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Annual and seasonal mean temperatures in Finland during the last 160 years based on gridded temperature data

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ABSTRACT: The annual and seasonal mean temperature of Finland was calculated for 162 years based on spatially interpolated monthly mean temperature records. The spatial interpolation method, known as kriging, was used with the following forcing parameters: the geographical coordinates, elevation of the terrain, and percentage share of lakes and sea. Homogenised data was used, and thus the most important factor affecting the accuracy of the interpolated data was the uneven distribution of the available observation stations both in time and space. The uncertainty due to the homogenisation adjustments made earlier was notably smaller. In the mid-1800s, the uncertainty in the annual and seasonal mean temperatures was large, with a maximum in winter of over $\pm 2.0\,°C$, but the accuracy improved quickly with time as the number of the observation stations increased. At the beginning of the 20th century, the uncertainty related to the limited station network was less than $\pm 0.2\,°C$, in winter less than $\pm 0.4\,°C$. According to the data, the rise in Finland's annual mean temperature has been statistically significant during the last 100, 50 and 30 years. During the last 100 years the increase in the mean temperature was largest during spring, but during the last 50 years winters have warmed up the most. The temperature time series obtained are compatible with grid point values picked from the global temperature data grids starting from the 1880s, though the global data sets tend to smooth the extremes. Copyright © 2009 Royal Meteorological Society

1. Introduction

Accurate quantification of the climate's observed variability on time scales ranging from inter-seasonal to centennial is required for number of purposes, including the monitoring and quantification of the present climate, understanding the variations in the climate, and giving a reliable basis for the evaluation of numerical models of the atmosphere and climate. The existence of long homogeneous climate records is not only beneficial to meteorological research but also to a number of other branches of environmental research, e.g. for constructing past climate from proxy records such as tree rings or shells and mussels (Guyette and Rabeni, 1995; Helama *et al.*, 2004; Holopainen *et al.*, 2006; Grudd 2008; Helama *et al.*, submitted) as well as borehole temperatures (Bodri *et al.*, 2001).

The spatial interpolation of climatic data is common and the techniques are various. A number of global gridded surface temperature data sets consisting of monthly land and/or sea surface temperature data for over 100 years have been developed and analysed, e.g., by Jones (1994), Parker *et al.* (1994), Jones and Moberg (2003), Smith and Reynolds (2005) and Brohan *et al.*

(2006). Casty *et al.* (2007) investigated the European pattern climatology over 230 years, based on monthly independently reconstructed gridded fields. A shorter but temporally higher resolution global data set has been created by Caesar *et al.* (2006), including global daily maximum and minimum temperatures for the period 1946–2000.

The spatial resolution of the global data sets is typically a few degrees in latitude and longitude. For gridding projects concentrating on a smaller area, e.g. for Europe or only for a particular country, a higher spatial resolution, typically from 5 to 50 km, is used, and the data sets may contain several variables (Perry and Hollis, 2005). However, usually these do not cover more than approximately 50 years of data, mainly because of the weak data coverage in earlier years. Mitchell and Jones (2005), however, offered monthly variations of nine climate variables for the period 1901–2002 at a 0.5° resolution covering the global land surface. Efthymiadis *et al.* (2006) developed a gridded monthly precipitation data set for the Greater Alpine Region for 1800–2003 at a spatial resolution of 10 min of arc.

The aim of this work is to produce the longest possible time series of annual and seasonal mean temperature values for Finland using a spatial interpolation method called kriging. Instrumental temperature records from Finland, Sweden, Norway and Russia are used as input. The time series are calculated starting from the year 1847. Until now, long Finnish temperature series have been based

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on direct averages of station observations (Tuomenvirta, 2004). A daily gridded climate data set starting from the 1960s was generated earlier by Venäläinen *et al.* (2005) using the same interpolation method as is used in this study.

This work assesses the ability of the kriging method to perform with low observation station densities, and addresses the accuracy of the obtained mean temperature analyses. The analysis of the time series obtained includes trend estimates for different time periods, with the statistical significance of the trends, and an assessment of the uncertainties.

The data and methods are presented in Section 2. The quality of the gridded data is covered in Section 3 and the analysis of the data in Section 4. Discussion of the results is given in Section 5 and the conclusions are drawn in Section 6.

2. Data and methods

2.1. Temperature observations

The weather observation data used in this study consists of monthly mean temperatures from Finnish weather stations, as well as from Swedish, Norwegian and Russian weather stations near the Finnish border. Finnish, Swedish and Norwegian temperature data were provided by the national meteorological institutes of each of the countries, and Russian data were obtained through the European Climate Assessment & Dataset (ECA&D) project (Klein Tank *et al.*, 2002) supported by the Network of European Meteorological Services (EUMET-NET). The monthly mean temperature values were calculated from daily mean values by direct averaging.

Only those Finnish weather stations were included for which the homogenised monthly mean temperature time series was available. Tuomenvirta (2001) performed the homogenisation using the Standard Normal Homogeneity Test (SNHT) (Alexandersson, 1986). The Finnish station data are quality-controlled and homogenised up to the year 2003. For the years 2003–2008, non-homogenised data have been used, but no significant changes in the measurement network or in the instrumentation have happened during that time. In Sweden, the homogenisation procedure has been very similar to the Finnish (Alexandersson and Moberg, 1997; Moberg and Alexandersson, 1997) and covers the years until 2006. From Norway and Russia non-homogenised data are used.

The annual number of observation stations is presented in Figure 1. In 1847, there were six stations recording monthly mean temperature, three of them being in Finland, and one each in Sweden, Norway and Russia. The number of observation stations stayed at around 10 until the late 1860s. In the 1870s the observation network was extended in western Finland, and in the 1880s many new stations were set up in different parts of southern and central Finland. The establishment of weather stations in the northern part of the country, Lapland, started relatively late compared to the other parts of the country:
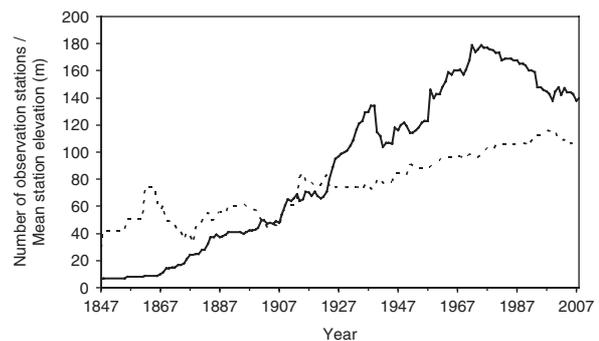

Figure 1. The number of observation stations (solid line) and the mean elevation (m) of the stations (dashed line) per year.

the first weather stations in Lapland were established in the early 20th century, and the longest homogenised temperature series from the area, from Sodankylä, does not start until 1908. Thus, data from Sweden, Norway and Russia are particularly important for constructing a fair picture of the annual and seasonal mean temperatures in the northern part of the country. The number of observation stations increased continuously until the 1970s. After the peak then of 180 homogenised stations, the number of observation stations has decreased, reaching a level of 130–150 in the 21st century.

The locations of the observation stations are presented in Figure 2. In each month the maximum amount of available observation stations was included in the analysis; thus, the number of stations varies, even between successive months.

2.2. Spatial interpolation method

Using spatial interpolation methods, information about climatic variables can be extended to areas with no observations. In this study, to interpolate the station data to a regular grid, a stochastic interpolation technique known as kriging was used (Matheron, 1963).

Vaajama (1966) already presented in the 1960s maps of maximum and minimum temperatures in Finland based on the highly smoothing method of averaging station values lying at a distance less than 50 km from the point under study. Later, in the 1980s, Ojansuu and Henttonen (1983) compared three different spatial smoothing methods for Finnish climatological data: a method of moving averages, a trend surface model and a model combining these two. The combination model was found to be the most suitable for calculating long time series, because it gave the most reliable results with respect to time.

The model applied in this study was developed especially for climatological applications by Henttonen (1991) following the theoretic approach by Ripley (1981). It is a combination of a trend surface model and a covariance function that is used to smooth the differences between the measured and the estimated values. The model has previously been used for research projects by Venäläinen and Heikinheimo (1997), Venäläinen *et al.* (2005), Vajda and Venäläinen (2003) and Vajda (2007).
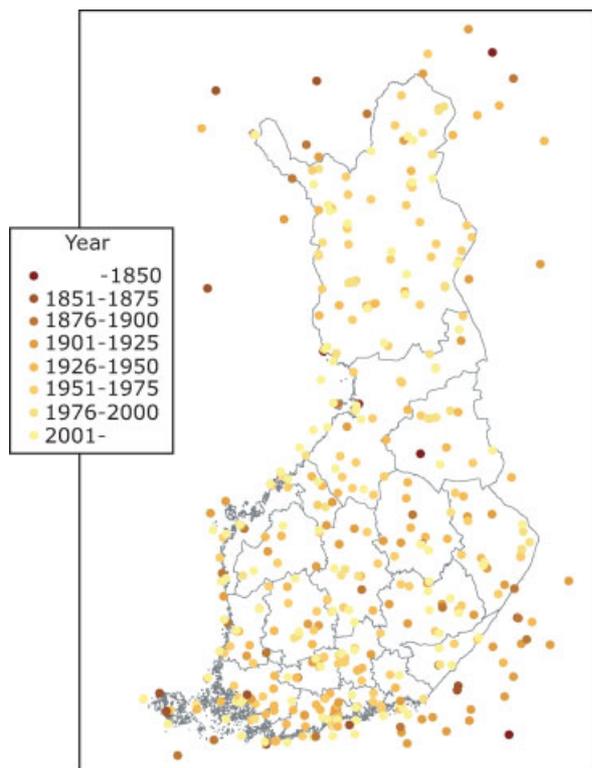
Figure 2. Map of the observation stations located in Finland, Sweden, Norway and Russia. This figure is available in colour online at wileyonlinelibrary.com/journal/joc

Kriging is a spatial prediction method that gives the best linear unbiased predictors of unobserved values, and also provides an estimate of the variance of the prediction error. In kriging, the predicted surface, i.e. the value of the analysed parameter (Z) at any location (x), is given as the sum of two terms:

$$Z(x) = m(x) + e(x) \qquad (1)$$

where $m(x)$ (often called the 'trend' or 'drift') describes the broad-scale features of the interpolated variable, and $e(x)$ is a spatial stochastic process, specific to the given position x, describing the small-scale random variation.

The advantage of kriging is that it can take into account external forcing parameters in the interpolation. In this study, such parameters as the geographical coordinates (x, y), elevation of the terrain (h), and percentage share of lakes (l) and sea (s) are included for each grid box. The parameters are specified for each grid box separately. The share of lakes and sea is defined as a percentage of the grid box area covered by water. From November to April the lake parameter is set to zero, as in winter the lakes are frozen and their influence on the climate is minor. The influence of the sea is taken into account throughout the year. The external parameters are included in the trend component $m(x)$. The plane coordinates are of the second order and the other parameters of the first order. Thus, the functional form of the quadratic trend surface is given by

$$m(x, y, h, l, s) = a_0 + a_1 x + a_2 y + a_3 x^2 + a_4 y^2$$
$$+ a_5 xy + a_6 h + a_7 l + a_8 s \qquad (2)$$

where the coefficients $a_0 \ldots a_8$ are obtained using the least-squares-fit method of the observed values.

The importance of the external forcing parameters – the geographical coordinates, elevation and the percentage share of lakes and sea – for mean temperature interpolation was studied by Vajda and Venäläinen (2003) in northern Finland. According to their study, the geographical position and local elevation had a significant influence on regional climate, while the effect of lakes and seas seemed to be secondary. However, the amount of lakes is considerably smaller in northern Finland compared with, e.g. the middle and eastern parts of the country, whereas the influence of the Baltic Sea is evident in coastal Finland (Laaksonen, 1977).

The resolution for the model was 10 km, giving a total of 3829 grid points covering the whole of Finland. With such a relatively coarse resolution, local effects may be masked, and for studies focussed on a smaller area, a higher resolution grid, e.g. 1 km, should and can be used (Vajda, 2007).

2.3. Validation of the spatial model

The interpolated temperature data were validated against the observed monthly mean temperatures. The calculated skill scores are the mean bias deviation (MBD), the mean absolute error (MAE), the root-mean-square error (RMSE) and the compound relative error (CRE), that measure the percentual bias (%), the absolute error (°C), the deviation (°C), and the correlation of the spatial model, respectively. All the quantities were calculated as modelled minus observed value. Validation was done using data from the period 1971–2000.

All the calculated skill scores are shown in Figure 3. The MAE and RMSE show larger errors and deviations in winter than in summer, deriving from the inability of the model to catch the lowest mean temperatures, although, according to the CRE, the model fits the observations slightly better in winter than in summer. The correlation (R^2) between the modelled and the observed values varies from 0.96 in July to 0.99 in November. According to the MBD, the model overpredicts (MBD positive) during winter and underpredicts (MBD negative) during summer. Thus, the estimated fields are smoother than the observations: the estimates are on average warmer than the observations in winter and colder in summer. The MBD shows a striking step change in the transition seasons. This is likely related to the lake effect that is set to zero from November to April in the model, while in reality melting and freezing of lakes does not take place all at once but gradually. Lakes in the southernmost Finland melt on average already in April. In autumn, freezing occurs in the northern Finland in October but in the southern parts of the country lakes do not freeze until the end of November or early December (Korhonen, 2005).

2.4. Calculation of the seasonal and annual mean temperature values

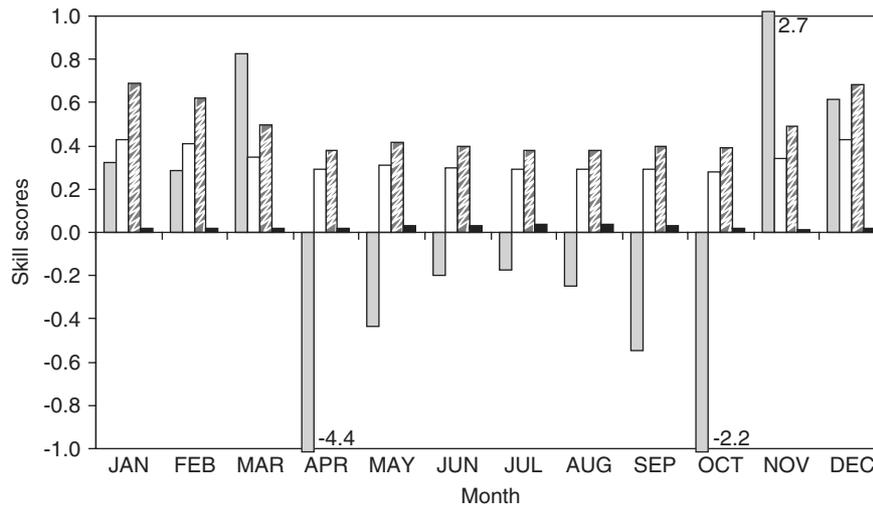The individual station values of monthly mean temperature calculated from the station's observations were first

Figure 3. Skill scores for the monthly data: mean bias deviation, MBD (grey), mean absolute error, MAE (white), root-mean-square error, RMSE (dashed) and compound relative error, CRE (black).

interpolated to a regular grid using kriging for each month starting from the beginning of 1847, and continuing until the end of February 2009. After interpolating the monthly mean temperatures, the annual and seasonal values were calculated by averaging the monthly grid values for each grid point separately. The annual and seasonal mean temperature values for the whole country were then calculated as averages over the total of 3829 grid point values.

3. Quality and uncertainties of the data set

The errors and uncertainties in the interpolated annual and seasonal mean temperature values are caused by the inaccuracies involved in the original instantaneous temperature measurements, as well as in the daily and monthly averaging methods, and also in the interpolation procedure of the monthly station values. In this study, homogenised data were used, and issues concerning instrument changes, relocation of the thermometers or the stations, changes in the environment (including urbanisation), changes in observing times, or changes in the formula used for calculating mean temperature have already taken into account in the station data (Tuomenvirta, 2001).

Thus, only two uncertainty factors are discussed here:

1. uneven distribution of stations in space and time;
2. homogeneity adjustments already made in the dataset.

3.1. The uneven distribution of stations

The most important factor contributing to the accuracy of the interpolated annual and seasonal mean temperatures is the number and distribution of the available station observations. The effect of the observation station coverage on the annual and seasonal mean temperatures was calculated by comparing mean temperatures for the period 1971–2000 based on certain station combinations imitating the station network in the early years of the time series

against the best combination of the station network, i.e. with all the available stations. This required the reasonable assumption that the spatial changes in temperature have not changed significantly during the study period. In order to make the comparison, only those stations that were operational in the reference period 1971–2000 could be included in the station combinations representing the earlier periods. However, in the final analysis, all available station observations were used and, therefore, the spatial coverage of the observations was somewhat better than the comparison would suggest. This is why the comparison is likely to give a slight overestimation of the error and uncertainty.

It was discovered that the error due to the limited station coverage is not random but mostly negatively biased, i.e. the interpolation gives too low estimates for mean temperatures compared with the reference data. Annual and December–January–February (DJF) mean temperatures constantly indicated a cold bias that followed a logarithmic distribution as a function of the number of observation stations. The interpolated values were therefore corrected according to this logarithmic function fitted to the median error [Figure 4(a)]. However, for the rest of the seasons, the distribution of the error was not as systematic, but varied between positive and negative values when the station density was low. Here, as a rough estimate, we applied the fit of a sixth-order polynomial [Figure 4(a)]. For March–April–May (MAM), June–July–August (JJA) and September–October–November (SON), for station numbers larger than 160, the median error was assumed to be zero. An interesting finding is that the cold bias also appears in the interpolated annual mean temperatures when the resolution of the station network is good, even with over 100 observation stations. Most of it results from the wintertime error.

In the spatial distributions, when the absolute error is set as less than 0.5 °C throughout the area, the required station coverage for the interpolation of the annual mean
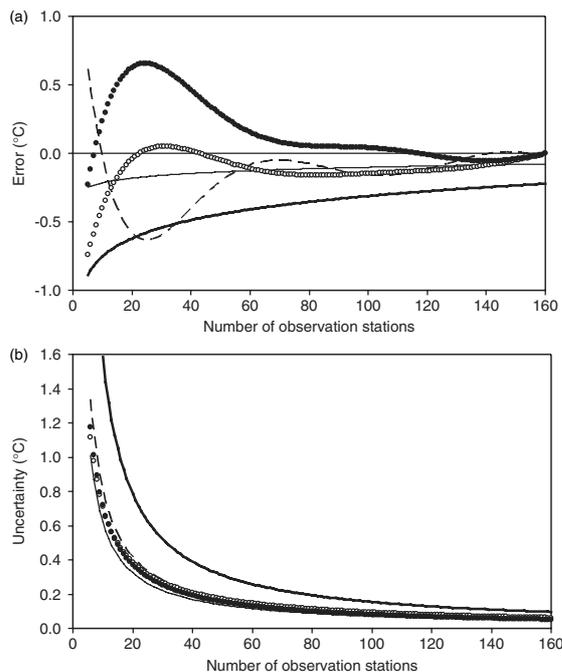
Figure 4. (a) Fitted curves for the median error (°C) of the interpolated annual and seasonal mean temperature as a function of the number of observation stations. Annual: thin solid line, MAM: open circles, JJA: filled circles, SON: dashed line, and DJF: thick solid line. (b) As (a) but for the 95% confidence limits (°C) of the annual and seasonal mean temperatures.

temperature is around 50, corresponding to the situation in the beginning of the 20th century. The same criterion is fulfilled with 40 stations in spring, 100 stations in summer, 80 stations in autumn, but not until 120 stations in winter, corresponding to the station network in the 1950s.

The quadratic trend function causes the resulting temperature surface to curve towards the edges in case there are observations only in the middle of the study area. In the mid-1800, lack of observations made the estimated temperatures decrease towards the western and eastern edges of the study area and increase towards north and south [Figure 5(a)]. These false spatial temperature trends were emphasised particularly in winter, when the average temperature differences between different parts of the study area are largest. The decrease in temperatures towards the eastern and western borders of the area dominated over the warm error in the northern part of the country, and made the areal average colder than that of the reference data.

The first observation stations in the western part of the country were established in 1873, and after that the error no longer shows as an erroneous spatial temperature trend there, but is more random. Near the eastern border the situation stays poor for longer [Figure 5(b)]. In that area, the winter mean temperatures can be up to two to three degrees too low at the turn of the 20th century. The cold bias near the eastern border remains right up to the mid-20th century, as especially in the north it was not until the 1960s to 1970s that the area was sufficiently covered by observations.

The first temperature series in Lapland, northern Finland, does not start until 1908. Before that the interpolation in that area is based only on the observations from the Swedish and Norwegian stations. The longest records
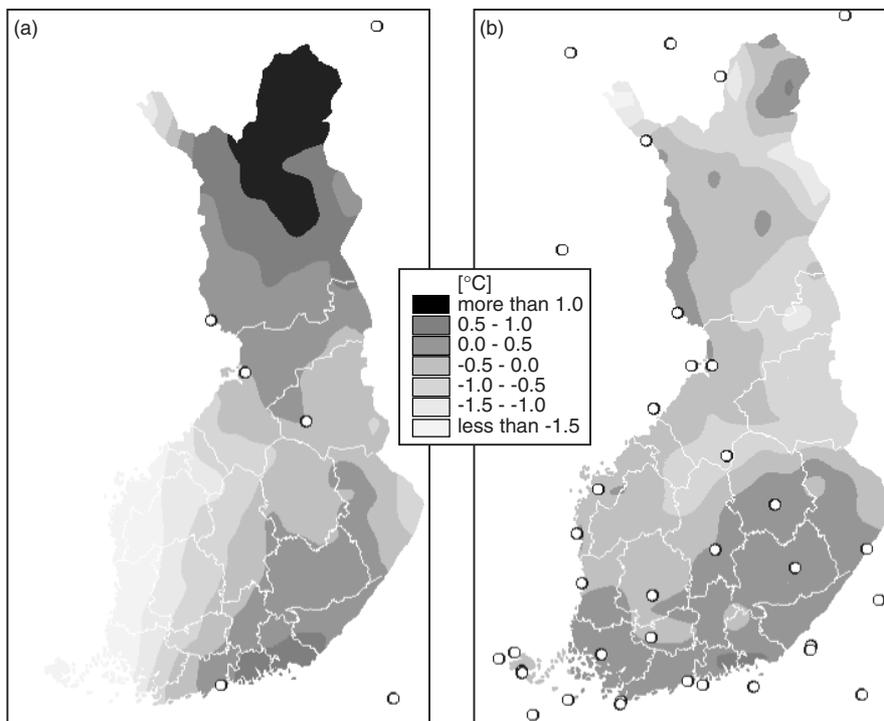

Figure 5. The average error (°C) in the interpolated annual mean temperature (a) in the 1850s with six observation stations (white circles) and (b) in the 1880s with 36 observation stations.

from the northern part of the study area, at the Norwegian stations of Vardö and Tromsö, represent warmer climatic conditions, because they are situated on the coast of the Norwegian Sea and the Arctic Ocean. Finnish Lapland represents a continental climate, and the temperature conditions there differ from those in the coastal areas, especially in winter when the average temperatures are several degrees Celsius lower inland compared to the coast. Thus, with an insufficient station coverage, the annual mean temperature in Lapland becomes at most over 2 °C too high [Figure 5(a)]. However, in the areal average, the warm error in Lapland is masked by the cold one in the western and eastern parts of the country.

The elevation of the available weather stations might become an important factor when the station network is sparse. Elevation, along with the geographical coordinates, plays a significant role in the kriging of climate variables (Vajda and Venäläinen, 2003). However, the influence of elevation is less than, e.g. the location or number of stations. Locally, in persistent inversion situations, higher altitude stations are expected to be warmer than lower ones. But, also in winter, mean temperatures decrease on average with increasing altitude. In the 19th century, especially in the 1860s, the mean station elevation was relatively high compared to the number of stations at that time (Figure 1). This might have contributed to the cold bias to some extent.

Elevation, as well as the other parameters included in the interpolation, works best when the observing network is dense. Without stations at various altitudes, the model cannot describe how the elevation affects the temperatures in different places. Similarly, if observations are concentrated only on the coast line, the model cannot describe temperatures correctly inland. For this reason, the parameters alone, without a sufficient station network, cannot produce correct estimates.

The uncertainty range for the best, corrected estimates of the annual and seasonal mean temperatures is obtained through the standard deviation of each tested station combination. The uncertainty range is set to satisfy the 2.5% and 97.5% probability limits, i.e. ±1.96 times the standard deviation of each station combination, assuming a normal distribution of the error. The standard deviation of the error, both in annual and seasonal mean temperatures, was found to follow a power law [Figure 4(b)]. Thus, the uncertainty of the interpolation is relatively large with the sparse station network of the mid-19th century, but it falls rapidly with an increasing number of observation stations. After a certain point, an increase in the number of stations does not significantly further improve the accuracy of the interpolation.

With 60–70 observation stations, the uncertainty in the interpolated annual mean temperature of Finland decreases to ±0.10 °C. This corresponds roughly to the station coverage in the 1910s. An uncertainty of less than ±0.20 °C is achieved with at least 35 stations, in the mid-1880s. The corresponding uncertainties in the seasonal mean temperatures of Finland with the same station combinations are approximately ±0.13 °C in MAM, JJA and SON and ±0.25 °C in DJF for 60–70 stations, and ±0.23 °C in MAM, JJA and SON and ±0.45 °C in DJF for 35 stations.

In the previous study by Ojansuu and Henttonen (1983), the standard errors of the May, June, July and August mean temperatures calculated with the combination model (a trend surface model with moving averages) were presented. The standard deviation of the monthly mean temperature estimates seemed to follow a power law, with values of 0.7–1.1 °C with less than 20 stations and 0.3–0.5 °C with over 100 stations. However, in Ojansuu and Henttonen (1983), the standard errors for different station combinations were calculated as an average of the residuals between the estimates and the values measured at the observation stations, whereas, in our study, the 95% uncertainty limits were calculated from the mean temperature residuals of the whole study area.

The geographical distribution of the available stations is fundamental both to the error and the uncertainty of the interpolation. It is, therefore, worth emphasising that the errors and uncertainties in the northern part of the area remain larger substantially longer than in the other parts of the country.

3.2. Homogeneity adjustments

The uncertainty due to the homogeneity adjustments is represented by the limit of how small inhomogeneities can be detected (Brohan *et al.*, 2006). The uncertainty related to the annual mean temperature of a single station after one homogenisation adjustment is set to ±0.15 °C. This value is based on results by Tuomenvirta (2001) who tested the homogeneity of Finnish annual mean temperature data during the period 1958–1993. For the seasons, arbitrarily chosen values of ±0.1 °C and ±0.25 °C for summer and winter, respectively, and ±0.15 °C for both spring and autumn were used in the calculations of uncertainties. The higher uncertainty in the winter arises from increased spatial and temporal temperature variance during the cold season.

The homogenisation uncertainty is assumed to be cumulative in reverse, so that the most recent segment is the most reliable. Each adjustment is considered to add an equal uncertainty to the station value, independent of the station network. So, when going back in time, after two adjustments the uncertainty of a single station is assumed to be doubled, after three adjustments tripled, etc. Finnish stations have been adjusted at most six times during 1847–2008. For all non-homogenised stations (Norwegian and Russian stations) and stations with no detailed information about the homogenisation adjustments performed (Swedish stations), an uncertainty of ±0.1 °C was assumed throughout the whole time span 1847–2008, both for annual and seasonal values.

The homogenisation adjustments uncertainty for the national mean temperatures is calculated as the root mean square of the station uncertainties, multiplied by $1/\sqrt{n}$, where n is the number of all (both adjusted and non-adjusted) stations. Thus, the homogenisation

adjustment uncertainty is very small when the number of the available stations is large, but increases backwards in time as the number of available stations decreases and the share of the more-than-once adjusted stations increases. During, roughly, the last 10 years, the uncertainty is zero, as there are no adjustments after the 1990s. The uncertainty in the mid-1800s varies between ±0.12 °C in summer and ±0.32 °C in winter. For the national annual mean temperature, the homogenisation adjustment uncertainty is at most ±0.19 °C (Figure 6).

The ability of a test method to detect a homogeneity break gets worse as the number of observation stations decreases. The uncertainty of the homogenisation adjustments thus becomes larger when going back in time, as the number of available reference stations for the homogenisation procedure decreases and the procedure becomes more inaccurate. At the very early years when it becomes impossible to detect homogeneity breaks, one would need to quantify the uncertainty level in a different way; one simple way might be to fit a curve to the uncertainties with time and extrapolate that to the earliest years. We have not taken the temporal change of the homogenisation procedure's ability into account in the uncertainty estimations because of its minor influence on the result. As shown in the previous section, it is the uncertainty due to limited station coverage that is the dominating source of error, when it comes to the estimates of the annual and seasonal mean temperatures of Finland.

4. Annual and seasonal mean temperature climatology in 1847–2008

The annual mean temperature of Finland in 1847–2008 based on the interpolated monthly mean temperature data is presented in Figure 7(a). Both the annual values and the smoothed values (calculated with a 21-point binomial filter) are given. The envelope curves give the 95% confidence limits due to the time series homogenisation adjustment and the limited observation station coverage. The corresponding graphs for the seasonal mean temperatures are presented in Figure 7(b)–(e).

Linear trends for the annual and seasonal mean temperature time series for different time periods were calculated using the least-square method (e.g., Spiegel, 1961).
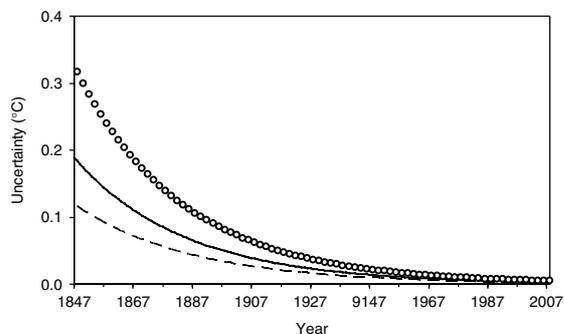
Figure 6. Uncertainty (°C) in annual and seasonal mean temperatures related to the time series homogenisation adjustment (annual: solid line, DJF: open circles, JJA: dashed line).
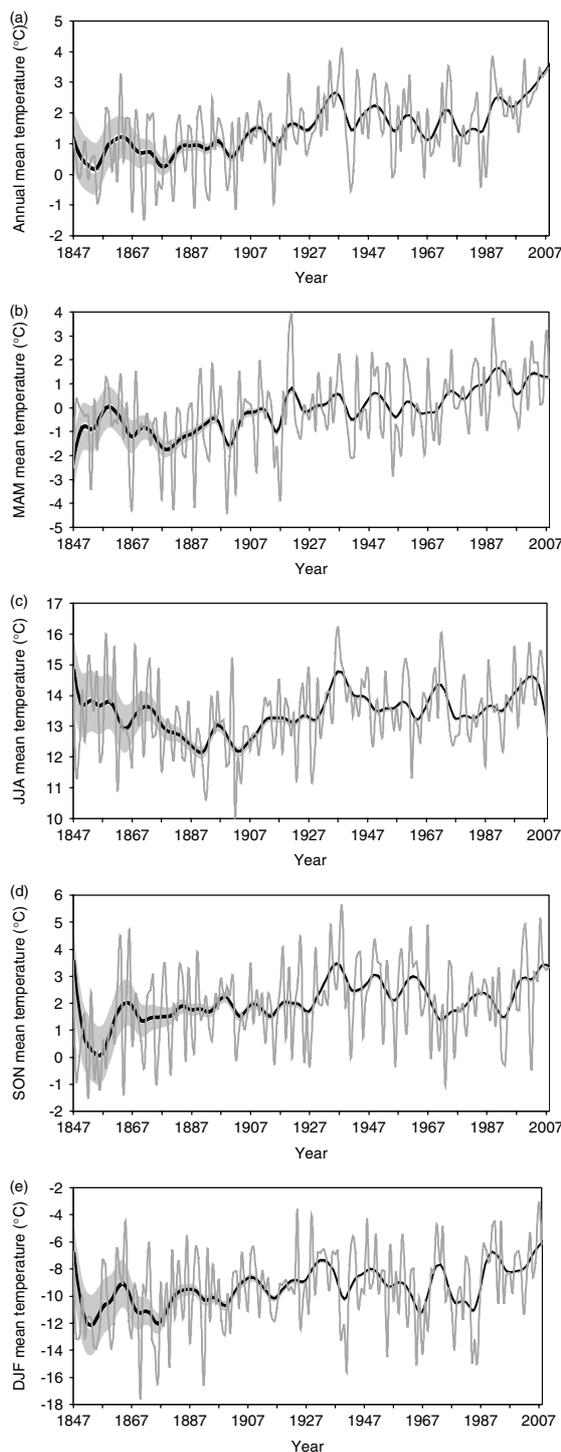
Figure 7. a) Interpolated annual mean temperature (°C) of Finland 1847–2008. Annual values are given by the grey line and smoothed annual values by the black line. The 95% uncertainty range due to the time series homogenisation adjustment (white envelope) and limited station coverage (grey envelope) are also shown. Same for b) MAM, c) JJA, d) SON, and e) DJF mean temperatures. The smoothed curves were created using a 21-point binomial filter giving neardecadal averages. The time series were extended backwards and forwards by reflecting the data across the end points before filtering to make the filtered series cover the whole time range.

The 95% confidence intervals for the slopes of the regression lines are calculated through the standard error of

the *y*-coordinate of the slope. The calculated trends with their confidence limits are given in Table I. The statistically significant trends at the 5% level according to the nonparametric Mann–Kendall test (Sneyers, 1990) are indicated in Table I with boldfacing.

During the last 100 years (1909–2008) the annual mean temperature in Finland has risen by 0.93 ± 0.72 °C. The rise in the mean temperature has been largest during spring, the 100-year linear trend being 1.59 ± 0.98 °C. However, for the last 50 years, the rate of warming has clearly been largest in winter, the rise in the DJF mean temperature from 1958 to 2007 being 3.47 ± 2.78 °C. During the last 30 years (1979–2008) the wintertime warming has been even more pronounced, the increase in the DJF mean temperature during the last 30 years being 4.28 ± 3.44 °C. Since 1988, winters in Finland have been mainly warm, the winter of 2007–2008 being a record mild one, while in the late 1970s and early 1980s there were several cold winters. This distinctive division into periods of cold and warm winters makes the 30-year linear trend in the winter mean temperature very striking.

The significance of the 100-year springtime trend was very high, at a level of 0.1%. For the annual mean temperature, the significance of the long-term trend was at the 1% level. During the last 30 years (1979–2008), the increase in the mean temperature has been statistically significant both annually and seasonally, excluding springtime.

The observed changes in the annual and seasonal mean temperature of Finland were previously studied by Tuomenvirta (2004), based on averaged temperature data of several station observations. According to Tuomenvirta (2004), the 100-year (1901–2000) linear trend for annual mean temperature was 0.08 °C per decade and was statistically significant. Based on the interpolated monthly mean temperature data used in this study, the decadal trend in 1901–2000 was 0.07 ± 0.07 °C while in 1909–2008 it was 0.09 ± 0.07 °C.

5. Discussion

The interpolation procedure is very sensitive to the station network when the station coverage is limited. Lack of observations caused the resulting surface to curve towards the edges. To improve the accuracy of the interpolated annual and seasonal mean temperatures, an essential task would be to develop the performance of the interpolation method when the observation station network is sparse. One possibility to limit the unrealistic curving of the surface might be to use a linear approach (instead of second order) for the plane coordinates when the station network is limited. However, this would not necessarily improve the temperature estimates, but the effect would be more aesthetic.

One could also follow the procedure of a reduced space optimal analysis for historical data sets presented by Kaplan *et al.* (1997). The technique has been applied at least in Schmidli *et al.* (2001; 2002) for precipitation

in the Alpine region. The basic idea of the method is to extract the modes of spatio-temporal variability from the high-resolution observations over a limited time period, and subsequently to estimate the coefficients of these modes from the sparser long-term data.

The sufficient station coverage varies between the seasonal and annual time scales. Generally, the errors and uncertainties in the seasonal mean temperatures are greater than those in the annual mean temperature [Figure 4(a) and (b)]. For monthly temperature data they are expected to be even larger, when going to a still shorter time scale. The errors and uncertainties are largest during winter, when the spatial variation of the average temperature is also largest. When using the interpolated mean temperature data the uncertainties should be kept in mind. The spatial annual or seasonal mean temperature data cannot be used in detail before a sufficient station coverage is reached, but instead can be used to give some general indicators of the climate at issue.

The phenological events of boreal trees, such as flowering and leaf bud burst, are showing an earlier occurrence that can be attributed to climatic warming. In Linkosalo *et al.* (2009), this phenomenon was used to verify direct observations of climatic warming. Long-term warming estimates derived from the phenological trends (between 1846 and 2005) suggested that the mean spring temperature increase in Finland has been 1.8 °C/century, which is close to the value 1.4 °C/century (between 1847 and 2008) indicated by this study. To eliminate the effect of urbanisation on local temperatures, phenological observations made in densely populated areas were rejected. This suggests that the urbanisation effect has been successfully removed from the actual temperature records too.

The seasonal mean temperature time series produced show similar patterns to the European ones (Casty *et al.*, 2007), but there are also differences. During spring, Finland's mean temperature has risen more or less steadily all the way from the 1880s, while on a European scale the springtime increase is concentrated in the first decades of the 20th century. Since the 1870s, winters had been warming in Finland, leading to temperatures higher than the 20th century mean from the 1920s to the late 1950s. European winters instead exhibit a slight cooling until 1900. The two severe cold winter periods in the 1960s and 1980s seem not on average to be as pronounced in Europe as in Finland.

We also compared the dataset with time series picked from the global temperature anomaly datasets: the NASA GISS Surface Temperature Analysis (GISTEMP) created by NASA's Goddard Institute for Space Studies (Hansen *et al.*, 1999, 2001) and the HadCRUT3 dataset by the UK Met Office Hadley Centre (Brohan *et al.*, 2006). Values representing Finland were collected from the global temperature anomaly grids, and the annual temperature anomalies in 1880–2007 were compared with the 10-km resolution data created in this work. The correlation coefficients between the 10-km resolution annual mean temperature anomaly time series and the global datasets were

Table I. Trends in annual and seasonal mean temperatures (°C) for different time periods with the 95% confidence intervals.

	1909–2008	1959–2008	1979–2008
Annual	**0.93** ± 0.72 (**0.09** ± 0.07)	**1.52** ± 0.96 (**0.30** ± 0.19)	**2.05** ± 1.07 (**0.68** ± 0.36)
MAM	**1.59** ± 0.98 (**0.16** ± 0.10)	**1.46** ± 1.19 (**0.29** ± 0.24)	0.78 ± 1.59 (0.26 ± 0.53)
JJA	**0.69** ± 0.70 (**0.07** ± 0.07)	0.63 ± 1.03 (0.13 ± 0.21)	**1.26** ± 1.21 (**0.42** ± 0.40)
SON	0.39 ± 0.99 (0.04 ± 0.10)	0.79 ± 1.40 (0.16 ± 0.28)	**1.55** ± 1.67 (**0.52** ± 0.56)
DJF	0.97 ± 1.87 (0.10 ± 0.19)	**3.47** ± 2.78 (**0.69** ± 0.56)	**4.28** ± 3.44 (**1.43** ± 1.15)

A trend per decade corresponding to each period is given in brackets (°C/decade) also with the 95% limits. For winters (DJF), the year corresponds to the first year of the season (e.g. winter 2008 = Dec 2008 to Feb 2009). Statistically significant trends (*p*-value max. 0.05) according to the nonparametric Mann–Kendall test are indicated with boldfacing.

high: 0.98 with GISTEMP and 0.97 with HadCRUT3. The extreme annual mean temperature values were found to be somewhat smoothed in the global datasets, especially in HadCRUT3, compared to the 10-km resolution data.

Globally, according to the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report, the 100-year linear trend (1906–2005) for the annual mean temperature is 0.74 °C, being over 0.1 °C larger than the corresponding value, 0.6 °C (1901–2000), in the IPCC Third Assessment Report (Trenberth *et al*., 2007). The corresponding value for Finland is virtually the same as the global value: 0.73 °C. Further, according to Trenberth *et al*. (2007), the global rate of warming during 1956–2005 is almost double that of the 100 years (1906–2005), i.e. 0.13 °C per decade in 1956–2005 *versus* 0.07 °C in 1906–2005. In Finland, the rate of warming has tripled during the last 50 years (0.30 *vs* 0.09 °C per decade, in 1959–2008 and 1909–2008, respectively), fitting the theory that the temperature increase is greater at higher northern latitudes (Trenberth *et al*., 2007). According to projections in Jylhä *et al*. (2004), the increase in the annual mean temperature in Finland during the next 30 years will continue the same (0.6 °C/decade) as the observed warming during the last 30 years (0.7 °C/decade). The warming is likely to remain strongest in winter.

6. Conclusions

With spatial interpolation, information about climatic variables can be extended to areas with no observations. In this way we can produce long, continuous time series that are useful, e.g. in calibrating climate proxies. However, interpolation estimates are very sensitive to the availability of the observations, especially when there are only a few active observation stations.

In this study, 162-year-long (1847–2008) time series of annual and seasonal mean temperatures were created for Finland. The most important factor causing inaccuracies in the data was the limited station coverage both in time and space. At the end of the 19th century, the uncertainties in the interpolated annual and seasonal mean temperatures exceed ±0.15 °C, the uncertainty in winter being as much as ±0.35 °C. The spatial distribution

of annual mean temperature remains inaccurate until the early 20th century, the seasonal distribution error already reaching a level of less than ±0.5 °C throughout the area in spring in the 1880s, in summer and autumn in the 1920s, but in winter not until the 1950s. Thus, to gain a longer time series of annual and seasonal mean temperatures with improved accuracy, especially in the spatial distributions, an interpolation method for low station densities needs developing.

However, it is important to note that from the beginning of the 20th century, the errors and uncertainties in the national averages based on this gridded data set are small, and thus there are now available over 100-year-long data sets of good quality for the annual and seasonal mean temperatures of Finland.

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