# Climate around year 1900

Scientific briefing from DCE – Danish Centre for Environment and Energy

Date: 17. Juni 2021 | **52**

AARHUS UNIVERSITY
DCE – DANISH CENTRE FOR ENVIRONMENT AND ENERGY

Danish Meteorological Institute

# Data sheet

Scientific briefing from DCE – Danish Centre for Environment and Energy

Category: Scientific briefing

Title: Climate around year 1900

Author: Flemming Vejen
Institution: DMI

Referee: Torben Schmidt, DMI
Quality assurance, DCE: Signe Jung-Madsen
External comment: There were no comments to the draft

Claimant: The Danish Environmental Protection Agency

Please cite as: Vejen, F.. 2021. Climate around year 1900. Aarhus University, DCE - Danish Centre for Environment and Energy, 42 s. – Scientific briefing no. 2021|52
https://dce.au.dk/fileadmin/dce.au.dk/Udgivelser/Notater_2021/N2021_52.pdf



Front page photo: Colourbox

Number of pages: 42

# Contents

# Summary

The work presented in this report is a part of a project with the purpose to calculate water balance and discharge of nutrients to open sea areas in Denmark around year 1900 (1890-1910). Based on digitized observations spatially distributed meteorological variables are provided for hydrological modelling of water discharge. Rain gauge observations have been subject to correction for different biases such as undercatch caused by wind turbulence, but not all the observations required for the bias correction were available. A climatological rain rate was assumed, precipitation type was calculated from dry air temperature, and wind speed at station level was calculated by interpolation. No information about station environment was available, and assumptions were therefore made about the general shelter conditions.

Discharge data are not available before 1917, thus corrected precipitation and modelled discharge around year 1900 cannot be verified. As discharge data are available after 1917, the period 1917-1950 is chosen as a validation period, and it is assumed that the conditions during this period are comparable with the period 1890-1910.

Regional and national water balance errors have been used to determine the most likely general shelter conditions. A general water balance error of 3 % for the verification period 1917-1950 suggests that the national values for corrected precipitation are realistic, but this is not the case at a regional scale. It is assumed that the spatial distribution of rainfall around year 1900 has not changed significantly over the years. The spatial distribution of amount of rainfall in a reference period 1989-2010 has therefore been transferred to the period 1890-1910 by using monthly climate factors (delta change) based on the relation in corrected rainfall between the two periods.

Around the year 1900, precipitation more often fell as snow, the annual correction level was generally higher at 22.9% per year compared to 15.0% for the reference period, all seasons were colder, and corrected rainfall amount was smaller. Around the year 1900, rain was measured with a Fjord rain gauge, and snow with a special snow gauge. In the 1910s, these gauges were replaced with the Hellmann gauge. The wind effect for the Fjord and snow gauge is unknown but is assumed to be the same as for Hellmann. It has not been possible to evaluate the significance of this difference.

Verification of climate data shows that monthly and annual values of temperature and precipitation are reasonably well in line with official climate values, but with a larger uncertainty at regional / local level. It is difficult to verify wind speed, but it is known that manual observation can be a source of uncertainty. In a parallel project manual wind speed observations have been corrected for homogeneity problems, and a trend adjustment has been applied using geostrophic wind speed. Based on uncorrected wind data, delta change was calculated at 0.931 per year (773 mm yearly corrected rainfall which is 60 mm less than 1989-2010). The corrected wind data show that the uncorrected wind around year 1900 is probably overestimated resulting in a 2.7 % too large value of delta change corresponding to 20 mm / year, which is at a reasonable level given the different assumptions and uncertainties.

# Preface

This project is financed by the Danish Ministry of Environment and food and is a part of a larger project investigating the transport of nutrients from land to the sea around Year 1900. The overall report will be available in 2021, but as the results on climate variables around year 1900 are used in various other projects, e.g as input to the marine models used in River Basin Management Plans 2021-2027, there is a need to provide documentation of the methods and results before that.

This research note present best estimates of spatially distributed data on temperature, wind speed and precipitation around year 1900. These data are used in the overall project as input to the DK-model, that model water runoff to the sea.

The data presented in this research note have been collected and analyzed during the main project and will therefore also be included as a chapter in a final report on the transport of nutrients from land to the sea around year 1900.

The final report will however undergo international peer review, which might result in modifications to the current text.

# 1 Introduction

The work described in this report is a part of a project with the purpose to calculate water balance and discharge of nitrogen and phosphorus to open sea areas in Denmark around year 1900, i.e., the period 1890-1910. The objective of this study is to provide this project with spatially distributed estimates of required meteorological variables to support the hydrological modelling of water discharge. The content of the present report will be included as a chapter in a final report on the transport of nutrients from land to the sea around year 1900. As no discharge data are available before 1917, it is not possible to validate modelled discharge around year 1900. Therefore, the period 1917-1950 is chosen as a validation period, and it is assumed that the condition during this period is comparable with the period around year 1900.

A limited number of digitized daily or monthly data in DMI's database has motivated digitization of a large number of climate data. The specific goal of the climate data activity is:

1) to collect and digitize historical meteorological data 1890-1950,
2) to develop an approach for bias correction of historical rain gauge data, an
3) to establish data series of bias corrected rainfall, air temperature and wind speed for a suitable number of inland stations evenly distributed in Denmark, also with data for Southern Jutland before 1920, and
4) to validate uncertainty of calculated climate variables around year 1900, and to compare results with official monthly and yearly climate values from the Danish Meteorological Institute.

Based on these data, 10×10 km$^2$ fields of bias adjusted precipitation and 20×20 km$^2$ temperature fields are established for calculation of evaporation which is climate data input to hydrological model run for the period 1914-1950 for evaluation of water balance (description of method and results will be published in the final project report). Finally, the goal is to refine or develop a new approach for bias correction of rain gauge undercatch if the modelled water balance is not sufficiently accurate.

### Measurements of precipitation

The measurement of precipitation is recognized as a challenging problem and uncertainties on rain gauge measurements have been widely reported. Precipitation measurements are affected by systematic errors, which lead to underestimation of the true precipitation for rain (e.g. Sevruk, 1979) and, especially, for snow (e.g. Groisman and Legates 1994; Yang et al. 1995). Field experiments have shown that wind speed is the most important factor for this undercatch on rain gauge observations (e.g. Sevruk and Hamon, 1984).

The interaction between a rain gauge, the wind flow and precipitation particles falling through the air is complex. The undercatch is related to the deformation of the wind field due to the geometry of the rain gauge (Sevruk et al, 2009), and the wind flow around the gauge opening is getting more turbulent causing a proportion of the particles not to be collected by the gauge. The installation height of a rain gauge is important due to the logarithmic wind profile, as was already demonstrated in the 1700s (Herberden, 1769).

The design and geometry of a rain gauge is of great importance for its aerodynamic properties and ability to measure precipitation, e.g., shown by wind tunnel experiments (Nespor, 1993), and modelling of the air flow around rain gauges also showed systematic differences related to the gauge geometry (e.g., Colli et al, 2018). Wind tunnel experiments have shown that even the opening area of a gauge and the length of the rim may affect the undercatch (Sevruk, Hertig and Spiess, 1989). Turbulens around the gauge opening is the primary cause of undercatch and it is not surprising that rain gauges of different construction installed at the same site frequently observe different precipitation amounts.

Precipitation measured by a rain gauge is subject to other systematic errors such as evaporation and wetting the magnitude of which depends on the type of rain gauge (Sevruk and Hamon, 1984). Evaporation loss is the water lost before the observation is made, and wetting loss is water adhered to the inner walls of a rain gauge that is subject to evaporation after a precipitation event and from the gauge after its emptying. The wetting loss from manual gauges varies with precipitation type, gauge type and the number of emptying times (Goodison, Louie and Yang, 1998).

Precipitation is an important parameter in hydrological modelling and studies of the water cycle, and sustainable water balance monitoring requires availability of accurate precipitation data. It is necessary to apply correction for the different losses on measured precipitation to acquire reliable calculations of the water balance (e.g., Plauborg et al, 2002).

Organised by the World Meteorological Organisation (WMO), great efforts are made to conduct field tests and establish models for correcting sources of error on measured precipitation. Three international rain gauge intercomparison projects have been carried out with measuring fields established in many countries. Different approaches have been developed for undercatch correction of liquid (Sevruk, 1982, Sevruk and Hamon, 1984) and solid precipitation (Goodison, Louie, Yang, 1998). DMI participated in the second international rain gauge comparison on liquid precipitation (1969-1984) and the third one on solid precipitation (1985-1998). A comprehensive correction model was developed which elegantly combines sub-models for rain, sleet, and snow in one and the same equation (Allerup, Madsen and Vejen, 1997) by which bias correction is conducted if information is available about wind speed, rain intensity, dry air temperature and proportion of precipitation fallen as snow.

An important part of the work is to apply bias correction to the observed precipitation, but several data required for this was not available around year 1900 such as shelter information at rain gauge stations and rain rate, wind speed was manually observed, and the observation frequency was low. A method is developed to come around these challenges. The corrected precipitation estimates that are included in the hydrological modelling are based on a range of meteorological data including manual wind speed as it was observed. Subsequently, it has become possible to correct the wind speed for a number of errors. The improved wind data has finally been used to examine the sensitivity of the corrected precipitation estimates.

# 2 Data and methods

## 2.1 Data

Monthly wind speed and daily temperature data for the period 1890-1950, and daily rain gauge data for the period 1913-1950, was published in analogue form in monthly or annual weather reports (DMI 1890-1950), and it is assumed that these data have been subject to quality assurance before publication, even though no documentation for methodology is found. Results based on these data will later in the report be evaluated against official climate values. Opposite to this, rainfall data from 1890 to 1913 are in the form of original observer reports (Rigsarkivet). From experience at DMI it is known that data of this form has not been subject to quality control. Quality assurance of such a huge amount of data is an extremely large task which is beyond the scope of the project, so only simple checks have been carried out, i.e., values exceeding certain thresholds are automatically flagged as suspicious or in error. It is assumed, that after this simple quality control these data are of required quality.

Since some of the stations are not registered in DMI's metadata database, the approximate position of the stations could only be determined from the station name. There is therefore a minor uncertainty in the position of some of the stations. It is considered that this is acceptable, as the alternative would be exclusion of data and increased uncertainty of estimated precipitation fields.

Daily values of maximum and minimum temperature for a limited number of climate stations[1], and daily values of measured precipitation for many rain gauge stations[2] have been digitized. It has not been possible within the resources of the project to digitize daily wind speed. Instead, monthly mean values have been digitized for a minor number of wind stations.

While the precipitation stations are evenly distributed, the 8-12 wind speed stations are primarily coastal stations. Until 1913 there are only 8 stations with wind speed, including one inland station. Later, the number increases to 11-12, of which approx. 6 inland stations are available. See example for 1900 and 1930 in Figure 1. Stations with temperature measurements are evenly distributed with both coastal and inland stations, and the number of stations is for almost the entire period approx. 20 with up to 9 inland stations (Figure 1). The number of digitized precipitation stations increases from approx. 80 in 1890 to about 130 in 1913 including stations from Southern Jutland, but already in 1919 there are about 270 stations, a number that is almost constant until 1950. In **Figure 2** is shown the spatial distribution of precipitation stations in the two periods 1890-1910 and 1914-1950. In the period 1890-1910 there is a lack of precipitation data in the northern part of Jutland, and there are regions with a relatively sparse rain gauge network. The rain gauge density is much higher in the period 1914-1950, but the network is slightly in-homogeneous with smaller areas of lower coverage.

[1] Up to 15-16 climate stations.

[2] Number of rain gauge stations increasing from approx. 80 in 1890 to 270 after 1919.

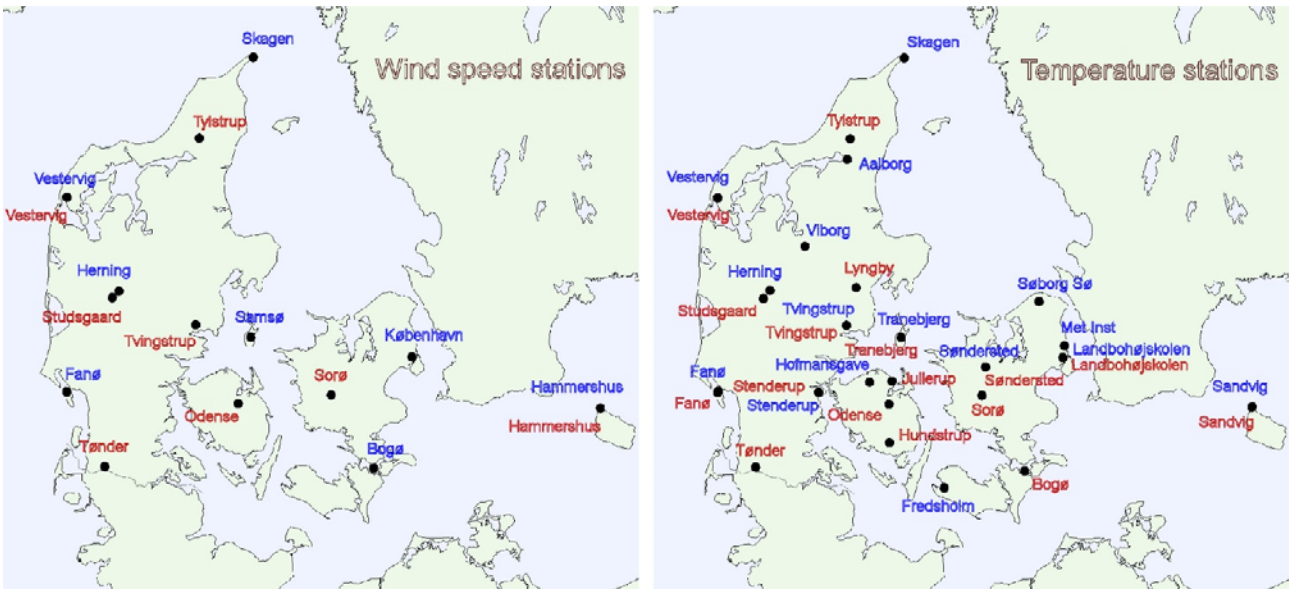



**Figure 1.** Stations with wind speed and temperature in 1900 (blue) and 1930 (red).

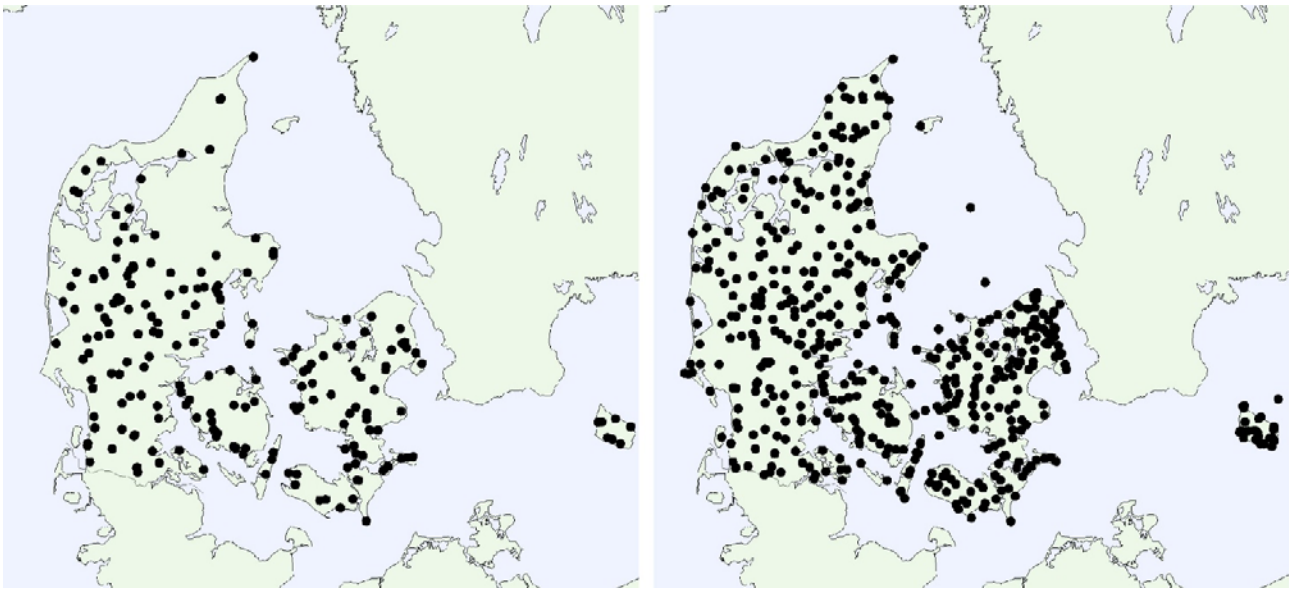


**Figure 2.** Stations with precipitation in the period 1890-1910 (left) and 1914-1950 (right).

Precipitation amount was measured with the so-called Fjord rain gauge up to the 1910s for measurement of rain, and for snow measurements, a zink bucket was used (hereafter called snow gauge). In the period 1910-1925 there was a gradual replacement with the Danish Hellmann gauge (Brandt 1994), which was the cornerstone of the DMI's rainfall network until 2011. For many years, it has been standard to measure the precipitation 1.5 m above ground level with the Hellmann gauge. However, observation practice in the beginning was to place the Fjord and snow gauge at 2 m height (DMI 1875). It is not known how long time this practice lasted, but it can reasonably be assumed that it gradually stopped during the transition to the Hellmann gauge.

The wind speed V is visually observed throughout the period. Until the end of 1910 it is given by the so-called Danish Land scale which has 7 levels (0-6) (Kristensen and Frydendal 1991). Hereafter, the well-known Beaufort scale (level 0-12) is used. At Meteorological Institute in Copenhagen the wind speed is given

in m / s for the period 1890-1893. When correcting for wind induced bias on precipitation measurement, V must be given in m / s. Therefore, a method is developed for conversion of Danish Land scale and Beaufort to m / s.

The official WMO Beaufort scale was adopted in 1946 and is called the WMO Code 1100. It is based on a scale defined in 1906 (Kaufeld 1981). Based on a large set of corresponding Beaufort observations and anemometer measurements of wind speed, WMO (1970) reported that this scale is biased; (1) at wind speeds > 8 (Beaufort), the scale indicates too high wind speeds, and (2) at wind speeds < 5 (Beaufort), the scale underestimates by 10-20%. These results are based on wind field analyzes over oceans. Therefore, a new scale CMM IV has been proposed. Officially, the WMO holds on to the WMO Code 1100 scale but recommends using the CMM IV scale in scientific studies (WMO, 1970). Intervals in m / s for both the two Beaufort scales and the Danish Land scale in Kristensen and Frydendal (1991) are shown in Table 1.

**Table 1**. Definition of the Beaufort scales WMO 1100 and CMM IV (WMO, 1970), and the Danish Landscale (Kristensen and Frydendal, 1991). Lower limit of each level is given in m/s.

| **Level** | **0** | **1** | **2** | **3** | **4** | **5** | **6** | **7** | **8** | **9** | **10** | **11** | **12** |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Beaufort WMO 1100 | 0 | 0,3 | 1,6 | 3,4 | 5,5 | 8,0 | 10,8 | 13,9 | 17,2 | 20,8 | 24,5 | 28,5 | 32,7 |
| Beaufort CMM IV | 0 | 1,4 | 2,8 | 4,6 | 6,7 | 9,0 | 11,4 | 13,9 | 16,5 | 19,3 | 22,5 | 26,1 | 30,1 |
| Danish Landscale | 0 | 2 | 6 | 10 | 15 | 20 | 30 | | | | | | |

## 2.2 Correction of rainfall measurement for wind induced bias

Many models have been proposed for correction of rain gauge observations for wind induced bias (e.g., Goodison, Louie and Yang 1998), but for correction of rainfall measurement in Denmark, the so-called "comprehensive model" is used, which elegantly combines the correction of solid, mixed, and liquid precipitation into one and the same model (Allerup, Madsen and Vejen 1997). According to this model, the correction factor $K_\alpha$ is given by:

$$K_\alpha = \alpha \cdot k_s(V,T) + (1-\alpha) \cdot k_r(V,I)$$

Here, $K_a$ = correction factor for the precipitation type $a$, where $a$ = an index indicating the proportion of precipitation fallen as snow. $k_s$ = correction factor for snow which is a function of the average during precipitation of wind speed V and air temperature $T$, while $k_r$ = correction factor for rain that is a function of V and rain rate I. In the exact expressions seen below, $\beta_0$, $\beta_1$, $\beta_2$, $\beta_3$ for snow (Allerup, Madsen and Vejen, 1997) and $\gamma_0$, $\gamma_1$, $\gamma_2$, $\gamma_3$, $c$ for rain (Allerup and Madsen 1980, Førland et al 1996) are empirical constants (values not shown here, but see the references):

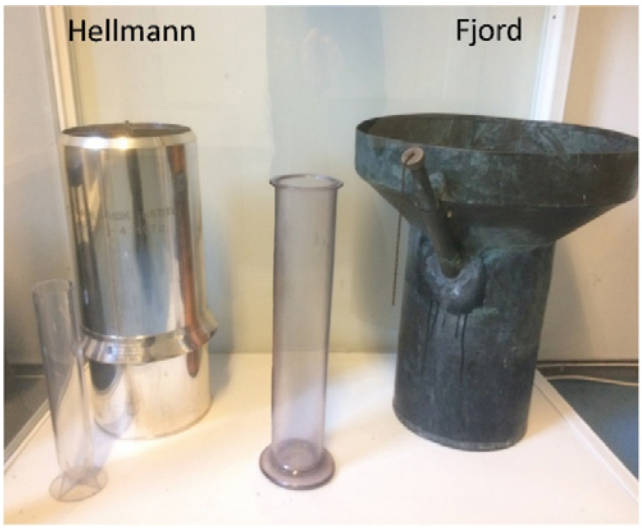
$$k_s = e^{\beta_0 + \beta_1 \cdot V + \beta_2 \cdot T + \beta_3 \cdot V \cdot T} \text{ and } k_r = e^{\gamma_0 + \gamma_1 \cdot V + \gamma_2 \cdot \ln I + \gamma_3 \cdot V \cdot \ln I + c}$$

It is not possible to calculate $V$ and $T$ during precipitation with the time resolutions available in the old data. Furthermore, daily values of V are not available in the digitized data set. Therefore, it is assumed that monthly values of $V$ and daily values of $T$ can be used, although this may give rise to systematic errors or larger uncertainty. For each rain gauge station, $V$ is interpolated from the network of wind stations (see later).

The design of a rain gauge is of great importance for its aerodynamic properties (Sevruk and Klemm 1989a), and the empirical constants are related to the type of rain gauge. The design of the Fjord and snow gauges is quite different from the Hellmann gauge, e.g., the collecting area is larger (1000 $cm^2$ opposite to 200 $cm^2$ for Hellmann) and the shape is different. In fact, their measuring ability is unknown. It is assumed that the two gauges have the same aerodynamic properties as Hellmann, and the same empirical constants as for Hellmann are used. Figure 3 shows the probably only existing copy of a Fjord rain gauge, and a drawing of the snow gauge, together with a Hellmann gauge.

## 2.3 Wetting and evaporation

The Hellmann gauge is designed so that the evaporation loss from the interior of the gauge is negligible, i.e., in the order of 1,5-2,0 mm per year. The undercatch for the Fjord gauge is assumed to be the same as for Hellmann, although its design probably causes a larger evaporation loss than for Hellmann. The average wetting loss is determined for the Hellmann gauge as a function of season and precipitation type in mm / day (Allerup and Madsen 1979, 1980, Vejen, Madsen and Allerup 2000, Elomaa FMI (Finnish Meteorological Institute) pers. komm.). It is assumed that these tabular values can be used for the Fjord gauge, although its wetting properties are unknown.



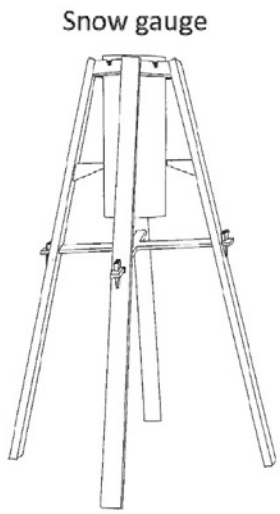




**Figure 3.** The Fjord rain gauge and the snow gauge were replaced by the Hellmann gauge during the period 1910-1925 (Brandt 1994), Photo: F. Vejen. Drawing: Brandt (1994).

## 2.4 Rain rate

In the rain gauge data 1890-1950 there is no information on rain rate, *I*. It is assumed that seasonal-dependent climatological values of *I* (Table 2), which are based on measurements of precipitation in Denmark over the period 1959-1974 at 4 stations (Madsen and Allerup pers. comm.), are representative of 1890-1950. It may be a source of uncertainty that changes in rain climate may have affected the level and frequency of rain intensities.

**Table 2**. Climatological values of rain rate [mm/h], *I*, based on measurements at 4 weather stations in Denmark over the period 1959-1974 (tabular values from Madsen and Allerup, pers. comm.).

| Month | J | F | M | A | M | J | J | A | S | O | N | D |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Rain rate | 1,12 | 1,21 | 1,18 | 1,38 | 2,01 | 2,46 | 3,01 | 2,90 | 2,26 | 1,71 | 1,37 | 1,26 |

## 2.5 Conversion of Danish Land scale and Beaufort wind speed to m/s

The model for bias correction of rainfall requires that the wind speed, *V*, is given in m/s at rain gauge level. *V* is the parameter that has the largest effect on rainfall bias, so it is particularly critical to solve the challenges with the manual observations of wind speed. Visual observation of *V* in Beaufort or Danish Land scale is primarily based on a human assessment of the wind's impact on smoke, flags, pennants, windmills, buildings, and vegetation. It is therefore uncertain which height above ground level *V* represents. Moreover, individual differences in the observers' experience will have an impact on the observations of *V*. To apply this type of observation, solutions to the basic challenges must be found; (1) conversion to m/s, (2) determining which height above ground level *V* represents, and (3) quantification of observation inhomogeneity.

The CMM IV and Danish Land scale have been analyzed to create a general model for conversion to m/s. It can be seen from Figure **4** that a 3rd order polynomial fits very well to these scales, where $a_1$, $a_2$, $a_3$ and $a_4$ are constants:

$$V_{m/\sec} = \alpha_1^3 V + \alpha_2^2 V + \alpha_3 V + \alpha_4$$

It is assumed that the model can provide a reasonable conversion between CMM IV and Danish Land scale and m/s for any value of *V*. Figure **4** shows the WMO Code 1100 scale to illustrate the over- and underestimation with this scale.

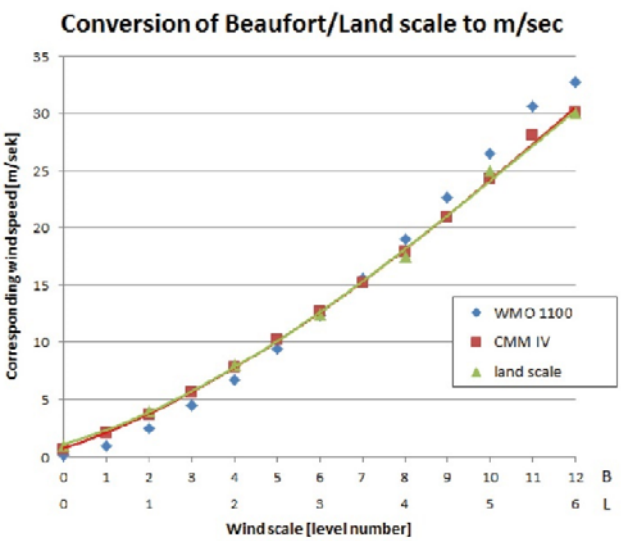



**Figure 4.** Conversion of Beaufort scale CMM IV, and Danish Land-scale to m/s. Red line: model for conversion of Beaufort CMM IV to m/s. Green line: model for conversion of Landscale to m/s. Blue, red and green dots shows interval center of each wind scale number. B = Beaufort level numbers, L = Landscale level numbers.

## 2.6 Control of observed wind speed for representativeness and homogeneity

Since the wind speed is manually observed there is no guarantee that data reflects the correct wind speed or that the long-term variation, or trend, is correct. Furthermore, it is uncertain what height above the terrain the observed wind speed represents. This may be critical since it is important for the bias correction of observed precipitation to know the measurement height of wind speed. It is not possible to control manual data against independent wind observations since parallel anemometer data do not exist for the period 1890-1950 and is generally not available until after 1953.

Figure 5 shows a comparison between Herning/Studsgaard and Karup, and the general wind conditions in the transition between the two periods 1890-1950 and 1953-2018 appear to be approximately the same. A slight difference between the two periods is seen which may be explained by the distance between the two sites. The same pattern is found at some but not all of the other stations. The difference in average wind speed at the transition between the two periods is close to a factor 1. If the offset is accounted for it is assumed that *V* in Beaufort and Danish Land scale represents 10 m above ground level at least for this station.

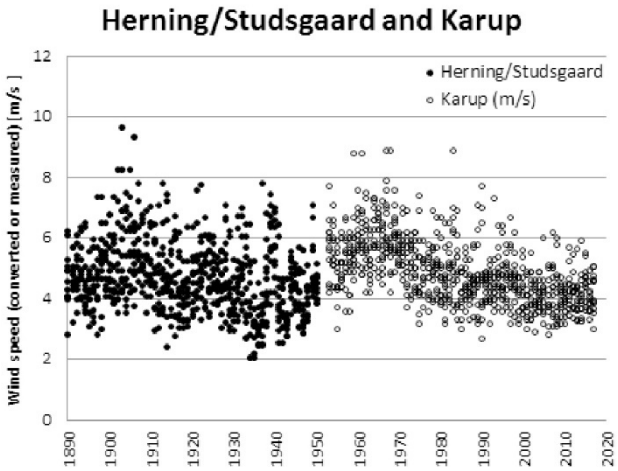



**Figure 5**. Comparison of monthly wind speed converted from Beaufort CMM IV for Herning/Studsgaard 1890-1950 (black) against wind speed measured with an anemometer (m/s) at Karup 1953-2017 (blue).

However, there are stations with a larger offset and one or more homogeneity breaks. Sequences of clear homogeneity breaks in the manual data series were identified for some of the stations. Homogeneous data series were found difficult to identify, and it may be too uncertain to use any of the series as reference data set for homogeneity control of other series.

A simple control of the reliability and subsequent adjustment of each wind series is conducted by; (1a) searching for abrupt changes in wind speed, (1b) data sequences with one or more well-defined and abrupt breaks are adjusted up or down to fit the wind level before and after the break, (2a) comparing with more recent anemometer data series from nearby weather stations to identify and quantify possible offsets in the general wind speed level compared to wind data after 1950, and (2b) adjust the series using an offset factor calculated from the average homogenized wind speed before and after 1950.

The justification of (2) assumes that the overall wind climate in the period before and after 1950 is approximately equal. Only few studies of historical wind climate in Denmark were found in the literature. The wind climate in the Limfjorden for the period 1897-2003 is reported in Christensen et al (2006), and the distribution of geostrophic wind velocity[3], $V_g$, reported for five 25-year periods over 1874-1987 generally shows only slight variations (Christiansen et al 2006). It is difficult to conclude on systematic trends based on their results. In Kristensen and Frydendal (1991), frequencies of exceedances of Beau-

[3] The geostrophic wind is the theoretical wind that would result from an exact balance between the Coriolis force and the pressure gradient force. The true wind almost always differs from the geostrophic wind due to other forces such as friction from the ground. Within certain uncertainties, geostrophic wind and surface wind are related to each other.

fort 6 and 9 have been calculated on a national level, but it cannot be determined from the results whether there is a systematic difference in wind speed in the periods before and after 1950.

Figure 6 shows examples of correction of wind series from Bogø. Clear homogeneity breaks are seen in 1910 and 1953, and an offset in the general wind level compared to anemometer series from Avnø are identified. The station environment of Avnø and Bogø is quite similar while the station on Omø is more exposed to the wind, thus Avnø is considered more representative of the station environment on Bogø. After correction for homogeneity break and offset has been applied the total series seems more consistent and fits quite well into the general level at the nearby station in Avnø.

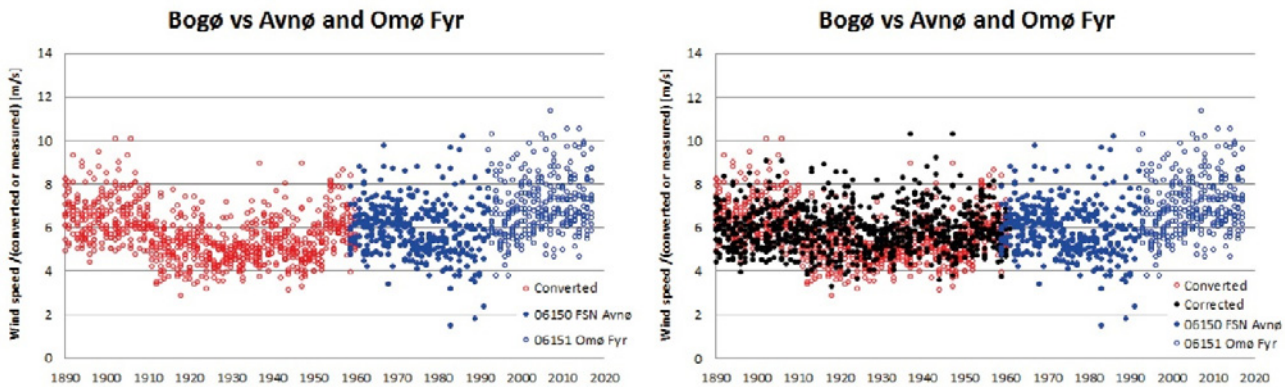



**Figure 6**. Left: raw monthly wind speed for Bogø converted from Beaufort CMM IV (red dots). Right: wind speed corrected for breaks and offset for Bogø (black dots). Wind speed for Bogø is compared with nearby anemometer measurements (m/s) at FSN Avnø (blue dots) and Omø Fyr (open blue circles).

Generally, the break and offset corrected values of $V$ are assumed to represent 10 m above ground, and $V$ is adjusted down to the height of the rain gauge (1,5 and 2,0 m) using the logarithmic wind law as recommended by the WMO (2008). Since there is no information on the actual height of the rain gauges at the individual stations, it is assumed that the Fjord and snow gauge were generally installed with the orifice at 2 m height[4] above ground level until 1915, and then replaced by Hellmann at 1.5 m height. There is no information on the properties of the ground surface, e.g. if there is snow lying or a grass cover, thus instead of the roughness parameter, the effective roughness length is used in the wind law in the form of a general value of 0.25 similar to that used in Refsgaard et al (2011).

Over the period 1890-1950, discrepancies can be found for some stations between the manual observations and the general trend of wind speed indicated by $V_g$, but it has been necessary to assume that wind data reflect the true local wind conditions, although this may not necessarily be correct for certain subperiods and series.

A more advanced method for controlling and possibly adjusting the old wind data would be to use $V_g$ as an independent data source for general trends in wind speed, but it was not possible within the resources of the project to develop a method for trend correction using $V_g$. After completion of the correction of rainfall and hydrological calculations, analyses and trend correction of

[4] Recall, that observation practice was to measure precipitation at 2 m level (DMI, 1875).

wind series were carried out in another project[5] and made available for this project. The results are used as supplement to evaluation of wind speed and corrected precipitation (3.6).

## 2.7 Correction for shelter conditions

It is common practice to adjust the wind speed for local shelter conditions. The shelter adjusted wind speed, $V_{shlt}$, is found by adjusting the wind speed at gauge height $h$, $V_h$, by the expression $V_{shlt}=\lambda V_h$, where $\lambda$ is a shelter correction factor given by $\lambda=1-c\cdot\eta$ (Sevruk 1988, WMO 2008). Here, $\eta$ = the average height angle to the top of the shelter given in degrees, and $c$ = a constant ($c$=0,024).

At DMI, the shelter conditions are indicated by height angles in eight compass directions, $\eta_I$, where $i$ = {1, … ,8} corresponds to the wind directions N over S to NW. A weighted average height angle, or shelter index, $\eta_{wgt}$, is then calculated by weighting each $\eta_I$ with the statistical frequency of winds from direction $i$.

DMI first began to measure height angles at rainfall stations in the 1960s. Since shelter conditions can have great significance for the local wind speed and thus the correction level, it has been necessary to make assumptions about height angles for the period 1890-1950.

Already in the 19th century, the wind's effect on precipitation measurements and the importance of shelter was well known (Brandt 1994). It is therefore assumed that in the period 1890-1950 the same practice was used as today to ensure good shelter conditions at rain gauge stations (neither too open nor overprotected).

## 2.8 Adjustment of wind speed to represent days with precipitation

A monthly wind speed value, $V_m$, probably does not represent the conditions during precipitation periods since it includes dry days as well as days with stable weather and low wind speeds. Presumably, the monthly wind speed value for precipitation days $V_{mw}$ is different from and higher than $V_m$, thus use of $V_m$ in the comprehensive bias adjustment model probably will result in too low corrections.

It is examined using data from 1989-2010 whether there is statistical evidence for conversion of $V$ to a value that is more representative for rainfall days. Monthly values of $V_m$ and $V_{mw}$ were calculated based on daily values of $V$ weighted by the amount of precipitation. The results show (Figure 7) that monthly values, $V_m$, can be corrected for bias and converted to $V_{mw}$ using a seasonal dependent adjustment factor, or ratio, $k = V_{mw}/V_m$. A seasonal dependent tendency has been found that k decreases with an increased number of precipitation days. Even though this relationship is noisy, it may make sense to incorporate the number of monthly precipitation days in the calculation of k in future improvements of the method. This was not done in the present project.

[5] Supported by DHI.

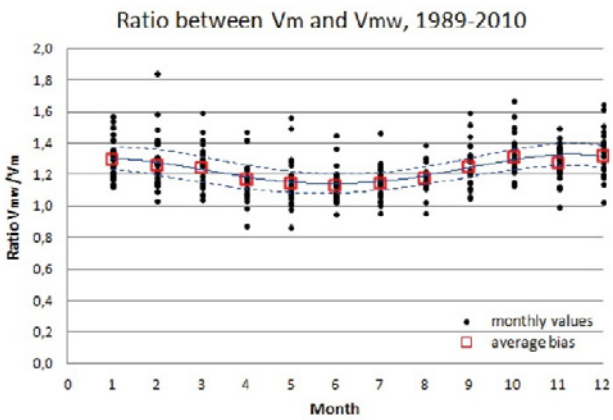



**Figure 7**. Calculation of bias between monthly value of *V* for all days, $V_m$, and monthly value for rainfall days, $V_{mw}$. The black dots show the value for the individual months, the red symbol shows the mean bias, while the dotted line shows the 1x standard deviation of the monthly values of the ratio *k*. The analyses are based on wind and temperature data for the period 1989-2010.

## 2.9 Calculation of average temperature and precipitation type

Daily observations of precipitation type t are not available in the historical data set. A widely used model for the determination of t uses the air temperature $T_a$ as an indicator of precipitation type, assuming that $t$ = snow if $T_a \leq 0$ °C, $t$ = rain if $T_a > 2$ °C, and otherwise sleet, although this method may cause bias between estimated and observed t (e.g., Feiccabrino et al 2013). $T_a$ is calculated as daily average temperature, $T_a = T_{avg} = (T_{max}+T_{min})/2$.

It has been argued that a more realistic probability function for t can be obtained by including air humidity (e.g., Harder and Pomeroy 2013), but this parameter is not available, so only the simplified model is possible. Snow index $a$ is calculated from $T_a$ by $a$=1 for snow, $a$=0 for rain, and $a=-0.5\ T_a +1$ for mixed precipitation.

## 2.10 Calculation of 20×20 km$^2$ fields for wind speed and temperature

For consistency reasons, the same interpolation principles are used as in KlimagridDK, where the weighting is calculated in relation to $1/r^2$. Here, $r$ = the distance between a grid point and a weather station (Scharling 1999).

For the period 1890-1950, however, the number of stations with wind and temperature data is quite limited, and especially for wind there are very few inland stations. Since it is very uncertain to calculate spatial distribution of *V* and *T* from so few stations, the interpolated values of *T* and *V* are adjusted with a method inspired by Olesen et al (2000), in which distributed variables are calculated even when the spatial resolution of the observation material is limited. The method combines observations with fields that for each month describe the relative spatial variation of *V* and *T*. These fields are incorporated into the interpolation technique by which daily 20×20 km$^2$ fields of *T*, and monthly of *V*, are calculated.

The established relative normal fields, $\mu$, in a 20×20 km$^2$ resolution are based on data for the period 1989-2017. The relative values, $\mu(i,j)$, are used to adjust interpolated values of temperature, $T(i,j)$, or wind speed, $V(i,j)$, in the formula below given as $F(i,j)$. In practice, the interpolated surface is lowered or raised depending on the pseudo-climatological value in an arbitrary point $(i, j)$. Especially for wind speed, it has the advantage that the interpolation, which is largely based on coastal stations, is forced to lower values inland.

$$F_{(i,j)} = \mu_{(i,j)} \sum_{g=1}^{N} w_{g(i,j)} F_g \Bigg/ \left( \sum_{g=1}^{N} w_{g(i,j)} \right)$$ , where the weight $w_{g(i,j)}$ is given by:

$$w_{g(i,j)} = \left( \frac{1}{r^2_{(i,j)}} \right)$$

Here, $w$ = a weighting function, $g$ = a gauge station $g$, $N$ = number of stations, $F$ = measured (or observed) value of wind speed $V$, or temperature $T$ at station $g$, $F(i,j)$ = value of $T$ or $V$ for a grid cell $(i,j)$, and $\mu(i,j)$ = the weighting function for the relative spatial normal value for grid cell $(i,j)$.

The weighting for the relative spatial distribution is given by:

$$\mu_{(i,j)} = \sum_{g=1}^{N} w_{g(i,j)} \frac{R_g}{R_{(i,j)}} \Bigg/ \left( \sum_{g=1}^{N} w_{g(i,j)} \right)$$

Here, $R_g$ = the relative climate value at gauge station of temperature $T$ or wind speed $V$, $R(i,j)$ = the relative climate value of $T$ or $V$ for a grid cell $(i,j)$, and $w$ = the weighting function previously defined.

It is assumed that the relative spatial distribution based on data 1989-2010 is consistent with the period 1890-1950, since the spatial distribution is determined by physical and meteorological factors as well as terrain conditions which are considered relatively unchanged over the period. Even though systematic changes in urbanization and vegetation over the period are seen, the overall meteorological conditions are assumed to be relatively constant. However, the use of monthly normal fields for simulating spatial variations at daily level increases the uncertainty of the interpolated values.

## 2.11 Bias correction of rain gauge observations

Correction of daily precipitation values from a specific rain gauge station uses $V$ and $T$ from the nearest 20×20 km$^2$ field. The snow index $\alpha$ is estimated by $T$. It is assumed that within a certain uncertainty margin interpolated and climate-weighted values of $V$ and $T$ represent the weather conditions at the precipitation stations, and that this approach therefore does not cause any significant systematic bias on the adjusted precipitation.

## 2.12 Calculation of 10×10 km$^2$ fields of observed and bias corrected precipitation

Almost the same interpolation method as previously described is used as for wind and temperature. The only difference is that rainfall is not adjusted with a relative normal field. The interpolated field value for precipitation sum, $P$, is given by:

$$P_{(i,j)} = \sum_{g=1}^{N} w_{g(i,j)} P_g \Bigg/ \left( \sum_{g=1}^{N} w_{g(i,j)} \right)$$ , where the weighting function $w_{g(i,j)}$ is given by: $w_{g(i,j)} = \left( \frac{1}{r^2_{(i,j)}} \right)$

# 3 Analysis and results

This section analyzes and evaluates the calculations of temperature, wind and precipitation, and the basic assumptions for bias correction of precipitation. The evaluation is a challenge as there is no independent data for test of the daily results. However, there are official national values of the meteorological variables (DMI database), thus the evaluation can be done by examining whether the interpolation of the meteorological variables can roughly reproduce the official monthly and yearly national climate values over the period.

At the Danish Meteorological Institute, official national values of $T$ and $P_m$ was previously (from the 1950s up to 2006) calculated as a simple average of station values with data from Jutland weighted with 7/10 and data from the islands with 3/10, and after 2006 based on grid interpolation of station data (Cappelen 2019). Before the 1950s the methods used have not been published. Since grid values calculated during this project are spatially distributed, in contrast to the official national values before 2006, minor differences between official values and grid calculations are expected. The possibility of assessing the calculations of $P_m$ and $T$ relative to climate values are greater than for $V$, since values for $K_m$ and $T$ are historically based on a larger network of stations than for $V$, e.g., the official normal for $T$ for 1886-1925 is based on 30 evenly distributed stations (Det Statistiske Departement 1964). The normal, or pseudo normal, for V 1931-1960 and 1961-1990 is based on a relatively limited number of coastal stations (Lysgaard 1969, Cappelen 2000), and official national values are not found.

In calculation of national averages of the meteorological variables in this project, grid cells in coastal regions are given lower weight depending on the fraction of land area.

## 3.1 Evaluation of temperature

The monthly grid values of $T$ are generally close to the official values (Figure 8). Table 3 provides details of the regression between the two data sets. While the official values are calculated based on many stations available at that time (Cappelen, 2019), the grid method is based on a considerably smaller number of stations. The interpolation method can reproduce the official monthly values with a high correlation, but it should be noted that grid values before 1920 are biased towards higher values, and for the period 1920-1950 towards slightly lower values as illustrated in Figure 8. The explanation of the jump in the general temperature level is that the relatively warm southern part of Jutland was not a part of Denmark before 1920, but it is included in the estimates of grid temperatures.

Despite the difference in calculation methods, the results are reasonable on a monthly level and the grid values of $T$ are generally well in line with the official values except that the lowest $T$ values are slightly biased towards lower values. On the other hand, the official values are assumed to be the truth even though evaluation of these values is not available. However, larger uncertainties are expected on a daily level because pseudo-climatological values are used in the interpolation method.

**Table 3.** Statistics on gridded and official national values of temperature $T$ for the two periods 1890-1919 and 1920-1950; bias of $T$ given as $T_{bias} = \sum(T_{grid} - T_{official})$, absolute bias given as $T_{bias} = \sum(|T_{grid} - T_{official}|)$.

| | *Year* | | *Month* | | |
|---|---|---|---|---|---|
| **Period** | $T_{grid}$ | $T_{official}$ | $T_{bias}$ | $T_{absbias}$ | $R^2$ |
| 1890-1919 | 7,547 | 7,449 | 0,098 | 0,263 | 0,998 |
| 1920-1950 | 7,730 | 7,785 | -0,055 | 0,228 | 0,998 |

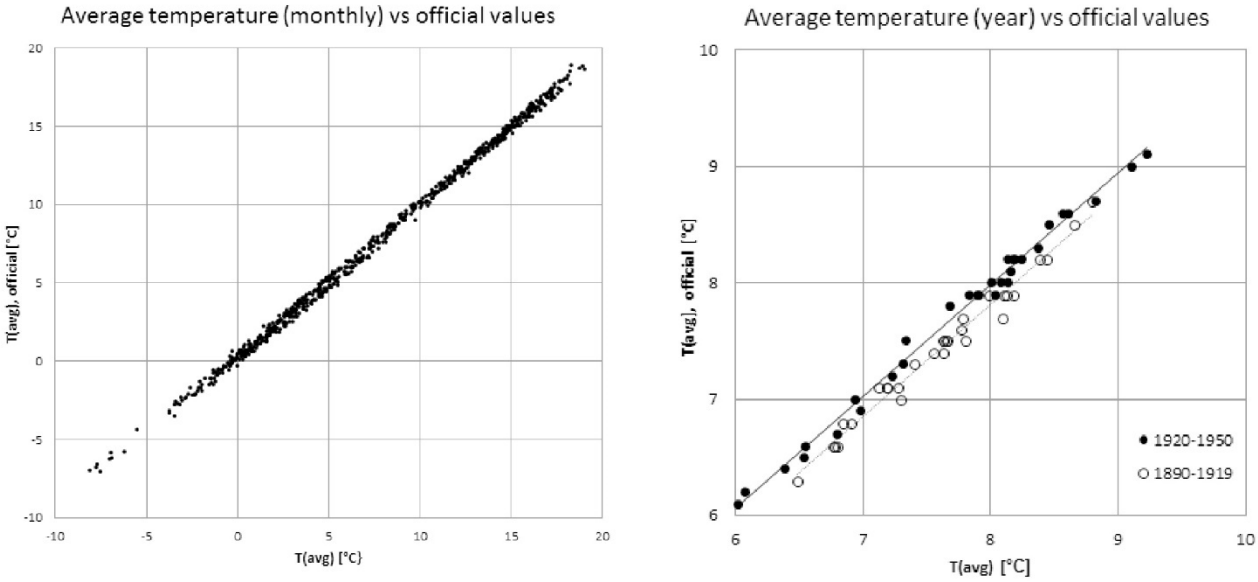
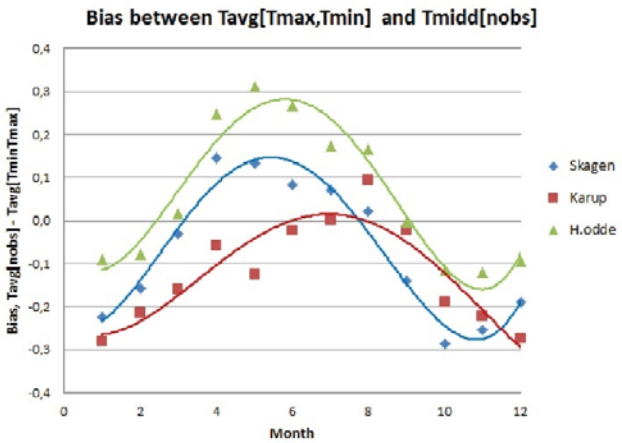



**Figure 8.** Left: national monthly grid values of temperature vs official climate values. Right: national yearly grid values of temperature vs official climate values

The method for calculating daily mean temperature probably gives rise to biased estimates. The daily variation of temperature depends on the solar radiation, which varies with the season, and distance to the sea for a given location. The method used for calculation of daily averages in this project, i.e., $T_{avg}=(T_{max}+T_{min})/2$, is evaluated using temperature data for the period 1989-2010. It is shown that compared to a mean temperature based on 24 daily observations, the method results in a seasonally dependent bias, at least for the analyzed stations shown in Figure 9.

For bias adjustment of precipitation this temperature bias is of importance only for months with probability of snow. Here, negative bias means that $T_{avg}$ will be slightly too low, and consequently the bias adjusted precipitation will be a bit too high in the winter months. Figure 9 also suggests a difference between coastal (Skagen, H. Odde) and land stations (Karup), which is plausible, cf. that the sea influences the daily variation of temperature. If it is assumed that Karup station is representative for inland stations, $T_{avg}$ will generally be underestimated in onshore areas, especially in the coldest months.

**Figure 9.** Average monthly bias between two methods for calculation of daily average temperature based on station data for the period 1989-2010. The lines for one inland and two coastal stations show the absolute difference between average temperature based on 24 daily observations and average based on daily minimum and maximum temperature.




## 3.2 Evaluation of measured precipitation

Figure 10 shows monthly and yearly gridded values of measured rainfall compared to official values. The scatter is probably due to difference in calculation methods. It is seen from the annual values (Table 4) that for the period 1890-1919 grid values marginally underestimate measured precipitation compared to official values by 0,562 mm per month and with an absolute bias of 2,9 mm.

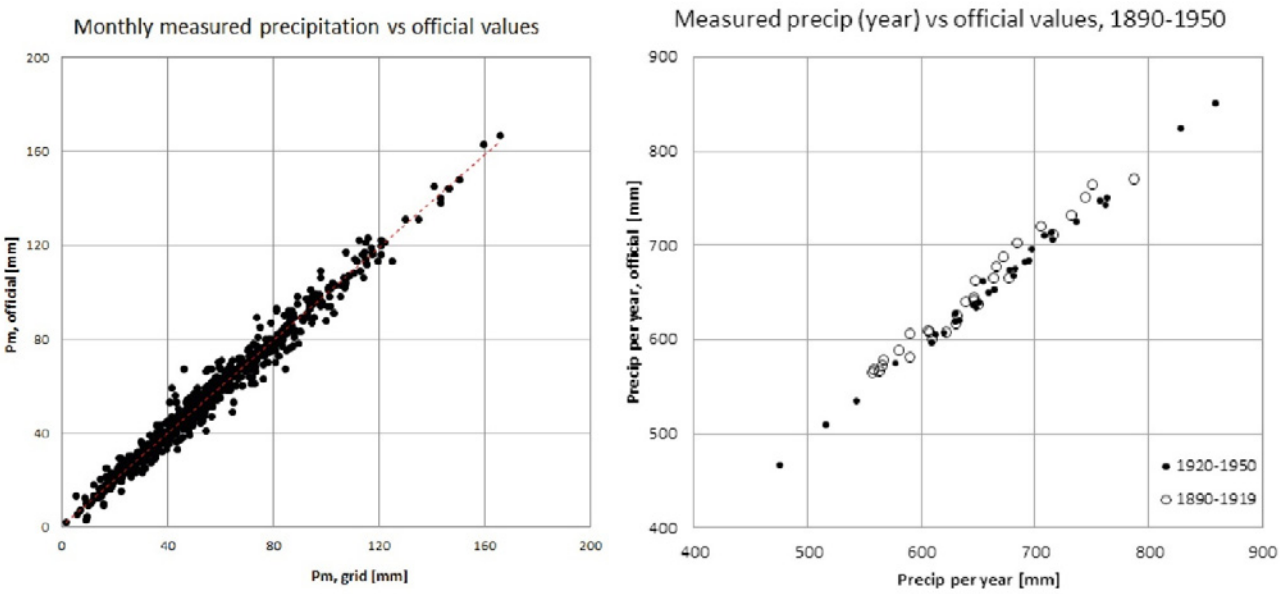



**Figure 10**. Left: national monthly grid values of measured precipitation vs official climate values. Right: National yearly grid values of measured precipitation vs official climate values.

Before 1920 the relatively wet Southern Jutland was not a part of the official values, and it was therefore expected that the grid values would be slightly higher than the official ones which is not seen in the results. For the period 1920-1950 the grid method results in approx. 10 mm more on an annual basis than the official values corresponding to a monthly bias of 0,739 mm and an absolute bias of 2,9 mm per month. The differences between grid values and official climate values may be caused by the different calculation methods, but the effect cannot be verified, as there is no documentation of the methodology of the time (Cappelen, 2019).

**Table 4.** Statistics on gridded and official national values of measured precipitation $P_m$ for the two periods 1890-1919 and 1920-1950. Bias of $P_m$ given as $P_{m(bias)} = \sum(P_{m(grid)} - P_{m(official)})$, absolute bias given as $P_{m(bias)} = \sum(|P_{m(grid)} - P_{m(official)}|)$.

| | *Year* | | *Month* | | |
|---|---|---|---|---|---|
| **Period** | $P_{m(grid)}$ | $P_{m(official)}$ | $P_{m(bias)}$ | $P_{m(absbias)}$ | $R^2$ |
| 1890-1919 | 643,0 | 646,8 | -0,562 | 2,865 | 0,980 |
| 1920-1950 | 670,4 | 659,8 | 0,739 | 2,909 | 0,978 |

In the absence of homogeneity tests for precipitation series, a larger uncertainty must be expected in the calculated rainfall climate around the year 1900, especially as the catch efficiency of the Fjord rain gauge is unknown compared to Hellmann. The number of digitized rainfall series is large (approx. 630) and it would be an insurmountable task to check and homogenize all of these within the project. However, homogeneity control would be particularly relevant for the transition between Fjord / snow gauge and Hellmann. In a study of historical trends in rainfall in Central Jutland, it has been shown from 4 long time series of precipitation that there are up to several homogeneity breaks and that there is only one homogeneous series out of the 4 series in the period 1890-1950 (Karlsson et al 2014). In three of the series there were homogeneity breaks during the period of transition to Hellmann.

## 3.3 Evaluation of wind speed

There are no official national values for monthly or annual wind speed, $V$. Prior to 1953, practically all wind data is given in Beaufort or Danish Landscale, and in the 1950s there are only a few stations with wind speed in m / s, so evaluation can be done only on the basis of a few long time series of $V$. Evaluation is done partly by looking at the continuity of $V$ at the transition from Beaufort to m / s, and partly by looking at the co-variation and trends in manual series in relation to geostrophic wind velocity, $V_g$.

For some of the old data series, the trend for manually observed winds does not perfectly match the general trend calculated as average for the most homogeneous series, at least for parts of the period 1890-1950. Evaluation of the wind data by inter-comparison is probably subject to considerable uncertainty as the homogeneity of many of the series is questionable and difficult to verify. Trend discontinuities and homogeneity breaks in the individual series may partly be explained by different experiences and observation practices of the observers, movement of stations, or by changes in the local environment. The obvious uncertainty of manual wind observations may influence the general estimates. Especially, wind speeds converted from Landscale is probably subject to considerable uncertainty due to the limited number of wind classes (0-6 levels).

A more advanced method for evaluation of the manual wind data is to use $V_g$ as an independent source for general trends and level of wind speed. After correction of precipitation and water balance was finished, external funding[6] made it possible to develop a methodology for trend correction and homogenization of all wind series in the period 1890-1950 using trend information calculated from $V_g$ (the method is developed at DMI, will be published later). Although uncertainty is associated with the calculation $V_g$ and it is not related

[6] DHI, www.dhigroup.com.

to surface winds, $V_g$ is assumed to give a reasonable impression of the general development of the wind climate and trends in the period 1890-1950.

Daily values of $V_g$ were calculated for a long period of time (1874-2018) using methods described in Alexandersson et al (1998). $V_g$ is based on pressure differences between three stations. In the period 1974-1970 the stations Hammershus, Nordby and Vestervig were used, and in the period 1961-2018 various stations were used (Hanstholm, Thyborøn, Kegnæs, Sæddenstrand Fyr, Rømø Juvre, Hammer Odde and Christiansø Fyr). Analyses of $V_g$ were conducted to get a picture of the overall trend in $V$ over the historical period, and to compare trends of $V_g$ with the corresponding Beaufort / Landscale trends.

Care must be taken when comparing geostrophic wind and surface winds. While the calculated $V_g$ represents a point, the surface winds shown in Figure 11 are averages for the whole of Denmark. The geostrophic wind is not a perfect measure of the surface wind, and e.g., stability, friction and gradient winds influence the relation between $V_g$ and the true wind near the ground (see e.g., Luthardt and Hasse 1981). There are also spatial differences between a point value of geostrophic wind and wind at individual stations.

Figure 11 shows monthly values of $V$ for Denmark with and without trend correction for 1890-1950 and values of $V$ for 1989-2010s together with four weather stations where $V$ has been measured with anemometer since 1953. The stations chosen are all inland stations while the grid estimates for Denmark are averaged over 20×20 km$^2$ grid cells. No obvious homogeneity break is seen at the transition from Beaufort to m / s in the 1950s. A comparison between trend corrected and uncorrected national values of 20×20 km$^2$ wind speeds seems to indicate that Danish Landscale (1890-1911) is probably overestimating and the Beaufort scale probably underestimating $V$, at least in parts of the period up to 1950 (Figure 11). This underlines the importance of trend corrections, as the uncertainty of manual wind observations would otherwise deleteriously affect bias correction of precipitation discussed later.

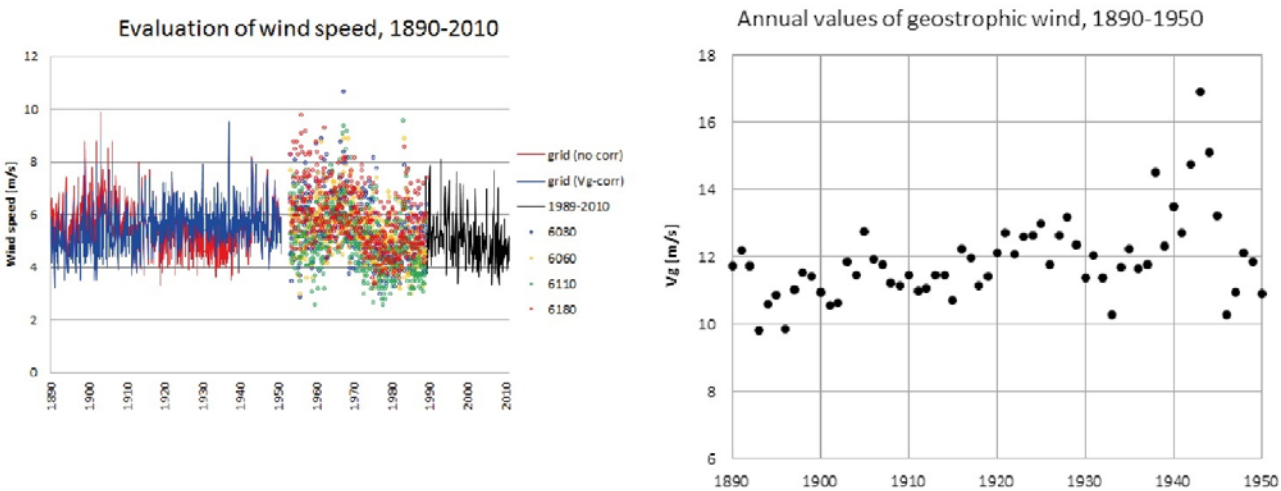



**Figure 11**. Left: monthly wind speed for the period 1890-2010, national averages of 20×20 km$^2$ grid values and selected station values. Red line: grid values not corrected for trends. Blue line: grid values corrected for trends using analyses of geostrophic wind velocity, $V_g$. Black line: grid values for the period 1989-2010. Coloured bullets: station values. Stations are 06030 FSN Aalborg, 6060 FSN Karup, 6110 FSN Skrydstrup, 6180 Copenhagen Airport. Right: annual values of geostrophic wind based on daily pressure data compiled in the EU project WASA (Schmidt et al 1997).

The results of analyzes of homogeneity breaks and geostrophic wind (Figure 5, Figure 6 and Figure 11) show that the trend-corrected winds appear to result in a better indication of the historical wind climate than the initial uncorrected wind data. The indication of gradually increasing geostrophic wind speed during the period is also reflected by the blue line in Figure 11 which is not the case with the red uncorrected line. However, it is important to emphasize that this must not be considered the result as the trend correction method should be subject to further refinement and tests, and manually observed winds be analyzed in more details in future studies to further optimize the conversion to m/s.

## 3.4 Evaluation of shelter assumptions

Of the parameters needed for bias correction of rainfall, the lack of information on shelter conditions is particularly critical, as the wind speed at gauge height must be adjusted by the shelter correction index, $\eta_{wgt}$, which is generally not known at rain gauge station until the 1970s. Assumption of a nationwide constant shelter index may result in significant over- or underestimation of bias corrected rainfall, locally, regionally, or nationwide. Streamflow values are calculated by the national water resource model, the DK model, for a number of catchments using input of bias corrected precipitation data. To get an idea of what would be the most appropriate index, the calculation of bias corrected precipitation is iterated over various assumptions about shelter index, $\eta_{wgt}$. The effect on hydrological modelling is evaluated using streamflow gauge stations, which are available over the evaluation period 1917-1950. The water balance error, which is the absolute difference between calculated and measured streamflow at several streamflow gauge stations, is calculated for a number of catchments and is compared for the different shelter index assumptions. The goal is to minimize the water balance error. A variety of shelter indices have been used ranging from nationwide values to more regionally variable indices:

1) $\eta_{wgt}$ =19 nationwide,
2) $\eta_{wgt}$=12 nationwide,
3) a 40×40 km$^2$ shelter index model based on analyses of shelter conditions at rain gauge stations over the period 1970-1995,
4) a pragmatic adjustment of the 40×40 km$^2$ shelter model, a so-called quick fix, where shelter values in certain regions have been modified to higher or lower values to account for local water balance errors.

Bias adjusted precipitation for the period 1989-2010 has been calculated for 10×10 km$^2$ fields (Vejen et al, 2014), and the estimates have been verified and approved in relation to water balance estimates in NOVANA (Det Nationale Overvågningsprogram for Vandmiljø og Natur). It is assumed that 1989-2010 can be used as a reference period against which the results for 1890-1910 and 1914-1950 can be compared and evaluated.

There is a clear seasonal variation in the correction level, $K_a$, due to generally higher wind speeds and increased frequency of snow during the winter season (Vejen et al 2014). Since $T$ and $a$ is causally related, and there is an indirect coupling between $T$ and $V$ (because of generally higher $V$ in winter and the opposite in summer), analysis of the relationship between $T$ and $K_a$ as function of shelter index, $\eta_{wgt}$, can give a reasonable impression of the overall correction level and can be used as a measure to find the most optimal value of $\eta_{wgt}$.

The Fjord and snow gauge was replaced by Hellmann in the period 1910-1925, and it is assumed that the Fjord and snow gauge makes up most of the station network at least until 1915. As the quality of precipitation data 1890-1913 and the error characteristics for the Fjord and snow gauge are unknown, evaluation of $\eta_{wgt}$ is carried out only for the period 1914-1950. This prevents that bias due to the properties of Fjord and snow gauge and quality issues until 1913 erroneously influences the conclusions. The evaluation is performed for scenarios of shelter and bias adjusted wind speed, $V_{mw}$, i.e., recall that $V_{mw}$ represents a monthly mean value for precipitation days.

The systematic relationship between monthly values of $T$ and $K$ is clear (Figure 12), and it is evident that a shelter index of about 12 causes the relationship between $T$ and $K$ to be closer to the verification period 1989-2010 than e.g., an index of 19. This suggests that the use of a nationwide shelter index = 12 appears to work quite well. In fact, the water balance error on a national level (1917-1950) is quite low for this scenario with a value of 3%, which is acceptable given the many assumptions.

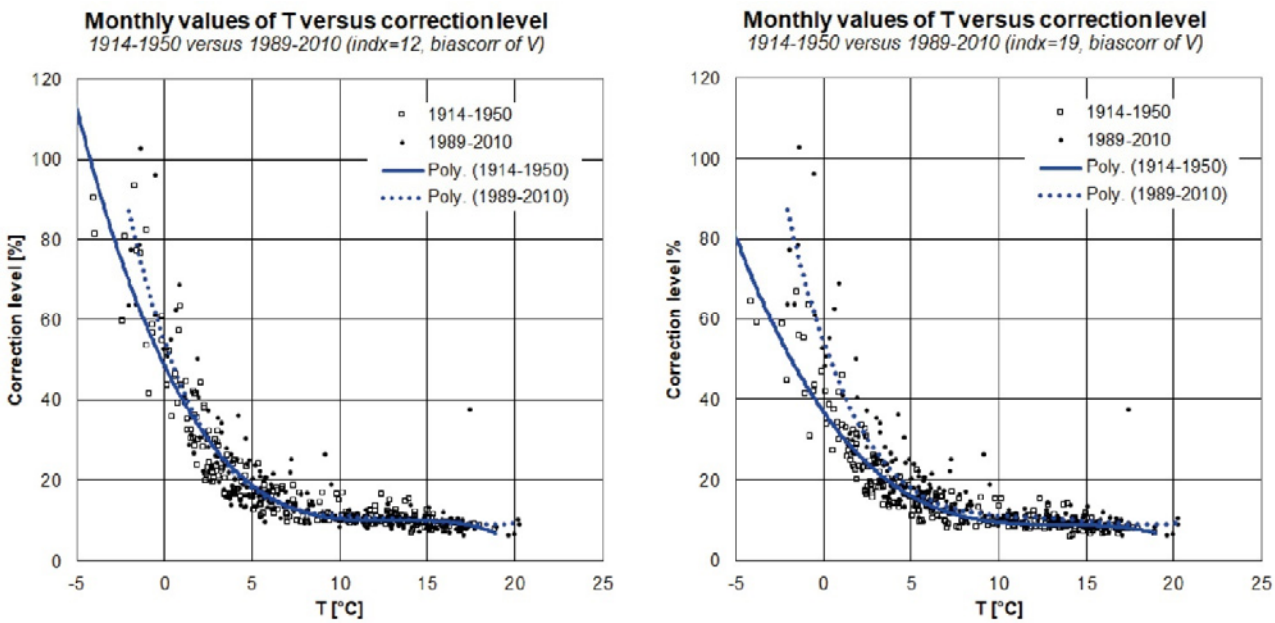



**Figure 12.** A comparison of temperature and correction level (%) for monthly values of bias corrected precipitation for the reference period 1989-2010 and the period 1914-1950. Results are shown for two scenarios: (left) shelter index 12 and bias correction of $V$, (right) shelter index 19 and bias correction of $V$. Here, bias correction of monthly values of $V$ is done by using $V_{mv}=kV_m$. The dotted line is a polynomial fit to the reference period, while the full line is the corresponding fit for scenario data.

However, despite the good fit on a national level, water balance calculations for the scenario show major errors on a regional scale with excessive water deficit in Western Jutland and excess of water in the eastern part of Denmark. The assumption of equal nationwide shelter conditions does not hold. Therefore, experiments have been carried out with 40×40 km$^2$ modelling of regional shelter variations, but these models still yield large regional water balance errors.

The shelter "index 12 model" was then combined with a so-called quick fix to obtain more reasonable regional estimates, e.g., a forced reduction of 10% is applied to the adjusted precipitation sum for eastern part of Denmark. Additionally, local problems with water balance estimates were found especially in Djursland and Als, and local changes of shelter index were applied in the attempt to come up with more trustworthy results for these regions. None of

the scenarios did fully solve the problems with the large errors in regional water balance.

## 3.5 Definition of delta change climate factor for precipitation ($\Delta\phi_m$)

It was concluded that further experiments are needed to come closer to a solution on the problems with regional water balance, but this is beyond the resources of the project, thus it was decided not to use the "shelter index 12 model" for calculation of precipitation climate around year 1900. Nevertheless, since the water balance error at national level was at an acceptable interval (%) for the period 1914-1950, it is assumed that the model assumptions and setups will work with the same level of uncertainty for the period 1890-1910 on national scale, but still with considerable uncertainty in water balance on regional scale.

A pragmatic approach was then to use national values of corrected precipitation to calculate monthly delta change climate factors, $\Delta\varphi_m$, defined as the ratio of historical (1890-1910), $P_{hist}$, to present (1989-2010), $P_{pres}$, values of corrected precipitation:

$$\Delta\varphi_m = P_{hist}/P_{pres}$$

It must be noted that $\Delta\varphi_m$ is based on corrected precipitation for 1890-1910 calculated using the same bias-correction method and the same model setup and assumptions as for 1914-1950. Historical and present precipitation is averaged over all grid cells which all have been given equal spatial weight.

If it is assumed that the relative regional variations of precipitation amount are identical for the two periods it may be reasonable to use the delta change climate factors for projection of regional variations of present climate to the period 1890-1910. This would imply that the temporal development in the climate is the same all over the country, which does not appear to be the case, since the stream discharges varies differently between regions with increase in discharge between 4 and 30% from 1935 to today (Transport on nutrients from land to sea, report in prep.). Thus, the use of a general delta change factor, $\Delta\varphi_m$, results in regional variations of the uncertainty in the calculated rainfall climate for this period, and probably also around year 1900. Despite the differences in response, it seems reasonable to use the general $\Delta\varphi_m$ to adjust the identified climate change.

## 3.6 Analyses on the effect of wind speed and shelter index on corrected precipitation

As previously shown, there is uncertainty associated with the calculation of wind speed and conversion from Beaufort and Danish Landscale to m/s. It is difficult to quantify how much this contributes to the uncertainty of corrected rainfall. Experiments have been carried out in which the converted wind has been applied a correction for trend estimated based on geostrophic wind, $V_g$. The results for $V_g$ suggest that $V$ is somewhat too high during the period of Danish Land Scale observations (1890-1911), while trend adjusted winds for 1912-1950 on average are close to the level without this correction, but with underestimation in certain periods (see section 2.6). Note, that all analyses in this section, grid cells in coastal regions have been given lower weight depending on the fraction of land area.

Sensitivity analysis has been performed where corrected rainfall has been calculated using this new wind. It is seen in Figure 11, that the trend corrected wind speed is partially underestimated compared to the uncorrected wind speed after 1912. It is clear, that higher trend corrected wind speed will result in increased amount of bias corrected precipitation, and that more precipitation will cause too high values of modelled discharge compared to the water balance errors of 3 % reported in section 3.4. By iteration it is now found that the resulting increase in corrected precipitation can be practically eliminated if the shelter index is changed from 12 to 15, i.e., the new index compensates for the changes in V and reproduces corrected precipitation close to the original values that resulted in a water balance error of 3%. As shown in Table 5 and Figure 13, this is true for the years after the 1910s, not for the periods 1890-1910 and 1900-1920. For scenario 2 (trend correction of $V$ and $\eta$=15) $P_k$ is much lower than for scenario 1 (no trend correction of $V$ and $\eta$=12). For example, for the period 1890-1910 $P_k$ is 793,1 mm for scenario 1, but only 758,5 mm for scenario 2. An explanation of these results may be related to the observation practice until the 1910s.

Until the 1910s, rainfall was measured 2 m above ground level (MI, 1875), but the shelter index assumes that the rain gauge is located at 1.5 m height as for Hellmann. The vertical increase of wind speed due to the logarithmic wind profile makes only a negligible contribution to the under-catch in measured rainfall. This effect is considered when converting wind speed to gauge height.

But the different measurement height may have caused the average shelter index to be lower than 15 up to the 1910s, and the rain gauge more exposed to the impact of the wind. It can be calculated theoretically that for a typical garden the shelter index, $\eta_{wgt}$, will change from 15 to 13 if precipitation is measured in 2,0 m level instead of 1.5 m. The result of using $\eta_{wgt}$ = 13 and trend corrected $V$ in the period 1890-1915 is shown in Table 5 and Figure 1 (scenario 3).

The question is whether the landscape was more open around year 1900, especially in the western part of Jutland, which would cause a lower shelter index. For example, we would expect a smaller number of plantations and lower height of the vegetation than today. Probably, the assumption of a constant shelter index in the whole period 1890-1950 does not hold. For example, it can be calculated, that $\eta_{wgt}$=10 would practically eliminate the quite large changes in corrected precipitation that are seen around the 1910s (Table 5).

***Table 5.*** *Measured annual precipitation ($P_m$) and results of corrected precipitation,* $P_k$*, based on scenario 1 (no trend correction of* V*,* η*=12), scenario 2 (trend correction of* V*,* η*=15) and scenario 3 (as scenario 2 but with adjustment of* η *for the period 1890-1915, see text for explanation. The difference between scenario 1 and 2, and 1 and 3 is also shown (mm and %).*

| ***Period*** | ***Measured precip, $P_m$*** | **Scenario** | | | **Difference** | |
|---|---|---|---|---|---|---|
| | | **1** | **2** | **3** | **1 vs 2** | **1 vs 3** |
| *1890-1910* | 634,1 | 793,1 | 758,5 | 769,7 | -34,6 (-4,5 %) | -23,4 (-3,0 %) |
| *1900-1920* | 645,5 | 794,5 | 772,7 | 781,2 | -21,8 (-2,8 %) | -13,3 (-1,7 %) |
| *1910-1930* | 674,9 | 801,8 | 798,8 | 801,9 | -3,0 (-0,4 %) | 0,1 (0,0 %) |
| *1920-1940* | 673,0 | 789,8 | 793,8 | 793,8 | 3,9 (0,5 %) | 3,9 (0,5 %) |
| *1930-1950* | 665,3 | 786,5 | 786,2 | 786,2 | -0,3 (-0,0 %) | -0,3 (-0,0 %) |
| *1916-1950* | 671,9 | 794,1 | 794,7 | 794,7 | 0,5 (0,1 %) | 0,5 (0,1 %) |

Analyzes of geostrophic winds indicate that Danish Landscale which was in use until 1911 is likely to overestimate *V*. The use of the lower trend corrected values of *V* in scenario 2 therefore causes $P_k$ to be substantially lower than for scenario 1 for the period up to the 1910s.

The trend correction of *V* probably more accurately reflects wind conditions than the uncorrected ones and are currently considered the best approach for calculation of *V*. In a study of historical trends for precipitation, Karlsson et al. (2014) detected homogeneity breaks in three long time series of precipitation out of 4. There is a temporal coincidence between the time of these breaks and the period 1910-1925 where the replacement of rain gauges took place, thus this may be a possible explanation of the homogeneity breaks identified.

For calculation of corrected precipitation, it is assumed that the aerodynamic properties of the old and new rain gauges are equal. However, if the catch efficiency and wetting and evaporation losses are different it would result in an unknown contribution to the uncertainty of corrected precipitation 1890-1915, and to rainfall climate around 1900. These aspects are discussed in the following.

The Fjord and snow gauge has a considerably larger orifice area (1000 $cm^2$) than Hellmann (200 $cm^2$), and the rim of the gauges is very narrow compared to the sharper and wider rim of Hellmann (Figure 3). Experiments in a wind tunnel have shown that the wind speed around a rain gauge appears to decrease as the orifice area increases in the interval 127 to 500 $cm^2$, and when the rim is shorter (Sevruk and Klemm 1989b, Sevruk, Hertig and Spiess 1989). Even though the opening area of the Fjord and snow gauge is larger than this interval, their results suggest a lower catch ratio for Fjord compared to Hellmann. However, the changes in *V* in the wind tunnel experiment were within a few percent and thus probably of secondary importance compared with the large difference in the physical shape between the two gauges.

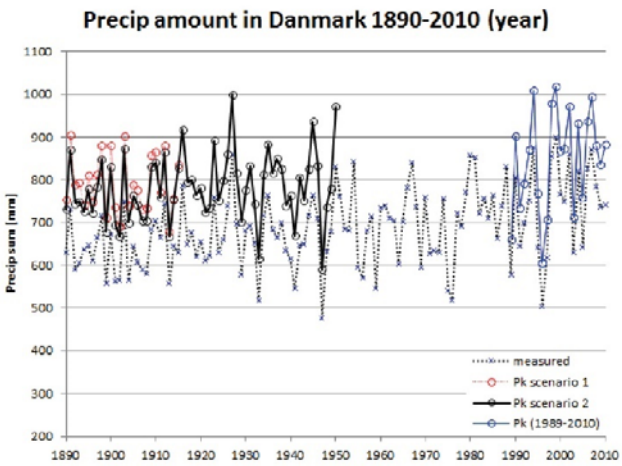



**Figure 13.** Measured and corrected annual rainfall in Denmark 1890-2010. Results of corrected precipitation are shown for scenario 1 (no trend correction of *V* and $\eta$=12), scenario 2 (trend correction of *V* and $\eta$=15) and scenario 3 (trend correction of *V* and $\eta$=13). Corrected precipitation is also shown for the period 1989-2010 (Vejen et al 2014).

The Fjord rain gauge has a larger surface of funnel and collection container than Hellmann, but its opening area is 5 times the size of Hellmann, thus the wetting loss for the two gauges is probably similar. The evaporation loss in the Hellmann gauge is negligible, as the precipitation is collected in a container inside the lower part of the gauge. This isolation of the collected rainfall is likely to dampen evaporation compared to the Fjord and snow gauges which are more "open", especially the snow gauge. The evaporation loss from

the Fjord and snow gauges is probably larger than for Hellmann, but it is difficult to calculate the losses theoretically, and it may require a field experiment to determine its magnitude.

Based on field experiments in a windy environment and on theoretical considerations Folland (1988) proposed that a rain gauge with form of a flat champagne glass and a large diameter would lead to better catch ratios than traditional rain gauges. Thus theoretically, a slightly better catch efficiency with Fjord gauges would be expected due to its form. On the other hand, the Fjord gauge is more angled shaped compared to the aerodynamic gauge referred to above and is equipped with a long spout which may further disturb the wind field. From the aspect of the design of the Hellmann and Fjord gauges it is difficult to find explanations as to why markedly less rainfall was measured before the 1910s. On the other hand, according to observation instructions the practice of measuring snow has been that, at high wind speeds, it might be necessary to turn the barrel of the snow gauge inversely down a flat snow surface for collection of the snowfall below (DMI 1880). Accurate measurements of snow may have been difficult, and critical statements about the snow measurements have been found in observer reports (Rigsarkivet).

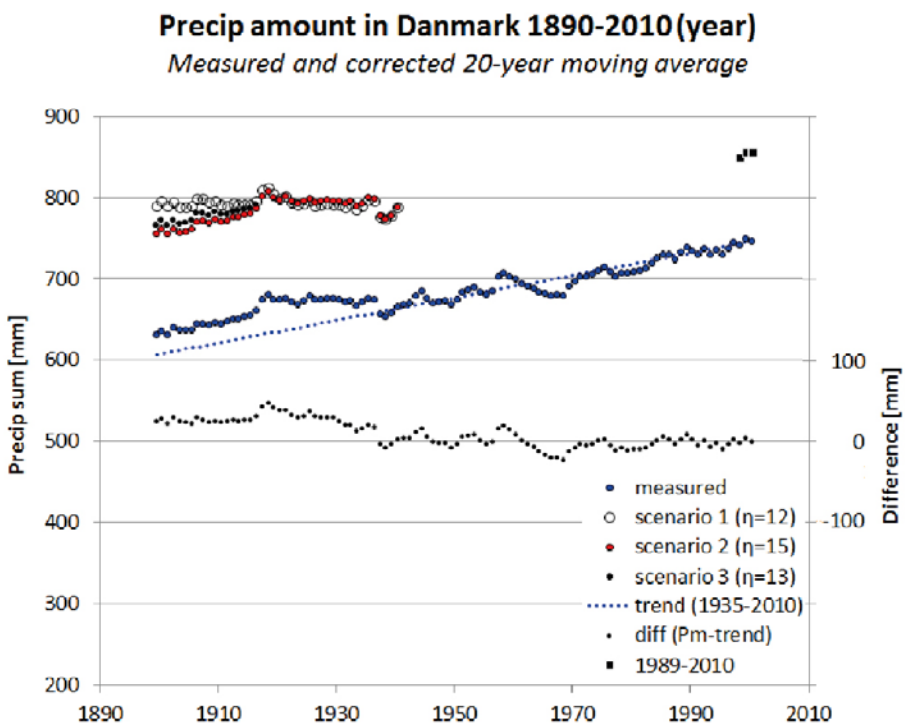



**Figure 14**. 20-years moving average of measured and corrected annual rainfall in Denmark 1890-2010. Note that each point indicates the middle of the time interval. Large blue dots: measured precipitation. Open black circles: scenario 1, corrected precipitation using shelter index 12 and $V_{mw}$. Red dots: scenario 2, corrected precipitation using shelter index 15 and $V_{mw}$ corrected for trends according to geostrophic wind. Large black dots: scenario 3, as scenario 2 except that η=13. Black squares: $P_k$ for 1989-2010. Small blue dots: trend line based on measured precipitation 1935-2010. Small black dots: difference between measured and expected precipitation according to the trend line.

Figure 14 shows a 20-year moving average of measured and corrected precipitation for shelter scenario 1, 2 and 3, and for recent years. The trend curve is based on measured precipitation 1935-2010 and is extended backwards in time to 1890. The overall trend in both measured and corrected rainfall does not appear to be linear over the period 1890-2010. The amount of corrected rainfall seems to be relatively constant 1915-1950 with a change towards a wetter climate in the years after 1950. Before 1915, the change in type of rain gauge and the lower shelter index causes a higher uncertainty in the level of corrected precipitation. Calculation of measured and corrected grid precipitation for the period 1951-1988 are missing, but it could be interesting in a climate perspective to analyze the changes during this period to see when corrected precipitation started to increase.

## 3.7 Summary of corrected precipitation

Figure 15 and Table 6 summarize the change in measured and corrected precipitation over the period 1890-2010 given as (unofficial) 30-year climate normal. Note, that in the calculation of annual precipitation grid cells in coastal regions have been given lower weight depending on the fraction of land area. . It is seen, that the measured amount of precipitation has increased by approx. 97 mm from around 1900 (1891-1920) until today (1989-2010), whereas the corrected value only has increased by 53 mm (scenario 1) or 70 mm (scenario 3). At the same time, the annual correction level of bias corrected precipitation has fallen from 22,9% for 1891-1920 over 18,1% for 1921-1950 to 14,6% for 1989-2010, which primarily is coupled to changes in wind, temperature, and rain climate.

The amount of corrected precipitation has been at a rather constant level, at least throughout the period between 1910s and 1950, although the measured amount has increased. This is partly due to a short warmer period, especially in the 1930s, with relatively little snow and lower correction level. The changes towards a wetter climate thus seem to have occurred after 1950. As pointed out by Førland and Hanssen-Bauer (2000), the omission of bias adjustment of precipitation entails a risk of interpreting on virtual climate change. On the other hand, the reported uncertainties of the converted wind speeds are critical for the corrected precipitation estimates, especially for the period 1890-1911 due to the use of the Danish Land scale.

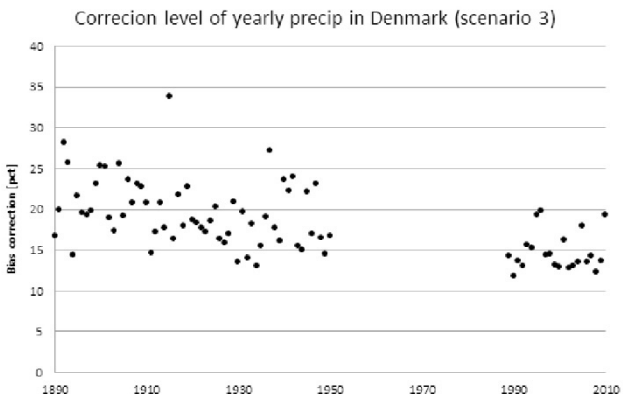



**Figure 15.** Annual correction level for Denmark 1890-1950 for scenario 3, and for the reference period 1989-2010.

The trend in bias adjusted precipitation and correction level is related to a change towards a gradually warmer climate with less snowfall. The changes in temperature climate and precipitation type do not occur gradually with a constant rate, but are complexly linked, giving large years to year variations in the correction level.

**Table 6.** Measured ($P_m$) and corrected precipitation ($P_c$) for scenario 1 and 3 (for explanation see **Table 5**), and correction level (%) for scenario 1 and 3, for Denmark for different 30-year periods. In calculation of national averages grid cells in coastal regions have been given lower weight depending on the fraction of land area.

| | 1891-1920 | 1901-1930 | 1911-1940 | 1921-1950 | 1931-1960 | 1961-1990 | 1989-2010 |
|---|---|---|---|---|---|---|---|
| Measured, $P_m$ | 643,8 | 657,4 | 670,3 | 670,9 | 670,1 | 711,5 | 740,9 |
| Corrected, $P_c$ [1] | 796,0 | 793,3 | 792,9 | 791,4 | - | - | 849,3 |
| Corrected, $P_c$ [3] | 779,4 | 786,5 | 794,9 | 792,3 | - | - | |
| $K_\alpha$ % [1] | 23,6 | 20,7 | 18,3 | 18,0 | - | - | 14,6 |
| $K_\alpha$ % [3] | 21,1 | 19,6 | 18,6 | 18,1 | - | - | |

## 3.8 The climate around year 1900 and calculation of delta change climate factor ($\Delta\phi_m$)

**Error! Reference source not found.** shows estimates of annual and monthly climate values around year 1900 (based on data 1890-1910) compared to the conditions 1989-2010. Measured precipitation has increased by 106,8 mm, but the corrected one is only increased by 65,8 mm, corresponding to a decrease in the annual correction level from 22,8% to 14,6%. This change is seen to be related to several changes in the climatic parameters. It has become warmer, which is particularly true during winter, where T for January and February has increased by approx. 2 °C. Opposite to this the temperature change of the three summer months is only +0.4 °C. The temperature change in the winter months is associated with a marked decrease in the proportion of precipitation falling as snow. Around 1900, almost 50% of the corrected precipitation fell as snow in the 3 winter months, while the proportion of snow in 1989-2010 was only approx. 26%. The climate has only changed marginally in the direction of lower wind speeds.

**Table 7.** The climate around 1900 (period 1890-1910) is compared to the reference period 1989-2010. Monthly and yearly values of measured and bias corrected precipitation, $P_m$ and $P_k$, and correction level $K_\alpha$ (%) are shown for scenario 1 ($\eta_{wgt}$=12). In addition is shown temperature $T_{month}$ and proportion of corrected precipitation fallen as snow (%), and trend corrected wind speed, $V$(*), for the period 1890-1910 using geostrophic wind data. The delta change factor, $\Delta\varphi_m$, of each month for scenario 1 used for calculation of water flow is also shown. Note that due to rounding of the monthly values, the sum of these is not equal to the annual rainfall sum. $K_\alpha$ (%) is calculated as average of individual months.

| | 1890-1910 (scenario 1) | | | | | | 1989-2010 | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | $P_m$ | $P_k$ | $K_\alpha$ (%) | $V$(*) | $T$ | snow% | $P_m$ | $P_k$ | $K_\alpha$ (%) | $V$ | $T$ | snow% | $\Delta\varphi_m$ |
| J | 44,9 | 71,7 | 59,6 | 5,8 | -0,3 | 52,0 | 61,0 | 74,4 | 22,1 | 5,9 | 1,7 | 18,3 | 0,963 |
| F | 33,1 | 54,6 | 64,6 | 5,6 | -0,4 | 58,1 | 50,6 | 66,9 | 32,3 | 5,9 | 1,7 | 32,2 | 0,816 |
| M | 41,6 | 58,7 | 41,1 | 5,6 | 2,1 | 32,9 | 46,9 | 57,1 | 21,7 | 5,6 | 3,3 | 12,8 | 1,028 |
| A | 40,1 | 48,3 | 20,4 | 5,3 | 5,8 | 3,2 | 37,7 | 44,2 | 17,1 | 4,9 | 7,0 | 1,7 | 1,094 |
| M | 46,1 | 51,8 | 12,3 | 4,9 | 10,7 | 0,0 | 44,4 | 49,2 | 10,8 | 4,6 | 11,2 | 0,0 | 1,053 |
| J | 50,2 | 55,7 | 10,8 | 4,8 | 14,7 | 0,0 | 61,6 | 67,6 | 9,6 | 4,6 | 14,2 | 0,0 | 0,824 |
| J | 61,7 | 68,0 | 10,2 | 4,7 | 16,3 | 0,0 | 61,4 | 67,0 | 9,1 | 4,3 | 16,8 | 0,0 | 1,014 |
| A | 88,1 | 96,8 | 9,8 | 5,0 | 15,5 | 0,0 | 77,5 | 83,9 | 8,2 | 4,4 | 16,8 | 0,0 | 1,155 |
| S | 53,6 | 59,6 | 11,3 | 4,8 | 12,6 | 0,0 | 70,8 | 77,4 | 9,3 | 4,8 | 13,4 | 0,0 | 0,771 |
| O | 69,6 | 78,1 | 12,2 | 4,9 | 8,4 | 0,2 | 79,9 | 88,0 | 10,2 | 5,1 | 9,2 | 0,1 | 0,888 |
| N | 49,4 | 60,0 | 21,3 | 5,0 | 3,9 | 12,3 | 66,9 | 77,5 | 15,9 | 5,3 | 5,1 | 6,2 | 0,773 |
| D | 50,1 | 69,6 | 38,9 | 5,6 | 1,1 | 35,9 | 63,0 | 77,0 | 22,2 | 5,2 | 2,1 | 21,1 | 0,905 |
| Year | 628,8 | 772,9 | 22,9 | 5,2 | 7,5 | 16,2 | 721,7 | 830,1 | 15,0 | 5,1 | 8,6 | 7,7 | 0,931 |

It is interesting to study the annual variation in rainfall. As of today, the spring months were also the driest around 1900, and the wettest months fell in late summer or autumn. Around 1900, August was the year's wettest month with as much as 98,0 mm (corrected), while October in 1989-2010 was top of the year with 90,8 mm. The spring months March-May were slightly wetter than today, while August had more precipitation. Almost all the other months were drier than in 1989-2010. These differences are reflected in the delta change climate factors ($\Delta\varphi_m$) shown in Table 7. As stated earlier, the magnitude of these factors may be affected by the uncertainty of wind speeds converted from Danish Landscale.

Probably, the differences in monthly and seasonally precipitation are related to changes in the atmospheric flows pattern. Periods of stable cold winter weather with limited rainfall were probably more frequent at that time, and summers were probably characterized by more unstable weather with fewer, and shorter, hot and dry periods, which is a weather type more often seen in recent years.

# 4 Evaluation of uncertainty

For the period 1890-1910, absent of discharge measurements makes it impossible to use hydrological modeling for evaluation of the precipitation estimates. But it is assumed that if the model setup and its assumptions work for 1914-1950, in which period discharge measurements are available for verification, it does so for 1890-1910.

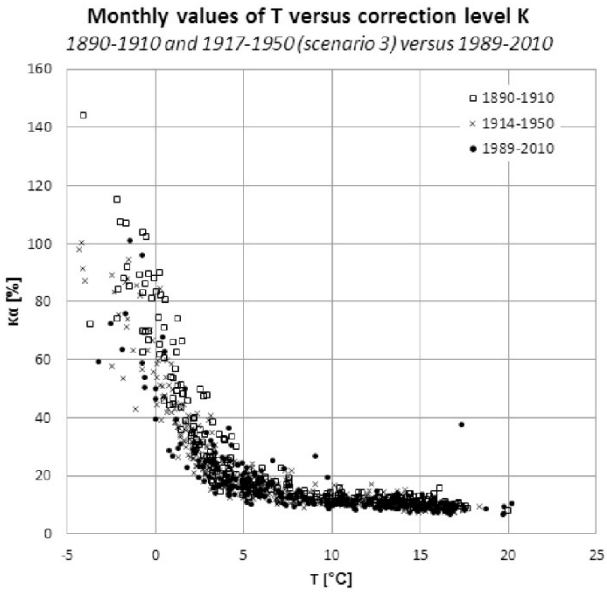



**Figure 16**. A comparison of temperature and correction level (%) for monthly values of bias corrected precipitation between the reference period 1989-2010 and the periods 1890-1910 and 1914-1950.

An unknown factor affecting the uncertainty for 1890-1910 is the use of the Fjord rain gauge since its error structure is unknown and its design is quite different from the Hellmann gauge (Figure 3). The correction factor is calculated using the equations for calculation of dynamic correction and since no information exists for the Fjord gauge, it has been assumed that the two gauges must be corrected identically. So, the Fjord gauge does not come into play in the calculation of correction factors.

Figure 16 shows a comparison between $T$ and $K_a$ (%) for 1890-1910 relative to the reference period 1989-2010. There seems to be a tendency for the correction level 1890-1910 to be slightly higher than for the other two periods. This is especially true at the lowest temperatures, that is, primarily during the winter months. It is also for this part of the year that any difference in the error structure of a rain gauge will be most evident, since the aerodynamic error related to snow is much higher than for rain. As previously shown, measurement with Fjord at 2 m height systematically gives higher corrections, which is the primary explanation for a higher correction level 1890-1910 than for the other periods.

Other factors that may affect the uncertainty are the basic assumptions for the calculations. This is especially true for the assumption for a nationwide shelter index of 12. Although it does cause problems regionally, it seems to work nationally for 1914-1950, which also is assumed to be the case for 1890-1910.

Analyzes of the controlling variable wind and precipitation seem to be quite well related to the official monthly climate values. One source of uncertainty

for 1890-1950 is the use of monthly values of $V$. It has not been possible within the project to study how this approach contributes to the uncertainty.

A measure of the uncertainty of corrected rainfall is composed of several significant contributions:

- Stochastic uncertainty on the correction model
- Spatial uncertainty (values of $T$, $V$ and $a$ are allocated to a precipitation station from nearest grid cell)
- Spatial uncertainty of gridded precipitation due to in-homogeneity of rain gauge station network
- Uncertainty of methods for calculation of meteorological variables
- Other sources of uncertainty, e.g., adjustment for shelter effect and trend correction of wind speed.

The uncertainty of the snow part of the correction model is for the range ±1×standard deviation approximately ±1% for common values of $V$ and $T$ and ±5% near the edge of the model (Allerup, Madsen and Vejen 1997), and for the liquid part of the model it is ±8% for high wind speeds (18 m/s) and ±2% for low wind speeds (4 m/s) (Allerup and Madsen 1986).

The standard use of the correction model is based on the input of on-site values of $V$, $T$, $\alpha$ and $I$, but none of these are measured at the precipitation stations. The effect on the daily correction level of the regional variation in Denmark of these parameters has been investigated (Allerup, Madsen and Vejen 2000). A basic criterion in this analysis was that the distance-related uncertainty should remain within the uncertainty of the correction model itself, i.e. lower that ±8%.

The analyzes led to isotropic distance relationships with simple rules for how far away from the precipitation stations daily values of $V$, $T$ and $\alpha$ during rainfall can be obtained without loss of confidence in the calculations of a daily correction factor. It was concluded that $V$ can be retrieved up to 50 km from the precipitation station without the uncertainty of $K_a$ exceeding ±8%. For $T$ and $\alpha$ it was found that the critical distance is larger than 50 km. Since the spatial resolution of grid values of $V$ and $T$ is 20×20 km$^2$, it is assumed that the contribution to the uncertainty of corrected rainfall is smaller than ±8%.

This may be surprising, but is since $V$, $T$ and $\alpha$ are calculated as mean values *during precipitation* and that it seems reasonable to assume that there are fairly isotropic conditions during rainfall, especially because of the structure of typical atmospheric pressure systems (Petersen et al 1981). However, the assumption of isotropy is more uncertain in coastal regions. Moreover, as this only applies to observations, the use of interpolated wind and temperature further contributes to the uncertainty.

Another contribution to the uncertainty comes from the interpolation method used for $P_m$, $P_k$, $V$ and $T$. Because of the limited number of wind and temperature stations, the interpolation was partly based on the average monthly spatial variability of these parameters for the period 1989-2010. The number of rain gauge stations the first years after 1890 were relatively limited, which also contribute to the uncertainty of interpolation. These uncertainties have not been quantified.

Due to the challenges of the shelter index, the delta change climate factor, $\Delta\varphi_m$, was used as a pragmatic approach to calculate corrected rainfall for 1890-1910. The result in Figure 17 shows the regional distribution of corrected rainfall around 1900 compared to more recent values, and in all regions are seen systematically smaller amounts. However, using a climate factor contributes to regional uncertainty, as the change in regional correction level depends on the magnitude of the corrected rainfall, as shown in **Figure 18**. The relative change is largest in the western parts and lowest in the eastern part of Denmark.

Analyses of geostrophic wind suggest uncertainty on wind data based on manually observed $V$. The results seem to show a general overestimation 1890-1911 with the Danish Land Scale, but only partially underestimating with Beaufort. While adjusted rainfall for 1914-1950 is calculated with an acceptable water balance error of 3% nationwide, the corresponding error for 1890-1913 is in principle unknown but is assumed to be on the same level. Generally, the corrected rainfall amounts appear to be lower for 1890-1910 compared to 1920-1950, but it is unclear whether this is due to uncertainties outlined above or whether the precipitation climate was drier around 1910.

The values of $\Delta\varphi_m$ in Table 7 is used for projection of regional variations of present climate to the period 1890-1910. The other scenarios for calculation of corrected precipitation, $P_k$, discussed in section 3.6 are probably more reliable estimates of the climate around 1900, given the various corrections of wind speed, $V$, and shelter index, $\eta$. Yearly values of $\Delta\varphi_m$ is calculated for the different scenarios and with and without sea-land weighting of coastal grid cells to get an idea of the uncertainty of delta change estimates. Introduction of land weighting to scenario 1 cause $\Delta\varphi_m$ to increase by 0,3 % from 0,931 to 0,934, while trend correction of $V$ in scenario 1 cause $\Delta\varphi_m$ to decrease by 1,5 % from 0.931 to 0,917. Due to the corrections of $V$ and $\eta$, scenario 3 would presumably result in a better estimate of $\Delta\varphi_m$ than scenario 1. In scenario 3, $\Delta\varphi_m$ is 0,906 or 2,7 % lower than in scenario 1 corresponding to approximately 20 mm / year which is an acceptable difference given the many assumptions and uncertainties.

The use of monthly delta change climate factors results in a change, or fraction, between the historical and recent precipitation varies from relatively high values in the eastern part of Denmark to quite low values in the western part, i.e., there are regional variations in the percentage change in precipitation amount. The use of a national delta change climate factor on a regional scale may result in systematic regional bias, which may have an impact on the uncertainty of runoff calculations on the spatial scale shown in the figure ($10\times10$ km$^2$). It may be necessary to average for larger regions and to redefine the delta change climate factor, e.g., by incorporation of regional variability.

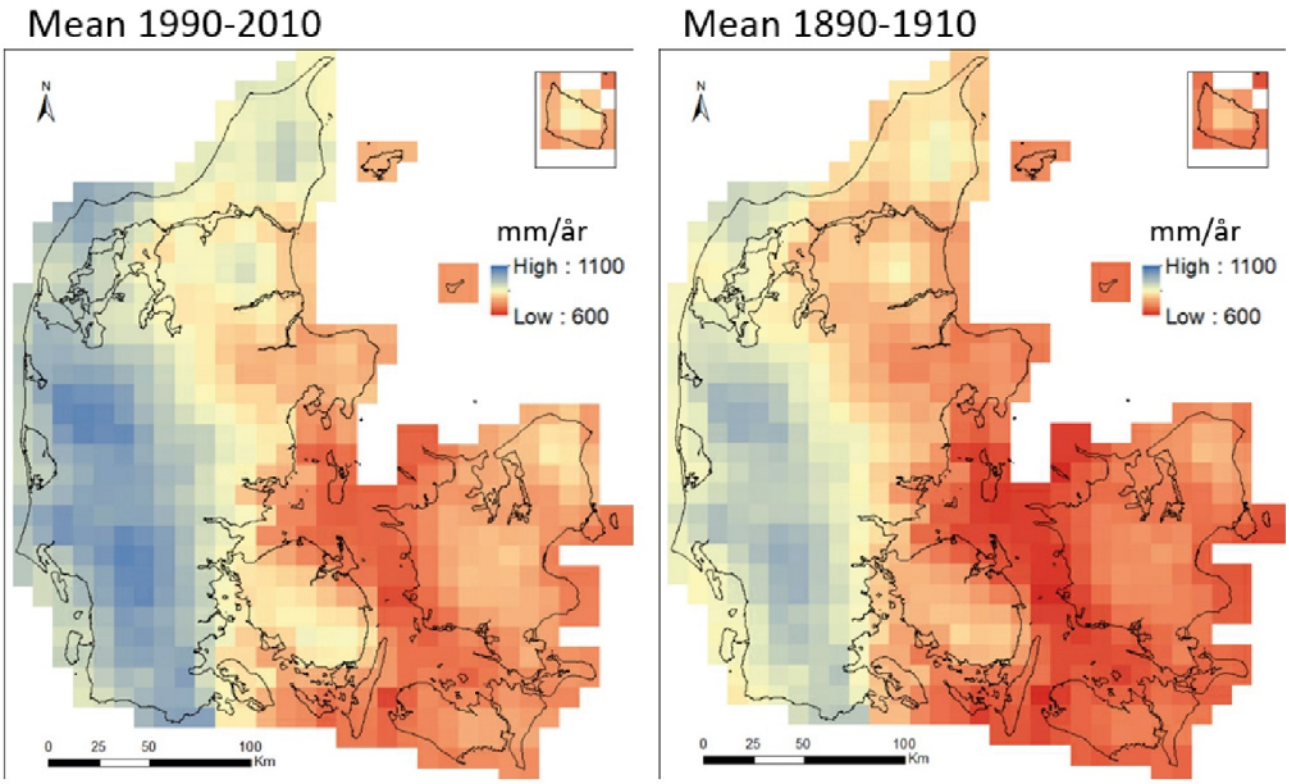
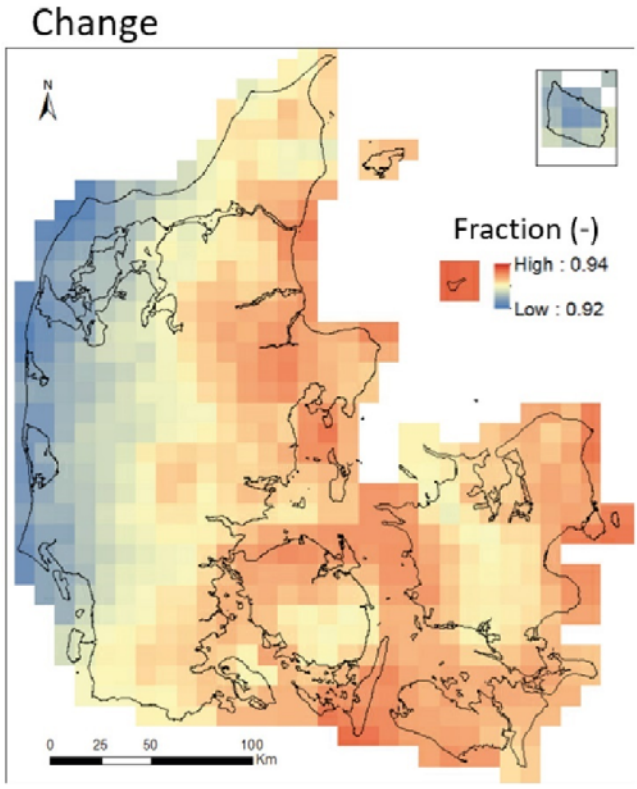



**Figure 17.** Left: mean precipitation per year 1890-1910 based on scenario 1. Right: mean corrected precipitation per year 1997-2017.

**Figure 18**. Mean change of corrected precipitation between 1890-1910 and 1997-2017.



# 5 Conclusion

The work described in this report is a part of a project whose purpose is to calculate water balance and discharge of nitrogen and phosphorus to open sea areas in Denmark around year 1900, i.e., the period 1890-1910. The objective is to provide this project with spatially distributed estimates of required meteorological variables to support the hydrological modelling. To make this possible, historical meteorological data over the period 1890-1950 have been digitized. Furthermore, an approach for bias correction of historical rain gauge data has been developed, and based on data series of bias corrected precipitation, air temperature, and wind data the climate around year 1900 has been calculated and validated. Various conditions contribute to the uncertainty of the corrected rainfall. Particular attention has been paid to the wind speed based on manual observations. Some of the wind series are affected by homogeneity breaks and deviation from the general trends in *V*. From analyses of geostrophic wind velocity, it was found that *V* was probably underestimated by the Danish Landscale which was in use until 1911, and that there was evidence of slight underestimation with the Beaufort scale at least in parts of the period 1912-1950.

Experiments were carried out with different combinations of shelter index and correction of *V* for homogeneity breaks and trends. Until 1910 the snow gauge and Fjord rain gauge was used and then during 1910-1925 gradually replaced by the Hellmann gauge. It has been argued that the error structure and catch efficiency of Fjord compared to Hellmann is only of secondary order compared to the measuring practice with Fjord being installed with its orifice 2 m above ground level. It was found probable that the measurement of rainfall at 2 m level until around 1915 probably caused the rain gauge to be more exposed to the wind than later. The resulting higher correction level is a reasonable explanation of the homogeneity break at the transition between the gauge types.

The calculations and assumptions are evaluated for the period 1914-1950 using water balance modelling of discharge. At national level, a water balance error of 3% has been found, but this covers large regional differences in error level. Of the many assumptions, the use of a general shelter index seems to have a quite large impact on the regional uncertainty. Analyses of the relationship between monthly temperature and correction level for the different periods support the impression of reasonable estimates on national level. However, it requires considerably more model experiments with fine tuning of the different basic assumptions to reduce the regional uncertainty to an acceptable level.

It has therefore been necessary to calculate corrected rainfall for 1890-1910 by using a delta change climate factor. In this approach national monthly correction factors were calculated based on corrected rainfall for 1890-1910 compared to 1989-2010. These national factors were then applied to the corrected rainfall for 1890-1910 to provide a spatially distributed daily time series of precipitation for the period 1890-1910.

Based on the assumptions in this study a delta change value of 0,931 per year was found which corresponds to approximately 773 mm per year, or 60 mm less than the precipitation amount in the reference period 1989-2010. Much

more of the precipitation around year 1900 consisted of snow which is related to a generally colder climate; 16,2% of the total amount was snow compared to 7,7% in the reference period. This also explains the generally higher correction factor of 22,9% per year compared to 15,0% for the reference period.

The uncertainty of bias corrected precipitation depends, among other things, on the reliability of wind speed data. Around year 1900 wind speed was observed with the so-called Danish Land scale which can be considered as a kind of half Beaufort. Manual wind observations are subject to different errors and larger uncertainty compared to modern automatic instruments. In the end of the project tests were made possible for correction of wind data for trend errors using a long time series of geostrophic wind speed. It was found that the wind speed used for bias correction was probably overestimated. As a result, corrected precipitation and delta change was a little too high. Using the corrected wind speed resulted in a delta change value of 0,906 (-2,7%) which corresponds to around 20 mm less precipitation per year compared to the original precipitation estimate which is an acceptable difference given the many assumptions and uncertainties.

# 6 Referencer


Alexandersson H., Schmith, T., Iden, K., and Tuomenvirta, H., 1998: Long-term variations of the storm climate over NW Europe. Global Atmos Ocean Syst 6:97–120.

Allerup, P. and H. Madsen, 1979: Accuracy of point precipitation measurements. Danish Meteorological Institute, Climatological Papers, No. 5, Copenhagen 1979, 84p.

Allerup, P. and H. Madsen, 1980: Accuracy of point precipitation measurements. Nordic Hydrology, 11, p. 57-70.

Allerup, P., and Madsen, H., 1986: On the correction of liquid precipitation. Nordic Hydrology, 17, p. 237-250.

Allerup, P., Madsen, H., and Vejen, F., 1997: A Comprehensive Model for Correcting Point Precipitation. Nordic Hydrology, vol. 28, 1-20.

Allerup, P., Madsen, H., and Vejen, F., 2000: Correction of precipitation based on off-site weather information. Atm. Res., Vol. 53, 231-250.

Brandt, M. L., 1994: The North Atlantic Climatological Dataset (NACD). Instrumenter og rekonstruktioner. En illustreret gennemgang af arkivmateriale. DMI Technical Report 94-19, København 1994, 74 p.

Cappelen, J., 2000: The Climate of Denmark - Key Climatic Figures 1990-1999. Technical Report, TR00-08, DMI, Copenhagen 1999, 47pp.

Cappelen, J., 2019: Denmark – DMI Historical Climate Data Collection 1768-2018. Technical Report, TR19-02, DMI, Copenhagen 2019, 112pp.

Christiansen, T., Christensen, T. J., Markager, S., Petersen, J. K., and Mouritsen, L. T., 2006: Limfjorden i 100 år. Klima, hydrografi, næringsstoftilførsel, bundfauna og fisk i Limfjorden fra 1897 til 2003. Faglig rapport fra DMU (Danmarks Miljøundersøgelser, Miljøministeriet), nr. 578, 2006, 85 pp.

Colli, M., Pollock, M., Stagnaro, M., Lanza, L. G., Dutton, M. and O'Connell, E., 2018: A Computational Fluid-Dynamics Assessment of the Improved Performance of Aerodynamic Rain Gauges. Water Resources Research, 54, 779-796.

Det Statistiske Departement, 1964: Folketal, areal og klima. Statistiske undersøgelser nr. 10, Det Statistiske Departement 1964, 271 pp.

DMI, 1875: Meteorologisk Aarbog for 1874. Udgivet af det danske meteorologiske institut, Kjøbenhavn 1875.

DMI (Danish Meteorological Institute), 1880: Maanedsoversigt over Vejrforholdene. Udgivet af det Danske Meteorologiske Institut. Published monthly and collected in a yearbook for 1880.

DMI (Danish Meteorological Institute), 1890-1950: Maanedsoversigt over Vejrforholdene. Udgivet af det Danske Meteorologiske Institut. Published monthly and collected in yearbooks, 1890-1950.

Feiccabrino, J., Graff, W., Lundberg, A., Sandström, N., and Gustafsson, D. (2015): Meteorological Knowledge Useful for the Improvement of Snow Rain Separation in Surface Based Models. Hydrology, 2015, 2, 266-288; doi:10.3390/hydrology2040266.

Folland, C. K. (1988): Numerical models of the rain gauge exposure problem, field experiments and an improved collector design. Q. J. Meteorol. Soc., Vol. 114, 1485-1516.

Førland (ed), E. J., Allerup, P., Dahlström, B., Elomaa, E., Jónsson, T., Madsen, H., Perälä, J., Rissanen, P., Vedin, H., and Vejen, F. (1996): Manual for operational correction of Nordic precipitation data. Nordic Working Group on Precipitation. First Edition. DNMI, Report Nr. 24/96, 66pp.

Førland, E. J., and I. Hanssen-Bauer, 2000: Increased precipitation in the Norwegian arctics: True or false? Climatic Change, vol. 46, 485-409.

Goodison, B. E. (ed), Louie, P. Y. T., and Yang, D., 1998: WMO Solid Precipitation Measurement Intercomparison: Final Report. Instruments and Observing Methods, Report No. 67, WMO/TD No. 872, 1998.

Harder P., Pomeroy J., 2013: Estimating precipitation phase using a psychrometric energy balance method, Hydrol. Proc. 27 (2013) 1901–1914, doi:10.1002/hyp.9799.

Heberden, W., 1769: On the different quantities of rain which appear to fall, at different heights, over the same spot of ground. Phil. Trans., 359-362.

Højberg, A. L., Windolf, J., Børgesen, C. D., Troldborg, L., Tornbjerg, H., Blicher-Mathiesen, G., Kronvang, B., Thodsen, H., and Ernstsen, V., 2015: National kvælstofmodel, Oplandsmodel til belastning og virkemidler. Revideret udgave september 2015. GEUS, 402 s.

Karlsson, I. B., Sonnenborg, T. O., Jensen, K. H., and Refsgaard, J. C., 2014: Historical trends in precipitation and stream discharge at the Skjern River catchment, Denmark. Hydrol. Earth Syst. Sci., 18, 595-610, 2014.

Kaufeld, L., 1981: The Development of a New Beaufort Equivalent Scale. Meteorol. Rdsch., 34, 17-23.

Kristensen, L., og Frydendal, K., 1991: Danmarks vindklima fra 1870 til nutiden. Havforskning fra Miljøstyrelsen, Nr. 2, 1991, Miljøministeriet, Miljøstyrelsen, 57 pp.

Luthardt, H., and Hasse, L., 1981: On the Relationship between Surface and Geostrophic Wind in the German Bight Area. Beitr. Phys. Atm., 54, 222-232.

Lysgaard, L., 1969: Foreløbig oversigt over Danmarks Klima, lufttryk, vindforhold, lufttemperatur, solskin, nedbørforhold og luftfugtighed hovedsagelig i perioden 1931-60. Det Danske Meteorologiske Institut, Meddelelser Nr. 19., 109 pp.

Olesen, J. E., Bøcher, P. K, and Jensen, T., 2000: Comparison of scales of climate and soil data for aggregating simulated yields of winter wheat in Denmark. Agric. Ecosys. and Env., Vol. 82, 213-228.

Olesen, S. E., 2009: Kortlægning af Potentielt dræningsbehov på landbrugsarealer opdelt efter landskabselement, geologi, jordklasse, geologisk region samt høj/lavbund. DJF intern rapport, markbrug nr. 21.

Petersen, E. L., Frandsen, S., Hedegaard, K. and Troen, I., 1981: Wind atlas for Denmark. Risø, Denmark.

Plauborg F., Refsgaard, J. C., Henriksen, H. J., Blicher-Mathiesen, G., Kern-Hansen, C., 2002: Vandbalance på mark- og oplandsskala. DJF rapport. Markbrug nr. 70.

Refsgaard, J.C., Stisen, S., Højbjerg, A.L., Olsen, M., Henriksen, H.J., Børgesen, C.D., Vejen, F., Kern-Hansen, C., Blicher-Mathiesen, G. (2011): Vandbalance i Danmark - Vejledning i opgørelse af vandbalance ud fra hydrologiske data for perioden 1990-2010. GEUS rapport 2011/77, 75p.

Scharling, M., 1999: Klimagrid – Danmark. Nedbør, lufttemperatur og potentiel fordampning, 20×20 & 40×40 km. Metodebeskrivelse. DMI Technical Report 99-12, Copenhagen 1999, 48 pp.

Schmidt, T., Alexandersson, H., Iden, K., and Tuomenvirta, H., 1997: North Atlantic – European pressure observations 1868-1995 (WASA Dataset version 1.0). DMI Technical Report 97-3, Copenhagen 1997, 13pp.

Sevruk, B., 1979: Accuracy of Point Precipitation Measurement. Laboratiory of Hydraulic Research, Hydrology and Glaciology, Federal Institute of Technology, Zürich, Switzerland. Commision for Hydrology, WMO 1979, 121p.

Sevruk, B. (WMO), 1982: Methods of Correction for Systematic Error in Point Precipitation Measurement for Operational Use (B. Sevruk). Operational Hydrology Report No. 21 (WMO-No. 589). Geneva.

Sevruk, B., and Hamon, W. R. (WMO), 1984: International Comparison of National Precipitation Gauges with a Reference Pit Gauge. WMO, Instruments and Observing Methods, Report No. 17, WMO/TD-No. 38, Secretariat of the World Meteorological Organisation, Geneva, Switzerland, 139 pp.

Sevruk, B. and Klemm, S., 1989a: Types of standard precipitation gauges. WMO, IAHS, ETH, International Workshop on Precipitation Measurement, St. Moritz, 1989 Switzerland, 227-232.

Sevruk, B., and Klemm, S., 1989b: Catalogue of national standard precipitation gauges. WMO, Instruments and Observing Methods, Report no. 39, WMO/TD-No. 313, 1989, 50 pp.

Sevruk, B., Hertig, J.-A., and Spiess, R., 1989: Wind field deformation above precipitation gauge orifices. Atmospheric Deposition (Proceedings of the Baltimore Symposium, May 1989), IAHS Publ. No. 179, 65-70.

Sevruk, B., M. Onda´ s, and M. Chvı´la, 2009: The WMO precipitation measurement intercomparisons. Atmos. Res., 92, 376–380.

Vejen, F., Madsen, M., og Allerup, P., 2000: Korrektion for fejlkilder på måling af nedbør. Korrektionsprocenter ved udvalgte stationer. 1989-1999. Danish Meteorological Institute, Technical Report No. 00-20, Copenhagen 2000, 53pp.

Vejen, F., Vilic, K., Jensen, H., and Kern-Hansen, C., 2014: Korrigeret Nedbør 1989-2010, 2011-2012 & 2013. Databeskrivelse & Resultater. Konsulentopgave udført for DCE – Nationalt Center for Miljø og Energi, Aarhus Universitet. DMI Technical Report 14-13, Copenhagen 2014, 228pp.

WMO, 1970: The Beaufort Scale of Wind Force. WMO Commission for Maritime Meteorology, Marine Sciences Affairs Report NO. 3, 22 pp.

WMO, 2008: Guide to Meteorological Instruments and Methods of Observation. WMO-No. 8, 7th edition. Secretariat of the World Meteorological Organization, Geneva, Switzerland.