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Applying a simple water-energy balance framework to predict the climate sensitivity of streamflow over the continental United States

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Abstract

The prediction of climate effects on terrestrial ecosystems and water resources is one of the major research questions in hydrology. Conceptual water-energy balance models can be used to gain a first order estimate of how long-term average streamflow is changing with a change in water and energy supply. A common framework for investigation of this question is based on the Budyko hypothesis, which links hydrological response to aridity. Recently, Renner et al. (2011) introduced the CCUW hypothesis, which is based on the assumption that the total efficiency of the catchment ecosystem to use the available water and energy for actual evapotranspiration remains constant even under climate changes.

Here, we confront the climate sensitivity approaches (including several versions of Budyko's approach and the CCUW) with data of more than 400 basins distributed over the continental United States. We first map an estimate of the sensitivity of streamflow to changes in precipitation using long-term average data of the period 1949–2003. This provides a hydro-climatic status of the respective basins as well as their expected proportional effect on changes in climate. Next, by splitting the data in two periods, we (i) analyse the long-term average changes in hydro-climatolgy, we (ii) use the different climate sensitivity methods to predict the change in streamflow given the observed changes in water and energy supply and (iii) we apply a quantitative approach to separate the impacts of changes in the long-term average climate from basin characteristics change on streamflow. This allows us to evaluate the observed changes in streamflow as well as to evaluate the impact of basin changes on the validity of climate sensitivity approaches.

The apparent increase of streamflow in the majority of basins in the US is dominated by a climate trend towards increased humidity. It is further evident that impacts of changes in basin characteristics appear in parallel with climate changes. There are coherent spatial patterns with basins of increasing catchment efficiency being dominant in the western and central parts of the US. A hot spot of decreasing efficiency is found

within the US Midwest. The impact of basin changes on the prediction is large and can be twice as the observed change signal. However, we find that both, the CCUW hypothesis and the approaches using the Budyko hypothesis, show minimal deviations between observed and predicted changes in streamflow for basins where a dominance of climatic changes and low influences of basin changes have been found. Thus, climate sensitivity methods can be regarded as valid tools if we expect climate changes only and neglect any direct anthropogenic influences.

1 Introduction

1.1 Motivation

The ongoing debate of environmental change has stimulated many research activities, with the central questions of how hydrological response may change under (i) climate change and (ii) under changes of the earth surface. These questions are also practically of high concern, because present management plans are needed to cope with the anticipated changes in the future. Therefore, robust and reliable estimates of how water supplies are changing under a given future scenario are needed.

The link between climate change and hydrological response, to which we will refer to as climatic sensitivity, is one of the central research questions in past and present hydrology. There are different directions to settle this problem. One direction of research tries to model all known processes operating at various temporal and spatial scales in complex earth-climate simulation models, hoping to represent all processes with the correct physical description, initial conditions and parameters. These exercises are compelling, however, it is hard to quantify all uncertainties of such complex systems (Blöschl and Montanari, 2010).

Another direction is to deduce a conceptual description valid for the scale of the relevant processes of interest (Klemes, 1983). For example the Budyko hypothesis has successfully been used as a conceptual model to derive analytical solutions to

estimate climate sensitivity of streamflow and evapotranspiration (Dooge, 1992; Arora, 2002; Roderick and Farquhar, 2011; Yang and Yang, 2011). A different conceptual approach has been taken by Renner et al. (2011), who use the concept of coupled long-term water and energy balances to derive analytic solutions for climate sensitivity. This concept is a theoretical extension of the ecohydrological framework of Tomer and Schilling (2009) who provide a simple framework to separate climatic impacts on the hydrological response from other impacts such as land cover change.

Before applying any method for the unknown future, it needs to be evaluated by using historical data. Preferably for the case of streamflow sensitivity, the data are at the spatial scale of water resources management operations, the data should be homogeneous, consistent, and cover a variety of climatic and hydrographic conditions.

1.2 Hydro-climate of the continental US

We found that the situation in the continental US fulfils many of these points and the agenda to publish data with free and open access clearly supported our research. Here, we employ data of the Model Parameter Estimation Experiment (MOPEX) of the US (Schaake et al., 2006) covering the second part of the 20th century in the US.

This period is particularly interesting, because significant hydro-climatic changes have been reported (Lettenmaier et al., 1994; Groisman et al., 2004; Walter et al., 2004). Most prominent is the increase of precipitation for a large part of the US in the 1970's (Groisman et al., 2004). Also streamflow records show predominantly positive trends (Lins and Slack, 1999), however, there are still open research questions regarding the resulting magnitudes and the causes of different responses to the increase in precipitation (Small et al., 2006).

Specifically, there is the need to quantify climatic impacts such as changes in precipitation or evaporative demand on streamflow. As Sankarasubramanian et al. (2001) note, there are large discrepancies in climatic sensitivity estimates, not only due to the model used, but also its parametrisation can obscure estimated links between climate and hydrology.

Furthermore, there is evidence of human induced changes in the hydrographic features of many basins, especially land use changes, dam construction and operation, irrigation, but also changes in forest and agricultural management practises are believed to have considerable impacts on the hydrological response of river basins (Tomer and Schilling, 2009; Wang and Cai, 2010; Kochendorfer and Hubbart, 2010; Wang and Hejazi, 2011). Yet, there is the difficulty to separate effects of changes in basin characteristics and those of climate variations, which operate on different temporal scales (Arnell, 2002).

1.3 Aims and research questions

This paper presents an evaluation of two conceptual hypotheses, the newly developed water-energy balance framework of Renner et al. (2011) and the Budyko framework, to estimate climate sensitivity of streamflow. We evaluate both frameworks by applying them to a large dataset describing the observed hydro-climatic changes within the continental US in the second part of the 20th century. We further aim to quantify the impact of climatic changes on streamflow under the concurrence of climatic variations and changes in basin characteristics in the US.

Specifically we address the following research questions:

1. Can we predict and attribute the streamflow changes to the respective changes in precipitation and evaporative demand?

2. How strong is the effect of estimated basin characteristic changes on (i) the change in streamflow and (ii) the sensitivity methods, which only regard climatic changes?

This paper is structured as follows. We first review the ecohydrological framework aiming to separate climate from other effects on streamflow and present the methods used to predict the sensitivity of streamflow to climate. The results are discussed in the light of the rich literature already existing for the hydro-climatic changes observed over

the continental US, but demonstrating the new insights gained from the application of the simple water-energy framework by Renner et al. (2011).

2 Methods

2.1 Ecohydrological concept to separate impacts of climate and basin changes

The approaches considered here, aim at the long-term water and energy balance equations at the catchment scale. Thus, we assume that interannual storage changes can be neglected.

The framework established by Tomer and Schilling (2009) represents the hydro-climatic state space of a given watershed by using two non-dimesional variables, relative excess water W and relative excess energy U. Both variables can be derived by normalising the water balance equation with precipitation (P) and the energy balance equation with the water equivalent of net radiation (R_n/L) (Renner et al., 2011):

$$W = 1 - \frac{E_T}{P} = \frac{Q}{P}, \quad U = 1 - \frac{E_T}{R_n/L} = 1 - \frac{E_T}{E_p}. \tag{1}$$

Relative excess water W considers the amount of water which is not used by actual evapotranspiration E_T and thus equals the runoff ratio (areal streamflow Q over P of a river catchment). Relative excess energy U describes the relative amount of energy not used by E_T. Note, that we use potential evapotranspiration E_p to describe energy supply by net radiation R_n/L. This has practical relevance, because E_p can be estimated from widely available meteorological data.

Tomer and Schilling (2009) analysed temporal changes in U and W at the catchment scale. With that they introduced a conceptual model, based on the hypothesis that the direction of a temporal change in the relationship of U and W can be used to distinguish effects of a change in land-use or climate on the water budget in a given basin. Three major hypotheses relevant for streamflow sensitivity to (a) climate and (b) changes in basin characteristics can be deduced:

1. climate change impact hypothesis (abbreviated as CCUW)

$$\Delta U/\Delta W = -1 \tag{2}$$

2. basin characteristics change impact hypothesis (BCUW):

$$\Delta U/\Delta W = 1 \tag{3}$$

3. a combination of both effects, where the change direction ϖ can be computed from the observed change signals of U and W:

$$\varpi = \arctan\frac{\Delta U}{\Delta W} \tag{4}$$

Thus the simple analysis of both, ΔW and ΔU can give a first-order guess to separate climate from basin characteristics changes (such as land cover change, land management changes, etc.).

2.2 Streamflow change prediction based on a coupled water-energy balance framework

The climate change impact hypothesis (CCUW) can also be applied to predict climate sensitivity of streamflow which is shown in detail in Renner et al. (2011). A central implication of the CCUW hypothesis is that the sum of the efficiency to evaporate the available water supply (E_T/P) and the efficiency to use the available energy for evapotranspiration (E_T/E_p):

$$\mathrm{CE} = \frac{E_T}{P} + \frac{E_T}{E_p} \tag{5}$$

is constant for a given basin. Any changes in CE, which we denote as catchment efficiency, are assigned to a change in basin characteristics. The fundamental assumption of constant catchment efficiency links water and energy balances. By using the total derivative of the definitions of W and U in Eq. (1) and combining with the CCUW hypothesis Eq. (2), the sensitivity coefficient of streamflow to precipitation can be derived (Renner et al., 2011):

$$\varepsilon_{Q,P} = \frac{P}{Q} - \frac{(P-Q)E_p}{Q(E_p+P)}. \tag{6}$$

The sensitivity coefficient $\varepsilon_{Q,P}$ describes how a proportional change in P translates into a proportional change of streamflow. The sensitivity is largely dependent on the inverse of the runoff ratio and the aridity of the climate. An analogue coefficient for the sensitivity to E_p is easily derived by the connection of both coefficients: $\varepsilon_{Q,P} + \varepsilon_{Q,E_p} = 1$ (Kuhnel et al., 1991).

The climate change impact hypothesis may also be used to predict absolute changes. Therefore, consider two long-term average hydro-climate state spaces ($P_0, E_{p,0}, Q_0$), ($P_1, E_{p,1}, Q_1$) of a given basin. Again, by using the definitions of W and U and applying the CCUW hypothesis, an equation can be derived to predict the new state of streamflow Q_1 (Renner et al., 2011):

$$Q_1 = \frac{\frac{Q_0}{P_0} - \frac{P_0-Q_0}{E_{p,0}} + \frac{P_1}{E_{p,1}}}{\frac{1}{P_1} + \frac{1}{E_{p,1}}} \tag{7}$$

2.3 Streamflow change prediction based on the Budyko hypothesis

The Budyko hypothesis states that actual evapotranspiration is primarily determined by the ratio of energy supply (E_p) over water supply (P), to which we refer to as aridity index (E_p/P). There are various functional forms which describe this relation. In this paper we use the non-parametric curve of Ol'Dekop (1911) and a parametric form of Mezentsev (1955), which are reported in Table 1. The parametric form introduces a catchment parameter (n) which is used to adjust for inherent catchment properties. The knowledge of the functional form $E_T = f(P, E_p, n)$ allows to compute the sensitivity coefficient of streamflow to precipitation (Roderick and Farquhar, 2011; Renner et al., 2011):

$$\varepsilon_{Q,P} = \frac{P}{Q}\left(1 - \frac{\partial E_T}{\partial P}\right). \tag{8}$$

Thereby, the first derivative of the respective Budyko function is used to derive the partial differential term $\frac{\partial E_T}{\partial P}$ which describes how E_T is changing with P. Further, Q is substituted via $P - E_T$ by the respective Budyko function. With that, the resulting sensitivity coefficients are functions of P, E_p and n, if the parametric form is used. The partial differentials and the respective Budyko functions can be found in Table 1.

Absolute changes in streamflow (dQ) by changes in precipitation or potential evapotranspiration can be predicted by (Roderick and Farquhar, 2011):

$$dQ = \left(1 - \frac{\partial E_T}{\partial P}\right) dP - \frac{\partial E_T}{\partial E_p} dE_p - \frac{\partial E_T}{\partial n} dn. \tag{9}$$

Note, that using Budyko approaches for predicting the effects of a change in aridity will also result in a change in catchment efficiency. This change is determined by the functional form and the catchment parameter as well as the aridity index of the basin (Renner et al., 2011).

Recently, Wang and Hejazi (2011) presented a method to separate and quantify impacts of climate change and basin characteristic changes on streamflow. In particular, they employ a parametric Budyko function, which is calibrated using data of a reference period and use it to estimate the climatic effect on streamflow. Then they use the difference to the observed change ΔQ_{obs} to quantify the effect of basin characteristic changes. For any details, please refer to Wang and Hejazi (2011). The method actually requires the same data as the approach of Tomer and Schilling (2009)

and thus allows a direct comparison. To compute the effect of basin characteristic changes using the CCUW hypothesis, one needs to assume that impacts of climate ΔQ_{clim} and basin characteristic changes ΔQ_{basin} on streamflow are separable and thus $\Delta Q_{obs} = \Delta Q_{clim} + \Delta Q_{basin}$.

3 Data

The approaches presented above are not very data demanding. Still longer time series of annual basin precipitation totals (P; mm yr^{-1}), river discharge data converted to areal means (Q; mm yr^{-1}) and potential evapotranspiration data (E_p; mm yr^{-1}) are needed. Further, the approach should be tested against a variety of hydro-climatic conditions and different manifestations of climatic variations. Therefore, we have chosen the dataset of the model parameter estimation experiment (MOPEX) (Schaake et al., 2006), covering the United States. It contains a large set of basins distributed over different humid to arid climate types within the continental US. The good coverage allows to describe the hydro-climatic state at a regional and continental scale of the US. A range of hydro-climatic and ecohydrological studies already used this dataset (e.g. Oudin et al., 2008; Troch et al., 2009; Wang and Hejazi, 2011; Voepel et al., 2011). The dataset covers 431 basins and can be freely downloaded from ftp://hydrology.nws.noaa.gov/pub/gcip/mopex/US_Data/. The catchment area of the basins ranges from 67 to 10 329 km^2 with a median size of 2152 km^2.

The dataset contains daily data of P, Q, daily minimum T_{min} and maximum temperature T_{max} as well as a climatologic potential evapotranspiration estimate ($E_{p,clim}$), which is based on pan evaporation data of the period 1956–1970 (Farnsworth and Thompson, 1982). Because a time series of E_p is needed, we use the Hargreaves equation (Hargreaves et al., 1985) to estimate daily E_p. The Hargreaves equation for potential evapotranspiration has minimal data requirements (T_{min} and T_{max}), but yields a good agreement with physically based E_p models (e.g. Hargreaves and Allen, 2003; Aguilar and Polo, 2011). Daily potential evapotranspiration is estimated by (Hargreaves and Allen, 2003):

$$E_{\mathrm{p,Hargreaves}} = a \cdot \mathrm{sd}_{\mathrm{pot}}((T_{\max} - T_{\min})/2 + b) \cdot \sqrt{T_{\max} - T_{\min}}, \quad (10)$$

where $\mathrm{sd}_{\mathrm{pot}}$ is the maximal possible sunshine duration of a given day at given latitude and two empirical parameters ($a = 0.0023$, $b = 17.8$). Keeping the Hargreaves parameters fixed, we find that the annual E_{p} totals estimated by the Hargreaves equation are generally lower than the climatological estimates included in the MOPEX dataset. An improvement could be made by calibrating the Hargreaves parameters, however, this would introduce further ambiguities, as these parameters tend to change not only with location but also with time and wetting (Aguilar and Polo, 2011).

Finally, all daily data, i.e. P, E_{p}, Q are aggregated to annual sums for water years defined from 1 October–30 September. The final dataset covers the period 1949–2003 with 430 basin series.

4 Results and discussion

4.1 Hydro-climate conditions in the US

The basins in the US MOPEX dataset cover a variety of hydro-climatic conditions, which can be seen in the mapping of long-term average variables (P, Q, $E_{\mathrm{p,clim}}$, $E_{\mathrm{p,Hargreaves}}$) in Fig. 1. The basins with most precipitation are found in the Northwest, the Southeast and along the east coast. The central part of the US receives considerable less precipitation, which is a continental climate effect intensified by the mountain ranges in the west and east, blocking west to east atmospheric moisture transport. Potential evapotranspiration obeys a north to south increasing gradient, which is modulated by the continental climate in the Central US. The bottom maps display the difference in E_{p} estimates, whereby the Hargreaves values, which are used further on, are less variable than the climatological E_{p} estimates.

Streamflow is naturally governed by precipitation input and follows the spatial patterns of precipitation. However, the arid conditions in the Central US result in lower streamflow amounts. This functional dependency can be seen in the Budyko plot in the left panel of Fig. 2, plotting the evaporation ratio E_{T}/P as function of the aridity index E_{p}/P. In general, the basins follow the Budyko Hypothesis, whereby Ol'Dekop's function explains 50 % of the variance. The aridity index of the basins ranges between 0.27 and 4.51, with most basins clustering around 1. The right panel of Fig. 2 displays the relationship of the nondimensional measures W and U, referred to as UW space. Note, that $W = 1 - \frac{E_{\mathrm{T}}}{P}$, whereby E_{T}/P is used in the Budyko plot on the ordinate. A thorough discussion of the relationship between both spaces can be found in Renner et al. (2011). The hydro-climatic data covers the UW space, meaning that there is a large variety of hydro-climate conditions in the dataset. W is ranging between 0 and 1, while U also shows a few negative values. This is due to two main reasons, (i) an underestimation of the energy supply, i.e. in E_{p} and (ii) a physical reason, where advection of heat into the basins leads to an additional input of energy for actual evapotranspiration. As we use the Hargreaves E_{p} estimates, the first reason is most likely.

4.2 Climate sensitivity of streamflow

Figure 3 provides a map of the climate sensitivity coefficients using the CCUW approach, i.e. applying Eq. (6) on the long-term average values of P, E_{p}, Q. For example, a sensitivity coefficient of 2 implies that a relative change in precipitation results in a two times larger relative change in streamflow. Within the US most coefficients range between 1.5 and 2.5. But there are also very high estimates, where a small change in annual precipitation would imply very pronounced relative changes in streamflow. This is due to the small amount of streamflow compared to precipitation and evapotranspiration in these predominantly arid catchments. Although there is some correlation of the sensitivity coefficient to the aridity index (Pearson correlation coefficient $\rho = 0.52$), we note that the inverse of the runoff ratio (P/Q) is the main controlling factor in determining runoff sensitivity to climate ($\rho = 0.99$).

To further illustrate this functional relationship, we plot $\varepsilon_{Q,P}$ in Fig. 4 as a function of the evaporation ratio, which is directly related to the inverse of the runoff ratio, but bounded between 0 and 1. From the left panel (black dots) we see that the estimate of the CCUW method ($\varepsilon_{Q,P;\mathrm{CCUW}}$) is primarily and nonlinearly determined by E_T/P. To estimate the uncertainty in estimation of $\varepsilon_{Q,P;\mathrm{CCUW}}$, we computed $\varepsilon_{Q,P;\mathrm{CCUW}}$ for each year of the annual time series and display the interquartile range (25–75 % percentile range) of all those annual sensitivity coefficients as vertical grey line. The uncertainty range increases with E_T/P. For values of $E_\mathrm{T}/P > 0.6$, the ranges get more apparent with about 25 % of $\varepsilon_{Q,P}$, which can be up to the order of $\varepsilon_{Q,P}$ for $E_\mathrm{T}/P > 0.8$. This implies, the smaller the runoff ratio of a given basin the larger is the sensitivity to climate variations and the uncertainty in its estimation. Moreover, the variability in climatic forcing of individual years or periods can have large impacts on the resulting streamflow.

The right panel of Fig. 4 provides a comparison of the sensitivity estimates of CCUW with the non-parametric Budyko approaches and the parametric Budyko function approach of Roderick and Farquhar (2011). We find that the non-parametric Budyko sensitivity approaches are determined by aridity only and show large differences to CCUW, already at medium values of E_T/P. The parametric Budyko function approach (Mezentsev) yields similar sensitivities as the CCUW approach for $E_\mathrm{T}/P < 0.9$. This is due to the parameter n, which inherently includes some dependency to E_T/P. However, it can be shown that there is an upper limit for the sensitivity coefficient which is set by $n+1$. Here, we estimated the largest value of n for the given dataset with $n = 9.1$ and the largest sensitivity with $\varepsilon_{Q,P,\mathrm{mez}} = 6.4$. In contrast, the sensitivity of streamflow to precipitation, estimated by the CCUW approach is not bounded and proportional to the inverse of the runoff ratio. This relationship is already expected by the more general definition of streamflow sensitivity Eq. (8). This behaviour is also discussed in Sankarasubramanian and Vogel (2003), Yang and Yang (2011) and observed by Chiew (2006) for Australian basins, using a hydrological model.

4.3 Assessment of observed and predicted changes in streamflow

Next, we evaluate the introduced analytical streamflow change prediction methods under past hydro-climatic changes in the contiguous US using data covering the water years from 1949 to 2003. As the approaches assume steady-state conditions, we evaluate the changes by subdividing the data into two periods, 1949–1970 and 1971–2003. This choice is in accordance with the recent study of Wang and Hejazi (2011). They justify their selection with a probable step increase in precipitation and in streamflow in large parts of the US around the year 1970 (McCabe and Wolock, 2002). First, we give an overview of the observed hydro-climatic changes and then evaluate the predictions of streamflow changes. Last, we employ the conceptual model of Tomer and Schilling (2009) to attribute impacts of climate and basin characteristic changes.

4.3.1 Hydro-climatic changes in the US

We describe the climatic changes by comparing long-term average data of the two periods 1949–1970 and 1971–2003. Analysing the difference of the average annual rainfall, we find an increase in P for most basins, whereby the increase is significant for 31 % of the basins ($\alpha = 0.05$, Welch two sample t-test with unknown variance, using (R Development Core Team, 2010, stats::t.test)). The topleft map in Fig. 5 displays the spatial distribution of changes in P, which are largest over the Mississippi River basin (>90 mm, excluding the Missouri River basin). Significant changes in precipitation are scattered over parts of the Mississippi basin and in the Northeast, however, there are hardly any significant changes in the peninsula of Florida and the West. The drastic increase in precipitation has already been discussed in many publications, (e.g. Lettenmaier et al., 1994; Milly and Dunne, 2001; Krakauer and Fung, 2008).

There are also significant changes in E_p, estimated by the Hargreaves equation. Here, we find a significant decline for 69 % of the basins. The topright map in Fig. 5 shows that the decrease is largest in the west of the Appalachian Mountain ranges

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(about −30 mm). These changes are directly related to a decrease in the diurnal temperature range, which is also reported by Lettenmaier et al. (1994).

Both, the increase in precipitation and the decrease in potential evapotranspiration should ideally lead to an increase in annual streamflow. We find that 31 % of the basins show a significant increase. The map in the bottomleft panel of Fig. 5 shows, that basins with significant increases in streamflow are predominantly found within the Upper Mississippi River basin and the Northern Appalachian Mountains and a few basins at the southern coast. These basins show an increase of about 41 % compared to the average of the first period. For most of the other regions, we find non-significant streamflow increases, while in the West there are mainly non-significant declines in annual streamflow. Please note, that we do not display basins when more than 10 yr of data are missing (78 basins) and that we removed 2 basins from further analysis, because the water balance was suspect ($Q > P$).

4.3.2 Evaluation of streamflow change predictions

In the previous section we described the changes observed in precipitation, potential evapotranspiration and streamflow comparing the long-term averages of two periods. Now we aim to predict the change in streamflow, using the climate sensitivity approaches of the CCUW method (i.e. application of Eq. 7) and the Budyko approaches. For the Budyko approaches, we use Eq. (9) and the functional forms of Ol'Dekop (1911) and Mezentsev (1955). Streamflow change predictions using non-parametric Budyko equations have e.g. been applied by Arora (2002), while Roderick and Farquhar (2011) used Mezentsev's equation.In particular we use the hydro-climatic state of the first period, described by P_0, $E_{\mathrm{p},0}$, Q_0, as well as the climatic states of the second period P_1,$E_{\mathrm{p},1}$ to predict the streamflow of the second period Q_1. Then we evaluate the accuracy of streamflow prediction by using the observed ΔQ_{obs} and predicted change ΔQ_{pred} signals. To evaluate the accuracy of the predicted streamflow changes, we use the relative mean absolute error (RMAE):

$$\mathrm{RMAE} = \frac{\sum |\Delta Q_{\mathrm{obs}} - \Delta Q_{\mathrm{pred}}|}{\sum |\Delta Q_{\mathrm{obs}}|}. \tag{11}$$

The RMAE is zero when predictions actually match the observations, while a value of 1 means a prediction error of 100 %.

The streamflow change prediction approaches are similar in performance, when considering the whole dataset (RMAE = 50 %, 51 %, 51 % for CCUW, Mezentsev, and Ol'Dekop, respectively). A scatterplot of predicted versus observed changes is shown in the left panel of Fig. 6, where dots close to the 1 : 1 line indicate good predictions. While most dots scatter around the 1 : 1 line, there is a considerable number of basins where prediction and observation are completely different. There is also no indication if one method is more realistic than the other.

However, when sorting the results by prediction accuracy, we find that CCUW is slightly better than the Budyko approaches. About 16 % (28 %, 55 %) of the basins have a RMAE prediction error smaller than 5 % (10 %, 20 %), while using the Budyko hypothesis with the parametric function of Mezentsev yields 14 % (26 %, 52 %) and Oldekop's function 12 % (26 %, 52 %). That means by only considering climatic forcing changes, we are able to predict streamflow changes with an error smaller than 20 % for about 55 % of the basins.

The bottomright map in Fig. 5 presents the predicted streamflow changes by the CCUW method. There is an eye catching similarity of the spatial pattern of precipitation changes shown in the topleft map. Comparing the predicted changes with the observed changes in streamflow in the bottomleft map, we see that there is a good coherence from the East to the Central US. Towards the Northwest, the predicted changes are significantly larger than the observed changes. However, the topleft map in Fig. 5 shows that although there has been an increase in precipitation in the West, hardly any of these changes have been significant. This may have important implications for the prediction methods considered here, which only deal with average climate conditions and disregard interannual variability. In fact, a non-significant change in precipitation eventually does not justify to assume a change in average annual precipitation, which is used as input to estimate some change in streamflow.

Next we assume, that the difference between observed and predicted (i.e. due to climatic changes) change in streamflow can be attributed to changes in basin characteristics. With that we get an estimate of impacts of basin changes from the CCUW method and compare it with the results obtained by the method of Wang and Hejazi (2011). This comparison is shown as scatterplot in the right panel of Fig. 6. The graph indicates that there is a general agreement between both estimates ($\rho = 0.97$), although they are derived from different theoretical frameworks. In general, basins with a low evaporation ratio tend to show negative changes due to basin change, while basins with larger evaporation ratios show positive changes in streamflow. The largest differences between both methods are found for basins with very high evaporation ratios. In this case CCUW predicts larger changes than the Budyko approaches, which was already discussed above. These changes are small in absolute values, but quite large when seen relative to the annual totals of streamflow.

4.3.3 Mapping the influence of climate and land-use impacts on streamflow

The ecohydrological framework of Tomer and Schilling (2009) is based upon the hypothesis that climatic impacts on streamflow can be separated from basin characteristic changes. Thus, the change direction ϖ, introduced with Eq. (4), can theoretically be used to assess the relative impact of both influences.

Based on the hydro-climatic states of the two periods, ϖ has been determined and is mapped in Fig. 7. The significance in the change in magnitude in the non-dimensional UW space is tested with a two sample t-test with unknown variance on annual values of $\sqrt{W^2+U^2}$. We find that 77 out of 351 basins show a significant ($\alpha = 0.05$) change in these states, which are predominantly found in the Central and Western US. In the eastern part, significant changes are rather randomly distributed. Further, almost all

basins with significant changes appear to be influenced by basin characteristic changes (i.e $\varpi \neq 135^\circ, 315^\circ$).

From the inlay histogram showing the frequency of observed change directions we see that the majority of basins is right of the positive diagonal. This implies, the reader may also refer to Renner et al. (2011, Fig. 1), that there is a climate trend towards decreased aridity (E_p/P) in 94 % of the basins.

Regarding catchment efficiency Eq. (5), there are three main change scenarios: constant CE, i.e. relative excess water W increases by the same amount as relative excess energy U is decreased (case i), increasing CE when $\Delta U > \Delta W$ (case ii), or a decline in CE when $\Delta U < \Delta W$. If we consider a segment of 45° centered at $\varpi = 315^\circ$ this would reflect roughly constant CE and valid conditions for the CCUW hypothesis. About 31 % of the basins are actually within this boundary. According to the map in Fig. 7, these basins are found mainly in the central part below the Great Lakes, along a band following the Appalachian Mountains, and a few single basins in the West. Basins with distinct climate impacts and improving CE (case iii, a segment of 45° centered at $\varpi = 270^\circ$) are most frequent (35 %) and found throughout the US. Almost all basins within the Great Plains and the West show constant or decreasing runoff and increasing E_T. This is in accordance with the findings of Walter et al. (2004), who detected positive trends in E_T but not in Q for western river basins (Columbia, Colorado and Sacramento River basins). These trends may be linked to intraseasonal changes in hydrology, triggered by higher winter temperatures and thus less snow, which is melting earlier (Barnett et al., 2008). Moreover, groundwater pumping for irrigation in the High Plains (McGuire, 2009) possibly contributed to the observed signals (Kustu et al., 2010).

From the map in Fig. 7 we see a transition of changes in CE over the Mississippi River basin. While the western part shows increasing CE, the central part remained mainly constant and the northern part shows large clusters of basins with a decline in CE. This transition may be primarily linked to the precipitation changes, which also show a west to east gradient (cf. map in Fig. 5). But agricultural cultivation, especially in basins of the US Midwest, may have amplified these trends. Most likely the additional rain could not increase evapotranspiration as a lack of soil water storage due to intensive tile drainage (up to 30 % of the total state areas in the Midwest are drained Pavelis, 1987). So, the intensive agricultural land management did not only increased streamflow on average, but also lead to immense nitrogen leaching of Midwestern soils (Dinnes et al., 2002), showing biochemical signals far downstream (Raymond et al., 2008; Turner and Rabalais, 1994).

Towards the East, changes in ϖ are spatially more heterogeneous. This is probably because topography and landuse are more diverse compared to the West, however, it is important to note, that the density of river gauge records is much larger. Most basins east of $-87°$ latitude show increasing CE (case ii, 40 %) and constant CE (case ii, 36 %). Basins with decreasing CE occur rather local (case iii, 16 %).

4.4 Prediction accuracy and the influence of basin characteristic changes

The discussion of the influence of possible changes in catchment efficiency above showed that there are distinct spatial patterns of the type of change, i.e. whether improving CE or declining CE under the general trend of increasing humidity.With that we hypothesise that the change direction ϖ has a strong influence on the accuracy and ability to predict streamflow changes by only a climate signal.

To verify this hypothesis, we group the dataset according to the change direction ϖ in UW space, computed by Eq. (4). We selected a bin width of 5° for aggregation and plot the bin average of observed and predicted streamflow changes against ϖ in the left panel in Fig. 8. The graph shows that all methods predict similar changes in streamflow, while the observed change can be quite different. However, at a change direction $\varpi = 315°$ there is good conformance with the observed changes. Further, at values of $\varpi < 315°$ we find overprediction, while for values of $\varpi > 315°$ the observed changes are underestimated. This effect can be explained by the concept of CE: i.e. when $\varpi < 315°$ ($\varpi > 315°$) CE improved (declined) and more (less) water is evaporated meaning less (more) for streamflow.

In the right panel of Fig. 8 we plot the RMAE against ϖ. We find that the smallest errors for all models are found in climate change hypothesis direction at $\varpi = 315°$. In this bin, the Budyko type sensitivity models have a larger error (Oldekop: 11 %, Mezentsev: 5 %) than the CCUW approach (1 %). This good agreement is a matter of the design of the CCUW approach. Outside this bin, the prediction error is quickly increasing for all methods with only marginal differences between the approaches.

To increase the dataset of observed streamflow changes, 5 comparing situations have been established by predicting 3 times forward and 2 times backward in time for different periods. The results are shown in Fig. 9, which is structured similar to Fig. 8. The left panel clearly reveals that ϖ has a strong influence on both, observations and predictions. Further, we note that there is a phase shift of 45° between the predicted streamflow changes and the observed changes in streamflow. The dependency of ΔQ_{obs} to ϖ can be explained by the definition of W. So maximal positive (negative) changes in observed streamflow are found at $\varpi = 0°$ ($\varpi = 180°$), that is the case when only W is changing and U is constant. Analogously, when $\Delta W = 0$, then there are only small changes in ΔQ_{obs}. The predictions are mainly driven by precipitation changes, which are also plotted in the left panel of Fig. 9. Consequently, both are synchronised in phase over ϖ. Following the hypothesis of Tomer and Schilling (2009), one would expect that changes in the aridity index (E_{p}/P) are largest in climate change direction (CCUW hypothesis) and lowest in basin change direction (positive diagonal, $\varpi = 45°$, 225°), e.g. refer to Renner et al. (2011, Fig. 1). This theoretical assumption is nicely reproduced in Fig. 9 and provides empirical evidence for the validity of the framework of Tomer and Schilling (2009) to separate effects of climate from basin changes.

The relative absolute error in the right panel of Fig. 9 reflects the phase differences observed in the left panel. So in climate change directions, the error of the prediction methods is lowest. In basin characteristic change directions at 45° and 225°, the relative error in prediction is about 100 % of the observed change. The largest relative absolute errors are found, when the observed changes in streamflow are very small and the climatic changes are large ($\varpi \approx 90°$, 270°). In this case, impacts of change in basin characteristics compensate for the expected changes in streamflow due to climate changes.

4.5 Uncertainty discussion

4.5.1 Limitations due to observational data

Both climatic sensitivity approaches are based on long-term average data. These input data are spatially aggregated to river basin averages from point data and evaporative demand and E_T are only indirectly observed. For example, Milly (1994) showed by an uncertainty analysis of input data to their Budyko based water balance model that uncertainties in input data may explain the deviations from observed and modelled discharge and evapotranspiration.

Another issue is that net energy supply, i.e. net radiation balance data, is ideally required. However, direct observations of net radiation are not available for the purpose to estimate long-term catchment averages throughout the US. Therefore, a practical choice is to use potential evapotranspiration models, which provide an estimate based on available meteorological data. Here, we used the Hargreaves equation, which only requires data of minimum and maximum daily temperature. Therefore, our estimates of change in evaporative demand are entirely based on the trends in diurnal temperature ranges and mean temperature over the US.

There may be other causes of the change in energy supply which are not reflected in the trend in diurnal temperature ranges. For example, changes in net long wave radiation as reported by Qian et al. (2007) or changes in the surface albedo due to land cover changes. While the latter can be attributed to basin characteristic changes, the former requires better high resolution radiation and energy balance estimates (Milly, 1994). These estimates may be available by using remote sensing products or reanalysis products for past periods. This is, however, out of the scope of this study.

Still, we believe, that the main conclusions regarding the retrospective assessment of hydro-climatic changes and their regional patters will not be altered significantly by

using improved data for evaporative demand. First, the main signal is covered in the diurnal temperature range data and second, the observed changes in the partitioning of water and surface fluxes can be attributed to a much larger part to the change in precipitation.

4.5.2 Uncertainties due to inherent assumptions

While introducing the theoretical framework by Renner et al. (2011) and the Budyko framework, considerable assumptions have been placed which may be violated by measurement reality.

First, we have to regard the assumption that the storages of water and energy are zero, which may be violated but hard to discern. For example, Tomer and Schilling (2009) used very dry periods to separate the data for computing long-term averages. However, this relatively subjective method may also introduce other problems. Secondly, we assume steady state conditions. Several processes may violate this assumption, resulting in a trend of E_T over time (Donohue et al., 2007). Our results clearly show that any process related to a change in basin characteristics, may result in dynamic state transitions whose impacts on evapotranspiration and thus streamflow can be larger than impacts of climatic variations. So we found, that catchment efficiency has been widely increasing in the Western US. This represents a non-stationary transition in the water and energy balances towards increasing actual evapotranspiration on the cost of streamflow. Thereby, the effects of climate and basin characteristic changes on streamflow seem to be of equal magnitude and compensate for each other. In the companion paper we discussed the different assumptions on catchment efficiency and climate changes. While the Budyko functions inherently assume, that CE is changing with the aridity index, the CCUW method assumes CE to be constant. Here, we are unable to verify which assumption is correct, because of the multitude of possible other effects, especially the large impacts of basin characteristics change. However, the clear spatial distributions of the change direction ϖ is an indication that basin characteristic changes result in larger effects, than the definition of hydro-climatic feedbacks.

5 Conclusions

This paper demonstrates the applicability and usefulness of the coupled water-energy balance framework (CCUW) of Renner et al. (2011) for the problem of estimating the sensitivity of streamflow to changes in climate. To test and compare the CCUW framework with the Budyko framework we employed a large hydro-climatic dataset of the continental US, covering a variety of different climatic conditions (humid to arid) and basin characteristics, ranging from flat to mountainous basins with land cover types ranging from desert over agriculture to forested basins.

Based on long-term average hydro-climatological data (P, E_p, Q), we estimated and mapped the sensitivity of streamflow to changes in annual precipitation. The main distinction between the Budyko and the CCUW hypotheses is the functional dependency of the sensitivity coefficients. The sensitivity coefficients estimated by the Budyko framework depend on the aridity index and the type of the Budyko function only. In contrast the CCUW hypothesis, implies that climatic sensitivity of streamflow depends to a large degree on the inverse of the runoff ratio, which is already expected from the general definition of the sensitivity coefficient. This fundamental difference may result in large differences, which are most prominent for basins where runoff is very small compared to annual precipitation. However, for most of the other basins both approaches agree fairly well.

Further, we evaluated the capability of the climate sensitivity approaches to predict a change in streamflow, given observed variations in the climate of the second part of the 20th century. The combination with the conceptual framework of Tomer and Schilling (2009) to discern climate from basin characteristic changes impacts yield comprehensive insights in the hydro-climatic changes in the US. We can reinstate, that increased annual precipitation lead to increases of streamflow and evapotranspiration in general. However, our results provide evidence that changes in basin characteristics influenced how the additional amount of water is partitioned at the surface. Particularly the mapping of ϖ, describing changes in partitioning of water and energy fluxes at the

land surface, yields a quick overview of dominant impacts on streamflow. The resultant patterns are spatially coherent and in agreement with previous studies. The quantitative separation of impacts of basin changes on streamflow supports the hypothesis that humans directly and indirectly alter water resources at the regional and large basin scale. Most prominent are changes in the seasonality of climate due to global greenhouse gas emissions (Thomson, 1995; Barnett et al., 2008) and intensified agricultural landuse, especially by artificial drainage and irrigation. The results suggest that the direction and magnitude of human impacts distinctly vary with climate, soil, landuse and hydrographic conditions.

Last, we tested how concurrent effects of basin and climate changes on streamflow, encoded in the change direction ϖ, influenced the accuracy in predicting streamflow changes. We found that there is a clear dependency on prediction accuracy. So, generally all sensitivity methods yield minimal errors, when a climate change direction is evident, which is a solid argument for the applicability of the Budyko and the CCUW frameworks.

The results show that both frameworks agree in prediction accuracy, but the impact of basin changes within this dataset is too large to assess which theory is better than the other. However, we argue that the CCUW method is superior to the Budyko framework, as it can be applied to any reasonable hydro-climatic state without any parametrisation. Still, real changes in basin characteristics and uncertainties in data which are essentially attributed to basin characteristic changes, are not predictable. So we conclude, that these impacts play a role and one needs to know about these effects, when applying any kind of climatic sensitivity framework.

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Table 1. Budyko functions and their partial differentials as used in the text.

Budyko function	$\frac{\partial E_{\mathrm{T}}}{\partial P}$	$\frac{\partial E_{\mathrm{T}}}{\partial E_{\mathrm{p}}}$	Reference
$E_{\mathrm{T}} = E_{\mathrm{p}} \cdot \tanh\left(\frac{P}{E_{\mathrm{p}}}\right)$	$1 - \tanh^2\left(\frac{P}{E_{\mathrm{p}}}\right)$	$-\frac{P}{E_{\mathrm{p}}}\left(1 - \tanh^2\left(\frac{P}{E_{\mathrm{p}}}\right)\right) + \tanh\left(\frac{P}{E_{\mathrm{p}}}\right)$	Ol'Dekop (1911)
$E_{\mathrm{T}} = \frac{E_{\mathrm{p}} \cdot P}{\left(P^n + E_{\mathrm{p}}^n\right)^{1/n}}$	$\frac{E_{\mathrm{T}}}{P}\left(\frac{E_{\mathrm{p}}^n}{P^n + E_{\mathrm{p}}^n}\right)$	$\frac{E_{\mathrm{T}}}{E_{\mathrm{p}}}\left(\frac{P^n}{P^n + E_{\mathrm{p}}^n}\right)$	Mezentsev (1955)
	$\frac{\partial E_{\mathrm{T}}}{\partial n} = \frac{E_{\mathrm{T}}}{n}\left(\frac{\ln\left(P^n + E_{\mathrm{p}}^n\right)}{n} - \frac{\left(P^n \ln(P) + E_{\mathrm{p}}^n \ln(E_{\mathrm{p}})\right)}{P^n + E_{\mathrm{p}}^n}\right)$		Roderick and Farquhar (2011)

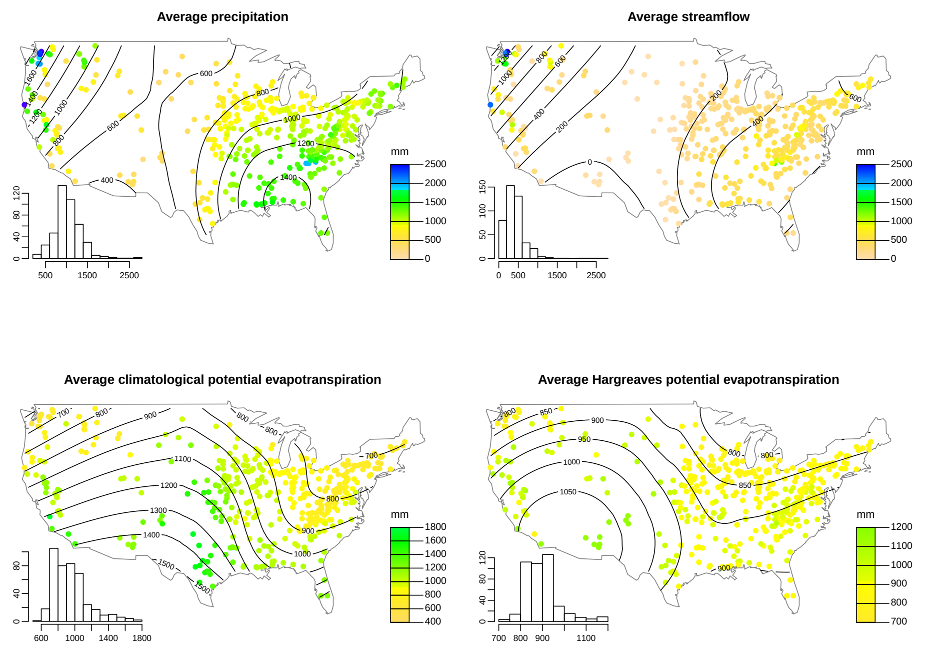

Fig. 1. Long term annual average of hydroclimatic variables of the US MOPEX dataset (1949–2003). The contour lines are derived from fitted polynomial surfaces (using R Development Core Team, 2010, stats::loess of the variables using the river gauge locations). The map of the US is taken from the maps package (Becker et al., 2011).

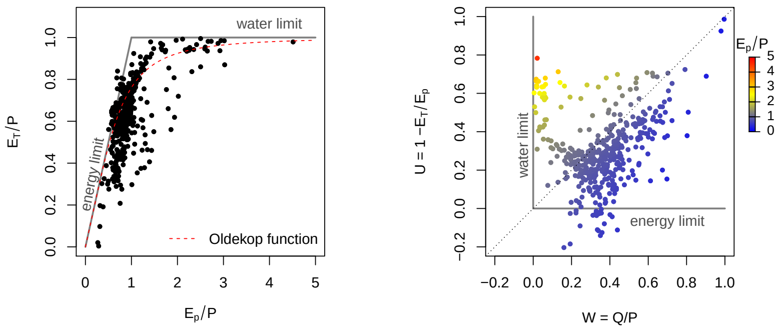

Fig. 2. Budyko (left) and UW space (right) plots of the period (1949–2003) of the MOPEX dataset. E_p is computed by the Hargreaves method. The 1 : 1 line in the UW space diagram separates areas with energy limitation ($E_p/P < 1$) and water limitation ($E_p/P > 1$). Grey lines indicate the water and energy limits.

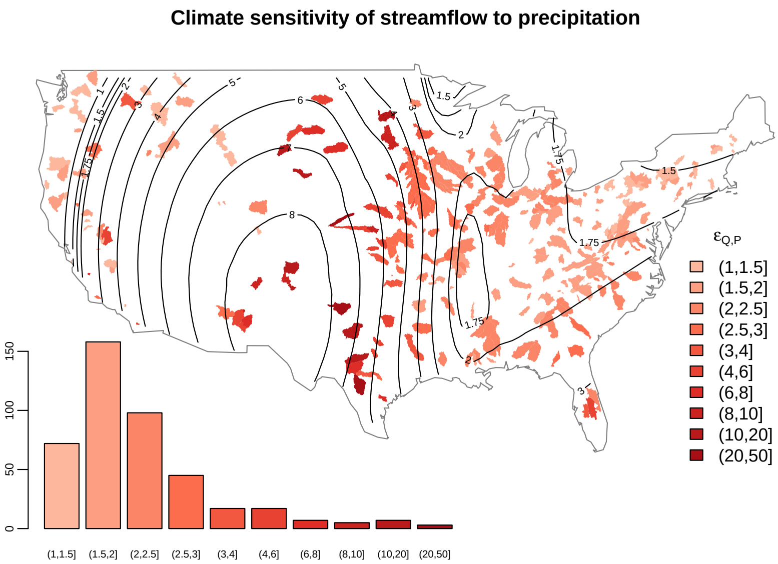

Fig. 3. Map of sensitivity coefficients of streamflow to precipitation estimated by the CCUW method based on long-term averages of (P, E_p, Q) for the period 1949–2003. The coloured polygons display $\varepsilon_{Q,P,\mathrm{CCUW}}$ for each basin using the watershed boundaries (retrieved from ftp://hydrology.nws.noaa.gov/pub/gcip/mopex/US_Data/Basin_Boundaries/, accessed in October 2011). An interactive version of this map can be found at http://wwwpub.zih.tu-dresden.de/~s1748291/Sensitivity.htm.

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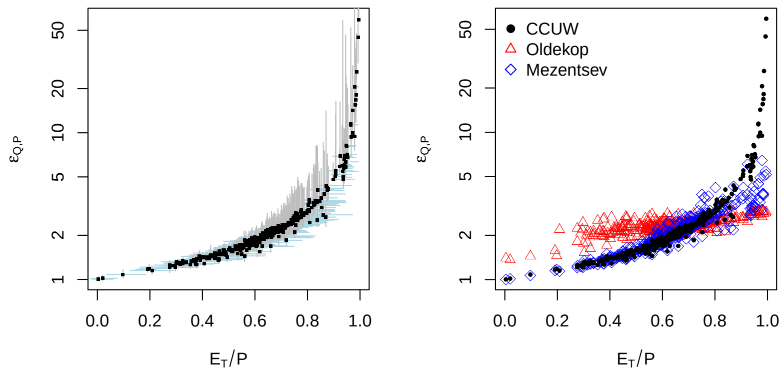

Fig. 4. Sensitivity coefficients of streamflow to precipitation as function of E_T/P. Left panel: $\varepsilon_{Q,P;\mathrm{CCUW}}$ computed for the CCUW method. Dots represent $\varepsilon_{Q,P;\mathrm{CCUW}}$ using long-term average data of the respective basin. Vertical grey lines depict the interquartile range of $\varepsilon_{Q,P;\mathrm{CCUW}}$ estimated for each year in the record, while lightblue horizontal lines show the interquartile range for E_T/P. Right panel: $\varepsilon_{Q,P}$ for different methods using long-term averages of (P, E_p, Q) of the period 1949–2003. Note, that a logarithmic y-axis is used for both plots.

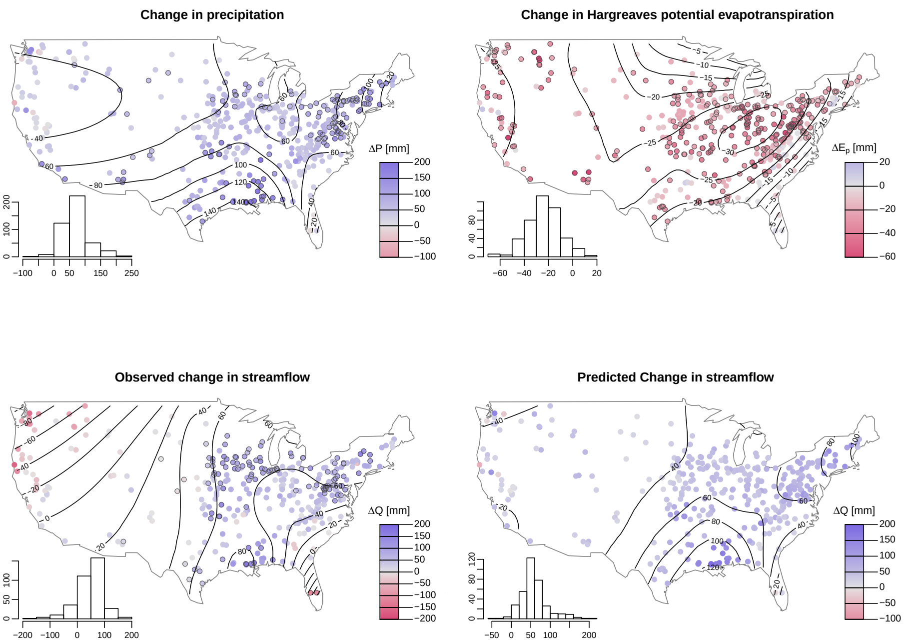

Fig. 5. Map of absolute change in hydro-climatic variables of the MOPEX dataset, comparing changes between the periods 1949–1970 and 1971–2003. Changes are given in (mm). Significant changes in the mean comparing both periods ($\alpha = 0.05$) are denoted with an open circle.

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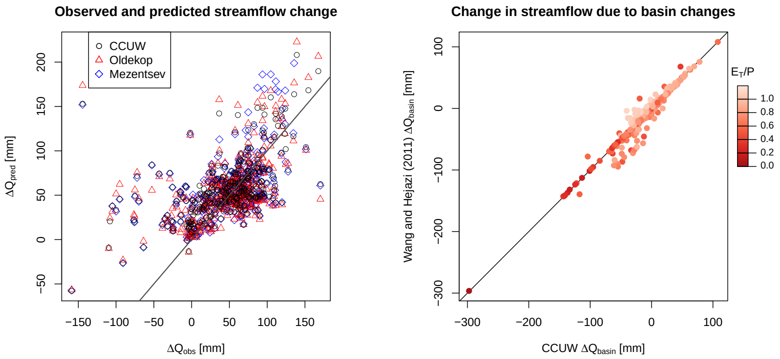

Fig. 6. Left: Scatterplot of observed vs. predicted changes in streamflow for MOPEX dataset without stations with missing data. The vertical difference to the 1 : 1 line depicts the deviation of the prediction to the observed value. Right: Estimated absolute change in streamflow, which is attributed to basin characteristic changes (ΔQ_{basin}). We compare the estimates of the method of Wang and Hejazi (2011) (employing Mezentsev function) with the CCUW estimates, i.e. the difference between observed and predicted changes. The colour of the dots represents the evaporation ratio E_T/P.

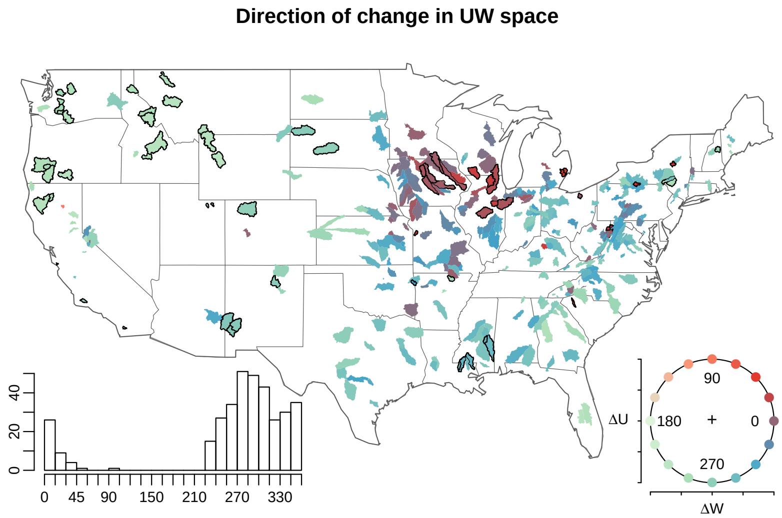

Fig. 7. Mapping of the change direction ω. The colour of the polygons indicates the value of ω with the corresponding wheel legend in the bottom right. Polygons with a black border indicate significant (t-test, $\alpha = 0.05$) changes of the absolute value in UW space. An interactive version of this map can be found at http://wwwpub.zih.tu-dresden.de/~s1748291/UWChangemapping.htm.

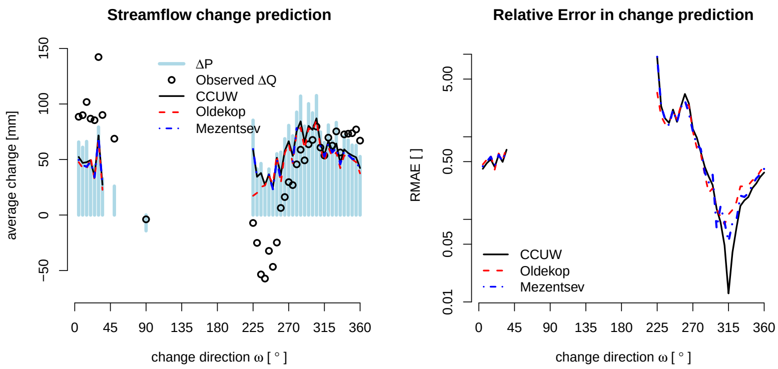

Fig. 8. Influence of change direction ϖ on streamflow change predictions. Left panel: Absolute bin average of changes in precipitation and streamflow prediction (CCUW, Oldekop, Mezentsev) and observation. Right panel: RMAE of predicted changes per bin using a logarithmic y-axis. The binning is based on a retrospective analysis of the change direction ϖ, whereby each bin has an angle width of 5°. Not all directions are observed.

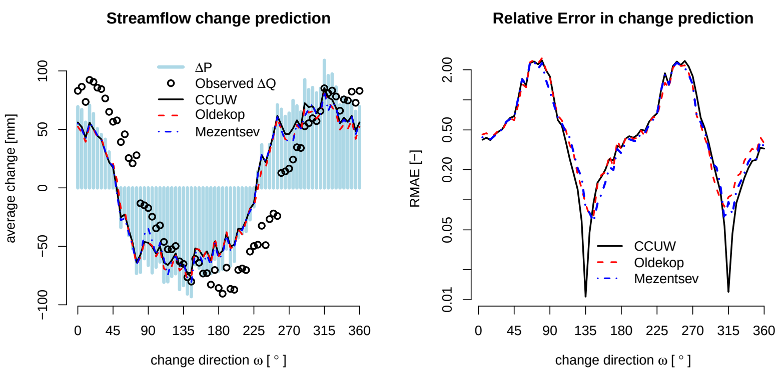

Fig. 9. Same as Fig. 8 but for this plot 5 prediction scenarios of the whole MOPEX dataset have been used. The periods have been 3 times forward and 2 times backward, where each period consists of 20 yr (water years). The start years of the first and second period are: 1951, 1971; 1951, 1981; 1961, 1981; 1981, 1961; 1981, 1951.