Aspects of DTR climatology in Finland

Fig. 4.1 shows the long-term average annual cycle of DTR at three stations in Finland filtered with G3 (approximately corresponding to a 10-day running mean). Among the Finnish stations, Utö Island (59°47'N, 21°23'E) in the Baltic Sea is an example of maritime conditions. DTR is depressed by the large heat capacity of the sea although it exhibits the same features of the annual cycle (e.g. spring and summer maxima and a minimum in October-November) as the other two stations. DTR at Jokionen Observatory (60°49'N, 23°30'E) is quite typical of an inland station in southern and central Finland. Both the minimum after the spring maximum and also the following summer maximum occur later at Sodankylä (67°22'N, 26°39'E) in northern Finland. All stations experience a clear drop from summer to autumn. The magnitude of mean monthly DTR from station to station varies according to continentality (by as much as 7°C within Finland) but also by up to 3°C between neighbouring stations due to local characteristics (topography, openness, influence of lakes, thermal conductivity of soil, etc.).

The annual DTR cycle is associated both with large-scale and local effects. For example, the spring maxima of DTR and the following decrease occur roughly at the same time over the whole of Finland (Fig. 4.1). They must therefore be caused by variations in the average large-scale atmospheric circulation and humidity in the period 1961-90. A local effect, that appears to be visible in DTR, is the "green up" when the onset of plant transpiration moderates the increase in DTR (Schwartz 1996, Durre and Wallace 2001a). This occurs during the first half of May at Jokioinen and about a month later at Sodankylä (Fig. 4.1).

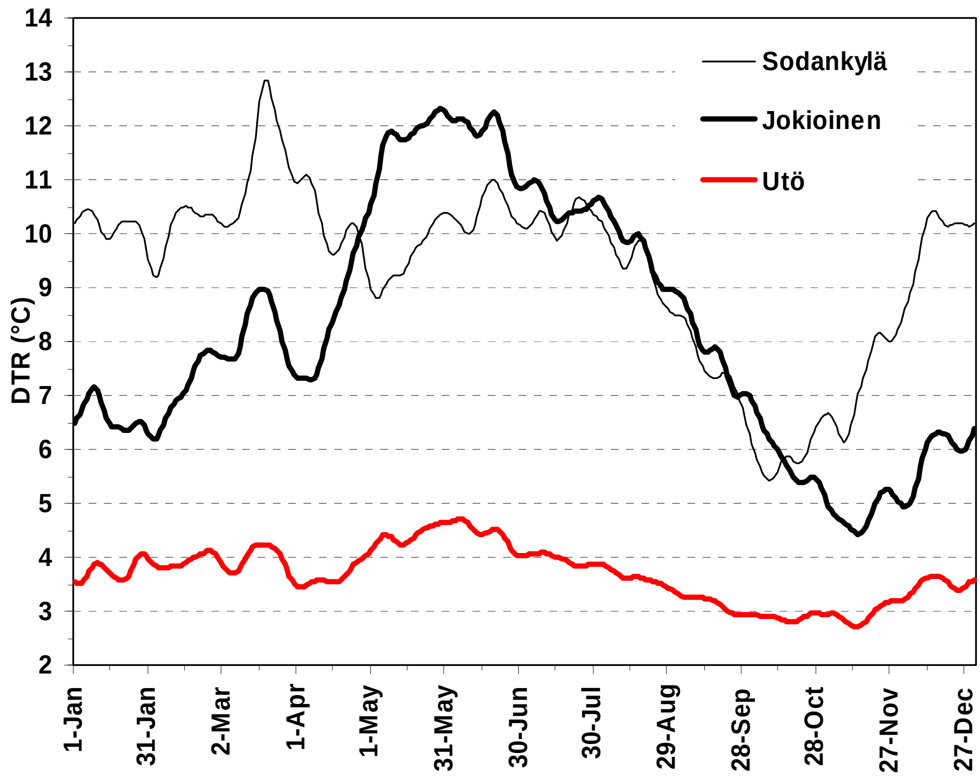

Fig. 4.1. G3 filtered (10-day) mean of DTR averaged over the period 1961-90 at Utö, Jokioinen and Sodankylä.

Daily synoptic data of several climatic elements from Jokioinen Observatory are analysed to separate mechanisms that are determining the annual cycle of DTR. Fig. 4.2 shows that the annual cycles of cloudiness and snow cover modify DTR. The average annual cycles of cloud cover and insolation largely determine the shape of the DTR annual cycle. The disappearance of snow cover coincides roughly with the spring minimum of DTR.

The highest DTR values have been related to changes of weather types in winter. Amplitudes larger than 30ºC have been recorded in Jokioinen during the 30 years. In spring and summer the daily temperature amplitude may reach values over 20ºC. The smallest DTR that have been observed are less than 1ºC during the period from autumn to spring. During summer variations at their smallest are about 2°C.

In order to evaluate possible relationships between DTR and other climatic elements, both simple (Pearson) and partial correlation coefficients were calculated between daily DTR and seven climatic elements at Jokioinen (Fig. 4.3). Daily means of dew-point temperature, cloud cover, amount of low cloud and wind speed were calculated from eight observations per day. Because of the huge sample size, even low correlations have high statistical significance.

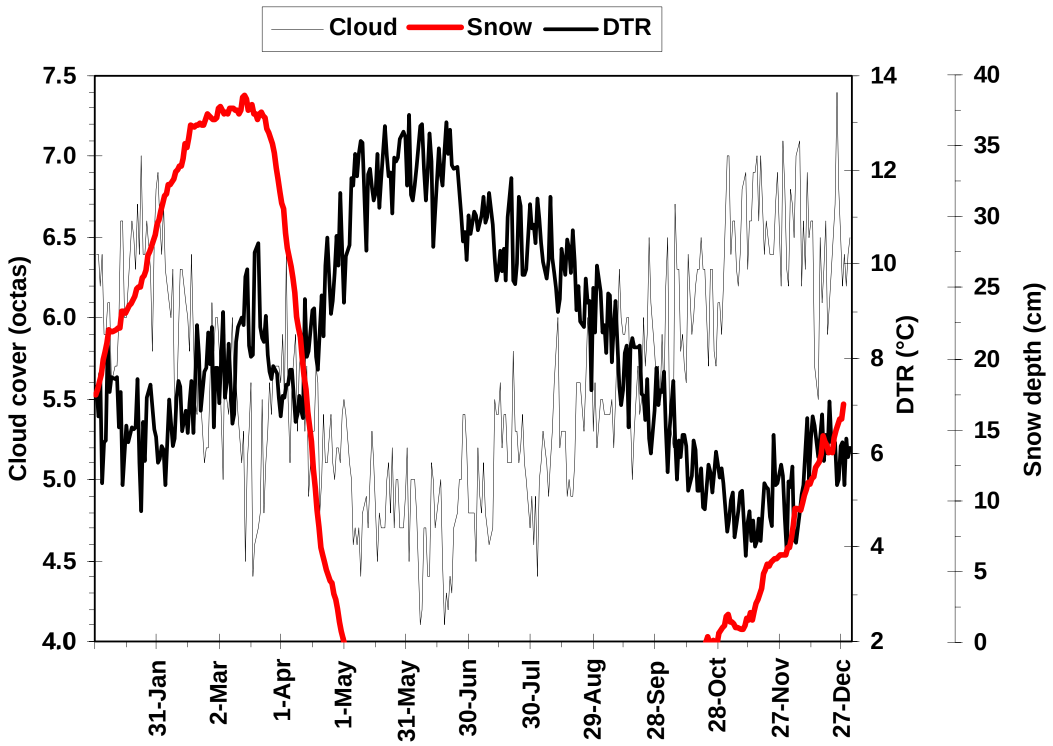

Fig. 4.2. Daily mean cloud cover, snow depth and DTR averaged over the period 1961-90 at Jokioinen (60°49'N, 23°30'E).

In order to measure the linear dependency between DTR and one particular variable, the influence of the other variables was partialled out (eliminated) by linearly regressing DTR on the other variables (Fig. 4.3). It should be cautioned that non-linear relationships will not necessarily be described satisfactorily using correlation analyses, and that the causal interrelationships cannot be correctly interpreted if important variables are missing.

Snow depth has low simple correlation coefficients (Fig. 4.3a). The partial correlation results suggest that after elimination of other variables, snow cover tends to reduce DTR during spring, but during winter snow cover is positively correlated with DTR (Fig. 4.3b). In Fig. 4.4 the DTR and snow depth series are fixed by the date of disappearance of permanent snow cover, which occurred at Jokioinen between 21st March and 10th May during the period 1961-1990.

The gradual disappearance of snow cover reduces albedo, which increases the amount of the downward short wave energy converted to sensible and latent heat flux. After the decrease of albedo, DTR rapidly increases with the increase of temperatures towards summer. However, before the total disappearance of snow, it seems that the melting of snow causes a dip in the otherwise increasing trend of DTR from late winter to early summer. The dip also seems to coincide with a narrow peak in cloud cover (Fig. 4.2).

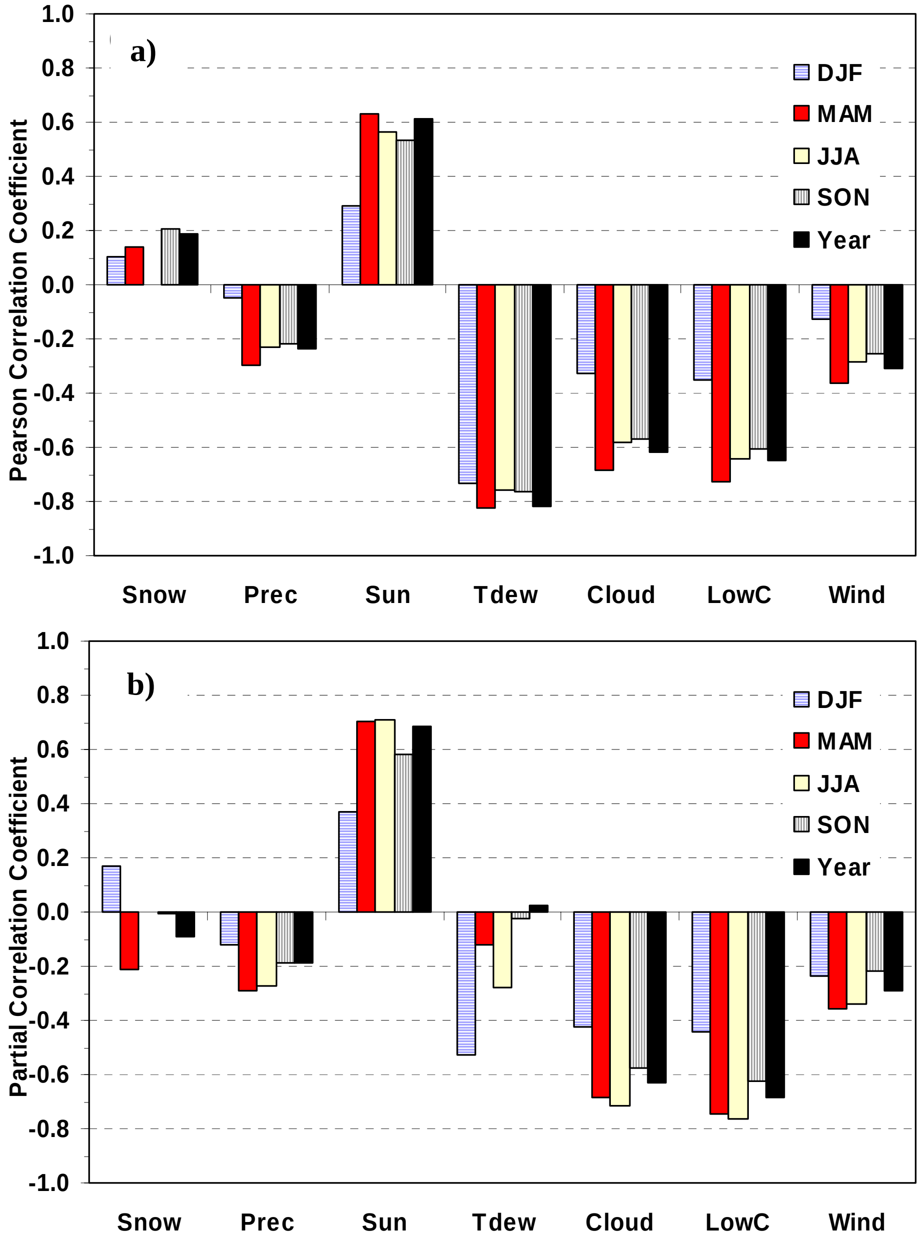

Fig. 4.3. Simple (Pearson) a) and partial b) correlation coefficients between daily DTR and snow depth at 06 UTC (Snow), daily precipitation (Prec), daily sunshine hours (Sun), daily mean dew-point temperature (Tdew), daily mean cloud cover (Cloud), daily mean low cloud cover (LowC) and daily mean wind speed (Wind) at Jokioinen (60°49'N, 23°30'E) during the period 1961-90.

Melting of snow and re-freezing of water are "extra terms" in the surface energy balance equation mainly effective in spring. Melting requires energy, which suppresses Tx. During cold nights re-freezing releases latent heat hence elevating Tn. Theoretically these processes (melting and re-freezing) lead to a decrease of DTR. More importantly, the remaining snow no longer insulates effectively, and heat fluxes between ground and atmosphere can be important. The snowmelt may also lead to a rise in evaporation and afternoon relative humidity that may lead to an increase in low-level clouds. However, without computations of all terms in the surface energy balance equation it is not possible to judge which processes are important in magnitude.

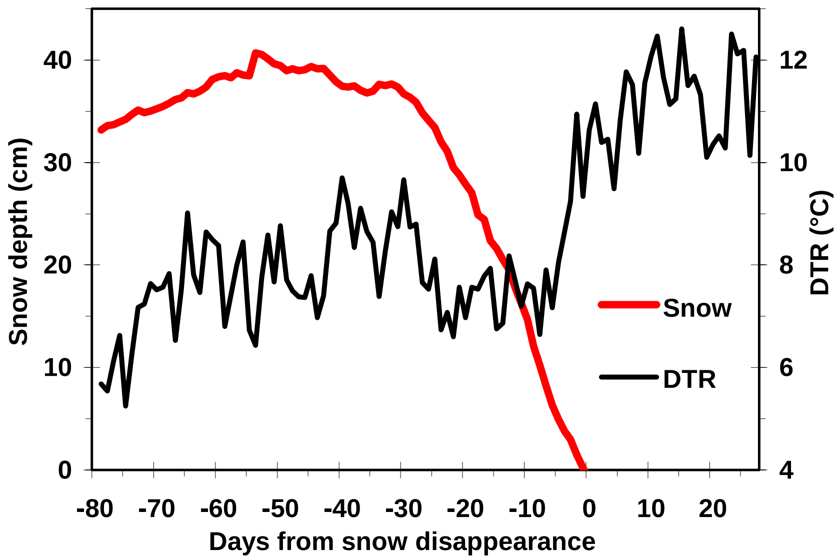

Fig. 4.4. Mean snow depth and DTR around the disappearance of snow (0 on the vertical axis) averaged over the period 1961-90 at Jokioinen (60°49'N, 23°30'E).

Precipitation is negatively correlated with DTR, especially during spring. According to Dai et al. (1999) soil moisture also reduces DTR in Kansas, USA. Soil moisture was not measured in Jokioinen. It is likely that in Finland soil moisture can play some role in spring and summer but not in autumn or winter.

Daily sunshine hours are strongly correlated with DTR, mainly through its influence on Tx. As expected, the lowest correlation occurs during winter, when short wave radiation has its minimum. Daily dew-point temperature shows strong negative (simple) correlation mainly via Tn, but the partial correlation coefficients are much lower except in winter, when dew-point temperatures are linked to weather types. The overall cloud cover and the amount of low cloud are very strongly correlated with DTR. Cloud cover, especially low clouds, reflects the sunlight and depresses Tx. Especially during wintertime the greenhouse warming effect of clouds on Tn is evident and narrows DTR. At Jokioinen, the effect of high wind speeds in reducing DTR is probably related to the prevention of surface inversions and hence avoidance of low minimum temperatures.

The results are in general agreement with those of Karl et al. (1993), Dai et al. (1999), and Durre and Wallace (2001b), but there are some differences due to the northern location of Jokioinen, which affects the annual cycle and the dependencies between climatic elements. DTR is sensitive to changes in the surface radiative, latent, and sensible heat fluxes. DTR is widely available and longer time series of DTR exist than of actual flux measurements. The interpretation of variations in DTR is complex due to its links with several factors whose importance varies geographically. In this section, there was no variable to describe the effects of advection of air masses with different characteristics. This is discussed in the following section and in paper II.

4.2 DTR dependencies in Fennoscandia

Besides the analysis of long-term temperature trends, linkages between different climatic elements were also studied in paper II. Tuomenvirta et al. (1998) studied relationships between area-averaged Fennoscandian temperatures (mean, mean daily maximum, mean daily minimum, highest and lowest of each month), temperature ranges (diurnal and extreme), atmospheric circulation indices (the North Atlantic Oscillation index - NAO, and zonal and meridional geostrophic flows as well as mean pressure over Fennoscandia) and cloud cover (see paper II for symbols and details about circulation indices and cloudiness). Simple (Pearson) correlation coefficients between the seasonal values of several climatic elements are presented in a matrix (Table 4.2). Inter-annual correlations are of most interest, but correlations between 5-year means are also shown if they are statistically different and significant.

A *t*-test can be applied to evaluate if a correlation coefficient differs significantly from zero. The confidence intervals for correlation coefficient (r) were estimated by transforming r to a normally-distributed variable (z) and utilising the *t*-test (Hald 1952). 95%-confidence intervals were calculated to examine whether correlation coefficients between 5-year means differed significantly from coefficients between yearly values (Table 4.2, below diagonals). In addition, it was required that at least either 1- or 5-year correlation coefficients be significantly different (95%-level) from zero. The possible effect of autocorrelation on the correlation coefficients was not taken into account.

The information in Table 4.2 allows some general remarks to be made about the Fennoscandian climate:

- As expected, 1-year correlation coefficients between T, Tx and Tn are very high and the correlation coefficients of 5-year averages are quite similar. Th and Tl exhibit somewhat lower correlation coefficients with T, Tx and Tn. The correlation between Th and Tl is at its lowest in summer, but is still highly significant.
- Temperature ranges generally show significant correlations with temperatures. Note that DTR correlates positively with both Tx and Tn in summer and negatively with both Tx and Tn in other seasons.
- Zonal circulation indices (NAO and ZI) describe the advection of maritime or continental air masses to Fennoscandia. NAO and ZI correlate mostly significantly with temperatures. In summer, temperature correlations with NAO are positive but are negative with ZI. Due to weak zonal atmospheric circulation, the westerly location of stations used to calculate the NAO index, and the large spatial scale of NAO, it has a different relationship to Fennoscandian temperatures than ZI. The temperature ranges are predominantly negatively correlated with zonal circulation indices because westerly winds often bring cloudy and windy weather.
- The meridional circulation index (MI) does not correlate highly with temperatures although some significant correlations can be found. For instance, in autumn the negative correlation coefficient between MI and DTR is highly significant, presumably reflecting the effect of humid cloudy weather that is often associated with southerly winds.

- Cloud cover anomaly (CA) correlates positively with temperatures in winter and negatively in summer. The negative correlation coefficient between CA and DTR is highly significant. Cloud cover effectively controls incoming shortwave radiation and outgoing longwave radiation. The strongest correlations with air pressure anomaly (PA) are also found for cloud cover, CA. Correlation coefficients between PA and temperatures are negative in winter and positive in summer as would be expected.
- In summer, the use of 5-year means reduces the magnitude of correlation coefficients between CA and temperatures but in spring correlation are increased due to rising trends in both CA and temperatures.

In paper II it was shown that a long-term decrease of DTR is the most dominant trend in the Fennoscandian data, consistent with large areas of the world (Easterling et al. 1997). A multiple linear regression model based on the results in Table 4.2 was constructed for Fennoscandian DTR in paper II. The simple model successfully identifies the link between decreases of DTR and increases in cloud cover and a strengthening of the zonal circulation. In paper II the basic physical relationships between DTR and the analysed variables are explained, but the treatment remains mainly statistical.

Table 4.2. Above grey diagonals: seasonal correlation coefficients between T, Tx, Tn, Th, Tl, DTR, intra-monthly ETR, NAO, zonal (ZI) and meridional (MI) circulation indices, cloud cover anomaly (CA) and sea-level pressure anomaly (PA) in Fennoscandia calculated for the period 1910-95 (see paper II for definitions of ZI, NAO, MI, CA and PA). Limits of statistical significance are ±0.21 (95%), ±0.28 (99%) and ±0.35 (99.9%) for the correlation coefficients. Below grey diagonals: seasonal correlation coefficients of 5-year means are shown if the difference between 1- and 5-year correlation coefficients is significant at the 95%-level (see text for details) and either of the coefficients is significant at the 95%-level. *(Table on following page)*

DJF	T	Tx	Tn	Th	Tl	DTR	ETR	NAO	ZI	MI	CA	PA
T		1.00	0.99	0.84	0.95	-0.68	-0.80	0.65	0.77	-0.07	0.40	-0.33
Tx			0.98	0.87	0.94	-0.62	-0.76	0.64	0.80	-0.14	0.33	-0.34
Tn				0.82	0.96	-0.75	-0.82	0.65	0.72	-0.00	0.44	-0.33
Th					0.74	-0.38	-0.42	0.51	0.81	-0.35	0.10	-0.27
Tl						-0.74	-0.92	0.65	0.67	0.03	0.41	-0.25
DTR					-0.84		0.78	-0.48	-0.20	-0.49	-0.67	0.20
ETR								-0.58	-0.43	-0.25	-0.50	0.17
NAO						-0.27			0.63	0.17	0.37	-0.43
ZI	0.66	0.69	0.59		0.51			0.77		-0.42	-0.00	-0.31
MI					0.34		-0.58		-0.23		0.67	-0.11
CA							-0.73			0.86		-0.35
PA				-0.03				-0.11	-0.03			
MAM	**T**	**Tx**	**Tn**	**Th**	**Tl**	**DTR**	**ETR**	**NAO**	**ZI**	**MI**	**CA**	**PA**
T		0.97	0.97	0.76	0.88	-0.30	-0.38	0.64	0.52	0.15	0.07	-0.07
Tx			0.90	0.82	0.80	-0.07	-0.24	0.60	0.47	0.09	-0.10	0.08
Tn				0.66	0.90	-0.51	-0.48	0.63	0.52	0.22	0.29	-0.19
Th		0.90			0.51	0.10	0.24	0.49	0.38	-0.01	-0.22	0.20
Tl						-0.47	-0.71	0.50	0.41	0.17	0.24	-0.16
DTR							0.61	-0.24	-0.26	-0.32	-0.85	0.59
ETR								-0.16	-0.16	-0.19	-0.44	0.35
NAO									0.50	0.30	0.05	-0.18
ZI										-0.15	0.15	-0.40
MI						-0.74	-0.42				0.26	-0.09
CA	0.43	0.34	0.59	0.25	0.59	-0.72				0.51		-0.58
PA	-0.32							-0.50		-0.39	-0.19	
JJA	**T**	**Tx**	**Tn**	**Th**	**Tl**	**DTR**	**ETR**	**NAO**	**ZI**	**MI**	**CA**	**PA**
T		0.98	0.93	0.84	0.68	0.66	0.53	0.42	-0.33	0.24	-0.66	0.51
Tx			0.87	0.88	0.60	0.77	0.61	0.37	-0.34	0.21	-0.73	0.60
Tn				0.71	0.79	0.35	0.33	0.49	-0.20	0.34	-0.36	0.30
Th					0.35	0.73	0.87	0.17	-0.36	0.20	-0.68	0.60
Tl						0.12	-0.16	0.46	-0.19	0.30	-0.22	-0.01
DTR							0.70	0.08	-0.40	-0.05	-0.91	0.73
ETR								-0.05	-0.27	0.05	-0.59	0.64
NAO									0.30	0.09	-0.14	0.05
ZI						-0.63	-0.46	0.53		-0.10	0.38	-0.16
MI								0.52	0.38		0.11	-0.18
CA	-0.43	-0.50	-0.07	-0.47		-0.83	-0.41		0.70			-0.69
PA				0.85			0.85				-0.45	
SON	**T**	**Tx**	**Tn**	**Th**	**Tl**	**DTR**	**ETR**	**NAO**	**ZI**	**MI**	**CA**	**PA**
T		0.99	0.99	0.70	0.86	-0.29	-0.47	0.62	0.38	0.44	0.31	-0.06
Tx			0.95	0.75	0.80	-0.14	-0.37	0.57	0.41	0.35	0.20	-0.02
Tn				0.64	0.88	-0.44	-0.52	0.64	0.34	0.52	0.43	-0.13
Th					0.40	0.11	0.22	0.25	0.38	0.04	0.02	0.07
Tl			0.81			-0.47	-0.81	0.70	0.23	0.57	0.37	-0.10
DTR					-0.28		0.56	-0.38	0.10	-0.64	-0.79	0.37
ETR			-0.31			0.13		-0.58	-0.00	-0.58	-0.38	0.15
NAO	0.49	0.42							0.42	0.53	0.32	-0.27
ZI	0.13	0.12	0.13				0.43	0.01		-0.22	-0.10	-0.36
MI	0.08	0.01	0.20	-0.27	0.36				-0.44		0.60	-0.13
CA				0.31		-0.64	0.28			0.21		-0.37
PA	0.30	0.36			0.23	0.54			-0.03		-0.54	

Table 4.2. (Caption on previous page)

According to Dai et al. (1999), the twentieth-century variations in DTR correlate with variations in cloud cover and precipitation in the USA, Australia, midlatitude Canada and the former U.S.S.R. In their study the annual cloud-DTR correlation coefficient in Europe (-0.4) is weaker than in other regions, and also weaker than the Fennoscandian coefficient in paper II (-0.8). The weak correlation coefficient may be due to the effect of averaging over the whole of Europe, where correlations between cloud cover and precipitation totals and NAO have a different sign in northern than in southern Europe (Hurrel and van Loon 1997). It may also be due to the poor quality of data used by Dai et al. During the last 5 decades, they report a decrease of DTR consistent with increasing trends of cloud cover and precipitation over land areas. DTR linkages with precipitation were not studied in paper II, but based on recent studies of precipitation (Heino 1994, Hanssen-Bauer and Førland 1994, Førland et al. 1996b and paper I), it seems that Fennoscandia follows this pattern.

Based on their analyses, Dai et al. (1999) concluded that the direct effect of an increase of greenhouse gases and aerosols on DTR reduction must be small. The anthropogenic influence, if there is any, is likely to be manifest through clouds, due to their mainly asymmetric forcing of the diurnal cycle of temperature. The amount, properties and lifetime of clouds are affected by changes in atmospheric humidity and loadings of aerosols. Thus increased levels of greenhouse gases may have contributed to changes in the hydrological cycle and cloud cover. Similarly anthropogenic emission of aerosols may indirectly contribute to a narrowing of DTR.

Recently, Forster and Solomon (2003) found a weekly cycle of DTR for many stations in the United States, Mexico, Japan and China. The size of the human-induced, weekend effect is comparable to that of the observed long-term trends in DTR. They conclude that the observed weekend effect is likely to be linked to aerosol-cloud interactions. Kalnay and Cai (2003) compare the observed temperatures in the continental United States over the past 50 years with the corresponding reconstruction from reanalysis data (Kalnay et al. 1996), which are insensitive to surface temperature observations. Their results suggest that about half of the observed decrease in DTR is due to urban and other land-use changes in the continental USA.

In Finland, Venäläinen et al. (1999) have shown that changes in land use may significantly change surface characteristics and the surface energy balance. This may have a significant effect on Tx and/or Tn and modify DTR. Paper II concludes that changes in land use (urbanisation, de- and reforestation, etc.) are mostly carried out at the local or regional scale, and are only likely to have provided a minor contribution to DTR reduction at the hemispheric or Fennoscandian scale. However, taking into account the recent results of Forster and Solomon (2003) and Kalnay and Cai (2003), it would seem necessary to perform further analysis on the causes of the observed narrowing of DTR in Fennoscandia. Recently, DTR has continued to decrease while at the same time anomalies of cloud cover have remained positive in Fennoscandia (Fig. 4.5; update of Figs. 6 and 8 in paper II to the year 2000).

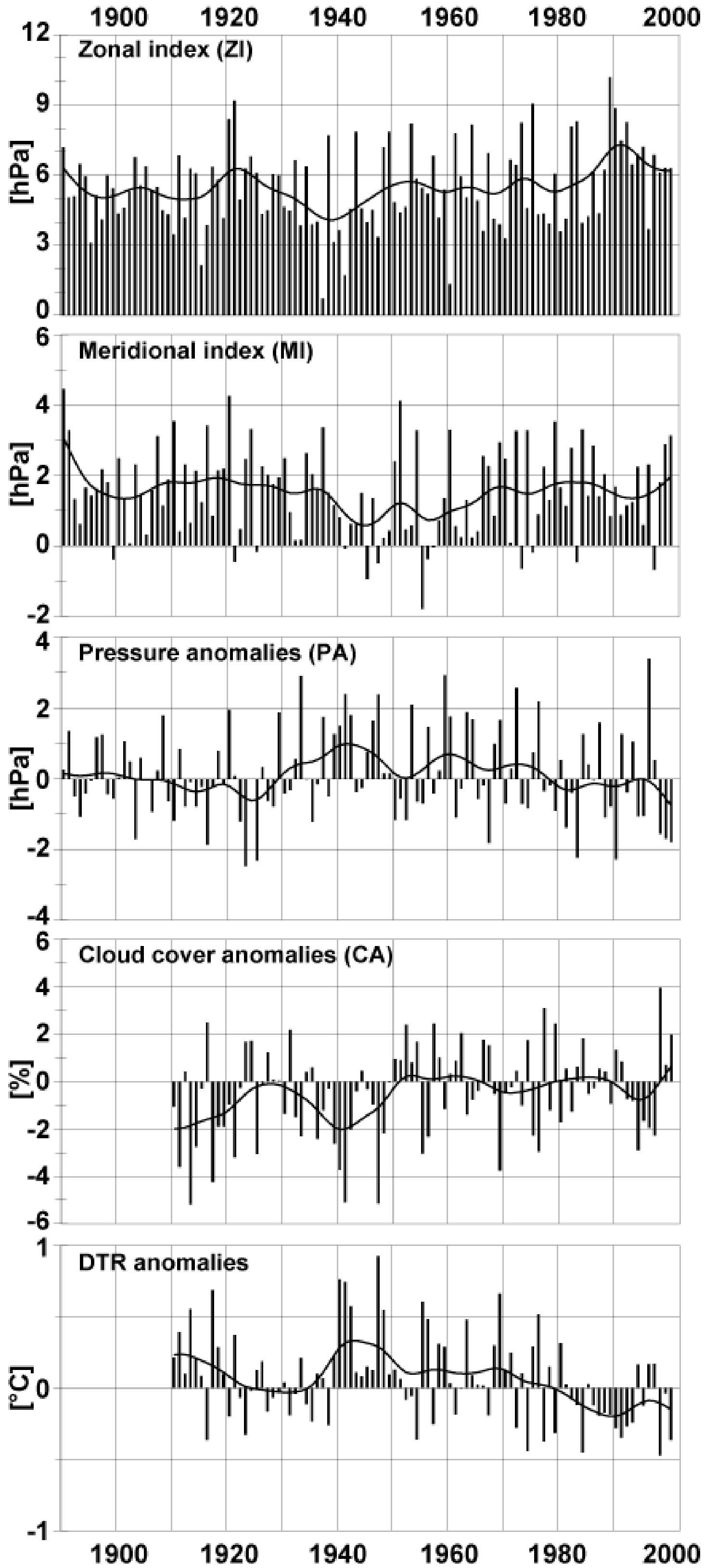

Fig. 4.5. Annual time series of the zonal index (ZI), meridional index (MI), pressure anomalies (PA), cloud cover anomalies (CA) and diurnal temperature range (DTR) anomalies for Fennoscandia. ZI, MI and PA are for 1890-2000, and CA and DTR for 1910-2000. Definitions of the indices are given in paper II. Bars show the annual values. The smooth curves display the G3 filtered series.

5. OBSERVED ANNUAL MEAN TEMPERATURE CHANGES IN FINLAND

In paper I, national average series were calculated from homogeneous temperature and precipitation series up to the year 1995. It was possible to update the temperature series to year 2002 based on 8 stations, but due to the diminished network this was not possible for the precipitation series based on 24 stations. The First Difference Method (FDM) approach for calculating unbiased national average time series was therefore developed. This approach is described, validated and applied in this chapter. In addition, the uncertainty of the mean temperature series is discussed, and annual mean temperature trends are analysed.

5.1 First Difference Method (FDM) for calculating unbiased area-average series

So far, large area-averages have been based on a few homogenised station series presented in papers I and II. In this section, an alternative approach using FDM is applied, in which nation-wide homogeneity breaks are adjusted, but other breaks are ignored. The justification for this approach comes from paper III and section 2 of this summary, where it is shown that for a large number of stations the inhomogeneities appear to be nearly random, i.e., the mean adjustment is close to zero (temperature) or one (precipitation). Concerning the nation-wide breaks in Finland, the formula for calculating daily mean temperatures has been changed on three occasions (in 1880, 1900 and 1926), thus causing systematic changes. However, these nation-wide homogeneity breaks can be handled based on earlier research. Heino (1994) determined adjustments for the entire temperature measurement network in Finland. Similarly, there are studies that can be used to adjust the precipitation gauge type changes in 1909 and 1981/82 (Solantie and Junila 1995, Heino 1994, Korhonen 1913).

In FDM, the differences between successive years at a station are used instead of the actual values. This allows the use of all data without defining common baselines for temporal anomaly calculations, as in the Climate Anomaly Method (CAM) that is used e.g. in Jones (1994) or the Reference Station Method (RSM) used by Vinnikov et al (1990). Peterson et al. (1998) have described and compared FDM, CAM and RSM. In the calculation of hemispheric temperature anomalies, all the methods produced quite similar results. FDM maximises accessible station records. In this study, the minimum length of time series that can be used is two years. For example, the year-to-year change of mean temperature, ΔT_i is

$$\Delta T_i = T_{i+1} - T_i \qquad (5.1)$$

where *i* refers to year. Thus by defining the annual mean temperature for a certain year it is possible to use a series of ΔT_i to construct time series of T or anomalies of T. Year-to-year changes in the annual averages are spatially conservative and, therefore, more suitable for area averaging than the actual values (Jones and Hulme 1996). Because area averaging is done for ΔT_i the use of sophisticated spatial interpolation methods is not needed. Besides temperatures, other climatic variables can be used in FDM (Peterson et al. 1998).

In the alternative approach to calculate national time series, the temperature and precipitation data described in section 2 are utilised. Series are established using the Finnish LPNN grid (see Figs. 2.3 and 2.4). The grid box value is the simple average of year-to-year differences at stations within the grid box. In area-averaging, each grid box is weighted with its land area within the borders of Finland. The few stations just outside LPNN grid boxes (e.g. Fig. 2.3) are included in the nearest grid boxes. If there are no stations within the grid box, simple bi-linear interpolation of the nearest grid box values is used in computing the missing value. The spatial methods used are simple and at a fairly low resolution, but they are sufficient for this study, because only one value per year, i.e. the average over Finland, is passed on to validation and climatic analysis.

5.2 Comparison of temperature series

Finnish annual mean temperatures are calculated for the period 1888-2002 from a series of ΔT_i using all available data. There are two versions: T_{ALL}(adj) from homogeneity tested and adjusted temperature series and T_{ALL}(ori+) from the original data corrected for averaging formula changes. The Finnish temperature series based on 8 stations (paper I) are updated and a version based on just 4 long-term stations (Fig. 2.4c) is compiled. Versions based on the original data are also shown for contrast. Comparisons are also made with data derived from the global data sets maintained at the CRU/UEA, England (Jones and Moberg 2003, Mitchell et al. 2002). T_{TYN}(FIN) is interesting because it has a fine spatial resolution and, therefore, the gridded data suits many climatic applications. T_{CRU}(FS) is an extensively-revised new version of the popular Climatic Research Unit (CRU) temperature anomaly data set (Jones 1994) used widely, e.g., in IPCC (2001). It can also be extended back to 1851 with a reasonable coverage over Fennoscandia (A. Moberg, 2002, personal communication). Table 5.1 gives details of the selected temperature series.

Because T_{ALL}(adj) is area-averaged and uses all available, homogenised temperature series, it is selected as the baseline for the comparison. Despite its merits, there is no indisputable way to prove that T_{ALL}(adj) is very close to the "true value", so the term "difference" is used instead of "error". Fig. 5.1 shows the annual differences between the other series and T_{ALL}(adj) during the common period 1901-2000. T_{ALL}(ori+) is very similar to T_{ALL}(adj) back to about the year 1950. Even during the early part of the 20th century the differences are mostly equal to or less than 0.05ºC. T_{ALL}(ori+) has a very small bias and the smallest RMS difference (Table 5.2). T_8(adj) differences remain less than ±0.3ºC, but are clearly larger than those for T_{ALL}(ori+); this can also be seen in the RMS difference in Table 5.2. There is also a systematic difference of about 0.15ºC during the three first decades of the 20th century. T_8(adj) is based on eight stations giving an area average that is biased towards southern Finland; differences can be expected, therefore. The origin of the bias can be climatic, e.g. due to different trends in northern and southern Finland, or artificial e.g. due to inaccuracies in the homogeneity adjustments. T_8(ori) is clearly biased, being about half a degree too warm at the beginning of the 20th century. The statistics of T_8(ori) in Table 5.2 are also the "worst". The bias is only partly explained by formula changes; most of it is due to individual station inhomogeneities. T_4(adj) has a slightly larger variability of differences than T_8(adj) but is otherwise quite similar; this is because there are three common stations in the series. T_4(ori) is severely biased during the two first decades of the century.

Table 5.1. Annual mean temperature series used in this study.

Symbol	Time period	Stations/Area	Homogenisation	Area averaging	Remarks /*references*
T_{ALL}(adj)	1888-2002	Ranging from 25 to 161 stations in Finland and neighbouring areas	Stations tested and adjusted with SNHT plus adjustments for formula changes	Area-averaged difference series in the LPNN grid	*Section 2*
T_{ALL}(ori+)	As T_{ALL}(adj)	As T_{ALL}(adj)	Original station data plus adjustments for formula changes	As T_{ALL}(adj)	As T_{ALL}(adj)
T_8(adj)	1901-2002	8 stations in Finland (see Fig. 2.4)	Stations tested and adjusted with SNHT plus adjustments for formula changes	Simple average of station values	Series differ insignificantly from *paper I*
T_8(ori)	As T_8(adj)	As T_8(adj)	Original station data	As T_8(adj)	As T_8(adj)
T_4(adj)	1847-2002	4 stations in Finland (Helsinki, Kuopio, Kajaani and Oulu, see Fig. 2.4)	Stations tested and adjusted with SNHT plus adjustments for formula changes	Simple average of station values	Series is based on only three stations for ca. 7 years*
T_4(ori)	As T_4(adj)	As T_4(adj)	Original station data	As T_4(adj)	As T_4(adj)
T_{TYN}(FIN)	1901-2000	Covers Finland; uses somewhat older version of station data than T_{CRU}(FS)	Homogeneity control	Area-averaged anomaly series in 0.5ºx0.5º latitude-longitude grid	Sophisticated spatial interpolation *Mitchell et al. (2002)*
T_{CRU}(FS)	1851-2000	Covers Fenno-scandia (land area within 55-70ºN and 5-30ºE)	Use of homogenised data sets and homogeneity control	Area-averaged anomaly series in 5ºx5º latitude-longitude grid	*Jones and Moberg (2003)*

* About 9 years missing from Kuopio during 1875-90 are interpolated from Tampere. About 14 years missing from Kajaani during 1873-86 are interpolated based on data from Oulu.

T_{TYN}(FIN) is a "true" area average over Finland. There is a small warm bias prior to the 1960s that increases to about 0.2ºC during the 1920s. T_{TYN}(FIN) has the largest bias of the homogenised series. The station data it is based on seem to contain inhomogeneities. For example, it appears that the change in the averaging formula adopted in 1926 has not been taken into account. T_{CRU}(FS) is an area average over Fennoscandia; one cannot expect it to closely follow just the temperature of Finland, e.g., it has the largest RMS difference. However, it is based on the best available long-term series for the region, and the series is unbiased compared to T_{ALL}(adj) (Table 5.2).

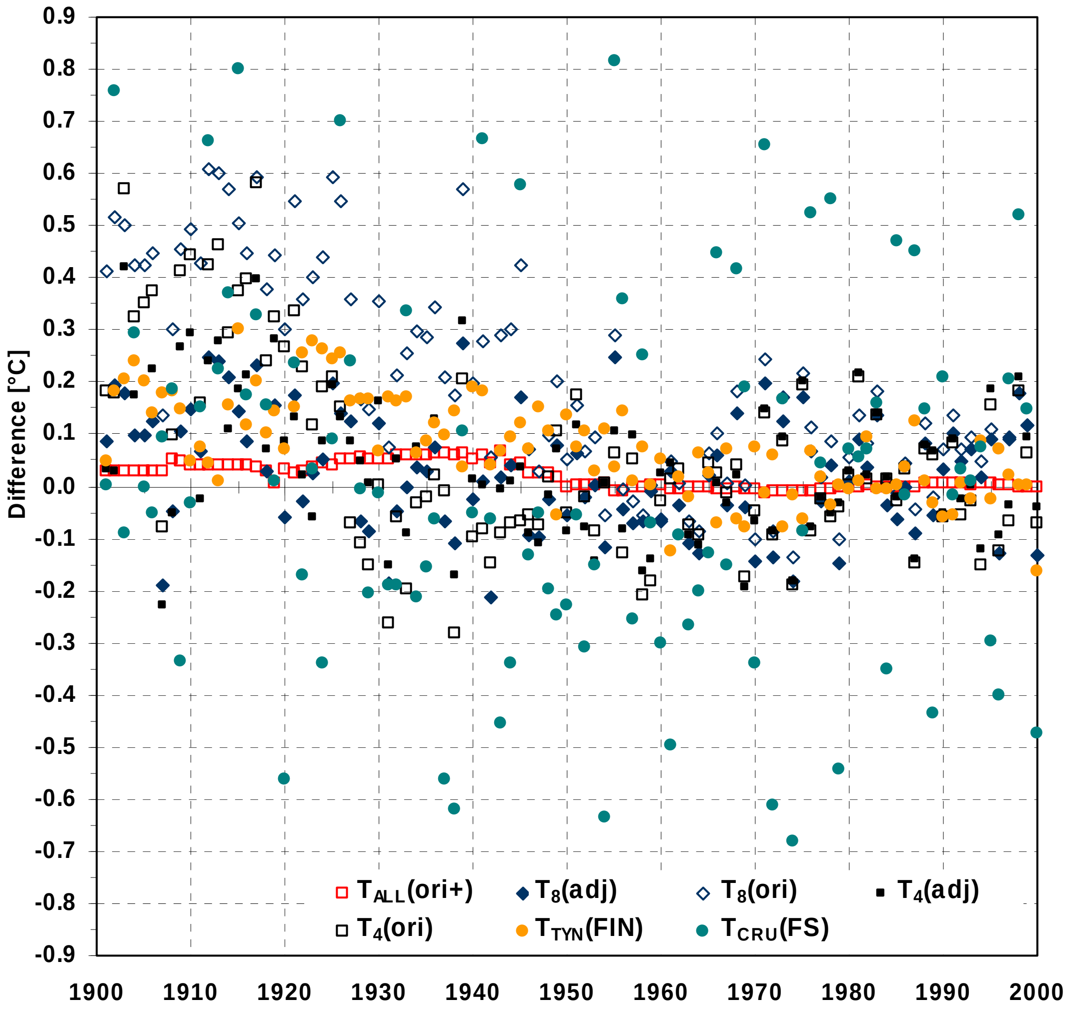

Fig. 5.1. Comparison of annual mean temperature series, 1901-2000. Differences between the other series and T_{ALL}(adj) (see Table 5.1 *for explanation of the series).*

In terms of interannual variability the series are quite similar. Table 5.2 contains the correlation coefficients of annual temperatures between T_{ALL}(adj) and the other series. T_{ALL}(ori+) has the highest correlation. Also T_8(adj), T_4(adj) and T_{TYN}(FIN) have very high correlation coefficients, although T_8(adj)/T_4(adj) is based on only 8/4 stations and T_{TYN}(FIN) has a clear bias. The unbiased T_{CRU}(FS) has a slightly lower correlation because it represents a larger geographical region. The non-homogenised series are also highly correlated with T_{ALL}(adj). The year-to-year variability is large compared to the magnitude of the inhomogeneities, and the correlation coefficient therefore remains high.

The magnitude of the variability for the various series is also quite similar. The standard deviation of the annual mean temperature of all the series representing Finland during the period 1901-2000 is almost the same, i.e. 1.05°C±0.02°C. Because T_{CRU}(FS) covers the whole Fennoscandia, its standard deviation is smaller, namely 0.85°C.

Table 5.2. Comparison of other annual mean temperature series with T_{ALL}(adj) during the period 1901-2000. Shown are the correlation coefficient, mean bias and root mean squared (RMS) difference with T_{ALL}(adj).

Period 1901-2000	T_{ALL}(ori+)	T_8(adj)	T_8(ori)	T_4(adj)	T_4(ori)	T_{TYN}(FIN)	T_{CRU}(FS)
Mean bias [ºC]	0.020	0.030	0.196	0.041	0.057	0.075	0.013
RMS difference [ºC]	0.031	0.116	0.282	0.137	0.192	0.120	0.346
Correlation coefficient	1.000	0.994	0.981	0.992	0.985	0.996	0.960

The series do have large differences in the linear trend estimates. As in section 3, both Sen's and the least squares method are used for linear trend estimation, and the Mann-Kendall test for statistical significance. The original series T_8(ori) and T_4(ori) underestimate the warming during the 20th century. The error is smaller in T_4(ori), where the urban warming of Helsinki (Heino 1994), fortuitously, partly compensates for the inhomogeneities in the other stations. In T_{TYN}(FIN) the warming trend is misrepresented due to problems in the early decades. Other trend estimates are within 0.061-0.076 ºC(10 yr)$^{-1}$, where the series using all available data in Finland give the largest warming. All homogenised series, except T_{TYN}(FIN), show significant warming at the 90%-confidence level, T_{ALL}(adj) and T_{CRU}(FS) even at the 95%-confidence level. In the case of the former, this is due to the largest warming; in the latter case it is due to the smaller variability in the large area average. Likewise, the approximately similar-sized warming of the global mean surface temperature is highly significant (at the 99%-confidence level; Folland et al. 2001).

Table 5.3. Linear trends determined using Sen's method (in bold), the Mann-Kendall test statistic u(t) (in italics) and the trend estimated using the least squares (LS) method (in normal font) for annual mean temperatures, 1901-2000. The significance levels of the linear trend are: 1.65 (90%), 1.96 (95%), 2.58 (99%).

Period 1901-2000	T_{ALL}(adj)	T_{ALL}(ori+)	T_8(adj)	T_8(ori)	T_4(adj)	T_4(ori)	T_{TYN}(FIN)	T_{CRU}(FS)
Sen Trend [ºC(10 yr)$^{-1}$]	**0.076**	**0.069**	**0.068**	**0.025**	**0.061**	**0.046**	**0.051**	**0.061**
u(t)	*1.98*	*1.83*	*1.92*	*0.65*	*1.65*	*1.28*	*1.53*	*2.11*
LS [ºC(10 yr)$^{-1}$]	0.076	0.070	0.067	0.023	0.060	0.045	0.054	0.063

T_{ALL}(adj) and T_{ALL}(ori+) are very similar during the period 1888-2002. The maximum difference of 0.07ºC is reached in 1943; this might be related to more or less systematic station relocations to cooler locations not taken into account in T_{ALL}(ori+). The systematic difference between the homogeneity-tested and the original series is quite small. Generally speaking, the series created with FDM do not differ greatly from the other properly homogenised estimates of Finnish mean temperature. It therefore seems a feasible approach to use FDM for annual precipitation sums, too.

5.3 Sources of error in temperature series

Comparison of series already gives an estimation of the inaccuracies in the annual mean temperatures. In this chapter some specific error sources are considered and the reliability of the series in the 19th century is pondered.

Wild (1881) had already resolved adjustments for some of the averaging formulas used in Finland. Later on, Heino (1974) compared several formulas across the country at 12 stations where hourly observations were available. Adjustments of monthly mean temperatures depend on climate, and Heino (1994) resolved corrections separately for land, coastal and lighthouse/island stations (Table 5.4). The adjustments in Table 5.4 have been used at Finnish stations included in T_{ALL}(adj), T_8(adj), T_4(adj) and T_{CRU}(FS). In T_{ALL}(ori+), the national average includes an adjustment of -0.17ºC from 1927 to 1926 and an additional adjustment of -0.11ºC from 1901 to 1900. The influence of the lighthouse/island stations on the national average is so small that the change in 1881/1880 was ignored. The corrections for T_{ALL}(ori+) are a mean value of national correction estimates derived from three estimates: differences between T_{ALL}(adj) and T_{ALL}(ori+), differences between T_8(adj) and T_8(ori), and a rough estimate of station corrections in the LPNN grid. There are factors that cause inaccuracies to the adjustments both at station and national level. The subjectivity involved in the classification of series between coastal and inland, local climatic factors and simple area averaging methods - at least these factors produce a downgrading effect. Based on the above three estimates, the inaccuracies in the national average annual adjustment are: ±0.03ºC (1901-1926) and ±0.06ºC (prior to 1900). The 1926 inexactness results in a ±0.003ºC(10 yr)$^{-1}$ uncertainty in the T_{ALL}(ori+) linear trend given in Table 5.3.

Table 5.4. Corrections of annual mean temperatures in Finland due to averaging and instrumental inhomogeneities according to Heino (1994).

[ºC]	**Inland**	**Coastal**	**Lighthouse/island**
Prior to 1880	-0.37	-0.19	-0.07
Prior to 1900	-0.37	-0.19	-0.03
1901-1926	-0.21	-0.14	-0.03

Homogeneity testing and adjusting attempt to remove biases from the data. The capability of SNHT to improve the reliability of series depends on the density of the station network. Thus, the older parts of series are less effectively tested than the modern parts. Also, simultaneous, roughly similar-sized breaks may remain undetected by relative homogeneity testing. Besides changes in the averaging formula, changes in thermometer screening are also a potential source of systematic discontinuities. Fortunately, in Finland changes in thermometer screening and ventilation have not occurred at the same time at all stations, but gradually (Heino 1994). SNHT should therefore be able to adjust for these types of breaks, if there are any.

In addition to the already-discussed formula and screening problems and the uncertainties in their adjustments, there are uncertainties related to every individual adjustment of a homogeneity break. Firstly, the quality of the reference series varies over time and space. Generally, the pre-1940s and northern Finland are less reliable. Secondly, as discussed in paper IV, although approximated to be constant in time, the temperature adjustments in fact depend on the weather type, thus varying from year to year. Thirdly, there are more sources of inaccuracy in the observations of the pre-1930s, or thereabouts, and, sometimes, the observations are characterised by larger error margins than in modern times. For example, during the 19^{th} century, stations were inspected infrequently, thermometer screening and heights were not standardised and observation times could not easily be synchronised.

Based on Fig. 5.1 and Table 5.2, the uncertainties in the adjustments and the history of temperature measurement errors (Heino 1994 and references therein) for the Finnish annual mean temperatures have been pieced together. The criterion used is that the error limits must include all conceivable errors. The subjective, time-varying error limits for T_{ALL}(adj) are: ±0.10ºC (>1949), ±0.13ºC (1927-49), ±0.18ºC (1926-01), ±0.30ºC (<1901).

Compared to T_{ALL}(adj), T_{TYN}(FIN) is created using a sophisticated spatial interpolation. Almost all annual differences between T_{ALL}(adj) and T_{TYN}(FIN) fall within the error limits of T_{ALL}(adj) during the last 50 years (i.e., the period without bias in T_{TYN}(FIN)). The limits are thus so wide that any future estimates of national temperature using e.g. advanced spatial interpolation and more exact adjustments will very likely fall within the presented thresholds. The error limits of the most recent period are somewhat arbitrary (and large), but there is a verifiable difference in quality between the data of the early 20^{th} century compared with those of the last 40 years or so; the latter are definitely better. The larger error bars in the 19^{th} century result from the sparse network and uncertainty related to observations. The subjectively-defined error limits for annual values (plotted in Fig. 5.3) are rather over-estimated than underestimated. For example, the standard deviations of the differences between T_{ALL}(adj) and T_{ALL}(ori+), and T_{ALL}(adj) and T_{TYN}(FIN) are only 0.02ºC and 0.09ºC, respectively, during the period 1901-2000.

Table 5.5 shows the correlation coefficients of annual mean temperatures between stations in T_4(adj) from two 30-year periods, namely, 1847-1876 and 1973-2002. All inter-station correlations are higher during the modern period than during the beginning part of the series. Some of the differences in correlation coefficients are statistically significant (statistical testing as in section 4.2). There is an increasing trend in the average, running 30-year correlation between the stations (not shown). However, it is still possible, but not very likely, that the variations in the correlation coefficients are due to natural climatic variability. Based on a knowledge of measurement history, one can conclude that the lower inter-station correlation at the beginning of the series in an indication of somewhat reduced reliability.

Table 5.5. Correlation coefficients of annual mean temperatures between stations belonging to T_4(adj). Values above the grey diagonal are for the period 1973-2002 while those below the diagonal are for the period 1847-1876 (in italics). If the correlation coefficient for the period 1973-2002 is statistically significantly larger (5%-level) than the corresponding coefficient for the 1847-76 period, it is shown in bold type.

Correlation coefficient	**Helsinki**	**Kuopio**	**Kajaani**	**Oulu**
Helsinki	1.000	0.968	**0.953**	**0.961**
Kuopio	*0.928*	1.000	**0.989**	**0.980**
Kajaani	*0.895*	*0.919*	1.000	0.984
Oulu	*0.906*	*0.921*	*0.962*	1.000

T_{CRU}(FS) and T_4(adj) are compared in Figure 5.2. T_4(adj) represents a sub-domain of T_{CRU}(FS), but the temporal smoothing with a Gaussian filter (G3, see section 3.1.1) brings out the common long-term temperature variations over the region. The correlation coefficient between the two low-pass filtered (G3) series shown in Fig. 5.2 is 0.98. It is higher than the high-frequency correlation between the series, as one would expect, because the year-to-year anomalies typically have a smaller spatial scale than the lower-frequency fluctuations. It is an unavoidable problem that there are less and less series available from earlier times, and their quality is to some extent questionable. This affects both T_{CRU}(FS) and T_4(adj). However, there is remarkably good agreement between the two series regarding variability on the scale of a decade.

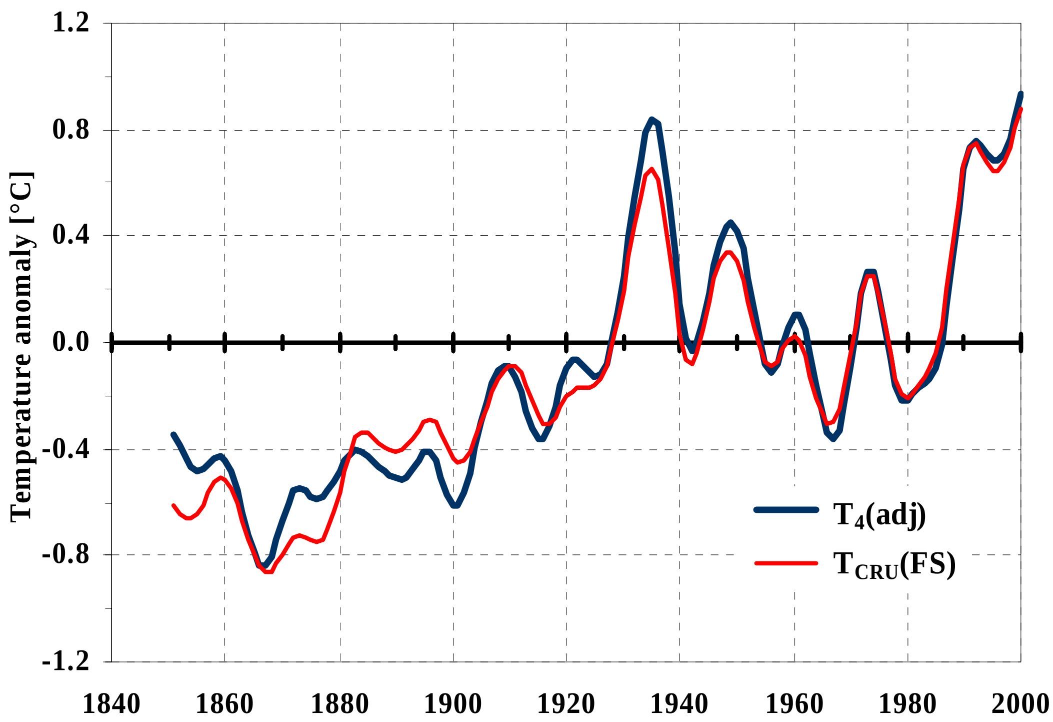

Fig. 5.2. Annual anomaly time series of T_4(adj) and T_{CRU}(FS) smoothed with G3 over the period 1851-2000 (reference period 1961-90).

5.4 Mean temperature variations and trends

T_{ALL}(adj) is the best estimate of the Finnish annual mean temperature series. The standard deviation of T_{ALL}(adj) rounds to 1.11ºC both for the whole series and for the 1961-90 period. The warmest year in the series is 1938 (Table 5.6). The next warmest years are from the most recent warm period. The top ten of warm years seem to be a random mixture of years since 1920. Even if one uses the T_4(adj) series (1847-2002) the same top ten years come up, but in a slightly different order. The coldest year in T_{ALL}(adj) is 1888, but 1915 was almost equally cold. In the top ten cold years of T_{ALL}(adj) only 1985 and 1987 are from the more recent half of the record. However, if T_4(adj) is examined (Fig. 5.4), 1867 stands out with such a margin that there is no doubt about the coldest year. The use of T_4(adj) changes the top ten cold years; seven years come from the 19th century with the most recent year in the top ten cold years being 1941.

Table 5.6. The coldest and warmest years in T_{ALL}(adj), 1888-2002, and the coldest year in T_4(adj) (in italics), 1847-2002 (Tm = annual mean temperature). Temperatures are anomalies from the period 1961-90.

[ºC]	Coldest years		Warmest years	
	Year	**Tm**	**Year**	**Tm**
	1867	*-3.4*		
1st	1888	-2.68	1938	2.36
2nd	1915	-2.60	1989	2.24
3rd	1902	-2.48	2000	2.17

Fig. 5.3 shows that there are large year-to-year variations. The G3 smoothed curve in Fig. 5.3 brings out fluctuations on the scale of a decade. The 1930s show up as the warmest 10-year period in the record, but the most recent decade has only been slightly cooler. The first decades of the series are the coolest. The G9 filtered curve simplifies the series into three segments, first a warming until 1940 or so, followed by a slight cooling until around 1970 and finally a warming trend. Folland et al. (2001) have determined the change points of the global temperature series. The same time periods also describe reasonably well the temperature trends in Finland. The annual mean temperature trends of T_{ALL}(adj) are compared with Northern Hemisphere (NH) and Global trends in Table 5.7.

The temperature trend over the 20th century in Finland is similar to the NH and Global warming trend (0.4-0.8ºC$(10\ yr)^{-1}$) given by Folland et al. (2001). The decade-scale fluctuations are large in Finland. Thus, the first warming period and especially the recent temperature increase are much steeper in Finland than globally. Comparison of the last two columns in Table 5.7 reveals that linear trends in Finland over a couple of decades are quite sensitive. Two additional years alter the trend significantly. Table 5.7 downplays the earlier warming trend in Finland, as one can judge from Fig. 5.3. With more optimal fitting at the beginning of the series, the early warming trend is of about the same magnitude as the recent warming.

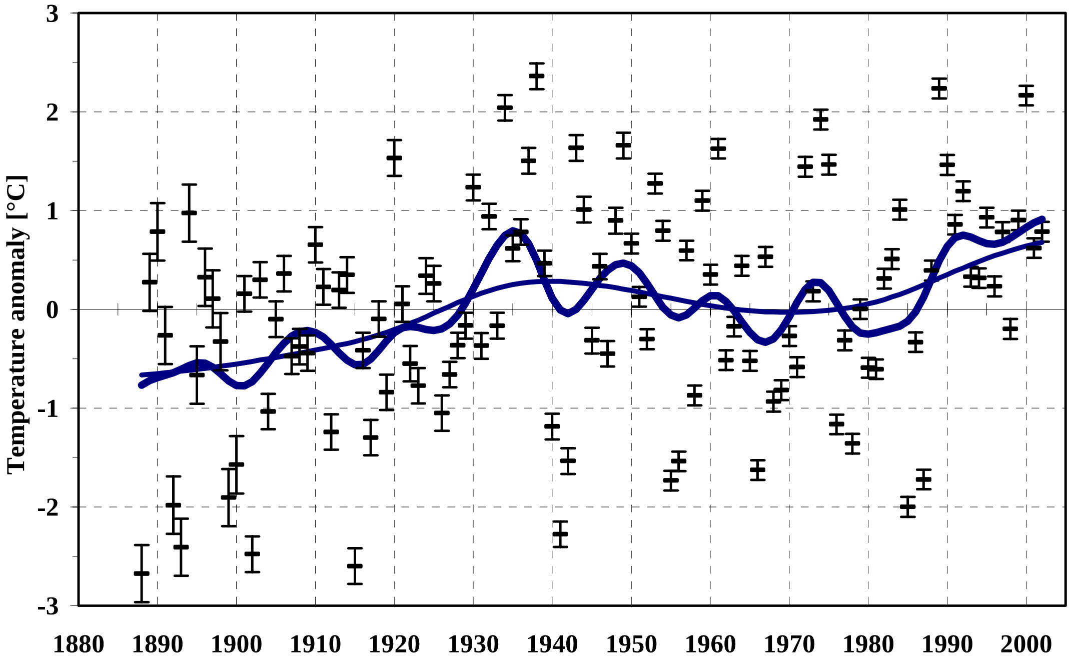

Fig. 5.3. Annual anomaly time series of T_{ALL}(adj), 1888-2002 (reference period 1961-90). Annual values have error bars marked with thin lines (see section 5.*3). The thick curve is smoothed with the G3 and the thin curve with the G9 filter.*

Concerning the statistical significance of the trends of annual mean temperature, two approaches are used. The first approach informs one if there is a trend in the observations that is statistically significant in the sense that it cannot be explained by random noise with no interannual autocorrelation. In the second approach, assuming that the coupled atmosphere-ocean general circulation models (AOGCMs) can describe the natural, internal variability of the climate system, one can evaluate whether the observed trends could occur by chance due to long-term climate variability. A similar approach is also used in Jylhä et al. (2004).

Table 5.7. Linear trends in annual mean temperatures. The linear trend in T_{ALL}(adj) is determined using Sen's method and the least squares method. Statistically significant trends (at the 5%-level) are indicated in bold type. Northern Hemisphere and Global trends are from the IPCC TAR (see Folland et al. (2001) for description of trend calculation). See also the text for the test results of the coupled atmosphere-ocean general circulation model unforced control simulations. Units: ºC(10 yr)$^{-1}$.

[ºC(10 yr)$^{-1}$]	**1901-2000**	**1910-45**	**1946-75**	**1976-2000**	**1976-2002**
T_{ALL}(adj)					
Sen Trend	**0.08**	0.29	-0.06	**0.81**	**0.70**
Least squares	**0.08**	0.24	0.00	**0.81**	**0.72**
Northern Hemisphere	**0.06**	**0.17**	-0.05	**0.24**	
Global	**0.06**	**0.14**	-0.01	**0.17**	

Firstly, linear trends are calculated with the non-parametric Sen's trend estimate and the least squares method (Gilbert 1987), and their statistical significances at the 95%-level are tested with suitable tests (the Mann-Kendall test and the *t*-test, respectively). In this case both tests give practically identical results (Table 5.7). Secondly, two 1000-year coupled atmosphere-ocean general circulation model (AOGCM) control simulations are utilised to describe unforced internal variability. The HadCM3 climate model is described in Gordon et al. (2000) and the CGCM2 in Flato and Boer (2001). From the 1000-year simulations with the two AOGCMs, linear trends of area-averages over the land grid boxes representing Finland are determined with the least squares method. A Gaussian distribution is fitted to the histogram of trend values. In the millennial control runs of HadCM3 (CGCM2) about 80% (88%) of hundred year trends are smaller in magnitude than ±0.08ºC(10 yr)$^{-1}$. It turns out that 94% (96%) of 27-year trends (compare with the period 1976-2002 in Table 5.7) in the HadCM3 (CGCM2) simulation are smaller than ±0.72ºC(10 yr)$^{-1}$.

Räisänen and Alexandersson (2003) compared the observed mean temperature of Sweden with the corresponding temperatures from the simulations of 19 AOGCMs (CMIP2 project, Meehl et al. 2000). They found a low probability (about 6%) for the observed annual mean temperature increase (+0.8ºC) from 1961-1990 to 1991-2000 to occur by chance (natural variability). They estimate that the anthropogenic forcing raised the probability of the observed temperature changes to occur to 23%.

The validation of model-produced internal climate variability at time scales of several decades or more is difficult because instrumental data sets are too short. Collins et al. (2001) have analysed the temperature variability of a HadCM3 millennial control run and find it to be reasonably realistic on time scales shorter than those examined in this study. According to Jylhä et al. (2004), the year-to-year temperature variability in HadCM3 and CGCM2 is reasonable when compared with observations in Finland.

There is a statistically significant trend in the annual mean temperatures in Finland during the 20th century, but in the light of the natural, internal climate variability simulated by AOGCMs, this is not uncommon. The recent (1976-2002) quite large increasing trend in Finnish annual mean temperatures is more unusual in long-term unforced AOGCM simulations than the warming over the whole 20th century.

The longest time series of Finnish mean temperature, T_4(adj), presented in this study is 156 years long. Although it does not contain any stations from northern Finland, it correlates fairly well with the more exact estimates on an annual level. At the seasonal level, there are error margins of a few tenths of a degree or so, related to single seasonal values during the 19th century. However, at least the variability on time scales of a decade or longer is believed to be represented realistically. Fig 5.4 displays annual and seasonal anomaly series for the reference period 1961-90. The variability is at its largest during the winter; G3 smoothed DJF fluctuations can therefore also be identified from the corresponding annual curve. The G9 curves give an impression of trends in the series, though the last 15 years or so may change when new years are added. The smoothed curves in Fig. 5.4 resemble the high-quality temperature reconstructions from Sweden (Moberg and Alexandersson 1997) and Norway (Hanssen-Bauer and Nordli 1998).

The most recent 40 years of T_4(adj) are selected to represent a period of increased atmospheric concentration of greenhouse gases and aerosols having an influence on climate (Mitchell et al. 2001). The first 40 years of the series are selected as a "baseline" period, i.e., a period of minimal anthropogenic influence on global climate. Other radiative forcings (solar, volcanic) have also been in operation, as well as the long-term natural variation of the climate system. The observed changes in T_4(adj) thus cannot be attributed to any specific forcing, but they may give guidance on "typical" changes. Table 5.8 complements the visual interpretation.

The statistical significance of the changes in the mean and standard deviation were evaluated with the *t*-test and F-test, respectively (Vining 1998). Heino (1994) has extensively studied the correlation of consecutive seasons, named "autocorrelation" in Table 5.8. The statistical significance of the differences between the correlation coefficients is tested as described in section 4.2. The degree of asymmetry and peakedness of the distribution were described with the coefficients of skewness, C_s, and kurtosis, C_k, introduced in section 3.3.

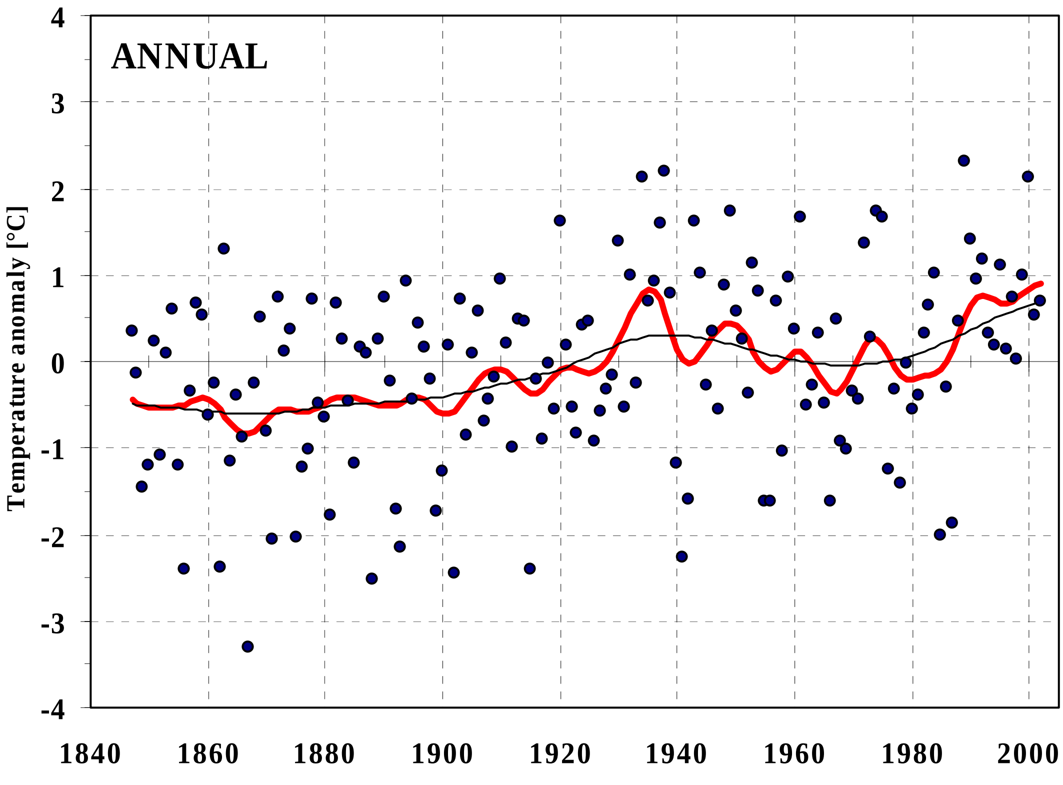

Fig. 5.4. Annual and seasonal anomaly times series of T_4(adj), 1847-2002 (reference period 1961-90). The annual values are marked with black dots. The thick curve shows smoothing with the G3 and the thin curve with the G9 filter. Note the different vertical scales of the DJF and MAM graphs. *(Continued on following page)*

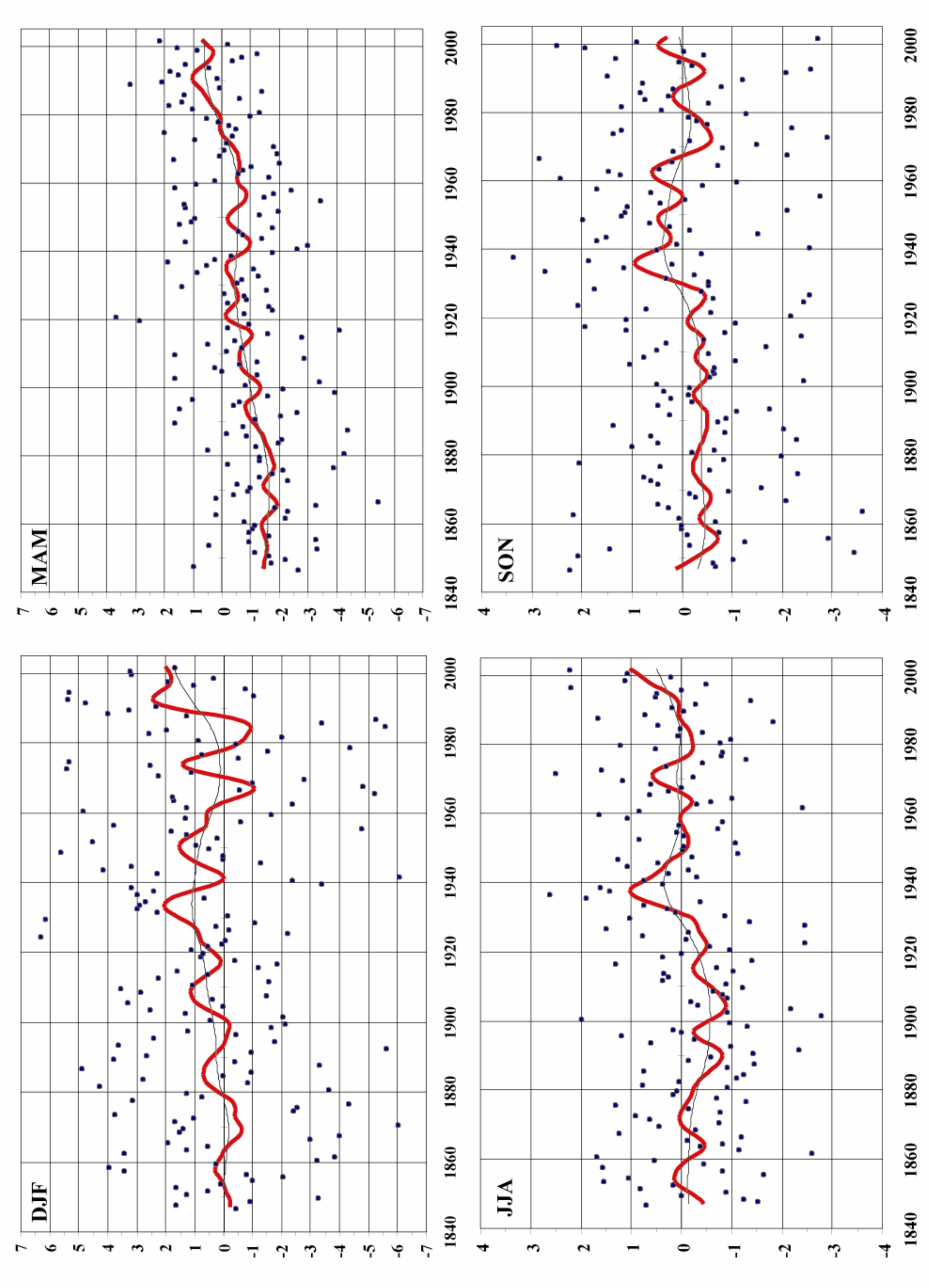

Fig. 5.4. (Continued)

The annual temperature increase from the period 1847-1886 to 1963-2002 is statistically significant. All seasons have warmed, but the main contribution comes from MAM, in which warming is significant at the 0.1%-level. The standard deviation has increased in DJF, but the change is not statistically significant. During the period

1847-86 the annual as well all seasonal distributions were negatively skewed, i.e., the distributions have a tail towards the negative anomalies (see section 3.3). There has been a shift towards a more positive C_s, except in DJF. During the past 40 years, the MAM and JJA temperatures have been slightly positively skewed. Other studies have also related warming to a loss of the coldest part of the distribution (e.g. Rebetez 2001, Heino 1994).

All the distributions in Table 5.8 are flat-topped (platykurtic) during 1847-86, except the MAM temperatures, which have a narrow central maximum (leptokurtic). The MAM distribution has undergone a nearly statistically significant change to flat-topped, while other seasons show no clear tendency in changes of C_k. The correlation between DJF and the following MAM is statistically significantly different from zero (the 5%-level is 0.312). A similar positive correlation exists between MAM and JJA in 1847-86, but not in the latter period. However, this, and the even larger correlation coefficient change in SON/DJF, are not statistically significant. Heino (1994) already noted significant correlation between winter and spring temperatures in Finland, and attributed it to a high frequency of blocking events in the atmosphere and the thermal heat capacity of the surrounding sea areas. Snow cover can also act as a "messenger" from winter to spring through albedo feedback. In southern Finland, cold winters may produce a long-lasting snow cover that, before its disappearance during the spring, reflects the sun's radiation, while after warm winters the snow cover may melt quickly, revealing ground with a low albedo. In northern Finland the amount of snow is more related to wintertime precipitation than to wintertime temperature.

Table 5.8. Comparison of the time periods 1847-1886 and 1963-2002 in the T_4(adj) anomaly series (reference period is 1961-90). Autocorrelation means correlation between consecutive seasons, e.g. column DJF contains the correlation between the DJF and the following MAM temperature anomalies. If the statistics of the time periods are significantly different (5%-level), they are in bold type. See section 3.3. for the description of testing the statistical significance of C_s and C_k.

	ANNUAL		**DJF**		**MAM**		**JJA**		**SON**	
	1847-1886	**1963-2002**	**1847-1886**	**1963-2002**	**1847-1886**	**1963-2002**	**1847-1886**	**1963-2002**	**1847-1886**	**1963-2002**
Temperature mean (reference 1961-90) [ºC]	**-0.54**	**0.20**	-0.08	0.52	**-1.61**	**0.21**	-0.16	0.18	-0.37	-0.09
Standard deviation [ºC]	1.04	1.05	2.59	3.12	1.33	1.28	1.02	1.01	1.43	1.40
C_s (skewness, eq. 3.8)	-0.56	-0.16	-0.31	-0.34	**-0.62**	**0.18**	-0.05	0.36	-0.30	-0.19
C_k (kurtosis, eq. 3.9)	-0.24	-0.48	-0.76	-0.74	0.57	-0.82	-0.73	-0.28	-0.17	-0.48
Autocorrelation			0.39	0.46	0.34	0.12	-0.01	-0.01	0.28	-0.14

The springtime warming is accompanied by a relative reduction of extremely cold MAM temperatures and a flattening of the distribution. As a results of these counteracting changes there is no change in the standard deviation. There are no drastic changes in other seasons, yet, there is a consistent temperature increase in all seasons. Table 5.8 draws attention to the non-gaussianity and non-stationarity of temperature distributions that may be of importance in applications of climatic "normals", for example.

For a second time millennial control simulations of AOGCMs are used to give an estimate of the long-term, internal climate variability. The focus is on the annual and MAM averages, because they are the only values that differ clearly (over the 1%-level) between the two time periods examined (Table 5.8). The standard deviation of 40-year temperature averages, STD(T), are calculated from the control runs with HadCM3 and CGCM2. Assuming a normal distribution, for the temperature difference between any 40-year periods to be statistically significant at the 5%-level, it must exceed the confidence interval $\pm\sqrt{2}\cdot 1.96\cdot \mathrm{STD}(T)$. The observed MAM temperature increase of 1.82°C is well outside the 95% confidence limits in both the HadCM3 and CGCM2 simulations. However, the annual mean temperature increase of 0.74°C is within the confidence limits of HadCM3 but outside those of CGCM2.

As stated in paper II and many other investigations of northern Europe (e.g. Jacobeit et al. 2001, Chen and Hellström 1999, Post and Tuulik 1999) and larger geographical areas (Hurrel et al. 2003, Thompson and Wallace 1998), westerly winds bringing moist air from the Atlantic typically result in mild wintertime and cool summertime temperatures, while continental air flowing from the east has opposite effects. The frequent westerly winds in the 1990s have contributed to the mild winter months in Finland. However, the warmth of the 1930s took place during a period of normal or below-normal wintertime zonal circulation over Fennoscandia (paper II).

6. OBSERVED ANNUAL PRECIPITATION CHANGES IN FINLAND

As shown in the previous section, FDM is well-suited for the building of reliable temperature averages over Finland. The temperature series based on adjusted and original data, T_{ALL}(adj) and T_{ALL}(ori+), are very similar. FDM is attractive because a large number of stations is needed for reliable calculation of area averages due to the large spatial and temporal variability of precipitation. The Finnish annual mean precipitation series, R_{ALL}(ori+), is therefore constructed from all available precipitation measurements using FDM for the period 1894-2002, i.e. replacing temperature with precipitation in (6.1). Changes in precipitation gauge design are adjusted, but SNHT has not been applied to detect and adjust single station series.

In this study, differencing is used in FDM for precipitation, too. Peterson et al. (1998) suggest that ratios might work better for precipitation than differences. However, first difference series are additive, whereas series of changes composed from ratios are multiplicative, making a ratio series quite sensitive to errors. Also, ratio series may not contain zero values (not a problem for annual sums in Finland). It turns out that there are also other severe limitations to the accuracy of the national precipitation estimate than the way FDM is applied. For example, the use of a $1^o x 1^o$ latitude-longitude grid and simple spatial methods are limitations to the accuracy. Moreover, there are gradients in the annual mean precipitation and its variability that are captured with varying accuracy due to the changing observing network.

6.1 Description of precipitation series

To evaluate the uncertainty in the estimates of mean annual precipitation over Finland, several time series have been constructed (Table 6.1). R_{ALL}(ori+) is based on all the available data. To evaluate sampling error caused by a sparse station network, two additional versions of R_{ALL}(ori+) have been created for the period 1970-2000 using stations mimicking the station network in 1894 and 1909, named R_{1894}(ori+) and R_{1909}(ori+). Fig. 2.3 shows the location of stations in 1894 and 1909. R_{1894}(ori+) consists of 34 stations throughout the series, i.e., missing years or stations closing down are replaced with data from a neighbouring station. R_{1909}(ori+) starts with 121 stations just as the network was in 1909, but no attempt is made to keep the network constant, and the number of stations drops to 84 by the end of the time period.

The main advantage of R_{24}(adj), presented in paper I, is that it is based on homogeneity-tested and adjusted data from 24 fixed stations. On the other hand, R_{24}(adj) is not very reliable in describing year-to-year changes in nation-wide precipitation due to its southern bias and the sparseness of the network. A version based on original station data is compiled for comparison (Table 6.1).

Reuna and Aitamurto (1994) have published precipitation values for a large number of Finnish drainage basins. Their data cover about 75% of Finland during the period 1921-81. The analysis performed at the Hydrological Office is mostly based on the same precipitation measurements used for R_{ALL}(ori+), except that the period 1941-50 is based on a somewhat smaller number of stations. The aim of the analysis has been to produce areal precipitation over drainage basins for hydrological purposes. According to Hyvärinen (1997), areal precipitation has been calculated with Thiessen polygons (1911-45), manual analysis of isohyets (1946-81) and computer-based gridding (1981-). The areal coverage of R_{HYD}(ori) is increased to about 90% by creating time series for additional drainage basins simply by averaging neighbouring drainage basin precipitation. Only some coastal areas are excluded from R_{HYD}(ori) (Table 6.1).

R_{TYN}(FIN) is extracted from Mitchell et al. (2002) who have published precipitation data for various countries. The fine spatial resolution data is based on work by New et al. (2000), who used precipitation data from Hulme (1994) and Dai et al. (1997b) that have not been adjusted for homogeneity breaks.

Knowing that measured precipitation is an underestimation of "true ground" precipitation, a third, corrected version, R_{cor}, based on R_{ALL}(ori+) has been created. R_{cor} is used to demonstrate the effects of climatic and non-climatic factors on the difference between "true ground" and measured precipitation. In cold and/or windy climates, this difference is mainly due to aerodynamical effects near the rim of the gauges, but is also due to evaporation from the gauge, adhesion of the water to the interior of the gauge, blowing/drifting snow and possibly due to measurement errors. Fundamentally, we are interested in the "true" precipitation. However, researchers must make do with a network of measured precipitation because there is not enough metadata and comprehensive weather observations to create reliable, long-term series of corrected precipitation. Recently, there have been initiatives to correct precipitation measurements operationally (Førland et al. 1996a).

Table 6.1. Annual precipitation series used in this study.

Symbol	**Time period**	**Stations/Area**	**Homogenisation**	**Area averaging**	**Remarks /*references***
R_{ALL}(ori+)	1894-2002	Ranging from 32 to 624 stations in Finland and neighbouring areas	Original station data plus adjustments for gauge changes	Area-averaged difference series in the LPNN grid	*Section 2*
R_{1894}(ori+)	1970-2000	34 stations as in 1894	As R_{ALL}(ori+)	As R_{ALL}(ori+)	*Sections 2 and* 6
R_{1909}(ori+)	1970-2000	Ranging from 121 to 84 stations	As R_{ALL}(ori+)	As R_{ALL}(ori+)	*Sections 2 and* 6
R_{24}(adj)	1910-1995	24 stations in Finland and neighbouring areas	Stations tested and adjusted with SNHT plus adjustments for gauge changes	Simple average of station values	*Paper I*
R_{24}(ori)	1910-1995	As R_{24}(adj)	Original station data	As R_{24}(adj)	*Paper I*
R_{HYD}(ori)	1921-1981	Areal precipitation for 18 large drainage basins	Original station data	Area-averaged precipitation covering about 90% of Finland	*Reuna and Aitamurto (1994)*
R_{TYN}(FIN)	1901-1998	Covers Finland	Original station data	Area-averaged anomaly series in 0.5ºx0.5º latitude-longitude grid	Sophisticated spatial interpolation *Mitchell et al. (2002)*
R_{cor}	1911-1981	As R_{ALL}(ori+)	Original station data plus correction for gauge undercatch at national level	Area-averaged difference series in the LPNN grid	*Section* 6, *Solantie and Junila (1995)*

In the climatological examination, complex correction formulas can be merged into one relatively simple formula that is a function of the type of precipitation gauge, type of precipitation, wind speed and sheltering of the gauge. Following the notation of Solantie and Junila (1995), the total correction factor (including error due to wind, evaporation and adhesion), $k_c^{'}$, for the measured precipitation is

$$k_c^{'} = \frac{1}{\overline{v} \cdot \overline{\alpha}} \, v \cdot \alpha \cdot k_c^{'}(\overline{p}) \qquad (6.1)$$

where α is the exposure factor, v is the wind speed and $k_c^{'}(\overline{p})$ is the correction factor as a function of $\overline{p}$, the proportion of solid precipitation. The overbars denote average conditions over continental Finland, i.e. $\overline{v}$ is 3.57 ms^{-1} and $\overline{\alpha}$ is 35% (Solantie and Junila 1995). Korhonen (1944) originally introduced the exposure factor, α. It is defined in such a way that the aerodynamic error is a linear function of α. The exposure factor can be determined as a function of the vertical angle subtended by obstacles around the gauge, η. A protected site has α=0% (η>18°) and at an exposed site α=100% (η<6°) (Solantie and Junila 1994). The values of $k_c^{'}(\overline{p})$ can be read for both the Wild and Tretyakov gauges used in Finland from graphs published in Solantie and Junila (1995). Subsequently, the corrected precipitation, R_c, is

$$R_c = k_c^{'} \cdot R_m \qquad (6.2)$$

where R_m is the measured precipitation. In sections 6.2 and 6.3 several variants of R_{cor} are produced using (6.1) and (6.2).

6.2 Determination of adjustments for precipitation gauge type changes

Heino (1994) gives a detailed description of the types of precipitation gauges that have been used in Finland. Until 1909 the so-called "Finnish gauge" was primarily used. The Wild gauge was in operation from 1909 to 1981/1982, excluding stations operating only during summertime (not utilised in this study). Since 1981/1982 the Tretjakov gauge has been the official precipitation measurement instrument at FMI. In 1992 a new measuring vessel was brought into operation, but this caused insignificant effect on the measured precipitation (T. Sankola, private communication, 2003).

In connection with instrument changes, there have been comparison measurements between the Finnish and Wild gauges, described by Korhonen (1913) and between the Wild and Tretjakov gauges described by Solantie and Junila (1995). All the above mentioned authors, as well as Heino (1994) and Tammelin (1984), have developed adjustments between gauge types and/or corrections to true precipitation. R_{24}(adj) uses the climatological adjustments (constant ratios for monthly sums) developed by Heino (1994) to convert precipitation totals measured with Wild gauges to corresponding totals that would have been measured with Tretjakovs. R_{24}(ori) and R_{TYN}(FIN) combine the measurements made using both Wild and Tretjakov gauges without any adjustments.

R_{ALL}(ori+), and all series derived from it, combine year-to-year changes measured with different gauge types. Thus, the problem of a change from one gauge type to another reduces to finding the adjustment(s) for the year(s) when the change(s) took place. Consequently, comparison measurements performed at selected stations around the country in those years when instrument changes happened can be used to calculate the year-to-year change with an unchanged instrument. Another estimate of the same year-to-year changes can be calculated with the help of correction formulas. Both approaches are used to define adjustments and their error limits for R_{ALL}(ori+).

Korhonen (1913) gives annual precipitation as measured with both the Wild and Finnish gauges in 1909. The mean difference of ten stations is 43 mm, i.e., 8%. Some summer months are missing, for which reason the percentage value gives an overestimate and the absolute value an underestimate. It is difficult to judge how representative the ten sites are (Fig. 6.1). For example, there is only one station from the northern half of Finland and some sites are quite exposed. Later on, Korhonen (1921) derived adjustments for the whole station network. These adjustments were applied and the precipitation measured in 1908 with Finnish gauges was adjusted to represent Wild sums. In this way another estimate, i.e., 32 mm, is calculated for the adjustment between 1909 and 1908. The mean value of these two estimates, 38 mm, has been used in R_{ALL}(ori+). In connection with the gauge type change, the number of stations almost tripled after a nearly stagnant period. Thus, besides the different instruments, 1909 is a milestone between two spatially quite different networks. For this reason, large error limits for the gauge type adjustment, namely ±12 mm, are used in further studies.

According to T. Sankola (personal communication, 2002), the switch from Wild gauges to Tretjakov gauges was accomplished mainly during the extended summer half (May-September) of the years 1981 and 1982. Half of the stations were dealt with during 1981, and this was done in a spatially random order, as was also the substitution in 1982. In order to determine the difference between gauges in windy conditions, most of the stations running comparison measurements were exposed sites. An upper limit for the adjustment can be calculated from parallel measurements at eight relatively open sites (including five airports) with complete records (Fig. 6.1). The average difference between the two gauges over the whole of Finland is calculated as a simple monthly average of the eight stations. To obtain the difference in annual sums, one has to weight the monthly differences with the proportion of gauges that had already been changed, e.g., the weight from October 1981 to May 1982 is 0.5 because half of the gauges were substituted during the summer of 1981. Thus, by selecting 31.12.1982 as the turning point from Wild to Tretjakov, one gets following estimates for the adjustments: 9 mm (Tretjakov) from 1983 to 1982, 24 mm from 1982 to 1981 (Wild), and 11 mm from 1981 to 1980 (Wild). These values supposedly give upper limits for the adjustments, because they represent open sites that have an exposure factor much above the national average (Solantie and Junila 1995).

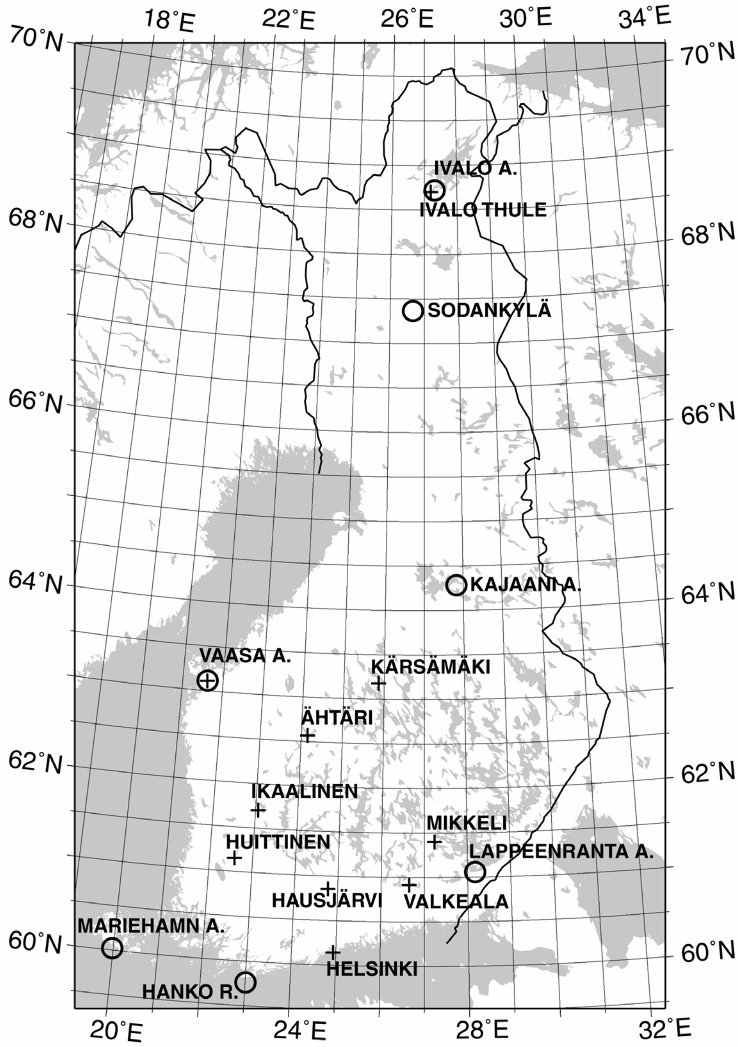

Fig. 6.1. Stations running comparison measurements in connection with precipitation gauge type changes. Crosses denote sites for Finnish and Wild gauge data in 1909 (Korhonen 1914). Circles denote sites for Wild and Tretjakov gauge data in 1981 and 1982 (Solantie and Junila 1995). The letter A refers to an airport. The station of Hanko is located on the island of Russarö. At Vaasa, the measurements in 1909 were not made at the airport, but at a more sheltered site.

A lower limit for the gauge type adjustments is obtained with the help of correction factors, measured precipitation and suitable assumptions. The measured precipitation sum, R_m, is

$$R_m = aR_W + (1-a)R_T \tag{6.3}$$

where R_W is the precipitation amount that would have been measured with Wild gauges, R_T the amount that would have been measured with Tretjakov gauges and *a* is the proportion of Wild gauges in the network. With equations (6.1-6.3), we can derive

$$R_W = R_m \frac{1+k'_T}{1+(1-a)k'_W + ak'_T} \qquad (6.4)$$

where k'_W and k'_T are the correction factors for precipitation measured with the Wild and Tretjakov gauges, respectively. During the period from November 1981 to April 1982 *a* is 0.5. Taking this into account, an equation for R_T corresponding to (6.4) becomes

$$R_T = R_m \frac{1+k'_W}{1+0.5k'_W + 0.5k'_T} \qquad (6.5)$$

Now we can calculate what would have been measured in 1981, if the network had remained instrumented with Wild gauges only. All stations are used to compute monthly precipitation sums in the LPNN grid in November and December 1981. The other calculations are also done in the LPNN grid. For k'_W and k'_T the percentage of solid precipitation, $\overline{p}$, is estimated in the LPNN grid based on 20 representative stations across the country. Making use of Solantie and Junila (1995), $\overline{p}$ is needed to define $k'_c(\overline{p})$ for both the Wild and the Tretjakov gauges. The diagrams for mainly protected stations on pages 50 and 54 of Solantie and Junila (1995) are used, i.e. α=25% (the average exposure factor for continental Finland is 35%). The measured Finnish annual mean precipitation in 1981 is 6 mm larger than the result of the calculation (using equation 6.4) described above. This is the lower limit for the adjustment of R_{ALL}(ori+) from 1980 to 1981 because of the two assumptions made. Firstly, a correction is used that assumes an exposure factor below the national average. Secondly, the minor influence of the Tretjakov gauges during the period May-October 1981 is neglected.

The adjustment of R_{ALL}(ori+) from 1981 to 1982 is estimated by calculating with (6.1), (6.2) and (6.4) what would have been measured in 1982, if the whole network had been instrumented with Wild gauges. Again, $\overline{p}$ is determined from observations. It is assumed that stations were better sheltered than they were in reality, thus, α=25% is used when calculating R_W for the months from January to April and from November to December. Simply neglecting the difference between R_W and R_T during May-October 1982 would give an unnecessarily small lower limit because on average about 75% gauges are already of the Tretjakov type (during summer 1981 the corresponding percentage was about 25%). Equation (6.1) is not well suited for liquid precipitation. Instead, parallel measurements at the stations (Fig. 6.1) are used to derive an estimate of the difference. After taking into account that about 75% of the gauges had been changed, in order to get a lower limit an additional factor 0.7 is used to lower the difference. Finally, one arrives at the result that the measured Finnish annual mean precipitation in 1982 is 17 mm larger than the result of the calculation with the low limit assumptions.

In the adjustment between 1983 and 1982, one needs to evaluate what would have been the precipitation in 1982, if only Tretjakov gauges had been in use. Equation (6.5) together with the assumption of a below-average exposure factor can be used from January to April. The difference during the liquid precipitation period (May-October) is taken from the simultaneous Wild and Tretjakov measurements. The calculated low limit estimate is 6 mm.

It is quite certain that the simultaneous measurements at open sites give higher adjustments than is needed on average for the whole Finnish network. The upper limits for the Wild-Tretjakov switch can be called "absolute", with the reservation that, because there are only 8 stations, random factors resulting from natural climate variability may cause these sites to be an unrepresentative sample of the national precipitation. An absolute low limit would have resulted from assuming that the Wild and Tretjakov gauges have identical wind, wetting and evaporation errors, and thus, measure about equal amounts of precipitation. This assumption would be worth considering at totally-sheltered sites, as shown by Solantie and Junila (1995), who studied four sites with α=0%. Despite assumptions to minimise adjustments, the calculated low limits may not be that far from the "true" adjustments. For a more accurate calculation of corrections one needs simultaneous observations on the wind and precipitation type at the stations, as well as exposure factors for all directions and the exact dates of gauge changes for all stations. The calculations become too laborious for the purpose of this climatological study. Instead, the "best guess" was chosen to be a combination of the low and high limits with weights of two and one, respectively. Table 6.2 summarises the annual adjustments and their "error limits" defined by quite different, partly subjective, methods.

Table 6.2. Adjustments used in R_{ALL}(ori+), R_{1909}(ori+) and R_{1894}(ori+), denoted "best guess". Also shown are the low and high limits for adjustments that are used in section 6.4.

Adjustment [mm]	**1908/1909**	**1980/1981**	**1981/1982**	**1982/1983**
Low **- "Best guess" -** High	26 - **38** - 50	6 - **7** - 11	16 - **19** - 24	6 - **7** - 9

6.3 Comparison of precipitation series

As an assumption, R_{ALL}(ori+) and R_{24}(adj) should give reliable estimates of annual Finnish mean precipitation. R_{ALL}(ori+) is selected as the baseline. In Fig. 6.2, R_{ALL}(ori+) is set to the same level as R_{24}(adj), in such a way that the mean values for the period 1961-90 are equal. It is easy to see that the two series have a fairly similar long-term behaviour, suggesting that the use of FDM has not introduced any major systematic bias into R_{ALL}(ori+). However, the largest annual differences are about ±40 mm. R_{24}(ori) and R_{24}(adj) mostly coincide during 1982-95. Because R_{24}(ori) simply joins the Wild and Tretjakov measurement series, differences with R_{ALL}(ori+) are negative, indicating that R_{24}(ori) underestimates precipitation during the period of (unadjusted) Wild measurements. Both R_{HYD}(ori) and R_{TYN}(FIN) are at a different level to R_{24}(adj), because R_{24}(adj) is biased towards southern Finland, giving a higher mean level. Before the 1960s, R_{HYD}(ori) has a clear trend relative to R_{ALL}(ori+) and R_{24}(adj). Visually the most striking feature of R_{TYN}(FIN) is the discontinuity in 1981 relative to R_{ALL}(ori+) and R_{24}(adj). R_{TYN}(FIN) seems to be based on the original data without adjustments for gauge type change. The estimated errors in the adjustments of

R_{ALL}(ori+) for gauge type changes are smaller than the deviations of other precipitation series from R_{ALL}(ori+).

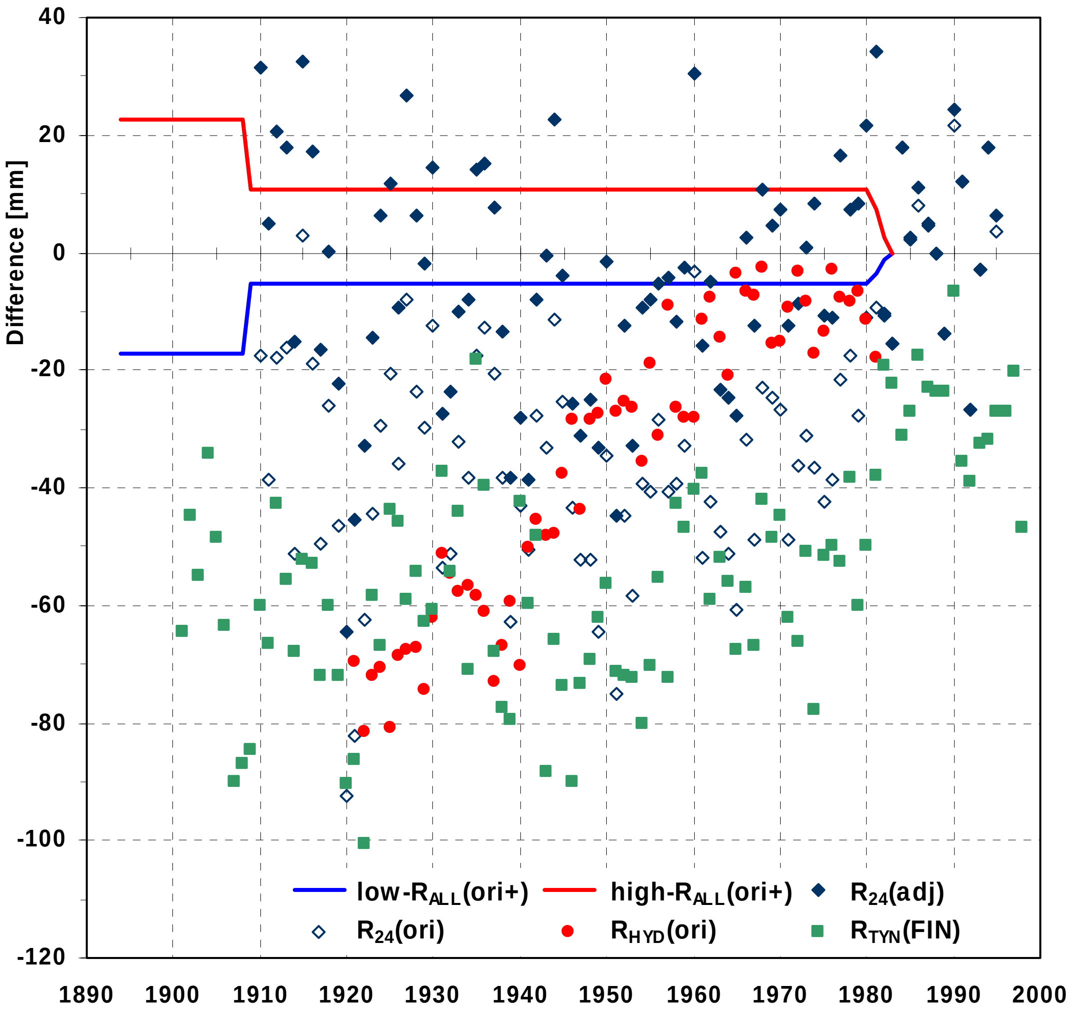

Fig. 6.2. Comparison of annual precipitation totals, 1894-1998. Symbols mark differences between the other series and R_{ALL}(ori+): high-R_{ALL}(ori+) and low-R_{ALL}(ori+) refer to Table 6.2; for the other series, see Table 6.1.

Because R_{HYD}(ori) does not cover the standard normal period 1961-90, the period 1921-81 is also examined in Table 6.3. The period 1921-81 is homogeneous in terms of instrument type and the station network is tolerable. The mean precipitation values for both time periods are about equal in R_{ALL}(ori+), except if the results of comparison measurements from open sites are used as adjustments. The differences in the mean value for the period 1961-1990 are related to a different spatial interpolation and to the use of different stations. In R_{24}(adj) the mean values of the two time periods are about equal, while R_{24}(ori) and R_{TYN}(FIN) show some increase from 1921-81 to 1961-90. The standard deviations of all series are quite similar but differ between the two time periods. The standard deviation can be used as measure of variability because the mean values are fairly close to each other. The commonly-used coefficient of variation is about 11-12% for the national yearly average series, whereas Heino (1994) reports values from 13% to 20% for single stations in Finland.

The mean bias is quite large in R_{24}(ori), R_{HYD}(ori) and R_{TYN}(FIN), but this results from the subjective choice of selecting the mean level of R_{ALL}(ori+) for the period 1961-90 as that for R_{24}(adj). The large bias influences the RMS differences, too. For the period 1921-81 the RMS differences are calculated so that the mean values are set equal (values in brackets). If the mean bias is removed, the RMS differences of R_{24}(ori), R_{HYD}(ori) and R_{TYN}(FIN) are of about equal size. This is a hypothetical arrangement, because in reality R_{ALL}(ori+) cannot simultaneously be at the same level as all three series.

The correlation coefficients in Table 6.3 have been calculated from the G3 filtered series. The beginning (end) of filtered series are influenced by the preceding (following) values; therefore, four years are cut off from both ends before the calculations. The variability on a time-scale of a decade or longer of R_{ALL}(ori+) and R_{24}(adj) shows high correlation. The correlation coefficients of R_{24}(ori) and R_{TYN}(FIN) are somewhat lower. R_{HYD}(ori) correlates weakly, indicating quite a different low-frequency behaviour, but the high-frequency behaviour is very similar to that of R_{ALL}(ori+) (see also Fig. 3.1). The correlation coefficients of the high-frequency component (see section 3.1) are very high between R_{ALL}(ori+) and all the other series.

Table 6.3. Comparison of other annual precipitation series (Tables 6.1 and 6.2) with R_{ALL}(ori+) during the periods 1961-90 and 1921-81 (separated by "/" in the table). The mean, standard deviation (STD), mean bias and root mean squared difference with R_{ALL}(ori+) are shown as well as the correlation coefficients. The RMS differences in brackets are calculated after setting the mean bias to zero. Correlation coefficients are given for the G3 filtered series (low pass) and the differences between the original and the G3 filtered series (high pass) (separated by "/" in the table). Correlation coefficients are calculated for the period 1914-91, except those between R_{HYD}(ori) and R_{ALL}(ori+), that are for the period 1925-77 (in italics).

Period 1961-90/1921-81	**R_{ALL}(ori+)**	**Low R_{ALL}(ori+)**	**High R_{ALL}(ori+)**	**R_{24}(adj)**	**R_{24}(ori)**	**R_{HYD}(ori)**	**R_{TYN}(FIN)**
Mean [mm]	580/582	577/577	588/593	580/575	557/545	/546	536/523
STD [mm]	64/71	64/71	64/71	68/77	66/73	/73	63/69
Mean bias [mm]		-4/-5	7/11	0/-7	-24/-37	/-36	-44/-59
RMS difference [mm]		4/5	9/11	15/20 (19)	33/41	/44 (22)	48/62 (15)
Period 1914-91 *(1925-77)*							
Correlation Coefficients low/high pass				0.91/0.97	0.80/0.97	*0.42/1.00*	0.84/0.98

There are some differences in the low-frequency variability between R_{ALL}(ori+) and the other series. Some of the most severe dissimilarities originate from the homogeneity breaks in connection with the gauge type changes, e.g. R_{24}(ori) and R_{TYN}(FIN). R_{HYD}(ori) spans the period of a Wild gauge network of increasing density and also shows a different low-frequency behaviour to R_{ALL}(ori+). The spotlight is now turned on the period 1921-80 that from an instrumentation point of view serves as a homogeneous interval for comparisons.

During the period 1921-80, the linear trend (least squares method) of R_{ALL}(ori+) is -0.74 mm(10 yr)$^{-1}$. This trend does not statistically significantly differ from zero at the 5%-level. The trends of R_{24}(adj), R_{24}(ori) and R_{TYN}(FIN) range from -0.55 mm(10 yr)$^{-1}$ to -0.65 mm(10 yr)$^{-1}$; they are not significantly different from the trend in R_{ALL}(ori+). In contrast to the other series, R_{HYD}(ori) shows increasing precipitation, with a trend of +0.59 mm(10 yr)$^{-1}$. This differs significantly from the trend in R_{ALL}(ori+). What causes the different behaviour of R_{HYD}(ori) during the period 1921-80?

Knowing that measured and true precipitation can have different trends (Førland and Hanssen-Bauer 2000), the possibility that R_{HYD}(ori) could reflect true precipitation, e.g., by harmonisation with river discharge or snow measurements, is now addressed. Equation (6.1) can be used to calculate a rough estimate of the corrected precipitation. Because the aim is not to produce an exact estimate but rather to explore a range of possible trends, the calculations are performed for the national averages. For some of the parameters in (6.1), annual average values for Finland are needed. Hyvärinen et al. (1995) have determined the monthly proportion of solid precipitation, $\overline{p}$, over 13 drainage basins for the period 1961-90. After transforming this information on to the LPNN grid, it can be estimated that the annual average $\overline{p}$ over Finland is about 38%. Lemmelä and Solantie (1977) reported $\overline{p}$=39% for the period 1931-60; the 30-year average of $\overline{p}$ has thus remained fairly stable. Solantie and Junila (1995) determined the mean value of annual wind speed, v=3.57 ms^{-1}, omitting coastal and island stations. At island stations, the annual mean wind speed, measured at a height of 10 metres, is 5-6 ms^{-1}; at coastal sites it is already 1-2 ms^{-1} lower. The year-to-year variation is not large. At both Helsinki and Sodankylä, the difference between the highest and lowest annual mean wind speed over 50 years or so is about 1 ms^{-1} (Heino 1994). In the Finnish precipitation network, there are stations that have an exposure factor of 100% (open) and those that have one of 0% (totally sheltered). There are no studies on changes over several decades of average exposure factor. According to Solantie and Junila, the average exposure factor over continental Finland is 35%. If one excludes the open and relatively open sites, the average drops to about 25%. Because in this study coastal and island stations are also included, a value of 40% is used as the national average. Table 6.4 summarises the parameters used in R_{cor}. The linear trend in R_{cor}, -0.92 mm(10 yr)$^{-1}$, is even slightly more negative than that in R_{ALL}(ori+) during the period 1921-80.

Equation (6.1) is used to explore the magnitude of non-precipitation related factors on the trend in measured precipitation. What would have been measured if a change in wind speed, proportion of solid precipitation or exposure of stations had happened? The changes in R_{cor}(wind), R_{cor}(snow) and R_{cor}(exp) (Table 6.4) are too large to be realistic. R_{cor}(max) is constructed in such a way that reasonable maximum changes of wind, snow and exposure are used. However, R_{cor}(max) is also artificial, because in reality there have not been large changes either in the mean wind speed or the proportion of solid precipitation (Heino 1994, paper I), nor has the sheltering of the station network changed systematically. The changes described in Table 6.4 are displayed in Fig. 6.3. To ease the comparison of trends, the series start from the same point and only low-frequency variability is shown.

Table 6.4. Parameters used in (6.1) to create artificial precipitation series for the period 1921-1980. R_{cor} is derived from R_{ALL}(ori+) with the correction procedures described in Solantie and Junila (1995). Varying the parameters in (6.1) linearly produces the other series. For example, in R_{cor}(wind) the wind speed is linearly reduced from 5.5 ms^{-1} in 1921 to 3.5 ms^{-1} in 1980.

Equation (6.1)	$\mathbf{R_{cor}}$	$\mathbf{R_{cor}}$**(wind)**	$\mathbf{R_{cor}}$**(snow)**	$\mathbf{R_{cor}}$**(exp)**	$\mathbf{R_{cor}}$**(max)**
Wind speed, v [ms^{-1}]	3.57	5.5 → 3.5	3.57	3.57	4.0 → 3.5
Proportion of solid precipitation, $\bar{p}$ [%]	38	38	40 → 23	38	40 → 35
Exposure factor, α [%]	40	40	40	45 → 30	45 → 38

R_{24}(adj), R_{24}(ori) and R_{TYN}(FIN) start from the same point in Fig. 6.3 and their low-frequency variability is qualitatively similar. The series differ from each other at the most by about 10 mm. The largest differences between the above-mentioned three series and R_{ALL}(ori+) are about 20-30 mm.

R_{HYD}(ori) is the uppermost curve in Fig. 6.3. R_{ALL}(ori+) and R_{cor} (the black curve with dots) start from the same point in 1921 but they end up about 70-80 mm lower level than R_{HYD}(ori) in 1980. R_{HYD}(ori) diverges from R_{ALL}(ori+) and R_{cor} throughout the time period; in particular, before the 1950s R_{HYD}(ori) behaves in a different way. In R_{cor}(wind) the mean annual wind speed is reduced down linearly by 2 ms^{-1} in sixty years, but even that does not produce as positive a trend as that seen in R_{HYD}(ori). If more precipitation falls in liquid form, it increases the amount of measured precipitation. A large systematic change of 17% (of the total) in $\bar{p}$ still produces a negative linear trend. A slight positive trend results from the (unrealistic) assumption that the sheltering of gauges at stations has improved by 15% in terms of α. R_{cor}(wind), R_{cor}(snow) and R_{cor}(exp) demonstrate the sensitivity of the precipitation measured to v, $\bar{p}$ and α according to the work of Solantie and Junila (1995). R_{cor}(max) attempts to illustrate the maximum reasonable combined effect of all three factors. The conclusion is that the low-frequency variability of R_{HYD}(ori) differs from that of R_{ALL}(ori+) and R_{24}(adj) more than can possibly be explained by changes in factors biasing the long-term measured precipitation (wind speed, proportion of solid precipitation and exposure factor) in Finland.

Hyvärinen (1997) evaluates the weaknesses and shortcomings of the drainage basin precipitation series that have been used to create R_{HYD}(ori). There are two <u>probable</u> explanations for the different low-frequency behaviour of R_{HYD}(ori). Firstly, from the beginning of the 1960s, possible even earlier, there has been a tendency to set up stations at higher altitude sites to compensate underestimation of areal precipitation. Secondly, knowledge of this underestimation may have affected the manual analysis. Both factors have probably contributed to a rising trend in areal precipitation. There are also some <u>possible</u> explanations for the increase. The use of different, partly subjective, analysis methods may have contributed, but there is no evidence for this. Perhaps the training of observers and improvement in the instructions to observers have increased the amount of measured precipitation, as most of the errors in the measurements lead to underestimation.

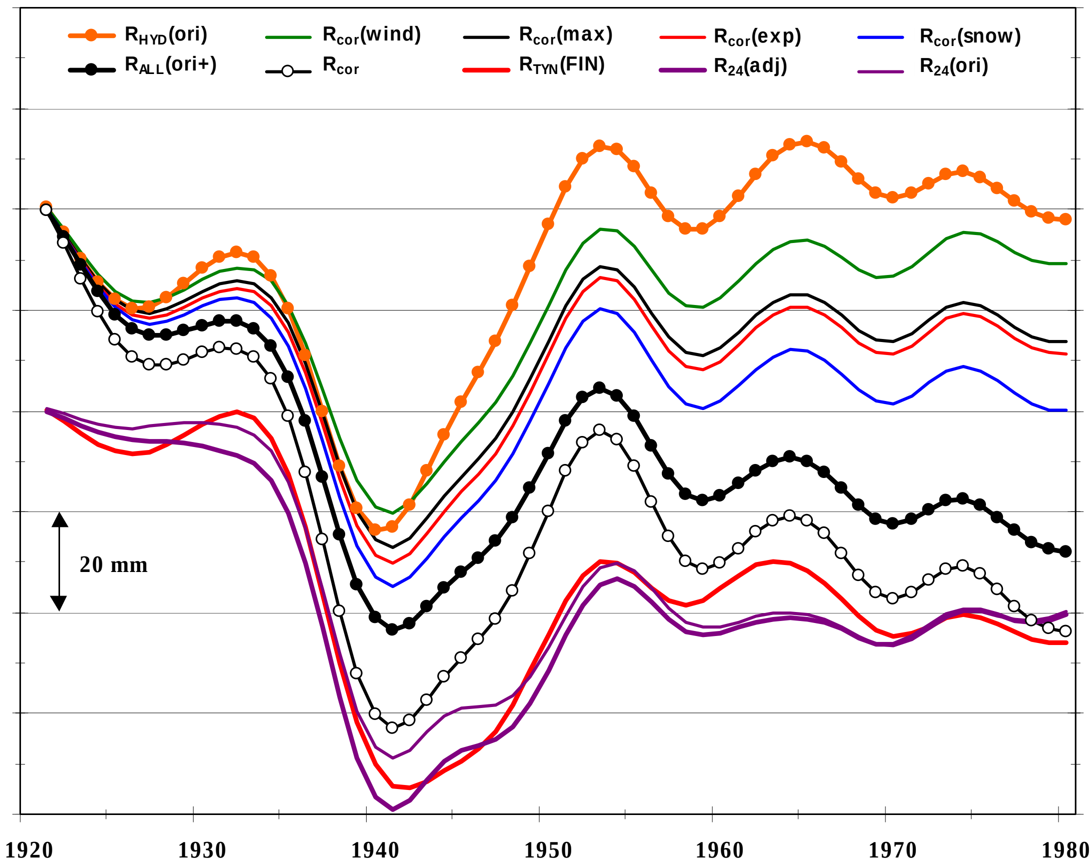

Fig. 6.3. Comparison of G3 filtered annual precipitation totals, 1921-1980. The series are artificially set to start at two points in 1921. The series refer to Tables 6.1 and 6.4.

The final conclusion is that R_{HYD}(ori) is biased and contains an artificial rising trend in precipitation. It must be noted that not all drainage basins reported in Reuna and Aittamurto (1994) suffer from inhomogenieties. As Hyvärinen (1997) points out, there are drainage basins where the station network has been representative, enabling studies of long-term changes.

6.4 Errors due to spatial sampling, and the total uncertainty in R_{ALL}(ori+)

At the beginning of R_{ALL}(ori+), the station network is sparse and southerly-biased (Fig. 2.3). It took until the 1930s before the number of precipitation stations reached two hundred (Fig. 2.2a). Because R_{ALL}(ori+) is area-averaged and uses all available precipitation series, it is selected as the baseline for the comparison. R_{1894}(ori+) and R_{1909}(ori+) are used to estimate the errors relative to R_{ALL}(ori+) due to incomplete spatial sampling during the early parts of the series.

The station network of R_{ALL}(ori+) is thinned out for the period 1970-2000 to calculate R_{1894}(ori+) and R_{1909}(ori+). R_{1894}(ori+) mimics the station network in 1894 covering about 30% of Finland on the LPNN grid. The network is kept practically unchanged through out the comparison period 1970-2000. In 1970 R_{1909}(ori+) imitates the network in the year 1909. However, no attempt is made to keep the network continuous. Hence some stations stop operation, and the network coverage is diminished from 80% to 72%. The average number of stations in R_{1909}(ori+) is about one hundred.

The comparison period mean precipitation is set equal for all three series before calculation of Table 6.5, thus, the mean bias is zero. The RMS difference of R_{1894}(ori+) is 25 mm, i.e. about 4% of the period mean (580 mm) and about 38% of the standard deviation (66 mm). It is considerably larger than that for R_{1909}(ori+). However, year-to-year changes in annual precipitation (first differences) are also captured relatively well by R_{1894}(ori+). The largest error in R_{1894}(ori+) is close to 50 mm, while for R_{1909}(ori+) it is about 10 mm. Incomplete spatial sampling causes the largest differences. Specifically, R_{1894}(ori+) misses the increasing trend that causes the large RMS difference in annual precipitation. R_{1894}(ori+) errors are large, and it seems difficult to extend the time series of Finnish mean precipitation back beyond the 1890s without considerable uncertainty. The R_{1909}(ori+) differences from R_{ALL}(ori+) are quite tolerable, considering the large variability of precipitation across a range of space-time scales. The comparison may give too optimistic a view of the accuracy of R_{ALL}(ori+), if the overall quality of precipitation measurements is worse during the first 40 years of R_{ALL}(ori+) than during the period 1970-2000.

Table 6.5. Comparison of the precipitation station networks of R_{ALL}(ori+), R_{1909}(ori+) and R_{1894}(ori+). Also shown are the root mean squared (RMS) difference of annual precipitation, RMS of year-to-year changes in annual precipitation, and the maximum and minimum differences from R_{ALL}(ori+), as well as the linear trend during the period 1970-2000.

Period 1970-2000	**R_{ALL}(ori+)**	**R_{1894}(ori+)**	**R_{1909}(ori+)**
Stations	318-624	34	84-121
Grid box coverage with ≥ 1 station [%]	94-100	30	70-82
Grid box coverage with ≥ 2 stations [%]	82-96	10	31-50
RMS difference of annual precipitation [mm]		25	5
RMS difference of year-to-year changes [mm]		10	6
Maximum difference [mm]		30	9
Minimum difference [mm]		-49	-13
Linear trend [mm/year]	2.3	-0.2	2.6

There is a systematic bias in the precipitation measurements due to the undercatch of the gauges. Even if this fundamental uncertainty is overlooked and the measured precipitation is analysed, there are still large differences between the national-average precipitation series examined in this study. A large number of stations is needed to overcome the uncertainty caused by the large spatial and temporal variability of precipitation. The uncertainty in the adjustment of annual precipitation due to the gauge changes in 1981/82 and 1909 is about 1-2% of the long-term mean measured precipitation. For solid precipitation the uncertainty is much larger and can reach 10% in individual winters. During the growth period of the station network there has been a systematic tendency to found new stations in regions with higher-than-average precipitation. Growing vegetation may also lead to an artificial increase in precipitation (paper IV and Hanssen-Bauer and Førland 1994). On the other hand, some of the long-term stations in cities have been moved to exposed sites at airports. Careful selection and adjusting is needed to retrieve unbiased station series.

The total uncertainty of the anomaly series of R_{ALL}(ori+) is subjectively estimated based on the results presented in this study: determination of gauge type change adjustments, comparison of series derived with different methods (e.g. spatial interpolation) and spatial sampling. The uncertainty in the anomalies of the annual precipitation sum over Finland increases with the age of data in a step-like manner as given in Table 6.6. Gauge type changes initiate the uncertainty. It also grows due to a diminishing number of stations and their spatially uneven distribution. The pre-1909 observations are assigned an extra uncertainty factor due to a possible imprecision of observation practices. The personal opinion of the author is that Table 6.6 overestimates rather than underestimates error limits around the national yearly precipitation anomaly. The error limits given in this study are by no means exact, but it is clear that the accuracy of precipitation estimates decreases before year 1910, due to poor spatial sampling and large measurement errors.

The low-frequency behaviour of R_{ALL}(ori+) is quite similar to the homogeneity-adjusted R_{24}(adj). The high-frequency variability of R_{ALL}(ori+) is similar to that of R_{TYN}(FIN) and R_{HYD}(ori) that are created with sophisticated spatial methods. It can be concluded that FDM, utilising all original measurements, seems to work well for precipitation, too. But it is not possible to evaluate the possible uncertainty inherent in the way FDM is used in this study, because of the deficiencies in the other national precipitation estimates.

Table 6.6. Estimated error limits for annual anomalies of R_{ALL}(ori+) for different time periods. Values are in mm.

1894-1896	**1897-1908**	**1909-1930**	**1931-1980**	**1981**	**1982**	**1983-2002**
±42	**±38**	**±26**	**±23**	**±21**	**±17**	**±15**

6.5 Variations in annual precipitation totals

Although there are large error margins related to the annual values of precipitation, 1941 is the driest year in all series by a margin of about 60-100 mm. The other nine years in the top ten dry years in R_{ALL}(ori+) differ by less than 30 mm (Table 6.7). Thus it is no surprise that in R_{24}(adj) and R_{TYN}(FIN) only about 7 of the top ten years are the same as in R_{ALL}(ori+) and with different ranks. The rainiest years in R_{ALL}(ori+) (as well as in R_{1909}(ori+) and R_{1894}(ori+)) is 1974, followed by 1981 with a 14 mm smaller annual total. However, in R_{24}(adj) and R_{TYN}(FIN) 1981 has larger sum than 1974 by 12 and 26 mm, respectively. The top ten wet years (Table 6.7) of R_{ALL}(ori+) contain about seven years that are included in the corresponding lists of R_{24}(adj) and R_{TYN}(FIN) (not shown).

It is worth pointing out that in principle the registration of droughts is not much affected by the type of the precipitation gauge. This is simply because the properties of the gauge have no influence if it does not rain. The scarceness of precipitation in 1941 must have been nation-wide, whereas the wettest years partly consist of convective precipitation that varies spatially. Quite a large number of stations are thus needed for reliable calculation of an area average over Finland.

Table 6.7. Top ten lowest and highest annual precipitation sums in R_{ALL}(ori+), 1894-2002. The values are deviations from the mean for the period 1961-90. Sums are arranged from the lowest to the highest value. Only the last two digits of years in the 20th century are given. N.B. The fourth highest value is for the year 1898.

YEAR - 1900																				
41	01	76	33	78	56	69	13	47	08	…	92	34	57	43	32	98	18 98	54	81	74
-186	-117	-113	-111	-109	-98	-97	-95	-90	-89	...	82	93	93	94	110	117	118	123	147	160
ANOMALY from the 1961-90 mean [mm]																				

There are fluctuations over the period of a decade of about ±30 mm, as can be seen from the G3 filtered curve in Fig. 6.4. The 1910s and the years around 1940 stand out as periods of low precipitation whereas the 1920s is the wettest decade. The G9 smoothed curve shows only weak trends, such as an increase of about 10 mm (~1.7% of the 1961-90 mean value) from the 1940s to the 1990s. This is somewhat contrary to what has been reported from Fennoscandia and northern latitudes in general. Statistically significant increasing annual precipitation trends have been observed in Norway (5-18%/100-year; Hanssen-Bauer and Førland 2000), in Sweden (10-20%/100-year; Alexandersson 2001) and for an average of all land areas between latitudes 55ºN and 85ºN (about 12%/100-year; Folland et al. 2001).

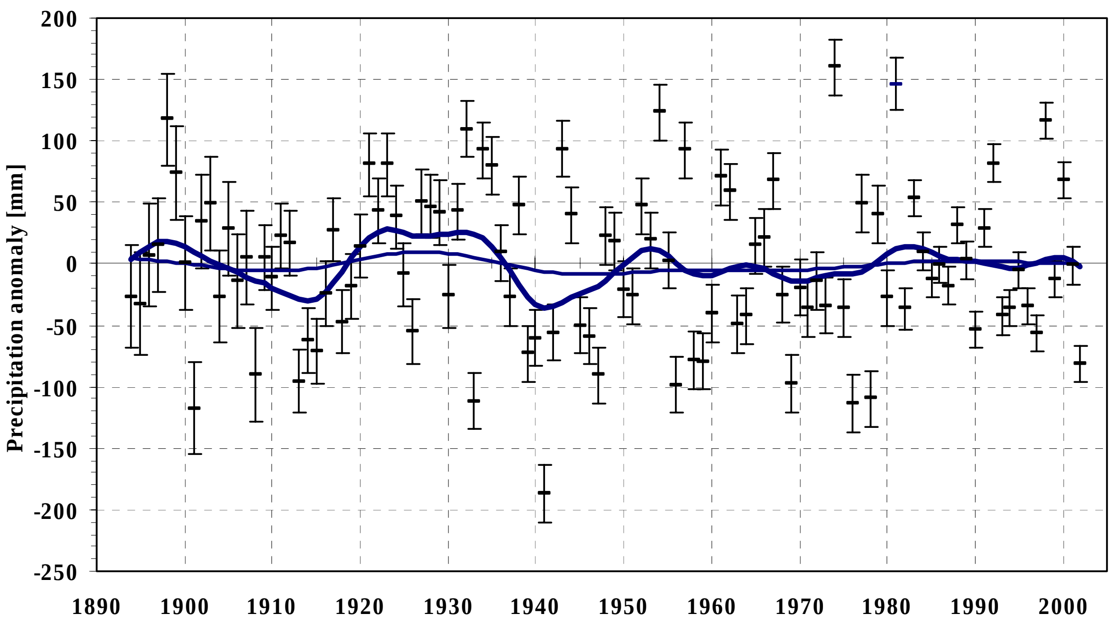

Fig. 6.4. Annual anomaly time series of R_{ALL}(ori+), 1894-2002 (reference period 1961-90). Annual values have error bars marked with thin lines (see Table 6.6). The thick curve is smoothed with the G3 and the thin curve with the G9 filter.

R_{ALL}(ori+) cannot be considered homogeneous in terms of instrumentation because it consists of time periods containing measurements with three different gauge types. Each gauge has its own characteristic errors. The Tretjakov gauge has a somewhat smaller aerodynamic error than the Wild gauge and a distinctly smaller error than the Finnish gauge. The ability of the Tretjakov gauge to measure more precipitation should lead to a larger variability in the annual sums than with the other gauges. This can be detrimental e.g. for the calculation of the long-term trends from R_{ALL}(ori+). In R_{ALL}(ori+), the standard deviation of annual precipitation is largest during the period of Wild measurements (1909-1981), even when STD is calculated from the high-frequency component. Thus, the natural variability of precipitation overshadows any possible increase in STD due to improved measurement accuracy.

No statistical analyses are presented for R_{ALL}(ori+) for three reasons. Firstly, Fig. 6.4 suggests that there has not been any systematic change in the annual mean precipitation over Finland. Secondly, large uncertainties compared to the long-term trends seem to be attached even to the annual mean values over Finland. Thirdly, data and methods do exist that can increase the reliability of long-term precipitation series. These will be discussed in the following section.

7. CONCLUSIONS

7.1 Main findings

The main findings of this study relating to the homogeneity testing and construction of climatic data sets are:

- SNHT is a practical and powerful tool for homogeneity testing and adjustment. The problem of homogeneous reference series can be solved by the iterative use of SNHT until relative homogeneity is obtained. The problem of detecting multiple breaks from time series can be avoided by testing and adjusting time series piecewise.

- Reliability can be much improved by using both statistical homogeneity tests and metadata information in the process of data sets construction. Statistical testing provides a method to detect unknown discontinuities in time series. However, relative testing methods may not be able to reveal roughly simultaneous changes in the observing network that can bias the whole data set. It is hoped that such changes have been documented as metadata. Metadata may also contain comparison measurements to adjust homogeneity breaks, and can offer physical reasons for breaks and exact timings to guide the homogenisation of a data set.

- Although SNHT is an objective method, subjective choices are required in the design of testing and adjustment procedures applied to data. For example, there are several ways to make use of metadata, to select confidence levels, and to construct reference series.

- It is useful to collect long-term time series of several climatic elements because there are many strong physical linkages between elements that can be used to verify homogeneity questions. For example, in this study, the regional averages of visual cloud cover observations get support from DTR, through their strong correlation, and vice versa.

- In the Finnish mean temperature and precipitation series there are simultaneous, nation-wide homogeneity breaks that bias the original series. Methods exist for adjusting these discontinuities (Heino 1994, this study sections 5 and 6).

- Averages based on a large number of stations can be homogeneous even without relative homogeneity testing and adjustment. The use of the First Difference Method (FDM) enables one to maximise the number of stations.

- There are very few long-term stations that meet the strict criteria of absolute homogeneity. Without adjustments erroneous conclusions may be reached. Therefore, it is important that homogeneity testing and adjustment is used to improve the quality of climatic data. Reliable Finnish time series make a relevant contribution to the studies of climate changes and their impacts, both nationally and internationally.

- Some of the global data sets evaluated in this study are based on original unadjusted data and, therefore, contain biased estimates for Finland. The problem is acknowledged (New et al. 2001) but not yet solved satisfactorily.

The main conclusions concerning the observed climatic trends in Finland and the Nordic region are:

- The 20th century precipitation (and temperature) climate of Finland is characterised by notable interdecadal variability. No significant, nation-wide precipitation trends were found.

- There are differences between various annual precipitation estimates over Finland. The reliability of precipitation analysis is relatively lower than that for temperatures.

- Statistical tests show that there has been a significant increase in the national annual and MAM mean temperatures during the last 150 years or so. The MAM temperature increase has been quite linear and is more significant than the annual mean temperature increase. The statistical significance of the annual mean temperature increase depends on the calculation period.

- Based on millennial, unforced control runs of coupled atmosphere-ocean general circulation models describing natural internal climatic variability, it seems that the most unusual feature of the Finnish annual mean temperature changes is the recent (from the mid-1970s) rapid increase in temperature. In the same model runs the 150-year MAM warming is outside the 95% confidence level.

- In Finland as in Fennoscandia, the mean maximum and minimum temperatures have increased during the period 1950-1995. At the same time, they have been decreasing in West Greenland. These trends are manifestations of the strengthening of the NAO. However, DTR has been significantly decreasing in both regions.

- In Fennoscandia, trends in the mean maximum and minimum temperatures over the period 1910-95 are not statistically significant, except the springtime increase in mean minimum temperatures. The decrease of DTR is significant, and can be explained primarily by cloud cover increase and a strengthening of the westerly flow bringing more humid marine air masses to Fennoscandia.

7.2 Pathways for further research

During the course of the work presented in this thesis, certain data acquisition requirements have appeared and topics for further research have arisen. This section contains a collection of issues to be concentrated on in the future work.

- It would add to the accuracy and spatial coverage of climate analysis if all climatic data series from the 19th century listed in Heino (1994) could be located from the archives and digitised. Holopainen (1999) and Holopainen and Vesajoki (2001) have compiled and analysed the 18th century observations in Finland. The gap between the data analysed in this study and the early instrumental observations of the 18th century should be bridged.

- There is also a demand for long-term daily time series of meteorological elements. Especially there is an interest in climatic extremes (e.g. Heino et al. 1999 and Frich et al. 2002) because society and the environment are often vulnerable to extreme events.

- During the period 1922-1965, the FMI maintained a network of stations measuring precipitation from May to September. At its maximum sometime in the 1940s there were about 500 stations in operation and in total there are about 1200 series. Recently this material has been digitised (A. Drebs, 2003, personal communication). Future analysis of liquid (or growing season) precipitation requiring either good long-term or spatial coverage should utilise these data.

- Concerning precipitation we are ultimately interested in the "true" precipitation amounts; measured precipitation is just a surrogate that may give ambiguous results, e.g., concerning trends (Førland and Hanssen-Bauer 2000). Besides compiling data sets containing all relevant observational data (measurements of precipitation, snow, wind, river discharge, etc.) there is a need to further develop and validate methods of correcting precipitation measurements.

- Analyses of past homogeneity breaks suggest valuable advice on the management of a climate observing network. Firstly, climatological observing networks should be designed to withstand relocations, observer changes and other often, but not always, random-like discontinuities, which will eventually affect all stations. Secondly, comprehensive comparison measurements are needed when new instruments, techniques or practices are introduced into the observing network.

- FDM gave promising results in this study. However, there are many new possibilities/questions concerning its use for the production of seasonal/monthly and gridded anomalies. For example, FDM can be used to transform homogeneity adjustment into a problem of spatial smoothing.

- Warming during the spring (March, April, and May) has been going on now for more than a century in Fennoscandia. What causes it? What is the role of the snow-albedo feedback? What consequences has the warming had?

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