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Discussions

Evaluating the influence of long term historical climate change on catchment hydrology – using drought and flood indices

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Abstract

This study uses a 133 yr data set from the 1055 km^2 Skjern River catchment in a western Danish catchment to evaluate: long-term past climate changes in the area; the capability of a conceptual hydrological model NAM to simulate climate change impacts on river discharge; and the occurrences of droughts and floods in a changing climate. The degree of change in the climatic variables is examined using the non-parametric Mann-Kendall test. During the last 133 yr the area has experienced a significant change in precipitation of 46 % and a temperature change of 1.3 °C leading to (simulated) increases in discharge of 103 % and groundwater recharge of 172 %. Only a small part of the past climatic changes was found to be correlated to the climatic drivers: NAO, SCA and AMO. The NAM model was calibrated on the period 1961–1970 and showed generally an excellent match between simulated and observed discharge. The capability of the hydrological model to predict climate change impact was investigated by looking at performances outside the calibration period. The results showed a reduced model fit, especially for the modern time periods (after the 1970s), and not all hydrological changes could be explained. This might indicate that hydrological models cannot be expected to predict climate change impacts on discharge as accurately in the future, as they perform under present conditions, where they can be calibrated. The (simulated) stream discharge was subsequently analyzed using flood and drought indices based on the threshold method. The extreme signal was found to depend highly on the period chosen as reference to normal. The analysis, however, indicated enhanced amplitude of the hydrograph towards the drier extremes superimposed on the overall discharge increase leading to more relative drought periods.

1 Introduction

Climate change is likely to result in significant changes in the hydrological regimes (IPCC, 2007b) in the future; however climate change projections and their impacts are

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very uncertain. Major sources of uncertainty are related to uncertainty in climate models and uncertainty on the capability of hydrological models to make predictions under a different climate than the one where they can be calibrated (Refsgaard et al., 2012). Previous studies show that predictions of hydrological models for precipitation regimes different from the period where they are calibrated results in reduced performance (Refsgaard and Knudsen, 1996; Donelly-Makowecki and Moore, 1999; Seibert, 2003). However, these studies have been made on dry and wet periods that are results of short-term climate variability rather than long-term climate change. To test hydrological models' capability to predict climate change impacts on hydrology there is a need for long time series showing climate change. Problems arise, however, as most long time series with discharge data are disturbed in different degrees by anthropogenic impacts such as river regulations and water abstractions, the influence of these changes should therefore be considered with care when trying to disassemble the climate change impact signal.

In past historical time climate change has also occurred leading to changes in precipitation; this has been recognized in several studies e.g. from Sweden, Norway (Tuomenvirta et al., 2001) and the United Kingdom (Jones and Conway, 1997); and also leading to changes in runoff as found in Wilson et al. (2010). The Danish area has also experienced climate change during the last century resulting in a large increase of precipitation and temperature (Jeppesen et al., 2011; Thomsen, 1993). The increase has been unevenly distributed with the largest increase occurring in western Jutland (Jørgensen and Cappelen, 2006; Kronvang et al., 2006). Long term historical climate change presents a potential for model performance testing under climate change.

Extreme events as droughts and floods have a profound effect on both economy and ecology of a catchment, affecting both availability and distribution of water. In a future climate change perspective the extreme events are also important because of the profound effect on future risk planning and management of water resources. Several definitions of droughts have been proposed in the literature as described in the review of Mishra and Singh (2010) depending upon the purpose of the study; decreasing

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precipitation, lack of precipitation, crop failure due to water shortage, river flow decrease or shortage in the water supply system. Floods, however, are often defined as increasing discharge in streams and lakes possibly leading to bank overflow. Generally, drought categories are defined as meteorological, hydrological, agricultural and socio-economic droughts (Hisdal et al., 2001b). Meteorological droughts describe the main origin of droughts in the system as a deficit of precipitation, they are especially critical when combined with high temperatures leading to enhanced evapotranspiration. Hydrological droughts are defined as a deficit of water in the surface and subsurface water bodies of an area, reflecting the effect and impact of the meteorological drought and/or heat waves.

Drought studies based on the threshold method have been carried out at global (e.g. Fleig et al., 2006), regional scale (e.g. Hannaford et al., 2010; Hisdal and Tallaksen, 2003; Hisdal et al., 2001a) and catchment scale (e.g. Peters et al., 2006; Tallaksen et al., 2009). Usually data series are often short (30–60 yr), representing drought signal on a short time scale. The resulting drought signal of an analysis depends both on method, the period analyzed (Hisdal et al., 2001a) and on the choice of reference period (Stahl, 2001).

The objectives of our study are:

1. To analyze the magnitude of the recorded climatic and hydrological changes in a Danish river catchment since 1875; and to test the performance of a hydrological model during these conditions. This includes answering the questions:

    a. Can a parameter set based on a calibration period be considered representative outside the calibration period in a non-stationary climate.

    b. To what degree is this representation influenced by anthropogenic factors in the catchment.

    c. Can the recorded climatic change be explained by known climatic drivers.

2. To analyze trends and occurrences of extremes in a non-stationary climate using drought and flood indices for stream discharge. This includes answering the questions:

    a. Have streamflow floods and droughts increase or decreased in the area.

    b. To what degree does the choice of reference period influence the trend and occurrence of the extremes.

2 Study area and data set

2.1 Skjern River Basin

The Skjern River Basin (2378 km^2) located in the western part of Denmark is selected as study area with the Skjern River sub-catchment Alergaarde as the focus (Fig. 1). It is bounded to the east by the Jutland Ridge; to the west the Skjern River reaches the North Sea through the Ringkøbing Fjord. For the period 1961–1990 the average precipitation is 1041 mm yr^{-1} with an average of 155 precipitation days pr. year and an average temperature of 8.1 °C. The sub-catchment area is 1055 km^2, with a 96 km long river system flowing from east towards west and discharging an average of 15.8 m^3 s^{-1} (Larsen et al., 2003; Ovesen et al., 2000). Groundwater flow is generally also from east to west with an average hydraulic gradient of 0.001 (Stisen et al., 2011). The geology in the area is partly a result of Saale ice age and partly a result of the location of the younger Weichselian ice sheet front at the main ice advance (Main Stationary Line) at the Jutland Ridge (Houmark-Nielsen and Kjaer, 2003). Where the eastern part of the catchment consists mainly of the end moraine deposits of sand and clay from the Weichselian, while the mid and south are dominated by the wash-out sand and gravel from this advance; and relic Saale moraine hills are predominately found in the northwest and west of the catchment.

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At the end of the 1700, heath and marshes constituted four-fifths of vegetation in the whole Skjern River catchment; the rest consisted of forest, agricultural land and meadows/grass vegetation concentrated in the stream valleys. During the 1800 the process of enclosure and the industrialisation of the Danish agricultural practice resulted in an extensive cultivation of new land. In the Skjern River catchment this meant up to a halved heath area and substantial reduction of wet lands and marshes. The reclaimed land was primarily used for agriculture and forest plantations. At the end of the century the forest covered around 12 % of the catchment, hereafter almost all afforestation ceased, apart from smaller plantings and hedgerows. The reclamation of the land continued into the 1900 where any heath area reduction resulted in new agricultural land, the maximum extent was reached in 1938 with 76 % (Fritzbøger, 2009). Today the smaller Alergaarde catchment consists of 60.5 % agriculture, 17.4 % grass, 14.0 % forest, 6.3 % heath, and 1.8 % urban areas (Fu et al., 2011).

This means that for the time period investigated in this study, the main land use changes were primarily in the very beginning and before the calibration period; where the land use changes from the 1900 have been abating. However, other human changes have occurred gradually since the 1900 including use of fertilization, irrigation and drainage. The potential effect of these changes on stream discharge will be discussed. For the catchment daily precipitation and temperature data has been recorded since 1875 and discharge data since 1920, constituting an exceptional data set both in length and resolution. The area is of particular interest as it is part of the HOBE-hydrological observatory (Jensen and Illangasekare, 2011).

2.2 Climate data

Daily precipitation data is available from four primary stations going back to 1875 (Fig. 1). The catchment precipitation is calculated as a weighted average of the stations: 23050 (42 %), 23220 (28 %), 24180 (9 %) and 24500 (20 %), where the weights are estimated from Thiessen polygons. Some of the early precipitation data are recorded as accumulated amounts for subsequent days, sometimes holding

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information on the number of days the sum represent. When this is the case the accumulated precipitation has been distributed equally over the period. Since not all the series are complete seven additional stations are used to supplement missing time slices (21100, 21430, 23180, 24070, 24240, 25010 and 25140). The measured precipitation is also corrected for wetting and aerodynamic effects using the standard correction methods of Allerup et al. (1997). Further information on the method can be found in Stiesen et al. (2011). Information about metadata for the stations are scarce and therefore shelter class have been assumed to represent moderate lee conditions (B).

Since no temperature stations with suitable data coverage are available within the catchment, an average of three stations (21100, 25140 and 27080) placed south, north and east of the catchment is used (Fig. 1). Based on data for minimum and maximum daily temperature the mean daily temperature is estimated. No temperature data are available between 1 January to 6 February 2000, and 2 September to 8 October 2000. The missing values are obtained from the climate grid provided by the Danish Meteorological Institute (Scharling, 2000). The temperature stations are located somewhat far from the catchment (around 80–100 km), however, when comparing the three stations average with the grid data from DMI (1989–2007), even though the average temperature is slightly elevated for the stations due to their proximity to the coast, the correlation coefficient is 0.98 demonstrating that these three stations provide a fair approximation of the temperature in the catchment.

The discharge station (Alergaarde, 25005) is placed at the outlet of the sub-catchment and contains data from 1920–2007. Again the series are not complete and missing data are complemented by values from the nearby Gudenå catchment (Tvilumbro, 21001). A total of approximately 2 yr of data are missing.

The potential evapotranspiration can be calculated using different empirical formulas. As temperature is the only available input data back to 1875 no radiation-based calculations could be used. The Thornthwaite (1948) and Hamon (1963) methods are both based on temperature data and to test their performances they were compared to calculations based on Penman-Monteith method (Monteith, 1965; Penman, 1948), which is a more physically-based method. Data on net radiation, wind speed, ground heat conductance, air temperature and relative air humidity for the period 1990–2009 from an agricultural research station at Foulum situated 48 km northeast of the Skjern River basin (Fig. 1) were used for the analysis. Both with respect to average annual values, monthly distribution and correlation coefficient the Thornthwaite method performed better than the Hamon method and it was therefore chosen as the most appropriate method for calculation of potential evapotranspiration. The Thornthwaite method is given by (Shaw, 1994):

$$\mathrm{PE_m} = 16 \cdot N_\mathrm{m} \cdot \left(\frac{10 \cdot \overline{T_\mathrm{m}}}{I_\mathrm{y}}\right)^a \quad \text{if } T_\mathrm{m} > 0 \text{ else } \mathrm{PE_m} = 0 \tag{1}$$

where "m" is the month and "y" is the year. The Thornwaite calculations are preformed using a number of parameters including: monthly daylight factor: N_m – number of possible hours of sun/12.

$$\text{Annual heat index}: \ I_\mathrm{y} = \sum i_{(\mathrm{y,m})}; \tag{2}$$

$$\text{Monthly heat index}: \ i_\mathrm{m} = \left(\frac{T_\mathrm{m}}{5}\right)^{1.5} \quad \text{if } T_\mathrm{m} > 0 \text{ else } i_\mathrm{m} = 0 \tag{3}$$

$$\text{a (function of } I): \ a = 0.49 + 0.0179 \cdot I_{\mathrm{y,m}} - 7.71 \times 10^{-5} \cdot I^2 + 6.75 \times 10^{-7} \cdot I^3. \tag{4}$$

The performance of the Thornthwaite formula was further evaluated by comparing monthly values to estimates by the Penman–Monteith method (Monteith, 1965; Penman, 1948) (Table 1). A statistical test of the significance of the regression line between the two was carried out (Sect. 3.1.2). For the nine months with significant correlation, the Thornthwaite estimates for the period before 1990, corrected through the nine regression equations, were applied. For the remaining three months the averages of the Penman estimates for the period 1990–2009 were used.

3 Methods

3.1 Statistical methods

3.1.1 Trend analysis

Trends in the measured climatic components and in extreme events cannot necessarily be considered to be linear or to follow a normal distribution. Therefore, the non-parametric Mann–Kendall test is used to evaluate if the increases/decreases of the time series are significant (Salas, 1993; Hisdal et al., 2001a). For all statistical test in this study a significance level of $\alpha = 0.05$ (95 %) is used. For a time series, y, with n number of time steps, the values at time $t = 1, 2, 3$ ....., n and at time $t' = 1, 2, 3$ ....., $n-1$ where $t = t' + 1$ are considered. The value at each time step t' is compared with the value at the following time steps and used to create a data set z as follows:

$z = 1$ for $y_t > y_{t'}$
$z = 0$ for $y_t = y_{t'}$
$z = -1$ for $y_t < y_{t'}$.

The Kendall score is calculated from the z data set as follows:

$$S = \sum_{t'=1}^{n-1} \sum_{t=t'+1}^{n} z. \tag{5}$$

The Mann–Kendall test statistics u_c is calculated as:

$$u_c = \frac{S+1}{\sqrt{V(S)}} \text{ for } 0 > S \tag{6}$$

$$u_c = \frac{S-1}{\sqrt{V(S)}} \text{ for } 0 < S \tag{7}$$

where $V(s)$ is the variance of z and calculated as:

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$$V(S) = \frac{1}{18}\left(J(J-1)(2J+5) - \sum_{j=1}^{J} e_j\left(e_j-1\right)\left(2e_j+5\right)\right) \quad (8)$$

and J is the overall number of groups formed by sets of data with identical values (tied group) while e_j is the amount of data in the individual group, i. The test has been reported to be almost as strong as a parametric counterpart and has traditionally been used to examine both droughts (Hisdal et al., 2001a; Wilson et al., 2010) and discharge trends (Burn et al., 2002; Mitosek, 1995). In this test the hypotheses are: H0 – the null hypothesis is no trend in the data; and H1 – the alternative hypothesis is that there is a trend. H0 is rejected when the Mann–Kendall statistics $|u_c| > u_{1-\alpha/2}$, corresponding to a $1-\alpha/2$ quantile of the standard normal distribution (Hisdal, 2001a). With a significance level of $\alpha = 0.05$ and $n = 133$ annual values the $u_{1-\alpha/2}$ has a value of 1.96.

3.1.2 Pearson product moment correlation coefficient

The Pearson correlation coefficient (ranging from −1 to 1) represents the degree of linear coherency between two variables. The correlation coefficient is calculated as:

$$r = \frac{\sum_{i=1}^{n}\left(O_i-\overline{O}\right)\left(S_i-\overline{S}\right)}{\sqrt{\sum_{i=1}^{n}\left(O_i-\overline{O}\right)^2}\sqrt{\sum_{i=1}^{n}\left(S_i-\overline{S}\right)^2}} \quad (9)$$

where O_i and S_i are the two variables at time i, $\overline{O}$ and $\overline{S}$ are the means and n is number of time steps. The significance of the correlation coefficient value being different from 0 can be investigated using the formula:

$$T = r\sqrt{\frac{n-2}{1-r^2}} \quad (10)$$

where $n-2$ is the degree of freedom. For a significance level of $\alpha = 0.05$ and $n = 20$ data points (Table 1) T should be either ≥ 2.101 or ≤ -2.101.

3.2 Hydrological modeling

3.2.1 NAM

NAM is a simple lumped rainfall runoff model, representing the hydrological system on catchment scale (Nielsen and Hansen, 1973; DHI, 2009a). NAM consists of several storages with different connections and properties: surface storage, root zone storage, groundwater storage and snow storage. The input data consists of daily average catchment precipitation, monthly potential evapotranspiration and daily temperature.

3.2.2 Model calibration

Calibration of the NAM is carried out using the global optimization algorithm AutoCal (Madsen, 2000) available in the MIKE-11 NAM modeling system. The objective function in the auto-calibration is defined as an even trade-off between water balance and Root Mean Square Error (RMSE) (DHI, 2009b), given as:

$$\mathrm{WB} = \left| \frac{1}{n} \sum_{i=1}^{n} (O_i - S_i) \right| \tag{11}$$

$$\mathrm{RMSE} = \sqrt{\frac{1}{n} \sum_{i=1}^{n} (O_i - S_i)^2} \tag{12}$$

where O_i and S_i are observed and simulated discharge at time i, respectively, and n is the number of time steps.

The performance of the model after auto-calibration is evaluated using these two performance statistics and three others describing the overall agreement between simulated and observed discharge values: the Pearson correlation coefficient (Eq. 9), the

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Nash-Sutcliffe efficiency NSE (Eq. 13; Nash and Sutcliffe, 1970) and the flow duration curve error index EI (Eq. 14; Refsgaard and Knudsen, 1996), all calculated for daily discharge values.

$$\mathrm{NSE} = 1 - \frac{\sum_{i=1}^{n}(O_i - S_i)^2}{\sum_{i=1}^{n}\left(O_i - \overline{O}\right)^2} \tag{13}$$

$$\mathrm{EI} = 1 - \int |f_\mathrm{O}(q) - f_\mathrm{S}(q)|\,\mathrm{d}q \Big/ \int f_\mathrm{O}(q)\,\mathrm{d}q \tag{14}$$

where $f_\mathrm{O}(q)$ is the observed flow duration curve and $f_\mathrm{S}(q)$ is the simulated.

3.3 Stream flow drought and flood index method

The threshold method was originally proposed by Yevjevich (1967) and has since been used in a number of drought studies. The method evaluates a discharge series by comparing each measured valued with an appropriate threshold calculated as a percentile of the flow duration curve. The threshold can be annual, seasonal or daily based and incorporating different degrees of severity depending on how the abnormal situation is identified (Stahl, 2001). Studies have used different threshold values for different purposes (Hannaford et al., 2010; Hisdal et al., 2001; Tallaksen et al., 1997). In this study a daily time step is applied in order to evaluate extreme events occurring within years (seasonal) and of a resolution of less than a month (Stahl, 2001; Tallaksen et al., 1997). Flood events have been evaluated using both measurement of the annual maximum peak flow (e.g. Cannarozzo et al., 1995) and different versions of discharge percentiles (e.g. Thielen et al., 2009). However for consistency the Yevjevich threshold method is here also applied to the high flow data, yielding a flow excess value instead of flow deficit.

Droughts are here defined as a discharge below the 70 % percentiles and flood events above the 10–30 % percentiles. The percentiles define a threshold between

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extreme and non-extreme situations. For both droughts and flood a total of three thresholds are defined, with the 70 %/30 % percentile as the least extreme and the 80 %/20 % and 90 %/10 % percentile as the more extreme thresholds. The most extreme drought threshold, Q_{90} is thus calculated as the 90th percentile of the observed flow or the flow value corresponding to an exceedance probability of 90 % on the flow duration curve (Hisdal et al., 2001b). A threshold value is calculated for each individual day, thus creating a daily threshold time series covering a year. The threshold procedure compensates for different discharge regimes and any natural variations in discharge caused by seasonal fluctuation.

In order to increase the sample size and hereby decrease uncertainty an enclosing 21 days time-window is incorporated (Fig. 2a). Thus the percentile value for 1 July 2000 is obtained on the basis of measurement from the period 10 days before and after the sample day. From the time-window and the sample day the threshold value is calculated as a percentile of the flow duration curve representing dry or wet extremes (Fig. 2b). For extreme drought the 90 % percentile (Q_{90}) is used. Additional percentiles can be applied to examine severe and moderate droughts (Q_{80} and Q_{70}) or flood events (Q_{10}, Q_{20} and Q_{30}). When daily threshold values for all 365 days are obtained an annual threshold curve for the data set is produced (Fig. 2c). The threshold curves are then compared to the hydrograph of the station. In order to remove errors of minor and mutually dependent droughts the threshold method is combined with an 11 days moving average pooling (MA-method) of the hydrograph (Hannaford et al., 2010; Hisdal et al., 2001; Tallaksen et al., 1997). The same procedure is applied for floods, except that a five days moving window is used. Days with measurements below drought or above flood thresholds are categorized dry/wet days. This information is then compiled in a two-dimensional distribution diagram, a Stream flow Deficiency Periods diagram (SDP) for droughts and a Stream flow Excess Periods diagram (SEP) for floods that show the occurrences of dry/wet days with time. For more information on threshold types refer to Hisdal et al. (2001b) and Stahl (2001).

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Choice of reference period

In index studies the threshold curves are typically based on the whole data period or an appropriate reference period representing the transition between a normal and abnormal situation. However, when studying climate change the discharge values may change from the beginning to the end of the period. For an overall discharge increase this will make the low flow days in the end of the period appear relatively wet, when compared to the full period. Thus different reference periods will result in different extreme events.

When the threshold curve is based on the full record, the available data are exploited to its fullest, but this is problematic if the flow regime is non-stationary. One way to deal with this is to define an arbitrary "normal" or "reference" period. It is not always apparent what normal is, and it has to be based on assumptions. In climate change context the 30 yr period from 1961–1990 is often defined as the reference period. This period is characterized by having a close to complete data base and subject to relatively small measurement uncertainties. However, the conditions in this period may not be stationary, because CO_2-concentrations have increased dramatically since the 1960s. Therefore, it could be argued that the oldest data in the data set should be selected as reference. However, data from this period is affected by measuring errors to a larger degree, and data for stream discharge is sparse. Another approach is to remove the trend in discharge by de-trending the whole data set. When the data are de-trended the effect of averaging over older or younger flow periods is eliminated. This enables the registration of relative drought events in the last years of the data set that would otherwise have been masked. In this study the full, the 1961–1990 and the detrended reference period are evaluated.

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4 Analysis of trends

4.1 Precipitation

Figure 3a shows the changes in precipitation in the catchment during 1875–2007. The overall annual average rainfall has increased by 346 mm with a rate of 2.62 mm yr^{-1}, corresponding to an increase of 46 % (Table 2). This increase is very high and as such some additional investigations were carried out to evaluate the credibility of the increase. First, it was documented that the increase was not an artefact of the Thiessen averaging to catchment precipitation; as the four precipitation stations also individual show increases between 40–51 % during the period (not shown; see Fig. 4 for the increase in the period 1920–2007). Secondly; some bias on the regression line may be introduced by using supplementary stations for missing time slices in the four primary stations. However, the two stations with the largest coverage in the catchment (23050 and 23220) also require the least supplements; and so this bias is thought to be of less significance to the overall catchment precipitation result. Thirdly; a test of seven additional precipitation stations (21100, 25140, 26400, 21430, 27080, 6186 and 6193) in Denmark (Cappelen et al. 2008, 2009) was carried out to evaluate the spacial differentiation (not shown). The stations have fixed catch correction factors (the same as the Skjern data) and data from 1920–2007 (apart from station 26400 and 21430 that are missing 2 and 8 yr of data respectively). Relatively large differences in the development of precipitation are found; four out of the seven stations have increases around +14 %, where the last three are −4, +5 and +67 %. The four precipitation stations from this study do show large increases (+14, +35, +39 and +41 %), but they are exceeded by station 26400 in southern Denmark (+67 %); and the increases can therefore not be dismissed as unrealistic. Looking at the special distribution of the stations there seems to be a tendency for the largest precipitation increases occurring for stations located in the most rain prone areas in Denmark; as is the case for the stations in this study.

Seasonally, the largest increase in the catchment occur in November, December, January, and February, while August shows a significant decrease in precipitation

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(Fig. 3, right panels). The other summer months show smaller changes, some statistically non-significant. Hence, precipitation changes have enhanced the seasonal difference between summer and winter, making winter a wetter season and summer relatively drier.

The amounts of snow-fall (precipitation amounts for temperatures below zero degrees) show no significant change over the analysed period.

Precipitation can also be analyzed with respect to the number of precipitation days. A precipitation day is here defined as a day with more than 1 mm of precipitation (Heino et al., 2008). As mentioned previously the precipitation data originates from four primary stations with additional stations to fill in missing data; these four stations were then averaged to obtain the catchment rainfall. The average rainfall is suitable for modeling as it should approximate actual average rainfall over the area; it is also useful when analyzing annual or monthly values. However, when looking at occurrences and strength of precipitation events on daily basis average precipitation is misleading as phantom events are created due to the averaging. Therefore, stations are in this case treated separately and all months that include averaged precipitation data are removed before the analysis.

The annual number of precipitation days in the catchment has increased significantly from an average of 80 precipitation days in the first five years to 141 in the last (Fig. 4a) corresponding to an increase of almost 1 precipitation day every second year. This increase is primarily caused by an increase in wet days during the winter months (Fig. 4d). In Fig. 5b the percentage trends of the regression line from 1875 to 2007 for five different categories of events as well as the total is plotted. The largest percentage increase is found for precipitation days with volumes from 1–5 mm, and from 15–20 mm. This indicates that previously non-precipitation days (days with precipitation below 1 mm) now receives precipitation and are categorized as 1–5 mm precipitation days contributing to the increase of precipitation events in general. The increase in the 15–20 mm indicates that more precipitation events have moved to this category making more events wetter. However, the 1–5 mm category accounts for around 23 % of

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the total volume of precipitation while the 15–20 mm only contributes with 12 %, hence indicating that the increase in precipitation is primarily due to the enhanced number of wet days and not to generally wetter days. It is unfortunately not possible to look at single rain event characteristics as only data with daily resolution are available; therefore it cannot be concluded whether or not intensity and occurrences of individual events have changed.

4.2 Temperature

Temperature in the area has increased by 1.4 or 0.01 °C yr^{-1} during the period (Fig. 3b). In contrast to precipitation changes the increase is distributed fairly even through the year, thus making all months on average equally warmer (non-significant change in variance).

4.3 Potential evapotranspiration

As evapotranspiration is calculated as a function of temperature it is not surprising that the signal resembles the temperature (see Fig. 3b and c). The increase in potential evapotranspiration during the period is statistically significant (Table 2) and amounts to 0.19 mm yr^{-1} or 25.1 mm (4 %) during the whole period. When looking at the individual months only March, April, and August shows significant trends, while October, November and December consist of Penman averages and therefore have no trend. Keeping this in mind it is clear that the significant trends are equivalent to the months with the highest temperature increase. The (non-significant) negative trend in June may be surprising as a temperature increase (even a low one) in this month should intuitively be reflected in an evapotranspiration increase. However, due to the formulation of Thornthwaites (Eqs. 1–4), the monthly temperature is related to the annual temperature through the index I_y. As the temperature increase in June is lower than the average annual increase, this result in a weak but negative evapotranspiration trend.

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5 Calibration results and model evaluation

Calibration of the NAM model was initially done for the period 1961–1970. The following five most important parameters were calibrated: U_{max}, L_{max}, CQOF, $\mathrm{CK}_{1.2}$ and CKBF. The U_{max} is the maximum storage capacity of the surface reservoir, while the L_{max} is the storage capacity of the root zone reservoir. These two parameters thus determine the amount of water hold in root and surface zones available for evapotranspiration. The CQOF is the overland flow runoff coefficient and it determines the proportion of water that infiltrates relative to the proportion that is routed as overland flow when U_{max} is exceeded. The $\mathrm{CK}_{1.2}$ and $\mathrm{CK}_{\mathrm{BF}}$ are both time constants for overland flow and baseflow, respectively, and determine the time it takes for the overland and baseflow to reach the stream. The specified parameter ranges were [5 : 35 mm], [50 : 400 mm], [0 : 1], [3 : 72 h] and [500 : 40 000 h], respectively, based on recommendations by DHI (2009b). The upper limit for the CKBF value has been assessed from Hansen et al. (1977) who analysed the parameter uncertainty of NAM for the Skjern River catchment by use of an automatic parameter optimization routine. Initial parameter estimates are listed in Table 3 and the results from the calibration period are listed in Table 4.

5.1 Model calibration

The model calibration results in a perfect water balance (Table 4) and also the performance parameters indicate an excellent overall performance. Visually the simulated hydrograph matches the observations very well (Fig. 5) with the exception of a distinct high peak in April 1970. This peak was generated by melting of an exceptionally large snowpack caused by a combination of a sudden shift to high temperatures and heavy rainfall. It appears that the degree-day snowmelt approach may not be suitable for this particular weather situation.

5.2 Parameter representation

The model performance for the individual 10-yr periods using the parameter estimates from the calibration period 1961-1970 is shown in Table 4. With respect to the water balance the model severely overestimate the volume of water in the last four decades, while there is a slight tendency to underestimation in the first four. This tendency is also reflected in the parameter values when calibrating on each of the nine 10-yr periods (Table 5). To compensate for the excessive simulated discharge in the four most recent decades the model attempts to balance this by increasing the capacities of the surface and the root zone storages (U_{max} and L_{max}) and in this way increase evapotranspiration. In all four cases the maximum values of L_{max} is reached and therefore the total discharge cannot be reduced sufficiently to obtain a small error.

The model ergo shows good performance for the earlier time periods based on model parameters from 1961–1970, while for the periods after the calibration period the model perform less satisfactory. When applying individual parameter sets estimated from each of the nine periods to simulate the full time series from 1875–2007 (Table 6) the four parameter sets representing the early periods shows water balance errors generally below 10 % before 1971 (upper left) while the periods after 1971 show large excess of water in the model (upper right). This indicates that a parameter set from any of the first five periods can be used as basis for simulation back to 1875 and the choice of the (1961–1970) calibration period is therefore maintained.

However, when looking at the parameter sets found for the four last periods (lower half of Table 6) it is clear that these results in large water balance errors. For the earliest periods the model underestimates the outflow while for the later periods too much water is produced.

The results indicate that a significant change is realised around 1970 and that some factors in addition to the climatic changes must be involved. The model is primarily driven by precipitation and temperature input and as such does not account for anthropogenic changes like irrigation and land use changes. However, the observed discharge may be influenced by human induced changes, as well as climate changes. Deviations between model simulations and observations outside the calibration period, especially for the periods after 1961–1970, could therefore be a reflection of anthropogenic impact.

5.2.1 Anthropogenic factors

As mentioned the anthropogenic changes during the considered period are relatively small compared to other parts of Denmark, but human activity has affected the area. The anthropogenic influences in the area can be classified in four overall categories; (1) conversion of moor/heath dominated areas to farmland and forest; (2) implementation of drainage systems; (3) large scale irrigation; (4) increasing use of commercial and livestock manure fertilization.

The land use changes in the area have occurred primarily before 1900, where large areas of the heath landscape were changed to agriculture and forest plantations. From the 1900s to the 1930s the development continued with a slowly ongoing shift towards more farmland. After 1930s only limited land use changes occurred. This means that the land use changes in the area most likely are not the direct cause for the sudden model deviations of the later periods, as the land use changes from 1960 and forward are only small and mostly gradual in nature (Fritzbøger, 2009).

Three types of drainage were done in the catchment: Drainage of fields, weed cutting in the stream and regulation of the stream. Draining of fields was originally done by digging smaller trenches and ditches (early 1800s), later replaced by underground drain pipes (mid 1800s). In the Skjern River catchment however little draining was needed in the permeable sand and gravel deposits; the drain system in 1861 in the fields in Skjern River catchment covered only 2 ‰ of the agricultural land); where the majority of the drainage took place on the old moraine hills to the northwest of the area. Due to this small size it is not consider relevant for the sizable water balance error. Weed cutting in the stream began already in the 1760 and was originally the responsibility of the landowner, but was later transferred to the municipalities and still is. The regulation

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of the stream itself by straightening and deepening was by no means small in the area, and newer records show that more than half of the stream network has be regulated (Ringkjøbing Amt, 2004). However for all three drainage types there is no indication that practices have changed significantly in the model simulation period.

The reason why the model performs worse in the later periods may be explained partly by the pumping for irrigation. Assuming that return flow is negligible, sprinkler irrigation will increase evapotranspiration with an amount equal to the pumping. The irrigation practices were intensified after severe droughts in 1975–1976. Clark et al. (1992) reported high irrigation use from the mid 1970s, increasing all the way to the 1980s where the current level was reached. No reliable data on amounts of groundwater abstraction for irrigation in the area are available, but previous studies have assumed around 15–30 mm yr^{-1} (Stisen et al., 2011), corresponding to 2–6 % of the simulated discharge amount for the last four periods. In Kronvang et al. (2006) the estimate of a water abstraction value for the area is based on the maximal pumping permissions (in 2000) yielding a total of 54 mm yr^{-1} corresponding to roughly 10 % of simulated discharge; however the water balance errors are in the range of 16–28 %. Therefore seemingly, groundwater abstraction for irrigation cannot alone explain the water balance error.

Clark et al. (1992) examined historical trends in precipitation, evapotranspiration and runoff in nine Danish catchments from 1920–1990. They also found indications of changes in the hydrological regime after 1960 and concluded that a possible explanation might be an increase in crop yield due to increasing use of fertilizers and pesticides, leading to an increase in leaf area index and in the efficiency of the leaf surface. These effects may in turn result in higher actual evapotranspiration. The increase in use of mineral fertilizers began in the 1950s (Clark et al., 1992) with a significant increase during the 1960s and the 1970s. Therefore, the described effect may have an impact on the water balance from this time. This could potentially have influenced the calibration of the parameters for the period 1961–1970 and thus lead to simulations of an excess of water in the earlier periods where fertilizer use is lower and thereby

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a relatively lower evapotranspiration. So the calibration period 1961–1970 may represent a transition between the earlier less influenced periods and the late highly affected decades.

This implies that even though the land use catchment is generally considered relatively constant for the last 150 yr, the minor anthropogenic activities in the area seem to have a profound effect on the catchment runoff after the 1970s.

5.2.2 Data biases

Apart from physical and management changes in the catchment, it is important to keep in mind that measured input data can contain errors or biases that can influence the model results.

An additional reason for the shortcomings of the model could originate in the bias correction of precipitation observations. Analyses reveal that the amount of precipitation falling as snow shows no significant trend since 1875, no matter if 0 or 1 °C is used as threshold factor for differentiating between snow/rain. The bias correction factors for the winter months have been derived from a combination of rain (low correction) and snow (high correction) representing the period 1961–1990 (Allerup et al., 1998). As the snow fall has been constant, while the total precipitation has increased significantly since 1875, the winter precipitation in the earlier days had a higher snow fraction; and thus a higher correction factor should be applied. The opposite applies for the most recent period 1991–2007, which has a lower snow fraction than 1961–1990. Inaccurate corrections factors can therefore be a contributor to the water balance error.

To evaluate the contribution to the water balance error as a result of the precipitation bias correction, a simple test using the dynamic correction model by Allerup et al. (1997) is applied. A constant wind speed of 5 m/s is assumed, as no wind speed data is available back to 1875. It is assumed that precipitation falling at temperatures below 0° is solid; while it is liquid above 2°; a linear interpolation is used between the two. When looking at the result of the simple test (not shown) divided into 30 yr periods (like the standard correction) is it shown that the periods generally follow the standard correction fairly well; except for the two periods 1875–1900 and 1931–1960. The deviation result in an annual linear slope of 2.29 mm yr^{-1} instead of 2.62 mm yr^{-1}, corresponding to overestimation of the trend of around 14 %. As it must be assumed that the standard correction best represent the period it is build on (1961–1990), precipitation is ergo underestimated back in time.

It is however not appropriate to used the dynamic correction throughout this study as no information is available on wind speed and as such a large uncertainty is associated with the dynamic correction. The simple test is therefore only used to evaluate the effect of the temperature increase on the correction values.

5.2.3 Additional error sources

Other sources of errors influencing the water balance both before and after the calibration period include bias in precipitation due to incorrect shelter class or changing shelter class during the period; uncertainties in the manual records and in the precipitation sums in the collection sheets; the assumption that the potential evapotranspiration based on data from Foulum can be used as representative of the Skjern River catchment and for fitting the Thornthwaite model. Errors are of course also introduced when using a hydrological model to represent a natural system as simplification of the system must be made during the process. However, these error sources are not necessarily influencing the simulation results in one specific direction but are more likely results in randomly distributed errors.

5.3 Analysis of trends – discharge and recharge

The time series of discharge starts in 1920 while the precipitation records date back to 1875. To supplement the missing data the model is used for simulating discharge based on the precipitation observations. The simulated discharge series show an increase in discharge of 2.2 mm yr^{-1} (Fig. 3d), while the observations only increase by 1.2 mm yr^{-1}. If the model deviations in the later periods after the calibration is a result

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of anthropogenic changes, the increase of discharge found in the simulations represent how the catchment discharge would have developed as a result of the climatic signal only without the human interference; but because of the human induced changes the observed discharge is lowered.

The distribution of the change in discharge with season suggests that winter and early spring have experienced the highest increase. However, even though precipitation has decreased in the summer months, discharge still increases in these months due to the buffering effect of the groundwater system in response to the recharge, which mainly occurs during the winter season (Larsen et al., 2003). The groundwater recharge increases by 1.97 mm yr^{-1} during the period and mainly during the winter season. The seasonal distribution reflects the distribution of precipitation increase.

6 Stream flow droughts and floods indices

The model simulations (based on the estimated parameters set) of discharge for the entire period, 1875–2007, are used in the analysis of extreme events. The impact of extremes varies with season, as some are more or less sensitive to extremes, but also with duration (time below the threshold), severity (deficit volume; amount of water *missing* to exceed the threshold into normal conditions) and intensity (severity/duration). Using the simulated discharge from the NAM model introduces some uncertainties; however here it is assumed that the model represent the *pre-1970* discharge signal well, however as it does not take recent human induced changes into account; the drought/flood signal will not represent actual changes in extremes after 1970.

6.1 Evaluation of reference period

The distribution of drought and floods in the catchment is evaluated using the SDP/SEP-diagrams (Stream flow Deficiency/Excess Period) (Figs. 6 and 7). Three different types of reference periods are evaluated for extreme occurrences all with Q_{90},

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Q_{80} and Q_{70} as limits for drought and Q_{10}, Q_{20} and Q_{30} as limits for flood. Reference periods are calculated using (A) full data set, (B) data from the period 1961–1990 and (C) full de-trended data set. The results are supplemented by the calculation of the annual dry/wet day count obtained by counting the number of days $< Q_{70}$ or $> Q_{30}$ for every year, this is also sometimes referred to as annual cumulated duration (corresponding to the ACD in Hisdal et al., 2001). This threshold is chosen to ensure that as many years as possible have extremes to improve the trend analysis (Wilson et al., 2010). This is also done on a monthly basis to evaluate the seasonal distribution of the wet/dry days (monthly cumulated duration), in addition to an analysis of monthly increase.

When the threshold are based on the whole data series the diagram shows the variation of extremes around the period mean (Figs. 6a and 7a). Here there is a tendency to fewer droughts with time and more floods; and relative to the whole period the 1885–1897 is extremely dry (more than 50 % dry days; Fig. 6-a2). The plots illustrate that due to the high non-stationarity of the catchment extremes at the ends of the period are obscured by the threshold representing an average over the full period.

Figures 6b and 7b show changes compared to the 1961–1990 situations, and are therefore an expression of changes in relation to a known situation. Again the drought occurrence seems to decrease and the flood occurrence increasing with time. The period of 1961–1990 is generally wet and as a consequence, the early years have a very high amount of drought episodes/few floods, when compared to this interval; as with the full reference period this is reflected in a blurred extreme signal, in this case especially in the old periods.

In the de-trended diagrams (Figs. 6c and 7c) the resulting extremes are an indication of changes in the amplitude of the hydrograph, meaning the relative extremes. This is reflected in a small increase in drought, while there is no significant increase in flood. This indicates that even though discharge generally has increased the amplitude of the hydrograph (for the low flows) has also increased giving rise to more relative droughts

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also in the later wetter years. The high number of drought occurrences in the beginning of the period has also been removed by the de-trending procedure.

The choice of reference period is important as both distribution (Figs. 6- and 7-a1/b1/c1) and the direction of the change (Fig. 6- and 7-A2/B2/C2) depends on the reference period used to calculate the threshold values. The de-trended reference type seems to reflect the extremes in the catchment better than the two previous, as a dry/wet day in a water rich environment (late periods) does not have the same definition as a dry/wet day in an arid (early periods); or as extremes are measured in comparison to the "normal" situation where the normal situation in this case is changing. Evaluation of the extremes will therefore from here on be focused on the results from the de-trended reference type method.

Even though the reference choice changes the extreme signal, it should be noted that some extremes are preserved in all plots. This indicates that some flood/drought extremes are stronger or more pronounced than others irrespective of the reference period. Examples of this are the 1932 and the 1947 drought and the 1995 and 2002 floods.

6.2 Occurrences of extremes

From the SDP- and the SEP-diagrams of the Skjern River catchment the extremes can be evaluated.

The most pronounced drought years (more than 2/3 days with dry conditions) are: 1933–1934, 1947–1948, 1964, 1996–1997 and 2003 (Fig. 6-c2); while wetter years are 1877, 1981–1984 and 1988 with a high number of floods (Fig. 7-c2). The distribution of dry/wet days over the season is fairly uniform (Figs. 6-c3 and 7-c3), where the significant changes in dry days with season show increase primarily occurring in June–October, and the significant changes in wet days are limited to January and July–August. As mentioned the ACD shows a significant increase in drought ($u_c = 2.8$; with $u_{1-\alpha/2} = 1.96$), while the flood change is non-significant ($u_c = 0.8$). However, this is including the anthropogenic *undisturbed* simulated discharge signal after 1970. When

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disregarding the last 37 yr the drought signal still shows a significant increase ($u_c = 5.2$; with $u_{1-\alpha/2} = 1.985$), and the decrease in the flood signal now becomes significant ($u_c = 3.9$).

The question of how good the indices and reference types represent the actual human registered drought/floods is difficult to answer as historical information on extremes is scarce. Little information on drought events is available, and information on flooding is even more difficult to obtain. Furthermore often flood events are only reported when they lead to overflows, and are therefore very local phenomena. In this section the little information on actually occurring extremes in Denmark is focused on the drought events. Locally in Denmark extreme dry conditions were reported in 1899, 1947, 1959, 1976, 1992 and 1995–1997. Common for the entire reported drought incidences are that they occurred during summer (May–August/September) and in combination with high temperatures and/or high sun hours. Forest fires, sand storms, lost crops and water scarcity have been reported in connection with the droughts (Feyen and Danker, 2009; Hansen, 1992).

One of the most pronounced droughts in Denmark is the 1947 drought. The drought was characterized by very little rain from the beginning of May and very high temperatures though June and partly July, followed by an extremely warm, dry and sunny August. The dry condition led to several smaller fires and was described as a catastrophe by the Danish minister of agriculture (Hansen, 1992). As mentioned, the event shows in the stream flow index (Fig. 6), were the dry conditions last from May to November, and again in most of December. The indices report this as the most prolonged and second most severe dry condition in the study period, lasting 195 days (below Q_{70}, not counting the December dries), with a severity of 37 mm and intensity of 0.19 mm day^{-1}.

Even though the simulated stream flow does not represent actual stream discharge after 1970, the drought indices capture the 1976, 1996 and 1997 droughts. The reason why all the early droughts (before 1970) are not captured on the top ten duration or severity, or scoring high on the ACD, may be due to errors in the methodology of identifying the droughts via the index method or it might indicate that even though the

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index captures some of the historical reported droughts; other factors than severity and duration of the event may influence whether or not an event is registered in the public (factors as timing, water demands and previous years water conditions).

7 Climate change drivers

According to IPCC anthropogenic climate change is mainly related to increased green house gas emissions (IPCC, 2007a), which has caused an increase in the global temperature after 1960. However the time frame of this study goes back to the 1870s before greenhouse gas emission really took flight. The long term historical change in temperature and precipitation for the Skjern River area must therefore be driven by other factors. Heino et al. (2008) studied the spatial patterns of historical precipitation change over the Baltic Sea region and found clear differences in distribution and signal of the changes. The study suggested that atmospheric circulation changes might explain the changes as areas exposed to humid westerly winds during winter experience large increases in precipitation. This was also found by Schmith (2001) who recognized a link between circulation patterns and variability of winter precipitation in North-western Europe. Therefore, the correlation of circulation patterns and the observed variables in Skjern River catchment were analysed. Both the correlation between the climate indices and (a) the climate variables (precipitation and discharge) and (b) the result from the extreme analysis of the stream flow (in the form of cumulated duration per month) were investigated.

One of the larger climatic drivers for the northern hemisphere is the North Atlantic Oscillation (NAO). The NAO describes a large scale weather system determining the strength of the westerly winds blowing towards Europe. The NAO-index represents the strength of the wind system measured as the normalized difference in pressure between Iceland and the Azores. During a period with positive NAO-index the pressure difference is large creating strong winds from the west bringing warm, moist air to northern Europe, while a negative NAO index indicates smaller pressure difference

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leading to weaker westerly winds resulting in colder air and less precipitation in Europe during winter (Stahl et al., 2011; Sutton and Hudson, 2003).

The NAO index from the 1900 to the 1930 was generally positive, making winters moist and warm. After 1940s and up to the 1970s the NAO index changed into a negative mode leading to colder and drier winters. From 1980 the NAO index has again changed direction to the positive mode (Hurrell, 1995a). After the early 1990s the index shifted to a negative or neutral mode; however the response in winter climate has been less clear until now leading to no obvious change towards colder, drier winters in Europe (Seidenkrantz et al., 2009). The lack of response is generally being attributed to the overlaying impact of the anthropogenic originated climate signal. Hannaford et al. (2010) examined the use of climate indices for drought forecasting in Europe, and found that Scandinavian droughts was primarily driven by the Scandinavian Pattern (SCA) first described by Barnston and Livezey (1987). The SCA describes a circulation pattern centred over Scandinavia. Another climate driver for the northern hemisphere is the Atlantic Multidecadal Oscillation (AMO), defined as a 60–90 yr change in North Atlantic Sea surface temperature. Debate on the actual nature of the driver is still ongoing and the question of whether or not it represents a true oscillation is still controversial (Knudsen et al., 2010). Warm phases (positive) of the AMO was prevailing in 1860–1880 and 1930–1960 and cold phases (negative) in 1905–1925 and 1970–1990, from the year 1995 and up to now the AMO has been in a warm phase (McGabe et al., 2004).

The correlation between the NAO index (data from Hurrell, 1995b) and precipitation/stream flow can be seen in Fig. 8a while the correlation to drought/flood can be seen on Fig. 8d. The correlation is generally stronger in winter than during summer, and the correlation in most of the winter months are significant (Eq. 9). This is not surprising as the NAO is primarily governing the winter climate. Generally the correlations to precipitation are better than to discharge, reflected also in the lower correlation between NAO and the stream flow extremes. When only looking at the NAO winter index and the DJFM extreme count (ACD with only the months DJFM) the correlations between NAO and the stream flow index are −0.30 for droughts and 0.37 for floods. The correlation between the NAO index and the drought and flood indices seems to indicate that there is a certain amount of influence from the NAO signal to the extreme response, however only 9–14 % (r^2) of the winter discharge extremes can be explained by NAO changes.

With respect to the SCA (data from NOAA, 2011) a positive correlation for drought and negative for flood are found, as expected since the sign of the dry condition for this index is the opposite to the NAO index (Fig. 8b and e). Like the NAO only some of the months have significant correlation (note as SCA are based on fewer values the limit for significance is higher). The influence of the SCA is more pronounced for precipitation in the winter and affects the discharge more in the late summer/fall. The SCA also seems to influence flood occurrences more than droughts, and the influence is more pronounced in the mid-winter and late summer/fall. For the AMO (data from NOAA, 2012) correlations are relatively weak (Fig. 8c and f), implying that the AMO does not contain much information that can explain drought and flood.

Newer studies have shown that Weather Type Classification (WTC) containing several climate drivers (or Weather Types) is a better representation of hydrological drought events in North-western Europe (Fleig et al., 2011) suggesting that individual climate drivers are not as good representatives of drought and flood as a compounded climate driver. Similarly, Schmith (2001) showed that the response in winter precipitation of the North-western Europe could be well simulated using the winter-mean mean sea level pressure (applying a principal component analysis with five leading WT components) while a simulation using only the NAO-index (one leading WT component) was not as good. Similar results were obtained by Kronvang et al. (2006) that found only 20 % (r^2) of winter precipitation in Denmark could be correlated to the winter NAO-index (here 17 %) and an even weaker link to discharge (8–10 %, here 7 %). This indicates that the NAO indices may hold too little information about the winter precipitation to be further linked to drought/flood occurrences in Denmark.

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8 Discussion

8.1 The 130 yr data series

In this study long series of historical data of precipitation, temperature and discharge were available. The discharge series were extended to cover the full 133 yr of the data set using the NAM model. The time series show significant climate change with 46 % increase in precipitation, 1.3 °C increase in temperature and 103% increase in discharge.

In Denmark the climate change with respect to precipitation is well known (Jeppesen et al., 2011), and trends in discharge series are also documented (e.g. Ovesen et al., 2000). Precipitation changes from Denmark are reported in Jeppesen et al. (2011) for the Copenhagen area (East Denmark) in the order of 129 mm yr^{-1} corresponding to 20 % (1850 to 2003); while the precipitation trend in Skjern catchment is considerably steeper. But as mentioned, it has previously been reported that the precipitation increase have been unevenly distributed in Denmark (Jørgensen and Cappelen, 2006; Kronvang et al., 2006); and preliminary investigations shows that the increase seems to be largest in the areas generally receiving the highest precipitation amounts.

Van Roosmalen et al. (2007) simulated future climate change for the Skjern River Catchment using the DK-model (MIKE SHE based) to evaluate consequences of climate change in the scenario period 2071–2100 (using the A2 and B2 scenarios). Their simulation showed precipitation increases of 12–16 %, and discharge increase of 13–20 %, with a tendency towards increasing discharge in winter and lower discharge in late summer and fall. Thus, the historical precipitation and discharge changes are much higher than the future expected, with the seasonal distribution of the change following the same pattern. This is interesting as the anthropogenic adaption response to these historic climatic changes have been non-dramatic.

The Skjern Catchment is one of the Danish catchments experiencing the largest historical change in precipitation and discharge, and additionally one of few with longer time series. This makes the data set unique in a Danish context, but also to our

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knowledge in an international sense. The data series are suitable for (i) evaluation of past climate change, (ii) test of climate models' capabilities to predict climate change; and (iii) test of hydrological models' capabilities to predict climate change impacts on river discharge.

8.2 Can climate change be explained by climate models?

The climatic changes in the Skjern River catchment have been significant and long term and cannot be explained by changes in greenhouse gasses, but should more likely be explained by changed atmospheric circulation patterns in Northern Europe. The known climatic drivers such as the SCA-, NAO- and AMO-indices can explain some of the change, but only a relatively small part of it. The rest we cannot explain with the present knowledge, and no climate model has so far been able to explain this historical climate change.

8.3 Can the recorded hydrological change be explained by the recorded climate change?

NAM is generally a suitable tool to analyze how variability in precipitation and potential evapotranspiration generates variability in discharge. NAM performed equally well as compared to more complex models as MIKE SHE in various differential split-sample tests of calibration on wet periods and validation on dry periods (Refsgaard and Knudsen, 1996). Furthermore, NAM was successfully used to distinguish between effects of climate variability and effects of land use change on runoff in Zimbabwean catchments (Lørup et al., 1998). The test in the present study evaluates the capability of the hydrological model to predict climate change impacts on runoff. It is different from the tests made by other studies (Refsgaard and Knudsen, 1996; Lørup et al., 1998; Seibert, 2003), which only tested the impacts of climate variability. Here we have tested the impacts on long term changes.

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The results show that the model performance is reduced for other periods compared to the calibration period. Within the period 1920–1970 most model runs performed well (WB < 10 %), regardless of the calibration period; and calibration parameters (from 1961–1970) was therefore acceptable for extension of the discharge record back to 1875. However, the NAM tests also indicate that we are not able to explain all recorded hydrological changes especially in recent time, even for a catchment with a relatively constant land use (for the last 150 yr). We have raised hypotheses that may explain the reason for this, including changes in the irrigation practice from the mid-1970s and increased evapotranspiration because of fertilization use. But it should be stressed that fully testing of these hypotheses would require further research. Nevertheless, the results imply that we cannot necessarily expect that hydrological models are able to project climate change impacts on runoff as accurately as they can predict the present situation. Reasons for this include ongoing anthropogenic changes both directly in the form of pumping, draining and so on, which may to some extent be possible to assess for future scenarios; but possibly also more unquantifiable though increasing plant productivity and land use change. Some of these changes may be driven by technological development while others are results of measures to adapt to the changes in climate. This aspect is most often ignored in the many studies of climate change impacts on hydrology.

Even for the period 1920–1970; where relatively few hydrological and anthropogenic changes are occurring and the model deviations are generally low; there is a tendency to increasing water balance error the further away from the calibration period, the model is used. These deviations may or may not be due to the use of calibration parameters used for precipitation regimes different from the calibration period. However, these deviations are completely obscured in the later periods by the much larger effects of the anthropogenic changes.

8.4 Extremes in a non-stationary climate

The analyses of the 133 yr time series illustrate how the drought and flood indices change significantly over time when climate is changing. Discharge has increased during the period, but even so the stream flow extremes also show an increase in number of relative droughts, while there is a significant negative trend in floods until 1970. Investigations on flood occurrences are scarce for Denmark, but for Sweden and Central Europe analyses have shown insignificant or no upwards trend in actual flooding events (Lindström and Bergström, 2004; Mundelsee et al., 2003). Only a few significant drought trends have been found for Denmark (Hisdal et al., 2001a). Stahl (2001) and Hisdal et al. (2001a) both reported an effect of reference period choice on extreme signal and occurrence. In this study results showed that this effect is even more important when the climate is highly non-stationary.

It should be stressed that the choice of reference period depends highly on the purpose of the study as all three reference periods hold information. However, for this study the very high degree of changes in the water regime advocates for a de-trended reference type (as both A and B will mask early or late periods respectively). Generally, this study points to using reference type C, when looking at a changing climate, however for future simulation type B might be more appropriate as the reference period is fixed to a known situation.

9 Conclusions

The Skjern River catchment in western Denmark provides a unique time series both with respect to length and resolution of the time series, minimal anthropogenic changes during the period considered and the largest Danish recorded climate change. Since 1875 the precipitation has increased by 46 % and discharge by 103 % (simulated), demonstrating the high non-stationarity of the climatic setting. The recorded changes are considerably larger than those expected from future climate change (van

Roosmalen et al., 2007). Only a small part of the past climatic changes can be explained by the climatic drivers SCA, NAO and AMO. The rest cannot be explained with current knowledge and no current climate model has so far been able to reproduce the historical climate change signal.

By investigating the performance of the well proven hydrological model (NAM) outside the calibration period a test of its capability to predict climate change impacts on river discharge was conducted. The results showed that the model performance deteriorated somewhat compared to the calibration period, indicating that not all hydrological changes could be explained. Possible reasons for the reduced model performance include enhanced crop yield leading to higher actual evapotranspiration, initiation of irrigation after 1975 and inadequate correction of precipitation data for undercatch. These results indicate that we cannot expect hydrological models to predict climate change impacts on discharge as accurately in the future as they can predict the present condition due mostly to anthropogenic changes.

Extremes in the non-stationary climate were evaluated using the simulated discharge from 1875–2007. The evaluation of the extreme signal and classification indicated that relative drought occurrences have increased in 1875–1970, while floods decreased. Most flood and drought indices depend on selection of a particular reference period, which is particularly problematic in case of a non-stationary climate. For studies aiming at analyzing present and past regimes we suggest to detrend the climate series, while use of a recent reference period is recommended for studies of future climate changes.

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Table 1. Evaluation of Thornwaite calculations compared to Penman as input of potential evapotranspiration data.

	Jan.	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug.	Sep.	Oct.	Nov.	Dec.
Correlation, r	0.56	0.59	0.73	0.56	0.66	0.70	0.52	0.64	0.47	0.03	0.31	0.29
Slope, b	0.6	0.4	0.8	1.0	1.3	1.9	1.2	1.2	0.8	0.0	0.3	0.2
Interception, a	7.5	10.8	19.0	16.5	−8.2	−104	−41.9	−38.9	−3.3	27.8	8.6	7.7
Test value, t	2.8	2.8	4.5	2.9	3.5	4.1	2.4	3.4	2.2	0.1	1.3	1.3
Table value, t_c	2.1	2.1	2.1	2.1	2.1	2.1	2.1	2.1	2.1	2.1	2.1	2.1
$\alpha = 0.05$	Yes	Yes	Yes	Yes	Yes	Yes	Yes	Yes	Yes	No	No	No

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Table 2. Overview of parameter average and statistics.

Variable	Unit	Average	Slope of regression line	Mann–Kendall, u_c	Rejection of H0?	% Increase*
Precipitation	[$mm\,yr^{-1}$]	925	2.62	7.5	Yes	46
Temperature	[$^{\circ}C\,yr^{-1}$]	8.1	0.01	5.6	Yes	20
Pot. evapotranspiration	[$mm\,yr^{-1}$]	609	0.19	2.7	Yes	4
Act. evapotranspiration	[$mm\,yr^{-1}$]	487	0.24	2.8	Yes	7
Discharge	[$mm\,yr^{-1}$]	435	2.2	11.5	Yes	103
Recharge	[$mm\,yr^{-1}$]	281	2.0	8.1	Yes	172

* The percent increase is calculated as the difference in the regression line from 1875 to 2007.

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Table 3. Initial parameter values for the NAM model.

Parameter	Description	Initial values
U_{max}	Maximum water content in surface storage [mm]	20
L_{max}	Maximum water content in root zone storage [mm]	120
CQOF	Overland flow run-off coefficient [()]	0.15
CKIF	Time constant for routing inter flow [h]	960
$CK_{1,2}$	Time constant for routing overland flow [h]	40
TOF	Root zone threshold value for overland flow [()]	0.1
TIF	Root zone threshold value for inter flow [()]	0
TG	Root zone threshold value for groundwater recharge [()]	0.4
CKBF	Time constant for routing base flow [h]	15 000
C_{area}	Ratio between groundwater and the topographical catchment [()]	1
C_{snow}	Constant degree day coefficient [mm °C^{-1} day^{-1}]	2
T_0	Base temperature (distinction between precipitation in rain/snow) [°C]	0

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Table 4. Model performance using parameter estimates found from calibration on 1961–1970 (WB – relative Water Balance; RMSE – Root Mean Square Error; NSE – Nash-Sutcliffe coefficient; *r* – Correlation coefficient; EI – Error Index).

Period	Average			Objective function		Performance parameters		
	Precipitation	Actual ET	Discharge	WB	RMSE	NSE	*r*	EI
Units	[mm yr^{-1}]	[mm yr^{-1}]	[mm yr^{-1}]	[%]	[mm day^{-1}]	[]	[]	[]
1921–1930	861	480	386	–6	0.25	0.74	0.87	0.94
1931–1940	877	490	382	–4	0.25	0.77	0.89	0.94
1941–1950	883	488	391	–2	0.30	0.59	0.77	0.96
1951–1960	944	480	465	–2	0.26	0.75	0.87	0.94
1961–1970	1004	498	479	0	0.29	0.70	0.84	0.98
1971–1980	1019	477	503	19	0.31	0.62	0.89	0.83
1981–1990	1099	492	609	28	0.39	0.54	0.89	0.78
1991–2000	1047	492	533	16	0.31	0.75	0.91	0.85
2001–2007	1092	539	544	17	0.28	0.73	0.91	0.86

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Table 5. Results from calibration on each individual 10-yr period.

Period	Parameters					Objective function		Performance parameters		
	U_{max}	L_{max}	CQOF	$CK_{1.2}$	CKBF	WB	RMSE	NSE	r	EI
Units	[mm]	[mm]	[]	[h]	[h]	[%]	[mm day^{-1}]	[]	[]	[]
1921–1930	16.2	65.2	0.15	49.8	7578	0	0.23	0.79	0.89	0.95
1931–1940	18.1	90.1	0.17	43.6	9057	0	0.22	0.81	0.90	0.96
1941–1950	14.9	106	0.12	39.6	12 370	0	0.28	0.62	0.79	0.96
1951–1960	19.6	160	0.15	44.1	7755	0	0.24	0.80	0.89	0.96
1961–1970	22.7	121	0.14	41.4	28920	0	0.29	0.70	0.84	0.98
1971–1980	16.8	400*	0.11	45.3	40 000*	6	0.26	0.73	0.88	0.89
1981–1990	18.5	400*	0.14	42.9	40 000*	17	0.34	0.64	0.86	0.83
1991–2000	31.8	400*	0.15	45.7	11 580	8	0.25	0.84	0.93	0.92
2001–2007	16.7	400*	0.15	56.6	40 000*	8	0.26	0.76	0.90	0.89

* The parameter has reached maximum value in the calibration.

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Table 6. Water balance results from using parameter values from the individual periods to drive the model.

Water balance for / Cali. from	Periods									Full period	Increase	
	1921-1930	1931-1940	1941-1950	1951-1960	1961-1970	1971-1980	1981-1990	1991-2000	2001-2010		Slope [mm/year]	% **
1921-1930	0%	5%	6%	12%	13%	36%	41%	31%	30%	19%	2.4	102%
1931-1940	4%	0%	2%	7%	9%	31%	37%	27%	26%	14%	2.4	106%
1941-1950	4%	1%	1%*	6%	7%	29%	37%	25%	26%	13%	2.4	107%
1951-1960	10%	6%	6%	0%	3%	23%	31%	19%	20%	8%	2.4	114%
1961-1970	8%	5%	3%	1%	0%	23%	32%	23%	20%	8%	2.2	103%
1971-1980	14%	12%	11%	11%	8%	13%*	21%	14%	13%	0%	2.2	108%
1981-1990	14%	12%	11%	11%	8%	13%	21%*	14%	12%	0%	2.1	106%
1991-2000	16%	13%	13%	8%	4%	14%	24%	9%*	12%	0%	2.2	115%
2001-2007	14%	12%	11%	11%	8%	13%	21%	14%	13%*	0%	2.1	107%

* The discrepancy from Table 5 to 6 is due to the fact that the model is building up a larger excess/loss of water when the complete time series (133 years) is run.
** The percent increase is calculated as the difference in the regression line from 1875 to 2007.

Blue - Areas with excess of water compared to observed (>10% = Dark; <10% = Light)

Red - Areas with lack of water compared to observed (>10% = Dark; <10% = Light)

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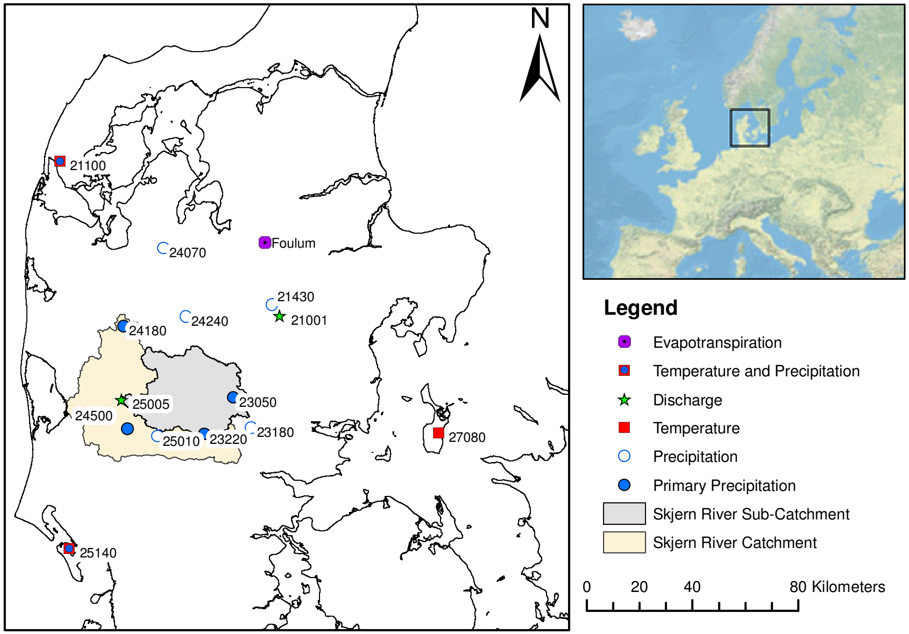

Fig. 1. Map of the study area with precipitation-, discharge- and temperature stations.

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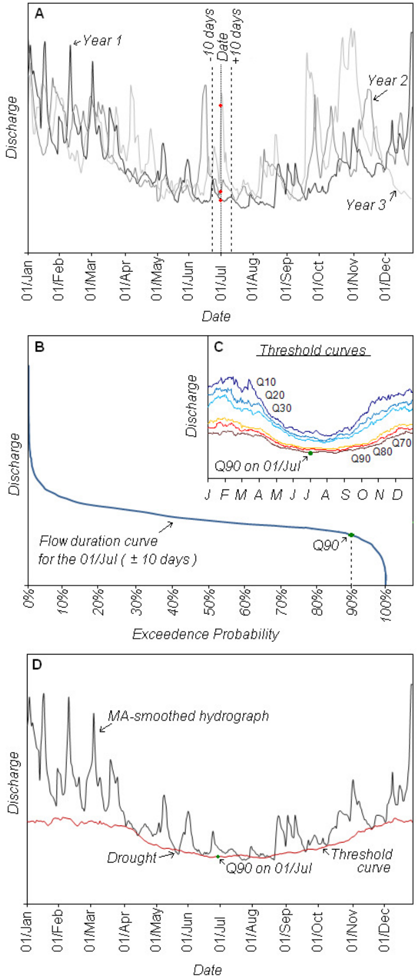

Fig. 2. Calculation steps from annual discharge series to a daily threshold curve.

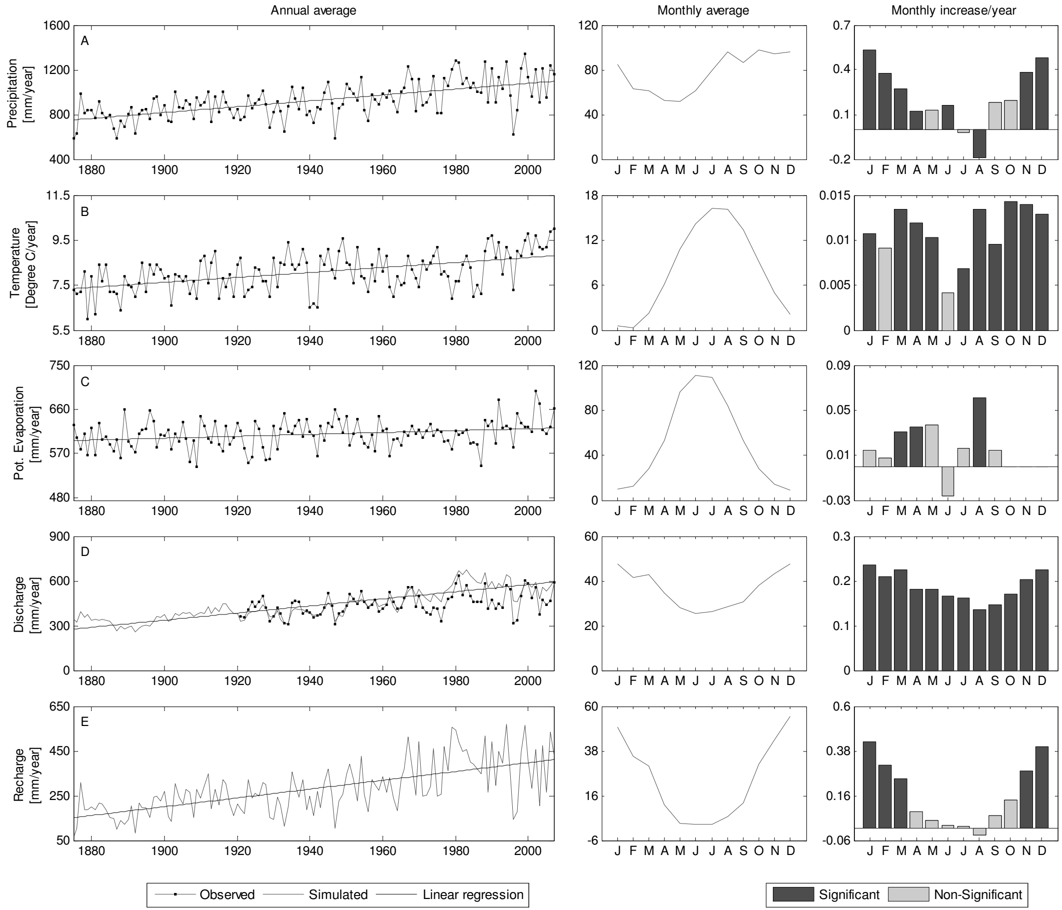

Fig. 3. Left panels: annual average of the components in mm. Centre panels: monthly average. Right panels: distribution of the in-/decrease for each month in mm yr^{-1}.

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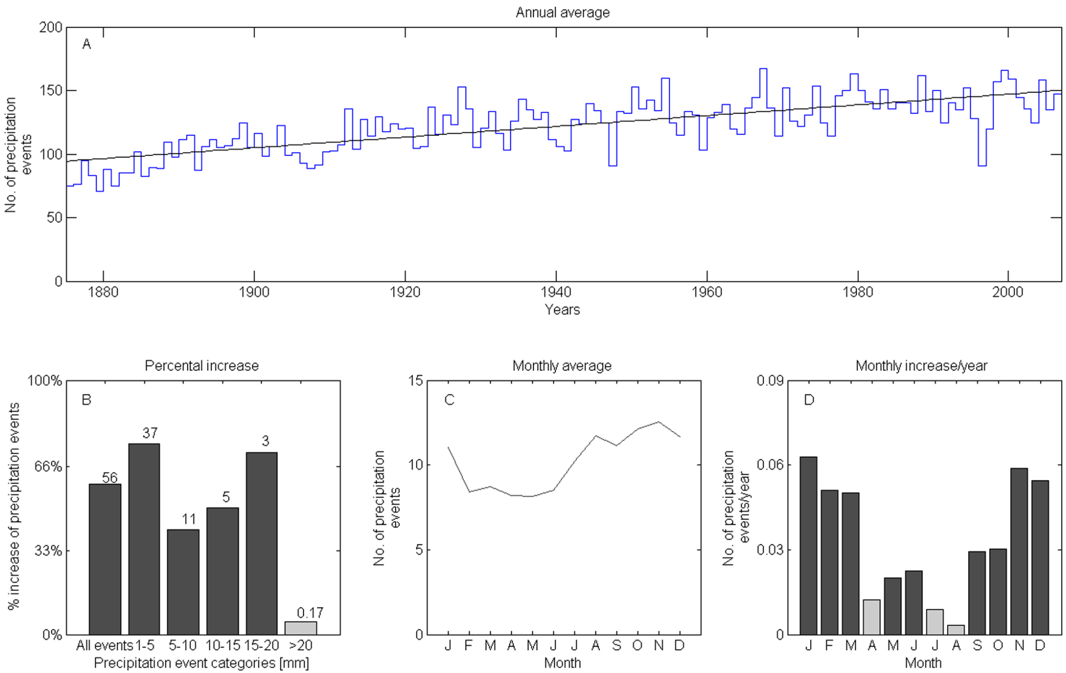

Fig. 4. (A) Count of all precipitation events > 1 mm within a year. **(B)** The percent increase as calculated as the difference in the regression line from 1875 to 2007 for all events total and for events divided into five different volume categories. The numbers above to columns represent the absolute number of events each class has increased with during the 133 yr. **(C)** Monthly average of events. **(D)** Distribution of the in-/decrease in events over the season.

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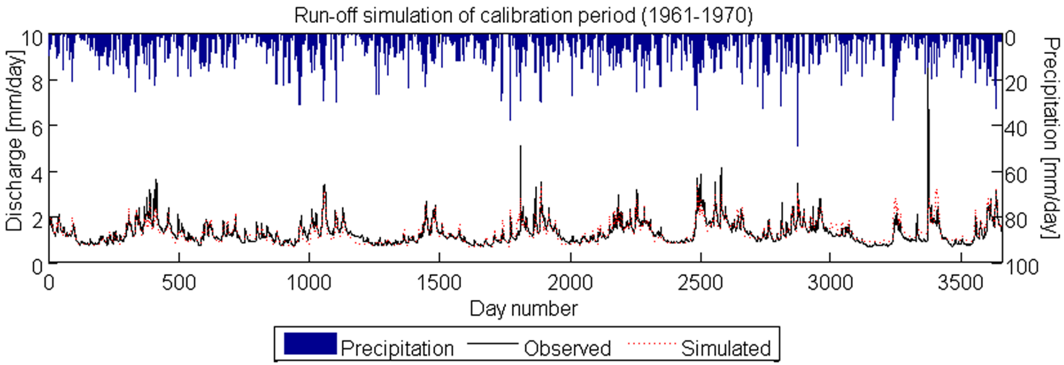

Fig. 5. Hydrograph for the calibration period with rainfall input.

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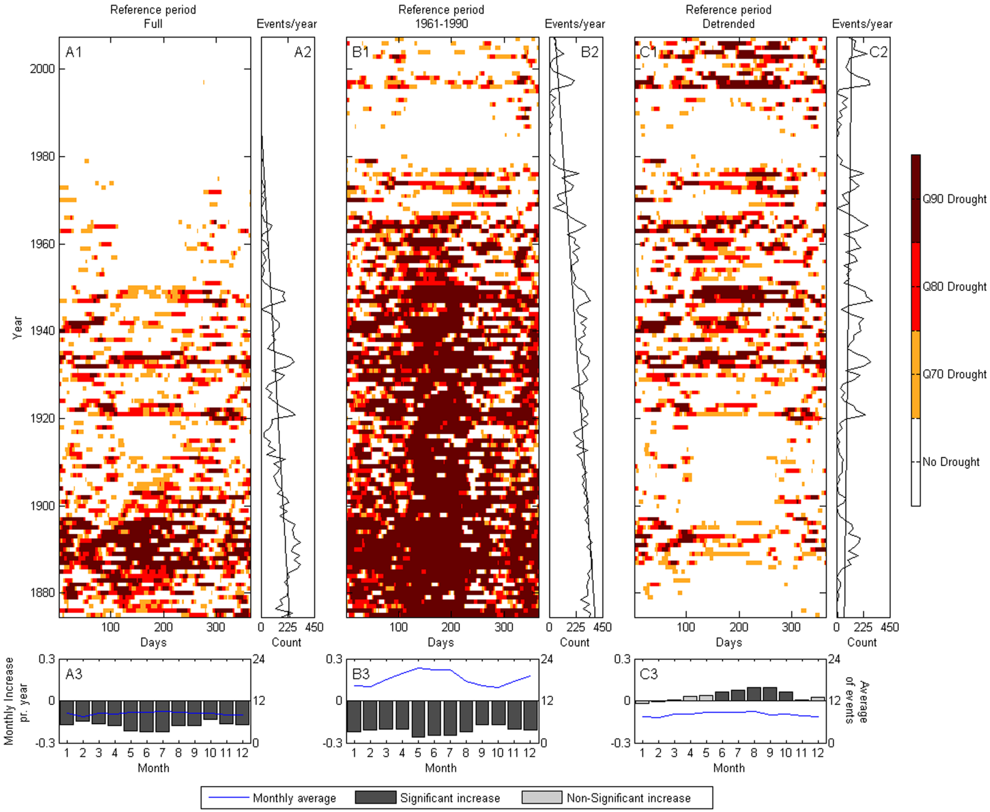

Fig. 6. Stream flow drought indices. **(A)** Reference period using the full data set as calculation basis. **(B)** Reference period only based on the 30 yr period from 1961–1990. **(C)** Reference period based on the whole data set after de-trending the discharge values. Plot 1: stream flow Deficiency Period diagram. Plot 2: count of dry days (Q_{90}–Q_{70}) per year. Plot 3: monthly average of all occurring dry days and distribution of the in/decrease in dry days over the season.

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Evaluating the influence of long term historical climate change

I. B. Karlsson et al.

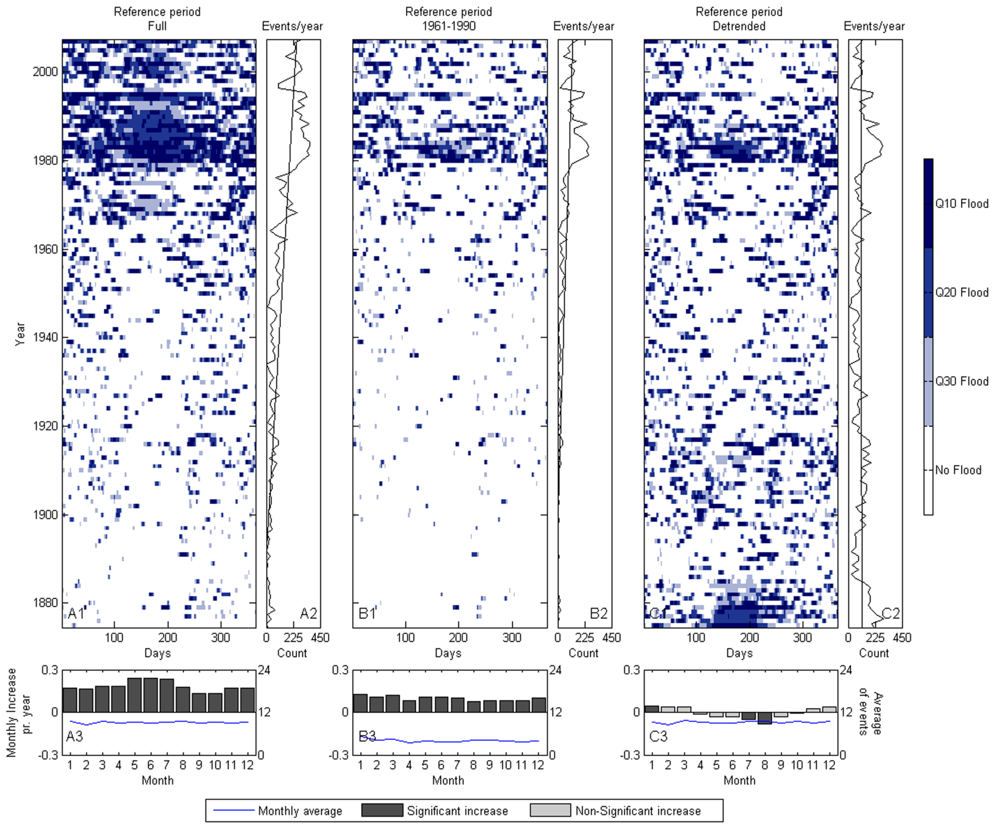

Fig. 7. Stream flow flood indices. **(A)** Reference period using the full data set as calculation basis. **(B)** Reference period only based on the 30 yr period from 1961–1990. **(C)** Reference period based on the whole data set after de-trending the discharge values. Plot 1: stream flow Excess Period diagram. Plot 2: count of wet days (Q_{10}–Q_{30}) per year. Plot 3: monthly average of all occurring wet days and distribution of the in/decrease in wet days over the season.

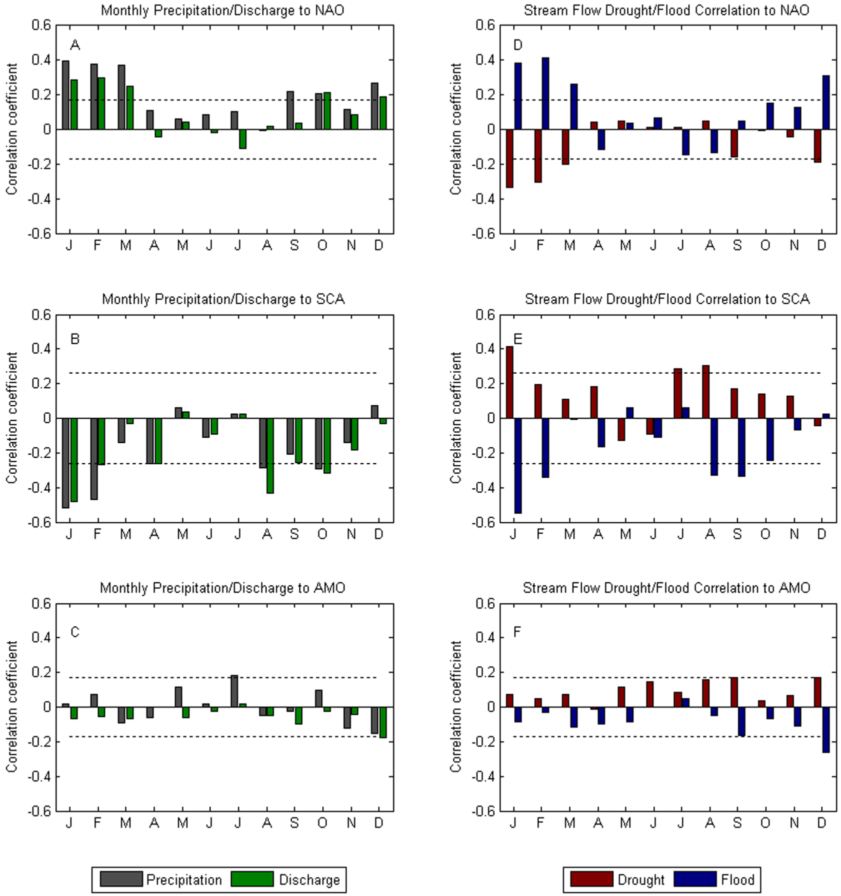

Fig. 8. Pearson correlation coefficient between precipitation/ discharge in the area **(A, B, C)** and the climate indices NAO, SCA and AMO; and between the stream flow indices and NAO, SCA and AMO **(D, E, F)**. Note that SCA information only covers 1950–2007, while the NAO and the AMO index cover 1875–2007. Dotted lines represent 5 % significance levels.