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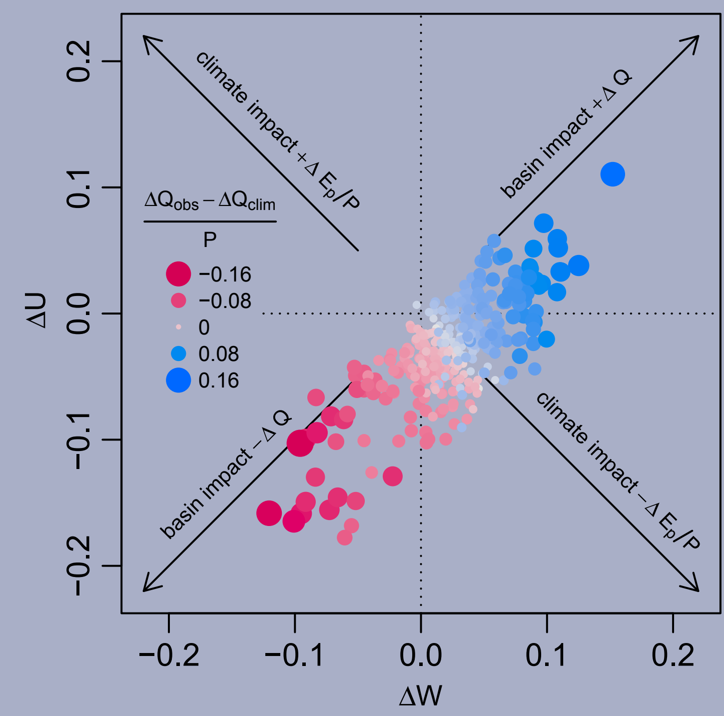

Maik Renner

Land use effects and climate impacts on evapotranspiration and catchment water balance

Tharandter Klimaprotokolle
Band 18

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TECHNISCHE UNIVERSITÄT DRESDEN

Fakultät Umweltwissenschaften

LAND USE EFFECTS AND CLIMATE IMPACTS ON EVAPOTRANSPIRATION AND CATCHMENT WATER BALANCE

EINFLUSS VON LANDNUTZUNG UND KLIMA AUF DIE GEBIETSVERDUNSTUNG UND DEN WASSERHAUSHALT VON FLUSSEINZUGSGEBIETEN

Kumulative Dissertation zur Erlangung des akademischen Grades
Doctor rerum naturalium (Dr. rer. nat.)

vorgelegt an der Fakultät Umweltwissenschaften
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ERKLÄRUNG DES PROMOVENDEN

Die Übereinstimmung dieses Exemplars mit dem Original der Dissertation zum Thema: **„Land use effects and climate impacts on evapotranspiration and catchment water balance“** wird hiermit bestätigt.

Ort, Datum

Unterschrift

Für Dani und Luisa

KURZFASSUNG

Die Verdunstung ist ein maßgeblicher Prozess innerhalb des Klimasystems der Erde, welche den Wasserkreislauf mit dem Energiehaushalt der Erde verbindet. Eine zentrale wissenschaftliche Herausforderung ist, zu verstehen, wie die regionale Wasserverfügbarkeit durch Änderungen des Klimas oder der physiographischen Eigenschaften der Landoberfläche beeinflusst wird.

Mittels einer integrierten Datenanalyse von vorhandenen langjährigen Archiven hydroklimatischer Zeitreihen werden die folgenden wissenschaftlichen Fragestellungen dieser Dissertation diskutiert:

A Haben beobachtete Änderungen der Landoberfläche und des Klimas zu nachweisbaren, instationären hydroklimatischen Änderungen geführt?

B Lassen sich die hydroklimatischen Auswirkungen von Klimaänderungen und Änderungen der Landoberfläche voneinander unterscheiden?

C Welche Faktoren beeinflussen die Sensitivität von Abfluss und Verdunstung auf Veränderungen der klimatischen und physiographischen Randbedingungen?

Hierbei fokussiert sich die Arbeit auf Änderungen im langjährige Mittel und im Jahresgang von hydroklimatischen Variablen auf der räumlichen Skala von Flusseinzugsgebieten.

Zur Untersuchung des hydrologischen Regimes wurde ein harmonischer Filter angewandt, der es erlaubt, die Eintrittszeit des Jahresgangs (Phase) zu quantifizieren. Diese klimatologische Kenngröße wurde für eine Vielzahl von Einzugsgebieten in Sachsen untersucht, wobei sich vor allem für die Gebiete in den Kammlagen des Erzgebirges signifikante Veränderungen ergaben. Es konnte gezeigt werden, dass die signifikante Phasenverschiebung der Temperatur seit Ende der 1980er Jahre zu einer verfrühten Schneeschmelze und dadurch zu einem Rückgang des Abflusses bis in die Sommermonate hinein geführt hat.

Desweiteren wurde eine modellbasierte Datenanalyse entwickelt, welche auf Massen- und Energieerhalt von Einzugsgebieten im langjährigen Mittel beruht. Das entwickelte Konzept erlaubt es, Auswirkungen von Klimaänderungen von anderen Effekten, welche z.B. durch Landnutzungsänderungen bedingt sind, abzugrenzen und zu quantifizieren. Die Ergebnisse einer Sensitivitätsanalyse dieses Konzeptes sowie die Anwendung auf einen umfangreichen hydroklimatischen Datensatz der USA zeigen: (i) Veränderungen im Wasser- oder Energiedargebot beeinflussen auch die Aufteilung der Wasser- und Energieflüsse. (ii) Die Aridität des Klimas

und nachgeordnet die physiographischen Faktoren bestimmen die Sensitivität von Verdunstung und Abfluss. (iii) Beide Faktoren beeinflussen die Stärke und Richtung der Auswirkungen von physiographischen Änderungen. (iv) Anthropogene Veränderungen der Landoberfläche führten zum Teil zu stärkeren Auswirkungen als klimatisch bedingte Änderungen.

Zusammenfassend zeigt sich, dass Änderungen von Landnutzung und Klima zu Verschiebungen im Wasserhaushalt führen können und damit auch die Annahme von Stationarität verletzen. Hydroklimatische Veränderungen bieten aber auch eine Gelegenheit zum Testen von Theorien und Modellen, um somit die grundlegenden Zusammenhänge zu erkennen, welche nicht durch Änderungen der Randbedingungen hinfällig werden.

ABSTRACT

Evapotranspiration (E_T) is a dominant Earth System process that couples the water and energy cycles at the earth surface. The pressure of global environmental changes foster the broad scientific aim to understand impacts of climate and land-use on evapotranspiration under transient conditions. In this work, the spatial scale of river catchments is addressed through data analysis of hydrological and meteorological archives with E_T classically derived through water balance closure.

Through a synthesis of various catchments with different climatic forcings and hydrological conditions, the core objectives of this thesis are:

A Did environmental changes in the past, such as climatic- or land-use and land cover (LULC) changes, result in detectable non-stationary changes in the hydro-climate time series?

B How can the impacts of climatic- from LULC changes on the hydroclimatology of catchments be separated?

C What are the factors that control the sensitivity of E_T and streamflow to external changes?

These research questions are addressed for the climatic scales of long-term annual averages and seasonal conditions which characterise the hydroclimatology of river catchments.

Illustrated by a rich hydro-climatic archive condensed for 27 small to medium sized river catchments in Saxony, a method is proposed to analyse the seasonal features of river flow allowing to detect shifting seasons in snow affected river basins in the last 90 years. Observations of snow depth at these same times lead to the conclusion, that changes in the annual cycle of air temperature have a large influence on the timing of the freeze-thaw in late winter and early spring. This causes large changes in storage of water in the snow pack, which leads to profound changes of the river regime, particularly affecting the river flow in the following months.

A model-based data analysis, based on the fundamental principles of water and energy conservation for long-term average conditions, is proposed for the prediction of E_T and streamflow, as well as the separation of climate related impacts from impacts resulting from changes in basin conditions. The framework was tested on a large data set of river catchments in the continental US and is shown to be consistent with other methods proposed in the literature. The observed past changes highlight that (i) changes in climate, such as precipitation or evaporative demand, result in changes of the partitioning within the water and energy balance, (ii) the aridity of the

climate and to a lesser degree basin conditions determine the sensitivity to external changes, (iii) these controlling factors influence the direction of LULC change impacts, which in some cases can be larger than climate impacts.

This work provides evidence, that changes in climatic and land cover conditions can lead to transient hydrological behaviours and make stationary assumptions invalid. Hence, past changes present the opportunity for model testing and thereby deriving fundamental laws and concepts at the scale of interest, which are not affected by changes in the boundary conditions.

CONTENTS

LIST OF MANUSCRIPTS

Renner, M. and Bernhofer, C.: Long term variability of the annual hydrological regime and sensitivity to temperature phase shifts in Saxony/Germany, Hydrology and Earth System Sciences, 15, 1819–1833, doi: 10.5194/hess-15-1819-2011, 2011.

Renner, M., Seppelt, R., and Bernhofer, C.: Evaluation of water-energy balance frameworks to predict the sensitivity of streamflow to climate change, Hydrology and Earth System Sciences, 16, 1419–1433, doi: 10.5194/hess-16-1419-2012, 2012.

Renner, M. and Bernhofer, C.: Applying simple water-energy balance frameworks to predict the climate sensitivity of streamflow over the continental United States, Hydrology and Earth System Sciences, 16, 2531–2546, doi: 10.5194/hess-16-2531-2012, 2012.

SYMBOLS AND ABBREVIATIONS

LIST OF SYMBOLS

Symbol	**Units**	**Description**
A_x		amplitude in the units of the measure x
C_E		catchment efficiency
E_p	mm	potential evapotranspiration
$E_{p,\mathrm{CRU}}$	mm	CRU TS 3.1 potential evapotranspiration estimate
$E_{p,\mathrm{Hamon}}$	mm	Hamon potential evapotranspiration
$E_{p,\mathrm{Hargreaves}}$	mm	Hargreaves potential evapotranspiration
E_T	mm	actual evapotranspiration
G	$\mathrm{W\,m^2}$	ground heat flux
H	$\mathrm{W\,m^2}$	sensible heat flux
L	$\mathrm{kJ\,kg^{-1}}$	latent heat of vaporisation
P	mm	precipitation
Q	mm	areal runoff
Q_{50}	d	date of 50% of annual flow (half-flow date)
ΔQ_{obs}	mm	observed change in Q
ΔQ_{clim}	mm	predicted change in Q due to climate variation
R^2	-	explained variance
R_n	$\mathrm{W\,m^2}$	net radiation
RR	-	runoff ratio
RR_3	-	three-monthly running runoff ratio
ΔS_e	$\mathrm{W\,m^2}$	change in heat storage
ΔS_w	mm	change in water storage
T_{coef}		slope of the linear regression of $\phi_{RR} \sim \phi_T$
TR	K	diurnal temperature range
U	-	relative excess of energy
W	-	relative excess of water
Y_x		Fourier transform of some series x (2.1)
n	-	catchment parameter
Φ	-	aridity index, $\Phi = E_p/P$

$\bar{\alpha}$		circular mean (2.4)
$\varepsilon_{E_T,n}$	-	sensitivity coefficient of E_T to the catchment parameter n
$\varepsilon_{E_T,P}$	-	sensitivity coefficient of E_T to precipitation P
ε_{P}	-	sensitivity coefficient to precipitation P
ε_{E_p}	-	sensitivity coefficient to potential evapotranspiration E_p
$\varepsilon_{Q,P}$	-	sensitivity coefficient of runoff Q to precipitation P
$\varepsilon_{Q,P;\mathrm{ccuw}}$	-	$\varepsilon_{Q,P}$ of the CCUW method
$\varepsilon_{Q,P;\mathrm{mez}}$	-	$\varepsilon_{Q,P}$ following the Mezentsev function
ϕ_{RR}	d	annual phase of runoff ratio
ϕ_{T}	d	annual phase of temperature
ρ_{cc}	-	circular correlation coefficient
ρ_{c-l}	-	circular-linear correlation coefficient
ρ_{snow}	-	circular-linear correlation between snow depths in March and ϕ_{RR}
σ_{α}^2		circular variance
ω	°	change direction in UW space, note that in chapter 2 α is used
ω_{obs}	°	observed change direction in UW space
ω_{Mez}	°	change direction in UW space computed following the Mezentsev function

LIST OF ABBREVIATIONS

Abbreviation	**Description**
CCD	climate change directions in UW space
CCUW	climate change impact hypothesis in UW space
CE	Central Europe
CHMI	Czech Hydro-meteorological Service
CRU	Climate Research Unit, University of East-Anglia
CUSUM	cumulative sums of standardised variables
DWD	Deutscher Wetterdienst, German Weather Service
FAO	Food and Agricultural Organization
HIGRADE	Helmholtz Interdisciplinary Graduate School for Environmental Research
IAHS	International Association of Hydrological Sciences
IQR	interquartile range
LULC	Land Use and Land Cover
MOPEX	Model Parameter Estimation Experiment
OK	Ordinary Kriging
RMSE	root-mean-square error
SRTM	Shuttle Radar Topography Mission
UW	the state space of relative excess of energy U vs. relative excess of water W

1 INTRODUCTION

Πάντα ῥεῖ - panta rhei
everything flows - alles fließt
(Simplicius of Cilicia, Heraclitus)

Everything flows, this saying of Simplicius of Cilicia is probably the shortest conclusion one can derive when studying earth system sciences. It is true for the water cycle, the flow of air in the atmosphere or the evolution of life. It also implies that interactions between elements are not fixed either.

An invisible but central element of earth system processes is the evapotranspiration of water, which, powered by the sun, lifts huge amounts of water into the atmosphere and keeps the water cycle running. Evapotranspiration is also highly interactive and links land surface properties with the flow of water and energy.

Evapotranspiration and its controls are thus a major research topic in past and present hydrological and atmospheric sciences. And as we are recognising that we are living in a world of global change, the International Association of Hydrological Sciences (IAHS) is about to announce a decade on predictions under change (PUC) (Thompson et al., 2011). Thus, making "panta rhei" to a central research topic[1].

The emphasis on the non-stationary behaviour of climate, land use and hydrology is challenging many current modelling tools which inherently assume no change of climate or land use. Hence predicting under change will also challenge our knowledge of earth system processes. This research aims to improve the knowledge on evapotranspiration and the roles of climate and land use under change. Therefore, methods are developed to identify non-stationary signals of hydrological and climatological records at the hydrological scale of river catchments.

1.1 MOTIVATION AND RELEVANCE

1.1.1 SCIENTIFIC IMPORTANCE OF EVAPOTRANSPIRATION

Under the term evapotranspiration we usually refer to the processes of evaporation and transpiration which transport water from the earth surface to the atmosphere.

[1] Demitris Koutsoyiannis on the discussion of the preliminary science plan of the IAHS at `http://distart119.ing.unibo.it/iahs/?p=264#comment-98`

E_T is a dominant flux of the surface water balance equation (1.1) and because the phase transition to water vapour requires energy, it is also part of the surface energy balance (1.2). This makes E_T an highly relevant process for both, the hydrological and the atmospheric sciences (Blöschl, 2005).

The simplified water and energy balance equations (in short water-energy balance) for a catchment can be written as (Dyck and Peschke, 1995):

$$P = E_T + Q + \Delta S_w \tag{1.1}$$

$$R_n - G = E_T L + H + \Delta S_e. \tag{1.2}$$

The surface receives water in the form of precipitation P, which is partitioned into actual evapotranspiration E_T, runoff Q and a water storage term ΔS_w. On the energy side, the latent heat flux $E_T L$, where L is the heat of vaporisation, describes the amount of energy required for the phase transition of liquid water into water vapour. The energy needed is supplied by the net radiation balance R_n at the earth surface. This energy is also partitioned into heating of the air, i.e. the sensible heat flux H, the heating of the soil G and some heat storage ΔS_e.

The water-energy balance equations show that E_T is mainly controlled by climate conditions, i.e. by the supply of water and energy. However, the evaporated water also feeds back to the atmosphere and E_T largely controls surface properties, which makes E_T an implicit variable (Bernhofer et al., 2002; Yang et al., 2008).

A further interesting property is the wide range of involved scales. The supply of water through precipitation is a discontinuous, event scale process ranging from minutes to days with seasonally changing patterns. The supply of energy is highly oscillatory, with diurnal and seasonal courses, but largely modified by the probabilistic nature of clouds, wind and lateral heat advection (Bernhofer et al., 2002).

Further, abiotic processes and feedback mechanisms, like surface evaporation of soils, lakes and interception are fast processes due to the relatively small water storage capacity (Savenije, 2004). The large amount of heat required for vaporisation $L \approx 2500 kJkg^{-1}$ quickly depletes the radiative energy supplied to a wet surface.

Biotic processes and feedbacks, in particular the transpiration of plants, are able to exploit the large water storage of the soil through roots - which thus increases the time scale and the spatial (vertical) scale compared to purely abiotic evaporation processes (Savenije, 2004). Further, the biosphere is able to convert photonic energy into chemical energy (Kleidon, 2010) which allows multiple pathways of energy and mass exchange and thus creates a rich structure affecting and controlling transpiration.

This apparent rich structure makes it quite complicated to describe (i) the full set of multiple processes at different scales and (ii) their interactions ranging from local to global scales. While many processes can be physically described, it is thus unclear how to upscale small scale processes given the large structural and functional heterogeneity? Furthermore, E_T is invisible for the human eye and until recently it can only be observed indirectly for a given area of interest (besides lysimeter data for grassland and crops). Thus, scientific approaches to understand areal E_T usually rely on measurements of the other terms in the water-energy balance equations (Brutsaert, 1982). Only recently larger data sets of directly observed E_T became available (FLUXNET) and are starting to be exploited (see, e.g., Jung et al. (2011)).

1.1.2 PRESSURE OF HUMAN DRIVEN CHANGES

Human activities are exposed to natural fluctuations of climate and hydrology. However, the apparent exponential growth of human population is based on exploitation of natural resources. Especially the use of fossil energy resources accelerated anthropogenic impacts (Crutzen, 2002) with some relevant listed below:

- anthropogenic greenhouse gas emissions, changing the long-wave radiative properties of the atmosphere
- rapid changes of land use and land cover (LULC), hydromorphology, intensive agriculture and forestry management, changing the land surface properties.
- environmental pollution in various forms (air pollution, acid rain, pesticides) leading to degeneration of ecosystems.

The sum of human activities is large enough to change natural equilibria / balance states at the local scale (e.g. land management), at the regional scale (e.g. environmental pollution), and at the global scale (climate change). Thus, there is an urgent need to understand how these changes effect specific processes and physiographic features of the earth system.

1.1.3 PRACTICAL IMPORTANCE OF EVAPOTRANSPIRATION

The flux of evapotranspiration is about 60% of the annual terrestrial water balance (Baumgartner and Reichel, 1975). Thus, E_T plays a dominant role, determining the amount of water available for humans and ecosystems.

Highly relevant for society is the management of water resources with regard to the variability of natural water supply and demand, which is also largely controlled by evapotranspiration. Thus, as E_T largely determines plant water requirements and yield, E_T is also important for agriculture and forestry. This brief consideration of the societal importance of E_T actually sets the scale where reliable quantifications of E_T are needed: While for agriculture the field and landscape scale is relevant, for water resources the catchment scale is important.

1.2 SCOPE

1.2.1 FOCUS ON THE CATCHMENT SCALE

In this work the scale of river catchments has been chosen. For a catchment we refer to the contributing area of a selected location where runoff occurs. So the water received over the catchment contributes to the lateral flow of water at the outlet of the catchment. Typical sizes of a catchment can range from a few hectares of a small stream up to millions of square kilometres for a large river basin. The runoff at an outlet represents an integral over the water fluxes within the catchment and thus integrates all involved processes. So external changes within the catchment must be consistent and large enough to leave detectable signals in the observed streamflow signal. While this helps to detect only relevant signals, it may hinder to detect compensating effects (Arnell, 2002; Brooks, 2003).

As elaborated above, this scale has practical relevance for water resources as we can assign some amount of fresh water to a certain area and period. Further river gauge observations allow to estimate the runoff term in the water balance of river catchments. By closing the water balance, usually for long term annual averages, actual evapotranspiration can be estimated for the catchment scale. Thus, in this work E_T will be approached through the catchment water balance. For this reason, the analysis of streamflow records and the hydrological response to external changes will be central.

1.2.2 CHANGES IN THE HYDROCLIMATOLOGY OF RIVER CATCHMENTS

The hydroclimatology of river catchments comprises the statistical characteristics of river flow at the climate time scale. Typically, we are interested in

a) long-term annual average conditions and year to year variability

b) seasonality, i.e. the river regime

c) extremes, such as floods and droughts.

All these characteristics are of high relevance for society and environment as they determine the general availability of fresh water, impacting a range of further issues such as water quality and fresh water ecology.

The pressure of anthropogenic changes is expected to change the hydroclimatology of river catchments (Bates et al., 2008) and there is evidence that these changes are already ongoing (Hamlet et al., 2007). On a global scale an intensification of the hydrological cycle due to global warming has been reported (Huntington, 2006). Yet there is evidence from pan evaporation records (Brutsaert and Parlange, 1998), signals in precipitation records (Zhang et al., 2007; Min et al., 2011) and streamflow records (Dai et al., 2009). Also changes in the annual cycle of temperature (Thomson, 1995; Stine et al., 2009) are believed to impact river regimes of snow dominated regions showing advances in timing (Stewart et al., 2005; Barnett et al., 2008). However, regional change signals of hydrological records can be quite different and these signals do not always correspond to climate signals (Milliman et al., 2008). Thus, it is important to stress that climate impacts on hydroclimatological characteristics are not the only concern. Other major concerns are LULC changes (Bronstert et al., 1995; DeFries and Eshleman, 2004; Foley et al., 2005; Legates et al., 2005; Zhang and Schilling, 2006) effecting landscape E_T. Further, reallocation of water resources with direct impacts on streamflow can result in much larger changes than by climatic changes only (Vörösmarty et al., 2000).

In this work the focus is set to changes in characteristics describing and controlling the annual average conditions as well as the seasonal course. These features describe the centre of a probability distribution and changes therein are likely to have more implications than changes on the extremes of a distribution being caused by single events. This means that in this work processes which are driven by extremes, such as droughts and floods are not covered.

1.2.3 HYDRO-CLIMATE DATA ANALYSIS

The very aim of studying changes in hydrology and climatology is to provide advice for stakeholders to adapt to the already existing or anticipated environmental changes of the future. However, before we do any predictions for the future, we have to make sure to be able to resolve past conditions, which is necessary as we are not in the position to do experiments. This puts emphasis on data collection through routine measurements, archiving and homogeneity analysis

– tasks which require great diligence. High quality data archives then allow to understand past variations and trends, whether through empirical or model-based data analysis.

Empirical data analysis Widespread routine measurements of meteorological and hydrological variables are available since about 1900. These archives allow useful statistical inference of changes. Thereby most studies present trend analyses (Lettenmaier et al., 1994; Dai et al., 2009; Stahl et al., 2010; KLIWA, 2003). And although trends are generally understood, conclusions solely on trends are limited, because the trend estimate is influenced by the period chosen and it is of limited value to extrapolate these trends (von Storch, 1995; Clarke, 2010) as there may be underlying periodicities and long-range dependencies in the data (Cohn and Lins, 2005; Koutsoyiannis, 2006). Also many studies report trends in single months or seasons of the year (Stahl et al., 2010) which indicate changes in the seasonal course (Déry et al., 2009), however, mostly without an holistic treatment of the annual cycle.

Rather than analysing trends of isolated variables many studies try to attribute observed changes to external changes. Mostly climate variations are linked to hydrological variations (Krakauer and Fung, 2008). Thus, for catchment scale analyses meteorological observations have to be interpolated. This prerequisite requires geospatial data processing and uncertainty treatment.

Most frequently applied are correlation based approaches linking climate and hydrological variables. While correlation provides evidence of statistical links, attribution of observed changes requires careful testing (Merz et al., 2012). Another problem can be the assumption of a linear relation and the possible colinearity of predictors in multivariate studies (Clarke, 2010).

A different popular statistical method trying to relate hydrological changes to climate is the elasticity concept (Sankarasubramanian et al., 2001; Zheng et al., 2009; Xu et al., 2012). Here, the anomalies are evaluated for similar variations. Although, this approach is useful for a quick scan of statistical links, usually problems are neglected such as colinearity of predictors or interannual changes in water storage.

Model based data analysis Since statistical approaches of change analysis have low or no a-priori assumptions on the nature of the link, they are usually not able to extract the physical cause of change. This is where physically based modelling of natural systems becomes important.

As the main research objective is a long standing problem in hydrology, there are numerous modelling studies available, see Bronstert (2006) for an introduction. The applicability of these models boils down to the question if we can transfer the model structure and its parameters to the conditions of future climate or LULC conditions (Klemeš, 1986)? This is the key question and is still under much debate (Merz et al., 2011). Is it better to transfer conceptual models which require calibration but have not so many parameters? Or should we use fully physical based models which do not require calibration but on the other hand upscale small scale processes through non-observable spatial and temporal heterogeneity?

This dilemma demonstrates the difficulty of the modelling task. It requires both, falsifiable hypotheses and the data to test them (Sivapalan, 2005; Kirchner, 2006).

1.3 OBJECTIVES AND RESEARCH QUESTIONS

The main title of this thesis opens up a wide field of interesting research and hence it is necessary to limit the scope and to derive specific research questions. As outlined in the scope section above, this work sets focus on catchment scale observations, with an emphasis on the hydroclimatology. The temporal scales considered are the (i) seasonal and (ii) the long-term annual average time scales.

In the following subsections, which are classified by the two temporal scales, the specific research questions of this thesis are formulated. The thematic research questions regard external impacts on the hydroclimatology of river basins.

These questions are approached through an analysis of hydrological and climatic data archives, describing the hydroclimatology of river basins over the last 50 to 100 years. However, suitable methods are needed on how to extract the signals of climate or LULC change from hydroclimatic records. Hence, a range of very interesting and challenging methodological questions emerge.

As guidance for the reader, both, the thematic and the methodological research questions formulated in this thesis are emphasised and numbered. In the summary chapter the specific questions will be revisited and answered.

1.3.1 SHIFTING SEASONS IN HYDROLOGY

Towards higher latitudes we observe that the annual cycle of solar radiation and thus temperature leads to distinct seasonal variations of hydrological variables such as streamflow. These seasonally varying meteorological forcings usually lead to high water availability during cold months and low availability during warm months when the demand for water is highest (with the exception of glacial river regimes). Thus, water resources managers early recognised the need for correct seasonal streamflow estimates to ensure water delivery (Loucks et al., 2005).

In many hydrological and climatology studies it is usually assumed that the annual cycle of temperature and streamflow is stationary (Shiklomanov, 1998) – an important assumption which simplifies many statistical methods.

However, there is increasing evidence from various data archives, that this assumption of stationarity, i.e. constant seasonal characteristics is invalid. The seminal paper of Thomson (1995) showed that there is considerable variability in the phase of the annual cycle of temperature. This means that the timing of the whole annual cycle can be different from year to year. Further, Thomson (1995) found that changes increase since the 1950s and he established a statistical link of the change in timing of temperature to the Keeling curve, i.e. the non-linearly increasing carbon dioxide concentrations in the atmosphere. This suggested that the anthropogenic induced climate change ultimately may also change the seasons.

Changes in the seasonality of temperature have impacts on vegetation supported by phenological studies e.g. Dose and Menzel (2004), impacts on the snow regime e.g. Mote et al. (2005) and thus on streamflow regimes e.g. Stewart et al. (2005); Barnett et al. (2008). The global investigation of the annual cycle of temperature by Stine et al. (2009) using the simple approach of fitting a sine curve to the temperature cycle supported the findings of Thomson (1995), but showed that the trends in the phase of temperature depend on regional different land surface characteristics such as the distance to the sea. Interestingly, the significant trends in the timing of global temperature have not been found in various GCM model results (Stine et al., 2009), indicating the still existing and eventually inevitable limits of such models.

The seasonal signal of hydrological records, also referred to as river regimes, is influenced by many processes, such as events of precipitation falling as rain or snow, the seasonal cycle of net radiation and the advection of heat driving evapotranspiration processes. Therefore, seasonal streamflow signals are spatially and temporally rather variable. This resulted in many different river regime classifications (Bower et al., 2004) which complicate the quantification of changes (Déry et al., 2009). Thus, most studies on hydrological data analysis avoid regime classifications and analyse each season or each month of the year separately. Studies focusing Central Europe such as Fiala (2008) and Stahl et al. (2010) found significant changes in streamflow. Often a change in the sign of the trend is reported with increasing flows in late winter, early spring and decreasing flows in late spring and summer. This suggests that there may be shifts in the timing of streamflow, but the statistics to test for such a shift are not sound when each month of the year is treated independently (Thomson, 1995). For this reason other timing measures of streamflow have been proposed. Most notable is the center of timing (Stewart et al., 2005), but this measure is dependent on the subjective choice of beginning and end of the year and there is no possibility to test the accuracy of this timing measure.

Objectives This part of the dissertation focuses on the analysis of the variability in the timing of hydrological regimes. Thereby, the following research questions are posed:

1. *How to estimate the seasonal timing of river flow?*

Measures are needed which allow inter-comparison between different hydrological regimes and thus do not depend on some kind of regime classification. Further, such measures should provide information on the accuracy of the estimate, which is not possible for existing streamflow timing measures.

Once we can quantify the timing of hydrological records the following questions can be addressed:

2. *Can we assume stationarity for the seasonality of hydrological records?*

This is important as many time series analysis methods assume a fixed timing of the seasonal cycle. Hence this assumption could be invalid if regime shifts in the seasonality are detected.

Further, the variability of the timing of hydrological records can be correlated with other observations such as the timing of the annual cycle of temperature to find out:

3. *What are the physical processes and where to expect changes in the timing of hydrological records?*

This understanding of why, where and how strong seasonal flow of rivers has changed in the past allows to identify regions where present and future climate changes may impact the availability of water for humans, vegetation and the environment.

1.3.2 Long-term annual average changes of evapotranspiration and streamflow

The pressure of climate and LULC changes requires robust estimates on how evapotranspiration and streamflow might change at the catchment scale. Thus, it is expected that external changes will modify the long-term average equilibrium of water and energy partitioning at the earth surface.

Given the high variability of future projections of precipitation and net radiation, the author thinks that first order estimates of changes in E_T and Q are sufficient with respect to the long-term annual time scale. However, such estimates should be based on fundamental principles, rather than purely empirical methods.

When analysing archives of climatological and hydrological records, we have to take account for the fact that the majority of observed gauge stations in influenced by both, climate and LULC changes. Thus, to avoid misinterpretation, e.g. attributing a change in streamflow to climate without testing for other causes (Jones, 2011), it is ideally necessary to separate the impact of both types of changes.

Objectives The problem of how to extract external change signals from hydro-climate data is approached through a model-based data analysis. Thus, first methodological questions need to be addressed:

4. *How to conceptualise the processes of water and energy partitioning of catchments at the hydroclimatic time scale?*

Here, we need to identify the driving fundamental principles and to find a model concept for the hydroclimatic time scale under the pressure of external changes. This also requires the definition of the meaning of potential changes in climate and LULC within this concept.

Variations in hydrological records are driven by variations of the climate forcings and by variations of the physical properties in a given catchment (Blöschl, 2005). It is expected that such external variation and changes result in specific patterns modified by the internal catchment processes (Sivapalan et al., 2001). Given this expectation we need to find out:

5. *How to identify and distinguish impacts of climate and land-use change from hydro-climate records?*

Climate change is considered to alter the hydrological cycle at global and regional scales. Many studies argue that changes in precipitation will have large impacts on the hydrology (Shiklomanov, 1998). Hence, there is the need to estimate the effect of changes in climatic variables, such as precipitation. However, it must be expected that climate changes are not translated linearly into hydrological changes (Risbey and Entekhabi, 1996) and impacts are considered to be affected by the climatic and hydrographic conditions of the respective basin (Piao et al., 2007).

6. *What determines the sensitivity of streamflow and evapotranspiration to changes in climate?*

This question regards the vulnerability of a certain river catchment to climatic changes and requires methods to predict the impacts of climatic changes.

7. *How large are basin change impacts compared with changes in climate?*

Although LULC changes are supposed to have impacts on the hydroclimatology of river catchments, these impacts are difficult to distinguish from climatic changes and hence quantifications are rather limited. For simplicity, changes in the physiographic properties of the catchment such as LULC change will be summarised as basin changes.

The proposed methods for separation of climate and basin changes are thoroughly tested with a large data set of hydro-climate data of the continental US. This allows a comprehensive analysis of the past hydro-climatic changes at the river catchment scale in the US in 20th century.

1.3.3 METHODOLOGICAL REQUIREMENTS: SIMPLICITY IS BEAUTIFUL

A central methodological focus of this work is to exploit hydrological and climatological data archives at the catchment scale. Thereby, the methods applied and developed adhere to the principles of parsimony and simplicity. This ensures generality, robustness, applicability, and comprehensibility. These points are important to stress, as e.g. experiences in the field of flood forecasting showed that it is better and more sustainable to train many people with a simple model, than to provide a complex model with few people understanding it (Werner et al., 2011).

This is in contrast to complex earth system models where many processes are being resolved. But for the research questions of this dissertation it would anyhow be necessary to aggregate the model output to the scales of interest. So, instead the author follows the spirit of Klemeš (1983) and search for appropriate conceptual models at the scales of interest.

1.4 STRUCTURE OF THE THESIS

The following three chapters represent independent articles, each with its own introduction, main part and conclusions. All papers have been published within the open access journal Hydrology and Earth System Sciences (HESS) of the European Geosciences Union (EGU). Each paper went through a peer-review process, whereby the manuscript is first published as Discussion paper in Hydrology and Earth System Sciences Discussions (HESSD). The core part of the review process is also open for everyone to follow [2].

In chapter 2 the seasonality of river regimes in Saxony/Germany is investigated (Renner and Bernhofer, 2011). The temporal scale of long-term annual average is central in chapter 3 and chapter 4. Chapter 3 presents an evaluation of two water-energy balance frameworks, developing the theory and presenting a thorough sensitivity study (Renner et al., 2012). The frameworks are then tested against a gradient of hydro-climatic conditions derived from the freely available MOPEX[3] dataset covering the continental U.S. (Renner and Bernhofer, 2012).

Chapter 5 then summarises the main results and draws final conclusions with regard to the objectives of the thesis.

[2] The discussion papers can be accessed through the paper websites:
Chapter 2: `www.hydrol-earth-syst-sci.net/15/1819/2011/`
Chapter 3: `www.hydrol-earth-syst-sci.net/16/1419/2012/`
Chapter 4: `www.hydrol-earth-syst-sci.net/16/2531/2012/`

[3] `ftp://hydrology.nws.noaa.gov/pub/gcip/mopex/US_Data/`

2 LONG TERM VARIABILITY OF THE ANNUAL HYDROLOGICAL REGIME AND SENSITIVITY TO TEMPERATURE PHASE SHIFTS IN SAXONY/GERMANY

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ABSTRACT

Recently, climatological studies report observational evidence of changes in the timing of the seasons, such as earlier timing of the annual cycle of surface temperature, earlier snow melt and earlier onset of the phenological spring season. Also hydrological studies report earlier timing and changes in monthly streamflows. From a water resources management perspective, there is a need to quantitatively describe the variability in the timing of hydrological regimes and to understand how climatic changes control the seasonal water budget of river basins.

Here, the timing of hydrological regimes from 1930–2009 was investigated in a network of 27 river gauges in Saxony/Germany through a timing measure derived by harmonic function approximation of annual periods of runoff ratio series. The timing measure proofed to be robust and equally applicable to both mainly pluvial river basins and snow melt dominated regimes.

We found that the timing of runoff ratio is highly variable, but markedly coherent across the basins analysed. Differences in average timing are largely explained by basin elevation. Also the

magnitude of low frequent changes in the seasonal timing of streamflow and the sensitivity to the changes in the timing of temperature increase with basin elevation. This sensitivity is in turn related to snow storage and release, whereby snow cover dynamics in late winter explain a large part of the low- and high-frequency variability.

A trend analysis based on cumulative anomalies revealed a common structural break around the year 1988. While the timing of temperature shifted earlier by 4 days, accompanied by a temperature increase of 1 K, the timing of runoff ratio within higher basins shifted towards occurring earlier about 1 to 3 weeks. This accelerated and distinct change indicates, that impacts of climate change on the water cycle may be strongest in higher, snow melt dominated basins.

2.1 Introduction

2.1.1 Motivation

In nival and pluvial catchments and river basins of Central Europe (CE) we observe a variable but distinct seasonal hydrological regime. This hydrological regime is a result of several processes induced by meteorological forcing and the properties of the receiving catchments. Looking at the water balance of typical basins in CE, precipitation has a small seasonal cycle compared to its variation and would alone not account for the distinctive seasonality of runoff. This is mainly introduced by basin evapotranspiration, resulting in lower flows during summer and early autumn. Also at catchments at higher elevations, snow accumulation and snow melt produce higher flows in late winter and early spring. Besides the local climate, catchment properties such as water storage in soils, evaporative demand of vegetation and human water management moderate the resulting hydrological regime.

Water resources management has to deal with the seasonality of hydrological regimes. Generally demand and water availability are out of phase, i.e. when the availability of water is lowest (summer) the demand is highest. This pressure on water delivery systems increases the need to correctly estimate the seasonality of hydrological processes (Loucks et al., 2005). Still, it is common practise to assume stationarity of monthly statistics and thus stationarity of the whole annual cycle in design studies for water resources management. However, there is increasing evidence for changes in the timing of the seasons from various disciplines. Earlier streamflow timing and snow melt have been reported e.g. by Stewart et al. (2005); Déry et al. (2009); Stahl et al. (2010). Further, phenological studies provide evidence of earlier spring season, e.g. Dose and Menzel (2004). Based on station and gridded data of surface temperature, Thomson (1995) and Stine et al. (2009) found tendencies of advanced seasonal timing.

Consequently, there is a need to estimate the timing of hydrological regimes, its variability and to check for long term changes, which could possibly violate the stationarity assumption of the annual cycle of hydrologic records. Furthermore, the sensitivity of the timing of hydrological regimes to changes in the phase of temperature needs to be assessed. This is especially important when considering the regional impacts of climate and land use change.

2.1.2 Seasonal changes in hydrologic records

Hydrological studies concerned with changes in streamflow within regions throughout CE usually analyse annual runoff and seasonal changes by monthly data (KLIWA, 2003; Fiala, 2008; Stahl et al., 2010). The majority of annual flow records in CE do not show significant trends, but spatially coherent trends in separate months have been reported. Remarkable is that positive streamflow trends have been found in winter months, which are followed by negative trends in spring (Stahl et al., 2010). Mostly it is concluded that these trends are a result of warmer winters which in turn lead to an earlier onset of snow melt. Mote et al. (2005) emphasise that the natural storage of water in snow affects greater water volumes than any human made reservoir and thus changes in snow pack directly affect river runoff.

There is a range of measures that can be used to directly estimate the timing of annual streamflow regimes, such as the timing of the annual maximum, the fraction of annual discharge in a given month or half flow dates (Court, 1962; Hodgkins et al., 2003; Regonda et al., 2005; Stewart et al., 2005). Even though these measures are relatively simple and easily understood, these metrics are only useful for hydrological regimes with distinct seasonality such as those dominated by snowmelt. Déry et al. (2009) note that synoptic events, e.g. warm spells in winter or late season precipitation may dominate such measures rather than long term changes in climate.

Relatively few studies have studied changes in the variability of the annual cycle, being the strongest signal in many climate records at mid to high latitudes (Huybers and Curry, 2006). By using a harmonic representation of the annual cycle, this variability has been studied for long records of surface temperatures (Thomson, 1995; Stine et al., 2009) and precipitation (Thompson, 1999). The resulting annual phases and amplitudes describe the timing of the annual cycle and its range based on the whole cycle instead of considering each month separately.

2.1.3 REGIONAL CLIMATE IN SAXONY

The Free State of Saxony is situated at the south-eastern border of Germany, covering an area of 18 413 km^2. In this study 27 river basins within Saxony have been analysed, which all belong to the Elbe River system. The climate is characterised by two main factors. First, there is a transition of the maritime western European climate to the continental climate of eastern Europe, which leads to a temperate warm and humid climate with cool winters and warm summers. Second, there is an orographic influence due to the mountain ranges at the southern border with elevations gradually increasing from 100 m up to 1200 m. Recently, the climate and observed trends have been described in detail by Bernhofer et al. (2008) and summarised by Franke et al. (2009). From the observed changes they deduce that climate change effects are more pronounced in Saxony than in other regions in Germany. They report long term shifts in observed global radiation and dependent variables such as potential evapotranspiration. These phenomena, also known as global dimming and brightening (Wild et al., 2005), have been very pronounced in Saxony due to reduced industrial and domestic emissions after German unification in 1990. Also with regard to air pollution, especially the ridge region of the Ore Mountains has been severely effected by tree die-off since the 1960s peaking in the 1980s (Kubelka et al., 1993; Fanta, 1997). With regard to precipitation Bernhofer et al. (2008) observed a positive trend in the number of droughts during growing season, combined with intensified heavy precipitation. These effects are partly compensated at the annual level by increased winter precipitation. However, at the same time winter snow depth and snow cover duration decreased, highlighting the effects of increasing temperatures especially during winter and spring.

2.1.4 OBJECTIVE AND STRUCTURE

The objectives of this paper are (1) to derive a climatology of the timing of the annual hydrological regimes for a range of river basins throughout Saxony; (2) to evaluate their interdecadal variability and trends; and (3) to determine the proximal processes governing the locally coherent patterns of the observed changes in timing.

To resolve these issues, a reliable measure for the timing, applicable for different hydrological regimes is needed. So instead of using streamflow records directly, we employ the basin runoff ratio, the ratio of discharge and basin precipitation. The emphasised seasonal fluctuation of runoff ratio is a direct measure of seasonal water availability and as being a normalisation, it makes different basins more comparable. The series of monthly runoff ratio are filtered for their annual periodicity by the harmonic method described in Stine et al. (2009) and the resulting annual phases represent a timing measure of the regime of the runoff ratio. The climatologic behaviour of the timing, being an angular variable, is then analysed by circular descriptive statistics. The interdecadal variability of the timing is being addressed by a qualitative method, namely cumulative departures of the average. Together with a correlation analysis to observed climate variables, such as timing of temperature, annual mean temperatures and monthly snow depths, we aim to identify the driving processes governing the changes in the timing of the runoff ratio.

2.2 METHODS

2.2.1 ANNUAL PERIODIC SIGNAL EXTRACTION

The aim is to estimate the timing of the annual cycle from a geophysical time series without a subjective definition of the seasons. Therefore, methods are necessary to extract the annual cycle signal from the data and to gain a time variant parameter, which defines the timing of the seasons.

In general, there are two ways to accomplish this task. First, there are form free models, which use some seasonal factor to describe a periodic pattern. This yields a good approximation to the periodic signal, at the cost, however, of estimating many parameters. The second approach are Fourier form models which are based on harmonic functions. These are generally defined by two parameters per frequency: phase (ϕ) and amplitude. These parameters are a natural representation of the seasonal cycle and are economic in terms of parameter estimation (West and Harrison, 1997). Using several long temperature series, Paluš et al. (2005) compared four different methods for estimating the temporal evolution of the annual phase (sinusoidal model fitting, complex demodulation via Hilbert Transform, Singular Spectrum Analysis and the Wavelet transform). They found good agreement between these methods and concluded that the annual phase is a robust and objective way to estimate the onset of seasons.

Recently, Stine et al. (2009) analysed trends in the phase of surface temperatures on a global perspective. They used the Fourier transform to compute annual phases and amplitudes:

$$Y_x = \frac{2}{12} \sum_{t=0.5}^{11.5} e^{2\pi i t/12} \, \tilde{x}(t + t_0), \tag{2.1}$$

where $\tilde{x}(t + t_0)$ are 12 monthly observations of one year with the series average removed. The offset t_0 denotes the middle of the month. Phase ϕ_x and amplitude A_x are derived for each year x, by computing the argument and modulus from Y_x:

$$\phi_x = \tan^{-1}\left(\mathrm{Im}\,(Y_x)/\mathrm{Re}\,(Y_x)\right) \tag{2.2}$$
$$A_x = |Y_x|\,. \tag{2.3}$$

Equation (2.1) is applied separately for each calendar year in the record to gain a series of annual phases and amplitudes. This method is based on the assumption that the annual cycle follows a sinusoidal function. Qian et al. (2011) note that such a-priori defined seasonal structures might underestimate nonlinear climatic variations and propose the usage of adaptive and temporally local methods such as empirical mode decomposition (EMD) and ensemble EMD. In an analysis of seasonal components of temperature records, Vecchio et al. (2010) showed that there has been a good agreement of the estimated phase shift of temperature using EMD and the estimate of Stine et al. (2009). Because of the simplicity of the method of Stine et al. (2009) and good agreement with more complex methods, this method has been used to estimate the annual phases of temperature and the runoff ratio in this analysis.

2.2.2 THE RUNOFF RATIO AND ITS ANNUAL PHASE

Generally, the runoff regime in Central Europe has distinct seasonal features, but it is not very balanced and can be quite different from an harmonic such as a cosine function. In contrast, the ratio of runoff and basin rainfall, the runoff ratio (RR) has a more distinct seasonal course. It represents the fraction of runoff observed at the basin outlet from the amount of precipitation for a certain period. The regime of the runoff ratio naturally reflects key processes of the basin water balance. Most important are the seasonal characteristics of precipitation, the actual basin evapotranspiration and the storage and release of water in soil or snow pack. Over the year these processes form a marked seasonal cycle.

Moreover, the runoff ratio is a direct measure of water availability of a basin, and thus an important quantity in water management. Lastly, the runoff ratio is a normalisation which allows to compare quite different hydrological regimes.

To illustrate the procedure of deriving a timing measure for the annual cycle of the runoff ratio, Fig. 2.1 (top) depicts monthly rainfall and runoff sums over a period of 5 years for an example basin. In the bottom graph, the resulting monthly runoff ratio (*Q/P*) is shown. Due to snow melt the ratio is larger than 1 in late winter time, also heavy rain events in summer may induce spikes in the record. Therefore, three-monthly running runoff ratios $\mathrm{RR}_3 = \mathrm{RR}_t = \frac{Q_{t-1}+Q_t+Q_{t+1}}{P_{t-1}+P_t+P_{t+1}}$ for each month t have been computed. These are generally smoother and better suitable for estimation of the annual cycle.

The estimated cosine functions are depicted in the bottom graph of Fig. 2.1. Geometrically the phase angle ϕ corresponds to the maximum of the cosine function. Further, ϕ is in fact a circular variable within the range of $[-\pi,\ \pi]$. For convenience the phase angles are transformed to represent the day of year (doy): $\mathrm{doy} = 365\,\phi/2\,\pi$.

Stine et al. (2009) note that monthly input data already gives accurate timing estimates. We verified this by using several temporal aggregation levels (daily, weekly, 14 days, monthly and 3 months windows). For temperature no essential change was found at all levels. For runoff ratio the annual phase estimates tend to a common annual phase estimate when increasing the aggregation window.

To compare the derived timing of runoff ratio, an independent metric, the half-flow dates (Q_{50}) have been chosen. Half-flow dates are for example used by Stewart et al. (2005) to

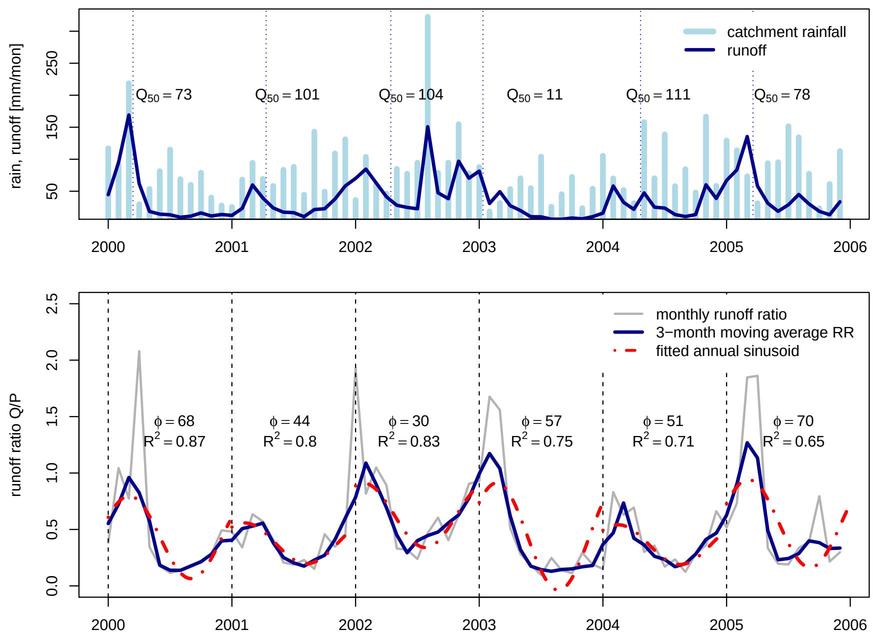

Figure 2.1: Top: monthly data of precipitation and runoff of a sample period from the station at Lichtenwalde. The vertical dotted lines depict the half-flow date (Q_{50}) of the respective year and its value is denoted as doy. Bottom: monthly runoff ratio, three-monthly moving runoff ratio and the resulting annual sinusoidal fits. The annual phases ϕ_{RR} are computed as doy and the annual explained variance (R^2) by the fitted sinusoids to the three-monthly running runoff ratios is given below.

analyse streamflow timing changes and their link to seasonal temperature changes in Northern America. To compute Q_{50}, the streamflow is accumulated over a period, e.g. one hydrological year, starting at 1 November. The half-flow date is defined as the day that 50 % of the annual sum have passed the river gauge. The derived half-flow dates of the illustrative example are shown in the upper plot of Fig. 2.1. Already for this short period, it can be seen, that both timing measures can have large differences in some years, whereby the phase estimate shows less fluctuations than Q_{50}.

2.2.3 DESCRIPTIVE CIRCULAR STATISTICS

To statistically analyse the timing estimates, one has to recall that the timing is a circular variable. Circular variables have certain properties, such as the arbitrary choice of origin and the coincidence of "beginning" and the "end". Therefore, linear statistics may be inappropriate and special treatment is needed to derive correct conclusions from the data (Jammalamadaka and Sengupta, 2001).

Considering a set of angular observations α_1, α_2, ..., α_n, the circular mean $\bar{\alpha}$ and variance σ_α^2 are computed as follows:

$$\bar{\alpha} = \arctan\left(\sum_{i=1}^{n} \sin(\alpha_i) \Big/ \sum_{i=1}^{n} \cos(\alpha_i)\right) \tag{2.4}$$

$$\sigma_{\alpha}^2 = 1 - \frac{1}{n}\sqrt{\left(\sum_{i=1}^{n} \sin(\alpha_i)\right)^2 + \left(\sum_{i=1}^{n} \cos(\alpha_i)\right)^2}. \tag{2.5}$$

To quantify the statistical relationship between the variability of the timing of temperature and the timing of runoff ratio, both being angular variables, circular correlation coefficients have been computed. The circular correlation ρ_{cc} between two circular vectors $\tilde{\alpha}$ and $\tilde{\beta}$ is defined as follows (Jammalamadaka and Sengupta, 2001):

$$\rho_{cc} = \frac{\sum_{i=1}^{n} \sin(\alpha_i - \bar{\alpha}) \sin(\beta_i - \bar{\beta})}{\sqrt{\sum_{i=1}^{n} \sin(\alpha_i - \bar{\alpha})^2 \sum_{i=1}^{n} \sin(\beta_i - \bar{\beta})^2}}, \tag{2.6}$$

with $\bar{\alpha}$ and $\bar{\beta}$ being the respective circular averages.

To compute the correlation ρ_{c-l} between a linear variable $\tilde{X}$ and a circular variable $\tilde{\alpha}$, Jammalamadaka and Sengupta (2001) suggest to transform the circular variable vector $\tilde{\alpha}$ into a linear variable vector $\tilde{w}$: $w_i = \cos(\alpha_i - \alpha_0)$. Whereby $\alpha_0 = \arctan(C_2/C_1)$, where C_1 and C_2 are the regression coefficients derived using using ordinary least squares from the expression:

$$\tilde{X} = M + C_1 \cos\tilde{\alpha} + C_2 \sin\tilde{\alpha}. \tag{2.7}$$

Then the transformed variable $\tilde{w}$ and the linear variable $\tilde{X}$ can be correlated using linear correlation measures, such as Pearson's product moment correlation coefficient.

Generally, significance testing of the derived correlation coefficients is based on the test statistic of the Pearson's correlation coefficient, which follows a t-distribution with $n-2$ degrees of freedom (R Development Core Team, 2011, function cor. test).

For a full treatment of circular statistics the reader is referred to Jammalamadaka and Sengupta (2001).

2.2.4 DETECTION OF NONSTATIONARITIES, TRENDS AND CHANGE POINTS

To analyse the variability of the estimated annual phase angles, it is necessary to check for nonstationarities, such as trends or structural changes of the mean or variance. As decadal changes of the mean may be expected from climatic variables, simple linear trends and significance testing may be not useful here. Another drawback is the sensitivity of the linear trend to the estimation period. Therefore, it is necessary to look for more complex trend patterns and analyse the low-frequency variability.

There are many simple graphical methods available for this purpose, with simple moving averages and cumulative sums of standardised variables (CUSUM) used in this paper. The CUSUM method is often used in econometric studies (Kleiber and Zeileis, 2008), where the focus is on the analysis of regression residuals and parameter stability over time. The method is also suitable to detect change points of a time series.

Zeileis and Hornik (2007) presented a general framework for the assessment of parameter instability, which is based on empirical estimating functions. These estimating functions, e.g. the mean of a series, must have the property that the sum of its residuals is equal to 0. To test for non-stationary behaviour, tests have been developed for these so-called empirical fluctuation processes (e.g. a CUSUM) based on Standard Brownian Motion or "Brownian bridge" processes, (Brown et al., 1975; Zeileis et al., 2002). Usually the resulting test statistic is a threshold level (dependent on the chosen significance level) that needs to be crossed by the CUSUM estimate to indicate a significant deviation from stationarity. However, Brown et al. (1975) state that these threshold levels "should be regarded as yardsticks", to emphasise that visualising the CUSUM lines may be more important than just applying the test.

To assess the stationarity of the mean of a circular variable the following steps are necessary to calculate the CUSUM. As the circular mean (Eq. 2.4) is the estimation function, we need to estimate the residuals of $\bar{\alpha}$. To fulfil the condition that the sum of the residuals must be 0, the angular deviations from $\bar{\alpha}$ are transformed into linear variables using the sine function:

$$y_i = \sin(\alpha_i - \bar{\alpha}), \text{ whereby } \sum_{i=1}^{n} \sin(\alpha_i - \bar{\alpha}) = 0. \quad (2.8)$$

Then the CUSUM C_i at time step i is computed as follows (Zeileis and Hornik, 2007):

$$C_i = \sum_{j=1}^{i} y_j \Big/ \left(\sigma_y \times \sqrt{n}\right) \qquad (i = 1, \ldots, n) \quad (2.9)$$

whereby σ_y is the estimated standard deviation of $\tilde{y}$ with length of the series n. In the case of linear data x_i the residuals of the series mean y_i are $y_i = x_i - \bar{x}$.

The estimated CUSUM C_i is a standardised, dimensionless quantity and is usually plotted over time. Some notes on how to interpret a CUSUM chart: a horizontal line fluctuating around 0, would imply a temporal stationary process. Segments of the CUSUM chart with upward slopes indicate above average conditions, while downward slopes indicate below average conditions. Peaks are an indication of the time of a change in the mean, which can be steady or abrupt. If the process under consideration changes positively, the residuals are negative and a negative CUSUM peak is shown, while under decreasing conditions a positive peak occurs. As the deviations from the estimation function (e.g. the long term average) are standardised, the magnitude and time of changes is comparable between different series. Last, a note to the sensitivity of the method to the choice of the interval. The method is sensitive to the starting and ending date, only with respect to the magnitude of the CUSUM, which is partly accounted for by the test statistic. However, the shape of the CUSUM line, i.e. the slopes and peaks remain at the same temporal positions, which allows for structural change testing without any sensitivity to the selected time interval.

2.3 DATA

The analysis comprises discharge series of 27 river gauges throughout Saxony and climatic data series such as precipitation, temperature and snow depth records. The station data used within this study have first been subject to a homogeneity test procedure, which has been used to detect possible structural changes in the series and to exclude anomalous series from the analysis. Further, climatic data such as rainfall, temperature and snow depths have been spatially interpolated to be able to compute river basin average values. A detailed description of these processing steps can be found in the appendix. All procedures are based on monthly data, as the method to filter the annual periodic components of the time series does not need higher temporal resolution data.

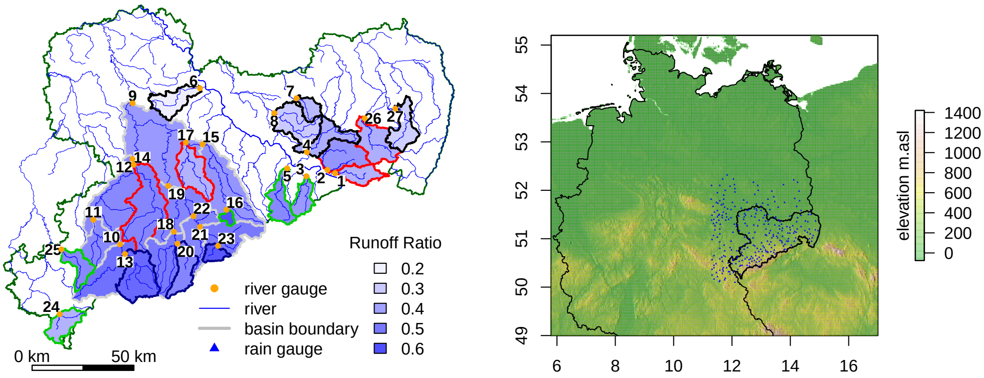

Figure 2.2: Left panel: map of the study area and long term average basin runoff ratios of the basins investigated. The bold numbers depict the id (cf. Table 2.1) of the river gauges (orange dots). The colour of the basin boundary refers to the 4 basin groups as used in Fig. 6. Grey boundaries indicate that the respective basin does not belong to any of these groups. Right panel: simple topographic map (geographical coordinates) of northern Germany with hillshading and terrain colours depicting elevation (Jarvis et al., 2008). The borders of Germany and Saxony are drawn as black lines and rain gauges used for interpolation are shown as blue triangles.

Due to extensive hydraulic engineering projects since the industrial revolution in the 19th century, a dense network of hydrologic gauging stations has been established in Saxony. We have chosen 27 river gauge stations, which almost fully cover the period 1930–2009. The stations cover large parts of Saxony, with catchment areas ranging between 37 and 5442 km^2. Most stations are within the Mulde River basin (15) or are tributaries of the Upper Elbe (6). Note that a range of basins are part of a common river network and are therefore physically and statistically not independent. However, 18 out of 27 are head water basins, which can be regarded as independent in terms of watershed properties. Detailed information may be found in Table 2.1 and the map in Fig. 2.2.

The discharge data have been converted to areal monthly runoff (mm $month^{-1}$) using the respective catchment area. Then the data have been subject to a homogeneity test procedure based on the catchments runoff ratio. Thereby, the Pettitt homogeneity test (Pettitt, 1979) has been performed on annual data as well as in a seasonal setting, where for each calendar month the test statistic has been computed separately, but only the largest test statistic of all months has been taken for significance testing. The significance levels have been determined by a Monte-Carlo simulation with normal $\mathcal{N}(0,\ 1)$ distributed random numbers. The details of 7 significant ($\alpha = 0.05$) inhomogeneous series are reported in Table 2.2. Note, that the reported year of the maximal test statistic does not necessarily identify the correct change point. However, in three cases dam constructions may be the probable cause of the inhomogeneity. For the other runoff series, no obvious reason has been found for the detected inhomogeneities. These are probably related to measurement errors (for example changes in the rating curve due to cross section changes) or the changes in catchment characteristics.

Table 2.1: River stations analysed over the period 1930–2009. The column elev denotes the mean basin elevation in meters above sea level, area denotes catchment area in km^2, RR denotes the long term average runoff ratio and miss gives the number of missing months.

id	station/river	major basin	upstream	elev	area	RR	miss
1	Kirnitzschtal/Kirnitzsch	Upper Elbe		381	154	0.36	0
2	Porschdorf/Lachsbach	Upper Elbe		378	267	0.43	0
3	Neundorf/Gottleuba	Upper Elbe		493	133	0.42	0
4	Elbersdorf/Wesenitz	Upper Elbe		317	227	0.37	0
5	Dohna/Müglitz	Upper Elbe		555	198	0.46	9
6	Merzdorf/Döllnitz	Upper Elbe		168	211	0.21	24
7	Koenigsbrueck/Pulsnitz	Schwarze Elster		274	92	0.34	26
8	Grossdittmannsdorf/Röder	Schwarze Elster		248	300	0.30	36
9	Golzern/Mulde	Mulde	10–23	481	5442	0.42	12
10	Niederschlema/Zwick. Mulde	Mulde	13	705	759	0.52	12
11	Zwickau/Zwick. Mulde	Mulde	10 13	631	1030	0.46	12
12	Wechselburg/Zwick. Mulde	Mulde	10 11 13 14	491	2107	0.46	0
13	Aue/Schwarzwasser	Mulde		742	362	0.54	0
14	Goeritzhain/Chemnitz	Mulde		410	532	0.47	0
15	Nossen/Freib. Mulde	Mulde	16	485	585	0.43	0
16	Wolfsgrund/Chemnitzbach	Mulde		629	37	0.60	2
17	Niederstriegis/Striegis	Mulde		374	283	0.36	13
18	Hopfgarten/Zschopau	Mulde	20	701	529	0.50	0
19	Lichtenwalde/Zschopau	Mulde	18 20–23	618	1575	0.47	0
20	Streckewalde/Preßnitz	Mulde		744	206	0.47	0
21	Pockau/Flöha	Mulde	23	688	385	0.50	0
22	Borstendorf/Flöha	Mulde	21 23	663	644	0.47	0
23	Rothenthal/Natzschung	Mulde		770	75	0.58	0
24	Adorf/Weiße Elster	Weiße Elster		599	171	0.36	35
25	Mylau/Göltzsch	Weiße Elster		518	155	0.46	12
26	Bautzen/Spree	Spree		357	276	0.37	24
27	Groeditz/Löb. Wasser	Spree		284	195	0.29	12

Table 2.2: Results of homogeneity tests of runoff ratio and information of larger dam constructions with the respective volume of the reservoirs given in hectometres (hm^3). The column Inhomogeneity reports the year and the month the maximal Pettitt test statistic and their respective significance levels.

Station	Inhomogeneity	Additional information
Streckewalde	annual: 1952***, seasonal: Jun 1970**	dam construction 1973–1976, 55 hm^3[a]
Goeritzhain	annual: 1953***, seasonal: Apr 1959*	
Niederstriegis	annual: 1962***, seasonal: Apr 1957**	
Neundorf	annual: 1967**, seasonal: Mar 1976*	dam construction 1976, 14 hm^3[b]
Groeditz	annual: 1948*, seasonal: Apr 1948*	
Pockau	annual: 1980*,	dam construction 1967, 15 hm^3[c]
Rothenthal	seasonal: Feb 1981**	

$^{*}\alpha = 0.05$, $^{**}\alpha = 0.01$, $^{***}\alpha = 0.001$.

[a] http://de.wikipedia.org/wiki/Talsperre_Pressnitz, [b] http://en.wikipedia.org/wiki/Gottleuba_Dam, [c] http://de.wikipedia.org/wiki/Talsperre_Rauschenbach

2.4 RESULTS AND DISCUSSION

2.4.1 ESTIMATION AND VARIABILITY OF THE TIMING OF THE RUNOFF RATIO

To gain some insight in the general spatial behaviour of the runoff ratio of the selected basins, a map of the long term average runoff ratio is presented in Fig. 2.2. There is generally a higher runoff ratio in southern mountainous basins, having a runoff ratio up to 0.6, which is mainly due to higher precipitation (up to 1030 mm annually). The basins in the hilly North have lower runoff ratios ranging between 0.2 and 0.4 and are characterised by lower precipitation (down to 630 mm), higher evapotranspiration and in contrast to the higher basins, larger bodies of groundwater due to unconsolidated rock.

As an example, a time series of the runoff ratio is shown for the gauge at Lichtenwalde in Fig. 2.3. The three-monthly running runoff ratio shows a distinct seasonal pattern, while the 2-year running runoff ratio exhibits some low-frequency variability. Looking at the spectra of the runoff ratio series, two distinct peaks are generally found, one at an annual and the other at the half-year frequency (not shown).

As a next step, annual phases and amplitudes have been computed for each basin by applying the method of Stine et al. (2009). The accuracy of the timing estimate is evaluated by comparing the estimated cosine fits with the original data. According to R^2 between 22 and 34 % of the variability of the monthly runoff ratio series are explained by this fit. However, as we are interested in the smooth seasonal signal, we used three-monthly running runoff ratios RR_3 to filter the annual cycle. Between 71 and 84 % of the variability of RR_3 is explained by the fitted cosines. Another positive effect is that the standard deviation of the annual phases estimated for monthly runoff ratios decreased from 18.5–27.6 to 12.7–20.1 days, when using three-monthly moving runoff ratios. This is mainly due to less extreme years, while keeping the overall phase average (54.4 to 55.3).

With regard to independence of the annual phase estimates, circular autocorrelation functions have been computed. We find that there are no significant ($\alpha = 0.05$) correlations in any series at lags from 1 to 10 years. Further the empirical distributions have been plotted vs. the "Von Mises" distribution (Jammalamadaka and Sengupta, 2001; Lund and Agostinelli, 2010), which showed no substantial deviations from the 1:1 line. The "Von Mises" distribution may be regarded as equivalent to the normal distribution for circular data. Thus, the timing estimates do not violate distribution and independence assumptions for trend and correlation assessment.

Next, average characteristics of the timing of runoff ratio over Saxony are analysed. We already discussed that the runoff ratio shows a north to south gradient corresponding to increasing basin elevation (Fig. 2.2), which is confirmed by the left panel of Fig. 2.4, depicting the relationship between the runoff ratio and basin elevation. We find that the annual phase is even more dependent on the basin elevation, cf. the right panel of Fig. 2.4, with a strong linear relation of 5.5 ± 0.3 days per 100 m elevation change. Naturally, lower basins appear to have an earlier timing than higher basins, which is due to earlier snow melt in winter/spring.

To generalize the results and because of the strong link to altitude, we chose to group the basins according to their average elevation, which resulted in 4 groups. For each group the phase average has been computed for each year using circular means (Eq. 2.4). Further, only non-connected basins are used, to achieve a set of independent basins. The respective height intervals and corresponding basins are presented in Table 2.3. Descriptive statistics such as circular average and standard deviations for each elevation group are given in column $\bar{\phi}_{RR}$.

As independent comparison to the annual phases estimated from RR_3, the basin average

Table 2.3: Average statistics of the river gauge stations, grouped according to basin average elevation without connected basins. The columns denote in order of appearance: the respective elevation interval, group member basin id, the phase average $\bar{\phi}_{RR}$ as calender day with respective circular standard deviation in days, the average half-flow date $\bar{Q}_{50}$, circular correlation coefficients ρ_{cc} between ϕ_{RR} and ϕ_T, the linear regression coefficient T_{coef} and its standard deviation and the circular-linear correlation ρ_{snow} between snow depths in March and ϕ_{RR}.

elevation	id	$\bar{\phi}_{RR}$	$\bar{Q}_{50}$	ρ_{cc}	T_{coef}	ρ_{snow}
160–320	4 6 7 8 27	11 Feb $\pm$ 14	28 Mar $\pm$ 19	0.23	0.93 $\pm$ 0.45	0.53
360–420	1 2 14 17 26	14 Feb $\pm$ 14	29 Mar $\pm$ 19	0.31	1.27 $\pm$ 0.45	0.64
500–620	3 5 16 24 25	25 Feb $\pm$ 13	1 Apr $\pm$ 20	0.55	2.01 $\pm$ 0.34	0.50
740–780	13 20 23	11 Mar $\pm$ 16	10 Apr $\pm$ 18	0.59	2.64 $\pm$ 0.41	0.72

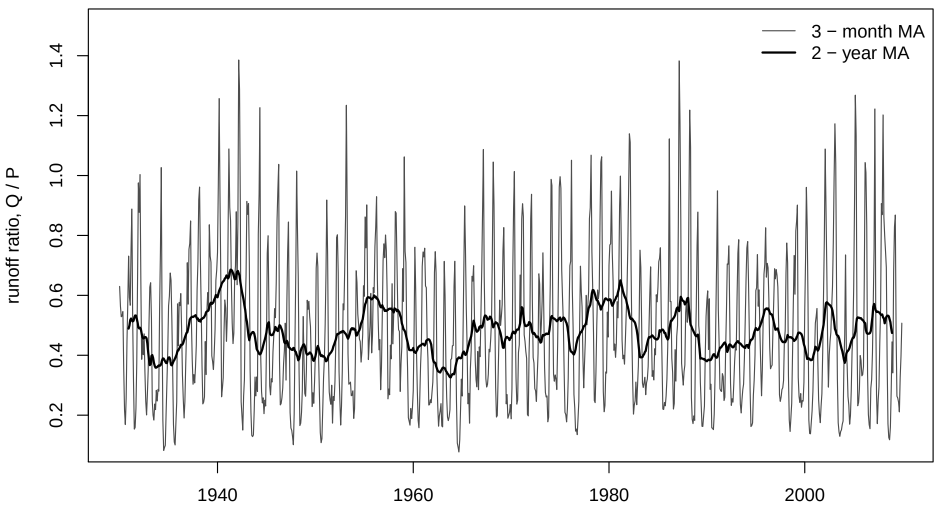

Figure 2.3: Time series of smoothed monthly runoff ratio at Lichtenwalde, Zschopau.

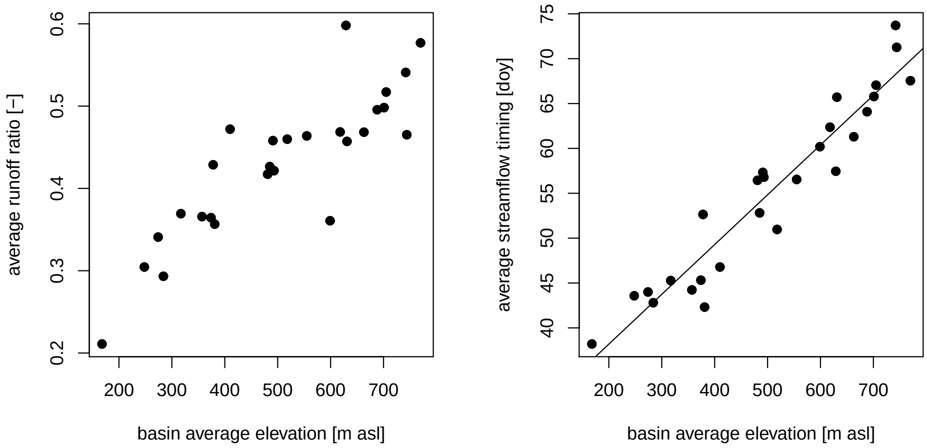

Figure 2.4: Height dependence of long term average runoff ratio (left panel) and dependence of the average streamflow timing (right panel).

half-flow date and its standard deviation are shown in column $\bar{Q}_{50}$ of Table 2.3. Generally, the half flow dates appear later than the annual phase estimates, with about 44 days in the lowest basin group and about 30 days in highest elevated basins. So half-flow dates do not show such a clear difference between high and low basins as annual phases do. Further, as lower basins do not have such a distinct seasonal pattern as higher basins, half flow dates are less able to discern the correct timing (Déry et al., 2009).

Comparing the different measures using the timing average for all series and years, the phase estimate for the runoff ratio is smallest ($\phi_{RR} = 55$), while the half-flow dates are largest ($Q_{50} = 93$). However, if we compute the phase directly from monthly streamflow we yield $\phi_Q = 70$, which is in between. So a part of the differences found between Q_{50} and ϕ_{RR} are due to the fact that half-flow dates are based solely on streamflow, while the phase estimate of the runoff ratio is normalised by precipitation. The other part of the differences is due to the different timing estimation techniques. There are also differences in standard deviation σ between both measures. While Q_{50} shows a σ of about 19 days, the timing measure using runoff ratios has a σ of about 14 days. The larger variability in Q_{50} can probably be attributed to larger uncertainties in its estimation, e.g. owing to single events (Déry et al., 2009).

2.4.2 TEMPORAL VARIABILITY OF THE TIMING

As the timing estimate has been computed for each calender year, it is now possible to investigate the high and low-frequency temporal variability.

Figure 2.5 presents annual phase estimates (converted into doy) of the runoff ratio of the lowest and the highest basin group, respectively. In general, there is a natural difference in timing between lower basins and more mountainous basins. However, from Fig. 2.5 it is apparent that these differences changed over time with much larger differences in the period 1950 to

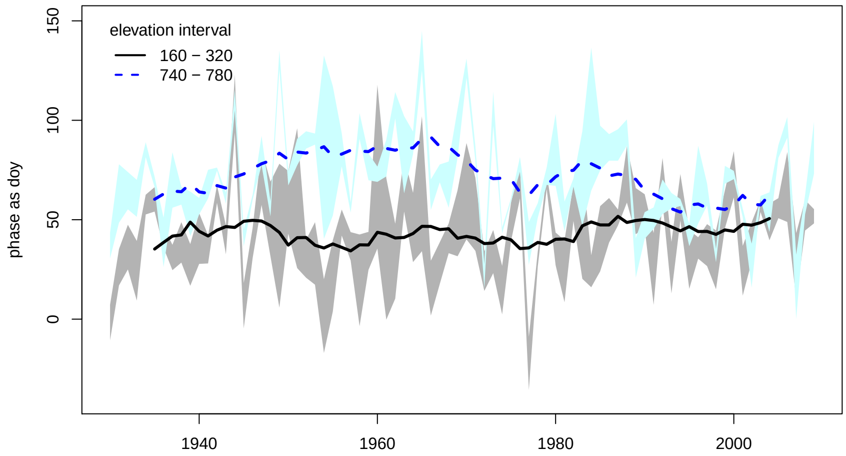

Figure 2.5: Time series of the annual phase of runoff ratio for two groups of stations at high and low elevations, respectively. The shaded area shows the within group range and the bold lines depict the 11-year moving average of the group average annual phase.

1980. Also the year-to-year variability of ϕ_{RR} is larger as well. In contrast to the low basins, there is a trend towards earlier timing in the higher basins since the late 1960s, decreasing the differences between low and high basins. We further note that the difference observed in the last two decades is now smaller than it has been observed before 1950. All other basins at medium elevations show a behaviour somewhere in between the both groups.

For CE Stahl et al. (2010) and for the nearby Czech Republic Fiala (2008) found eye-catching trends of increasing streamflow in winter, especially in March, while decreasing discharge is observed from April to June. These trends imply a change in the phase of the cycle towards earlier timing of streamflow. So, considering the same period (1962–2004) as Stahl et al. (2010), we can confirm a decreasing trend in the phase of runoff ratio in mountainous basins, see Fig. 2.5.

In the following, the decadal variability, trend patterns and change points of the timing of the runoff ratio will be analysed. Since we did not expect a linear trend prevailing over this long period from 1930–2009, we chose to analyse the data using CUSUM graphs, which display the low-frequency variability and structural deviations from the long term average over time. Figure 2.6 shows CUSUM lines based on the group average timing of the runoff ratio. The graph shows that with the beginning of the 1950s different trend directions in the particular basin groups have evolved. Basins above 500 m show upslope sections until 1971 and again from 1980 to 1988, exhibiting above average behaviour. This pattern is modulated by elevation and reveals that the low-frequency changes in timing are largest in the highest basins, where the CUSUM line hits the $\alpha = 0.05$ significance level of a stationary process in the year 1988. Another peak, although lower, is found in 1971. The peaks mark changes in trend directions

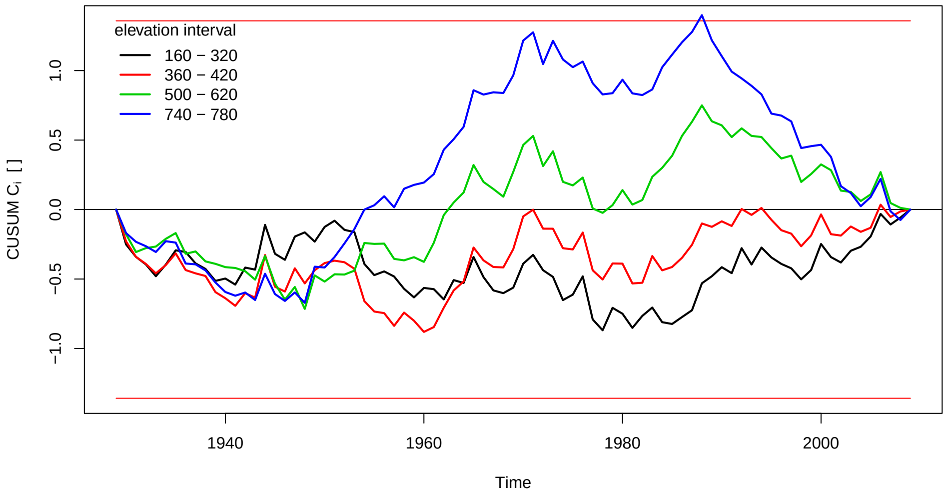

Figure 2.6: CUSUM Analysis of the annual phase of runoff ratio. Basins have been grouped according to their altitude and then a group average phase has been computed and used for the CUSUM analysis. The significance levels ($\alpha = 0.05$) for a stationary process are denoted as horizontal lines at the top and bottom of the graph.

and because they are positive, they reveal decreasing conditions. Both peaks are found in all CUSUM graphs, which is an indication that the low-frequency variability is mainly driven by a larger scale process.

Considering the year 1988 as a probable change point, the average shift in timing before (1950–1988) and after (1988–2009) is assessed. While the lowest basins show a delay of 7 days, there is no shift in the second elevation group. The basins above 500 m show negative shifts, i.e. an earlier timing of 10 and 22 days, respectively. Causes of the negative trend patterns will be discussed in the next subsections.

As there is evidence of non-stationary signals in the timing of runoff ratio, especially for mountainous basins (cf. Fig. 2.6), the usage of a fixed set of seasonal parameters, e.g. the monthly mean, results into a systematic bias and thus unnecessarily inflates the variance (Thomson, 1995). A standard design study based on data of the last 50 years will be biased due to the observed dynamics of the annual cycle of runoff. So, e.g. hydraulic design studies should acknowledge these structural changes of the annual cycle by using time variant, e.g. dynamic seasonal time series models.

2.4.3 DOES TEMPERATURE EXPLAIN TRENDS IN SEASONALITY OF RUNOFF RATIO?

Having analysed the phase of the runoff ratio, it is interesting to check for direct links to climatic variables, especially temperature. In Fig. 2.7 time series of the annual average temperatures for the lowest and highest basin groups are shown. The depicted range is drawn from single basin temperature series. The apparent temperature difference between the basin groups of about 2.5 K reflects the typical elevation gradient of temperature.

In the temperature series of basins at low elevations, a positive linear trend is detected in the order of 0.01 K per year (Kendall test p-value 0.02). Here, we applied the non-parametric trend test procedure for autocorrelated data suggested by Yue et al. (2002). In the higher basins the linear trend is weaker with 0.003 K per year (Kendall test p-value 0.22) and not significant. Instead, there is a period of lower temperatures from 1960 to 1990. Since then an increase is found.

The timing of the annual cycle of temperature for the basins investigated is also shown in Fig. 2.7. The differences between the basins are small (1.5 days between lowest and highest basins) compared to the standard deviation (on average 3.5 days). The long term variability of the annual phase of temperature is relatively constant. However, since the end of the 1980s, there is a decline in the average of about 4 days, concluding with the most extreme years (2006 very late, and 2007 very early) observed.

To visualise the temperature timing influence on the seasonality of runoff ratio, we classified the 80 years of data into early years, having annual phases below the first quartile (before doy 198) and late years, having phases in the last quartile (after doy 204). Then we used this classification to bin the series of runoff ratio for every month over the year. The resulting boxplots in Fig. 2.8 depict the seasonal runoff ratio distribution of each group over the year. As can be seen for the river gauge Koenigsbrueck, larger runoff ratios and larger variability from February to April are observed in late years, than in early years. At the river gauge Lichtenwalde, the differences between early and late years are even more distinct, with significant differences for the months April till August, with late years having an higher runoff ratio than earlier ones. The opposite is true for the months October till December. The average monthly temperatures superimposed in Fig. 2.8, reflect the actual differences between early and late years on temperature, which are larger during the first half of the year.

To quantify the link between these angular variables, circular correlation coefficients have been computed from the annual phase of the basin runoff ratio and the annual phase of the basin average temperature. The results are detailed for each basin elevation group in Table 2.3, column ρ_{cc}. The correlation coefficients tend to increase with elevation. A linear regression allows to assess the average effect of a change in the phase of temperature on the timing of the runoff ratio. The slope coefficient and its standard deviation of the regression line for each basin group is reported in Table 2.3, column T_{coef}. Note that in this case the timing has been treated as linear variable. The coefficient is also plotted against average basin elevation for all basins in Fig. 2.9. Again, there is a distinct height dependence, which is increasing with 0.36 ± 0.01 per 100 m basin height. For mountainous basins, we find a coefficient of about three in magnitude, which means that a decrease of the phase of temperature of 5 days, amounts to a decrease of 15 days in the timing of the runoff ratio.

The increased sensitivity of hydrological regimes to temperature at higher altitudes has been often cited in literature. E.g. Barnett et al. (2005) state that rising temperatures possibly lead to earlier timing of the hydrologic regime. We find indeed that the annual basin temperature is correlated with the annual phase of runoff ratio. In fact, there is a linear-circular correlation of -0.37 in the highest basins, which is linearly increasing with decreasing basin elevation to 0.13 in the lowest basins. However, this correlation is only half in magnitude compared to the phase of temperature. Considering the whole annual cycle, the timing relationship between temperature and runoff ratio is stronger and more relevant than the one with annual

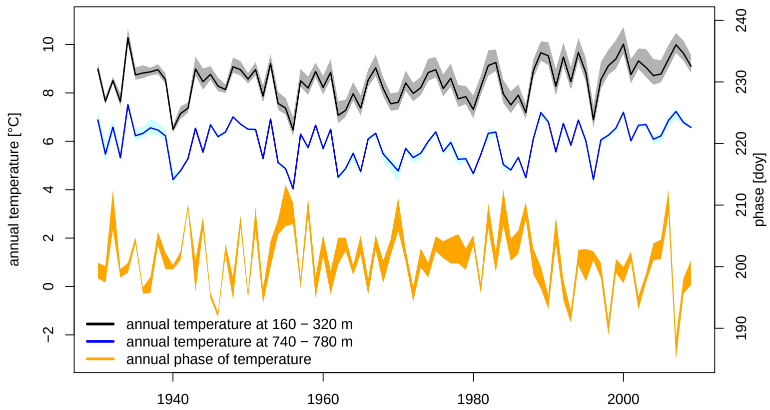

Figure 2.7: Range and mean of annual average temperatures of basins in the lowest and highest basin group. Orange shading: the range of annual phases of basin temperature ϕ_T of all 27 basins (units on right axis).

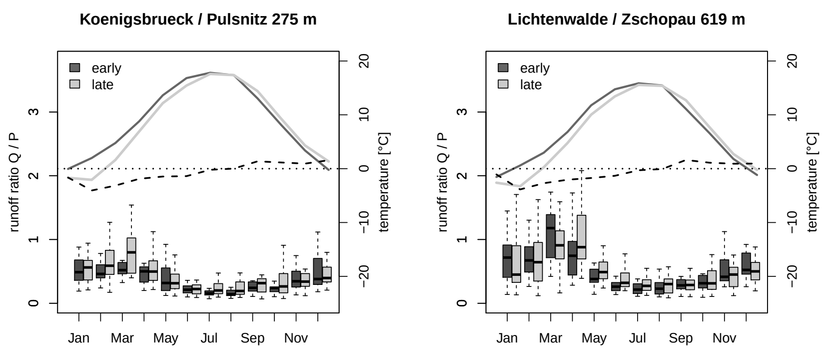

Figure 2.8: Box-whisker plots of monthly values of runoff ratio. The data are grouped according to early years with the annual phase of temperature below the 1st quartile and late years beyond the 3rd quartile. The bold grey and black lines denote the average monthly temperature for late and early years, with the corresponding axis on the right. The difference in temperature is shown as dashed line. The whiskers show the largest/lowest values within 1.5 – times of the interquartile range (IQR). Values outside 1.5 × IQR, if any, are denoted as outliers and are not shown for display reasons. There are quite a few outliers, but these are equally distributed among the groups.

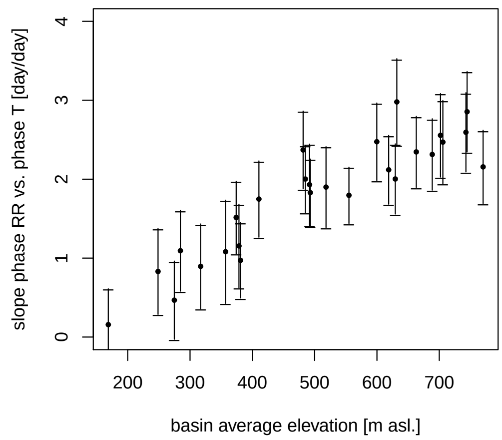

Figure 2.9: Height dependence of the regression slope coefficient (± standard deviation) between annual phases of streamflow and temperature.

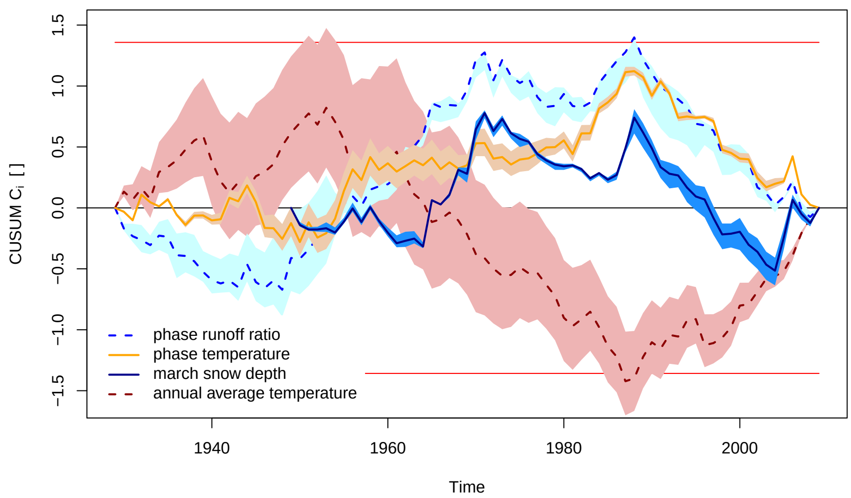

Figure 2.10: CUSUM Analysis of the 3 highest head water basins situated in the Ore Mountains. The bold lines denote the particular group average series, while the shaded areas depict the range of the single CUSUM lines. The significance levels ($\alpha = 0.05$) for a stationary process are denoted as horizontal lines at the top and bottom of the graph.

temperatures.

2.4.4 TREND ANALYSIS IN SNOW DOMINATED BASINS

In the previous section, we found that the correlation to the annual phase of temperature and its effect on the timing of the runoff ratio increases with elevation. It is clear, that temperature alone does not influence the runoff ratio, temperature instead acts as a trigger for snow precipitation and snow melt. Therefore, a detailed look at snow depth observations is interesting. Since 1950, a network of snow depth observations has been established in Saxony, which is dense enough to compute basin averages.

Winter average snow depths and snow cover are poorly correlated to the annual phase of runoff. For the basin Lichtenwalde and station data at Fichtelberg, ρ_{c-l} is not very high and also not significantly different from 0 at the $\alpha = 0.05$ level ($\rho_{c-l} = 0.2$ for winter average snow depths and $\rho_{c-l} = 0.29$ for snow cover duration). However, the average snow depth in March appears to have a significant correlation ($\rho_{c-l} = 0.55$). Therefore, for each basin the March average snow depth has been computed. Regarding the basin groups, we found positive and significant correlations, that are largest in the highest basins, see also Table 2.3, column ρ_{snow}.

Having identified the links of temperature, snow depth and runoff ratio in snow melt influenced basins, we investigated whether these variables might explain the trend patterns found in the phase of runoff ratio. As the low-frequency changes have been largest in the highest basins, we employed this group to depict the low-frequency changes of the phase of temperature, the annual mean temperature and snow depth observation for March. For each series the CUSUM graph can be seen in Fig. 2.10. While the group average series is used for testing, the shaded coloured bands depict the range of CUSUM lines of each basin group and thus reflect the general variability between the basins. On first sight there is a wide band for the annual mean temperature graph. This wide range can be mainly attributed to the low temperature station density before 1960. The band gets thinner, when the computation of the CUSUM line starts after 1960. This uncertainty, however, does not influence the general behaviour, with the negative peak in the year 1988, revealing a significant change in the mean towards higher temperatures. Also the phase of temperature shows a peak in 1988, but in opposite direction, indicating decreasing conditions. Besides, the low-frequency behaviour of the phase of temperature is not influenced by the change in station density, showing the robustness of this measure.

The CUSUM graph of the annual phase of runoff ratio (Fig. 2.10) also shows this peak in 1988, reaching the significance level $\alpha = 0.05$. Moreover, there is another peak apparent in 1971, where there is no indication from both temperature related series. But, we found a striking similarity of the CUSUM graph for March snow depths, which displays both peaks and even though the significance levels are not reached, they provide evidence, that late winter snow cover may also explain the low-frequency variability of ϕ_{RR}.

These statistical links underline the strong influence of late winter snow cover on the timing of ϕ_{RR} and subsequently on the annual hydrological regime. This influence naturally increases with elevation (cf. column ρ_{snow} of Table 2.3). Under the assumption that there is no limitation of winter precipitation, snow cover is to a large part controlled by temperatures below or above 0 °C. This argument might be an explanation that with increasing basin elevation, also increasing correlation and linear slopes (cf. Fig. 2.9) have been found between the timing of temperature and the timing of the runoff ratio (cf. Table 2.3). Still, late winter snow cover is a better predictor than the phase of temperature.

The coincidence of peaks in the CUSUM graphs in Fig. 2.10 indicates that the respective elements undergo structural changes at the same time. Especially the apparent 1988 change point in all series investigated may be related to distinct changes towards less air pollution with aerosols over CE since 1980 (Philipona et al., 2009). So probably, the increasing incoming short wave radiation resulted in increased temperatures, earlier snow melt and, eventually, also in the advance of the timing of temperature.

Air pollution also impacted forest vegetation, with subsequent tree-die off since the 1960s and with major clear cuts in the 1980s (Šrámek et al., 2008) at the mountain ridge in southern Saxony. Thus especially the headwaters of some rivers analysed here have been affected. Such dramatic changes in vegetation cover may have also influenced hydrologic processes and subsequently the timing of runoff ratio. However, quantifying such effects is out of the scope of this study, and remains open for further research.

2.4.5 UNCERTAINTY AND SIGNIFICANCE OF THE RESULTS

For the interpretation of the results it is necessary to list the sources of uncertainty and to examine their relevance.

First of all, there may be measurement errors or inhomogeneities in the observed runoff and rainfall series. When we assume that these errors lead to an abrupt but constant change of the mean at a given location, the cyclic behaviour and thus the phase is unlikely to be affected systematically.

In 7 basins inhomogeneities in the runoff ratio series have been detected. Without detailed information, it is impossible to correct for such changes. Therefore, these records have been kept in the dataset without a correction. We performed some cross checking by subsequently removing the suspect series from the computation of the group averages. The resulting differences to the original group average are of comparable magnitude than the standard deviation of the averaged series, but small with respect to the assessed correlations and long term shifts.

Another source of uncertainty is the estimation of basin precipitation. Apart from the spatial interpolation error, which is assumed to average out, we had to face the problem of changes of the observation network over time. To check for effects of this inhomogeneity, three different sets of input stations have been prepared (cf. the Appendix). When comparing the resulting annual phases of these different precipitation input sets, only marginal differences for the timing estimates have been found. Then there is some uncertainty in the estimation of the timing of the annual cycle using the approximation of a harmonic function to the data. We quantified this uncertainty by calculating the explained variance of the original series. This is not possible for traditional timing measures, such as half-flow dates. Further we showed that the year-to-year variability could be reduced by smoothing the data before applying the annual filter.

Finally, the overall uncertainty in the phase estimates and their low-frequency variability was assessed by grouping the data according to basin elevation. The range within a group is a measure of the accuracy of our estimates. As these ranges were generally smaller than the temporal variability, we can conclude that the averaging method was robust. Moreover, the main features are repeated within this large set of river basins distributed over several elevation levels. This last argument underlines that the change points found in the phase of runoff ratio are not random or catchment specific, but a result of changing climate conditions.

2.5 CONCLUSIONS

The timing of annual hydrological regimes of river basins throughout Saxony/Germany has been evaluated over the period from 1930 to 2009. We introduced a timing measure for hydrological time series, which is based on a harmonic filter of monthly data. The measure is applicable to all hydrological regimes found throughout Saxony and it is expected to work elsewhere, where a distinct annual cycle in a time series is apparent. Comparing with traditional streamflow timing measures such as half-flow dates, which can be easily biased by single events, we showed that the resulting standard deviation of the harmonic measure is generally lower and thus less influenced by single events.

A climatology of the timing of the dimensionless runoff ratio (RR) was established, covering 27 river basin at different elevation levels. Basin elevation was found to be the most important catchment characteristic, controlling (i) average timing, (ii) the magnitude of observed long-term shifts in timing and (iii) the apparent sensitivity to the timing of temperature. All mentioned characteristics increase with elevation.

Analysing the temporal variability, we observed a shift of the seasonal cycle towards occurring earlier in the year in basins being on average above 500 m, with the largest changes in the highest basins. This long-term shift in timing of runoff ratio represents a trend towards earlier timing of about 10 to 22 days in the last two decades, relative to the prevailing conditions between 1950 and 1988.

The interannual variability of runoff ratio timing records is in the same order as the apparent long term shifts, but independent from elevation. There is, however, a remarkable coherence of the year-to-year changes across all basins analysed. Also, the long-term change patterns revealed by a CUSUM analysis of the standardised anomalies showed a similarity in slopes and peaks between elevation groups. Presumably, the observed changes are driven by larger scale physical processes, which have similar effects at the annual, as well as at the decadal time scale.

As expected, a large fraction of the observed variability may be explained by the low and high frequency variability of temperature records. Indeed, the annual timing of temperature, which can be estimated with high confidence, showed significant positive correlations with the timing of the runoff ratio. Again, the correlation as well as the linear regression coefficient showed to be dependent on basin elevation. Moreover, the timing of the temperature cycle has more influence on the timing of the runoff ratio than the magnitude of annual average temperatures.

However, the apparent low-frequency variability of RR could not be explained by temperature observations alone. The main cause of the observed high and low frequency variability in higher elevated basins is the variability of late winter snow cover. It explained a larger fraction of the variability than the timing of temperature and matched the low frequent departures from the average of the timing of runoff ratio quite well.

The climatic changes observed by the temperature regime are most likely the major cause of the observed changes in hydrological variables. There is evidence of a structural change of the average behaviour of several observation variables in the year 1988. The CUSUM related stationarity test revealed (i) a significant shift in the timing of runoff ratio in high basins, (ii) a marked but not significant change in late winter snow depths, accompanied by (iii) a significant increase of annual temperature of about 1 K and (iv) a marked but not significant advance of the timing of temperature of 4 days.

We believe that this chain of changes has been triggered by the drastic changes in industrial and domestic air pollution, because the dimming and brightening of the atmosphere over Saxony resulted in remarkable changes in solar insolation. Further, this accelerated and distinct change in the timing of both, temperature and runoff ratio indicates that impacts of climate change on the water cycle are stronger in mountainous areas.

If the trends in the phase and average of temperature persist, a range of potential problems for water resources management will evolve. The most critical problem is that the delay between natural water supply and demand will increase and subsequently a larger artificial storage volume may be needed to maintain the same security level of supply. Next, the shift in both, mean and variability of monthly streamflow will alter traditional assumptions used for predicting seasonal water availability. This underlines the importance for maintaining and improving the existing observational network.

2.A PREPARATION OF BASIN INPUT DATA

2.A.1 PRECIPITATION

The geographical domain (11.5°–16° E, 50°–52° N) has been chosen for the spatial interpolation and station data selection.

The station network density has changed dramatically throughout time. Currently there is one station available in the database since 1858, 12 stations since 1891 rising up to 111 in the 1930s. Due to World War II only 20 stations were available in 1945. From the 1950s, the network has improved from 374 in 1951 to a maximum of 873 in 1990. Since 2000, the network density decreased to 354 in 2008.

To check for influences of the changing network, three data sets have been prepared. One set only with stations covering the full period without longer missing periods, another set which consists of all observations available at a time step and another set which has been used in the analysis. This last set is a compromise between the other two sets, meeting the requirement that the respective series covers at least 40 years, i.e. from 1950–1990. This set contains 368 stations.

Based on these stations a homogeneity test procedure has been conducted. Depending on available meta-data a part of these stations has been tested for known breakpoints using the Kruskal-Wallis rank sum test for changes in the location and the Bartlett test for changes in the variance. If all these tests reject the hypothesis of no change at the $\alpha = 0.05$ level, then the series has been flagged as suspect. Next, an iterative homogeneity test procedure has been done using a weighted series of about 5 reference stations. Reference stations are selected according to 4 criteria: (i) not inhomogeneous from previous test, (ii) best correlation of the differenced series (Peterson and Easterling, 1994), (iii) cover most of the record of the candidate station and (iv) are close to the candidate. Then the Alexandersson homogeneity test and the Pettitt test have been applied. If both tests reject the hypothesis of stationarity at the $\alpha = 0.01$ level, then the series has been flagged as suspect. Finally, a set of 299 precipitation series have been left for spatial interpolation, i.e. without any suspect series. The stations can be found in the right panel of Fig. 2.2. There are 83 stations during the 1930s, about 290 from 1950–1990 with 170 in the last decade.

Based on the station dataset a spatial interpolation for each month has been computed. First, a linear height relationship using a robust median based regression (Theil, 1950) has been established. Then the residuals have been interpolated onto an aggregated SRTM grid

(Jarvis et al., 2008) of 1500 m raster size using an automatic Ordinary Kriging (OK) procedure (Hiemstra et al., 2009). Monthly basin average precipitation is then computed by the average of the respective grid cells. The method of height regression and OK of the residuals has been chosen, as this method showed to have the lowest root-mean-square errors (RMSE) among other methods in a cross-validation based on monthly station data sets.

2.A.2 TEMPERATURE AND SNOW DEPTH DATA

The network of climate stations in the domain has also changed during time. Since 1930, 9 long temperature series have been available, this increased to 47 in 1961 and again reduced to 38 in 2008. A few snow depth observations are available from climate stations. Additionally, a dense network of snow depths has been established in the region since 1950. On average, 163 series are available. For both elements, the basin averages have been computed using the methods already described for precipitation in Sect. 2.A.1.

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3 EVALUATION OF WATER-ENERGY BALANCE FRAMEWORKS TO PREDICT THE SENSITIVITY OF STREAMFLOW TO CLIMATE CHANGE

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ABSTRACT

Long term average change in streamflow is a major concern in hydrology and water resources management. Some simple analytical methods exist for the assessment of the sensitivity of streamflow to climatic variations. These are based on the Budyko hypothesis, which assumes that long term average streamflow can be predicted by climate conditions, namely by annual average precipitation and evaporative demand. Recently, Tomer and Schilling (2009) presented an ecohydrological concept to distinguish between effects of climate change and basin characteristics change on streamflow. We relate the concept to a coupled consideration of the water and energy balance. We show that the concept is equivalent to the assumption that the sum of the ratio of annual actual evapotranspiration to precipitation and the ratio of actual to potential evapotranspiration is constant, even when climate conditions are changing.

Here, we use this assumption to derive analytical solutions to the problem of streamflow sensitivity to climate. We show how, according to this assumption, climate sensitivity would be influenced by different climatic conditions and the actual hydrological response of a basin. Finally, the properties and implications of the method are compared with established Budyko sensitivity methods and illustrated by three case studies. It appears that the largest differences between both approaches occur under limiting conditions. Specifically, the sensitivity framework based on the ecohydrological concept does not adhere to the water and energy limits, while the Budyko approach accounts for limiting conditions by increasing the sensitivity of streamflow to a catchment parameter encoding basin characteristics. Our findings do not support any application of the ecohydrological concept under conditions close to the water or energy limits, instead we suggest a correction based on the Budyko framework.

3.1 INTRODUCTION

In this paper we consider the question how variations in climate affect the hydrological response of river basins. Thus, we aim to assess climate sensitivity of basin streamflow Q and evapotranspiration E_T, (Dooge, 1992; Arora, 2002; Yang and Yang, 2011; Roderick and Farquhar, 2011). To do so, we need to consider the concurrent climate itself, because naturally the supply of water and energy is the main controlling factor of evapotranspiration (Budyko, 1974; Zhang et al., 2004; Teuling et al., 2009). Basin characteristics are also of high relevance: two basins with similar climate may have quite different hydrological responses (Yang et al., 2008). Spatio-temporal patterns of precipitation, soils, topography, vegetation and not least human activities have considerable impacts (Arnell, 2002; Milly, 1994; Gerrits et al., 2009; Zhang et al., 2001; Donohue et al., 2007).

Usually, one is tempted to represent such basin characteristics by conceptual or physically based hydrological models. However, the uncertainties arising from model structure and calibration may lead to biased and parameter dependent climate sensitivity estimates (Nash and Gleick, 1991; Sankarasubramanian et al., 2001; Zheng et al., 2009).

A remarkable paper of Tomer and Schilling (2009) introduced a conceptual model to distinguish climate change effects from land-use change effects on streamflow. They utilize two non-dimensional ecohydrologic state variables representing water and energy balance components, which describe the hydro-climatic state of a basin and carry information of how water and energy fluxes are partitioned at the catchment scale. The central hypothesis of Tomer and Schilling (2009) is that from the observed shift of these states, the type of change can be deduced. Their theory is based on experiments with different agricultural conservation treatments of four small field size experimental watersheds (30–61 ha). They observed that watersheds with different soil conservation treatments also showed different evapotranspiration ratios. Further, the shift within this hydro-climatic state space due to conservation treatments was perpendicular to the shift which was observed over time. They attributed this temporal shift to a climatic change characterised by increased annual precipitation.

The conceptual model proposed by Tomer and Schilling (2009) has great scientific appeal, because of its potential to easily separate climatic from land use effects on the water balance. Here, we aim to explore this potential of the framework to address the following research questions:

1. Can this concept also be used to predict streamflow/evapotranspiration change based on

a climate change signal?

2. What are the implications of such a model, given the range of possible hydro-climatic states and changes therein?
3. How does it compare to existing climate sensitivity approaches?

This paper is structured as follows. In the methodological section we embed the conceptual model of Tomer and Schilling (2009) into a coupled water and energy balance framework. With that we derive analytical solutions, which can be used to predict the sensitivity of streamflow to climate changes.

We then discuss the properties and implications of the new method. We compare our results with previous studies, namely those which employed the Budyko hypothesis for the assessment of streamflow sensitivity (Dooge, 1992; Arora, 2002; Roderick and Farquhar, 2011). In a second paper (Renner and Bernhofer, 2012), we will address the application of this hydro-climatic framework on a multitude of catchments throughout the contiguous United States.

3.2 THEORY

In this section we aim to derive a general framework for the analysis and estimation of long term average changes in basin evapotranspiration and streamflow. The theory is based on the water and energy balance equations, valid for an area such as a watershed or river basin. We revisit the conceptual framework by Tomer and Schilling (2009) and employ it to derive analytical solutions for (a) the sensitivity of a given basin to climate changes and (b) the expected changes in basin evapotranspiration and streamflow under a given change in climate.

3.2.1 COUPLED WATER AND ENERGY BALANCE

Actual evapotranspiration E_T links the catchment water and energy balance equations:

$$P = E_T + Q + \Delta S_\mathrm{w} \quad \text{and} \tag{3.1}$$

$$R_\mathrm{n} = E_T\, L + H + \Delta S_\mathrm{e}. \tag{3.2}$$

The water balance equation expresses the partitioning of precipitation P into the water fluxes E_T, streamflow Q (expressed as an areal estimate) and ΔS_w which is the change in water storage. The energy balance equation describes, how available energy, expressed as net radiation R_n, is divided at the earth surface into the turbulent fluxes, latent heat flux $E_T\, L$, where L denotes the latent heat of vaporization, the sensible heat flux H and the change in energy storage ΔS_e.

As we regard the temporal scale of long term averages and thus consider the integral effect of a range of possible processes involved, we can assume that both, the change in water and in energy storage, are zero. Dividing the energy balance equation by the latent heat of vaporization L, both balance equations have the unit of water fluxes, usually expressed as mm per time. Further, the term R_n/L, can also be denoted as potential evapotranspiration E_p, and expresses the typical upper limit of potential evapotranspiration (Budyko, 1974; Arora, 2002). With the above simplification we can write the energy balance equation as:

$$E_\mathrm{p} = E_T + H/L. \tag{3.3}$$

3.2.2 THE ECOHYDROLOGIC FRAMEWORK FOR CHANGE ATTRIBUTION

In the long term, actual basin evapotranspiration E_T is mainly limited by water supply P and energy supply E_p, which considered together, determine a hydro-climatic state space (P, E_p, E_T).

Regarding long term average changes in the hydrological states, these must be caused either by a change in climatic conditions, by changes in basin conditions or a combination of both, quietly assuming that our data is homogeneous over time. The conceptual model of Tomer and Schilling (2009) aims to distinguish between both types of causes. They employ two non-dimensional variables, relative excess energy U and relative excess water W, which can be obtained by normalizing, both the water balance and the energy balance by P and E_p, respectively:

$$W = 1 - \frac{E_T}{P} = \frac{Q}{P}, \quad U = 1 - \frac{E_T}{E_\mathrm{p}} = \frac{H/L}{E_\mathrm{p}}. \tag{3.4}$$

So, relative excess water W describes the proportion of available water not used by the ecosystem, which is in the case of a catchment the runoff ratio Q/P. Similarly, the remaining proportion of the available energy not used for evapotranspiration is expressed as relative excess energy U. Usually both terms are within the interval (0, 1], because E_T is generally positive, it cannot be larger than P and it is mostly smaller than E_p (excluding cases with a negative Bowen ratio). These limits are also known as the water and energy limits proposed by Budyko (1974). The relation of both terms is essentially a coupled consideration of water and energy balances, to which we will refer to as the UW space. So plotting U versus W in a diagram depicts the relative partitioning of water and energy fluxes of a given basin.

The long term average state expressed by W and U can be thought of as a steady state balancing water and energy fluxes through coupling between soil, vegetation, hydromorphology and atmosphere (Milne et al., 2002). Thus a shift in these two variables can be caused by changes within the basin (e.g. land cover change) but also by external environmental changes (e.g. climatic changes) (Tomer and Schilling, 2009). Deduced from observations, Tomer and Schilling (2009) proposed that the direction of change in relative excess water and energy (ΔW, ΔU) respectively, can be used to attribute the observed changes, e.g. in streamflow to a change in climate or basin characteristics such as land-use. The conceptual model by Tomer and Schilling (2009) is shown in Fig. 3.1. It displays shifts in W and U from a reference state.

The model can be explained as follows: assume that P and E_p are constant while E_T has changed over time. According to the model, this is a result of changes in basin characteristics, for example a change in land-use or land management. Such a case is displayed in the diagram (Fig. 3.1) by a change of ΔW, ΔU along the positive diagonal, i.e. a simultaneous increase or decrease in both W and U. Contrarily, a shift along the negative diagonal (i.e. $\Delta W/\Delta U = -1$) indicates effects of only climatic changes of long-term average P and E_p. As an example, consider a catchment where climatic variations may have led to a decrease in annual average P and leaving less water for both E_T and Q. Thus, the model would predict lower E_T, resulting in positive ΔU (increasing excess of energy) and in negative ΔW (decreasing excess of water).

One apparent problem is the definition of climate changes. This concept only refers to climate changes if long-term annual average precipitation or evaporative demand (E_p) are changing. Other climatic changes, such as seasonal redistribution or spatial changes in precipitation are not included in the model and can easily be mistaken as impacts of e.g. a change in land-use. Also, for example, an increase in atmospheric CO_2 concentrations, which supposedly effects

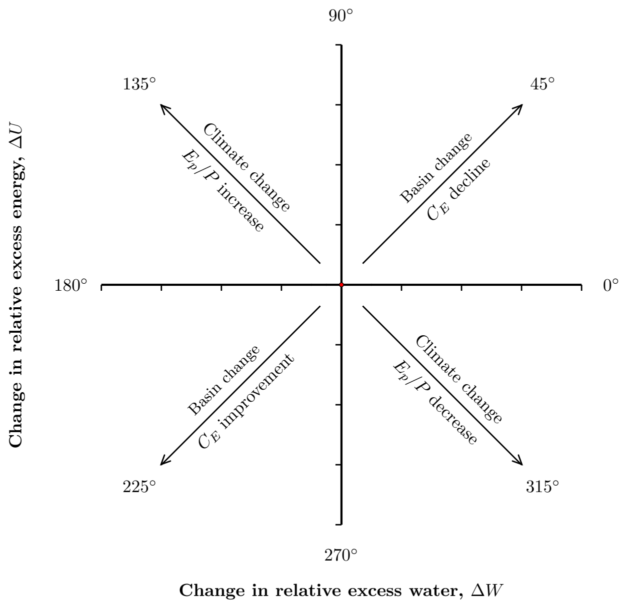

Figure 3.1: Illustration of the change attribution framework established by Tomer and Schilling (2009, after their Fig. 2). Considering climatic change effects, a change in either precipitation or potential evapotranspiration, will result in a change of both, relative excess water and energy but in opposite direction (change along the negative diagonal). Basin change effects, such as a change in vegetation or soils may lead to a change in evapotranspiration and thus in catchment efficiency (C_E, Eq. 3.8), which results in a deviation from the negative diagonal.

E_T (Gedney et al., 2006), can not be attributed to a climate change direction in Fig. 3.1. To avoid confusion, we will refer to climate changes, when P or E_p is changing, all other potential impacts on E_T are referred to as basin impact changes. A not so apparent problem is that this concept has been established for an area where P and E_p are of similar magnitude. Thus, we do not know if the approach is also valid under very humid or arid conditions.

The conceptual model of Tomer and Schilling (2009) states that climatic and basin characteristic changes lead to qualitatively different changes in the partitioning of water (W) and energy (U) at the surface. If we take this further and assume that the concept is invariant to the aridity index E_p/P of a given catchment, a quantitative hypothesis, relevant for the sensitivity of actual evapotranspiration and streamflow to changes in P, E_p, can be deduced:

$$\Delta U/\Delta W = -1. \tag{3.5}$$

We refer to Eq. (3.5) as the climate change impact hypothesis (abbreviated as CCUW).

A further interesting measure is the change direction in UW space α:

$$\alpha = \arctan \frac{\Delta U}{\Delta W}. \tag{3.6}$$

which allows to compare changes in the relative partitioning of surface water and energy balances of different basins.

3.2.3 APPLYING THE CLIMATE CHANGE HYPOTHESIS TO PREDICT CHANGES IN BASIN EVAPOTRANSPIRATION AND STREAMFLOW

Tomer and Schilling (2009) proposed to analyse shifts in W and U to retrospectively attribute changes in climate or in basin characteristics to changes in streamflow. Therefore, one only needs long-term annual average data of P, E_p and E_T, which may be derived from the water balance of a catchment ($P-Q$). However, the CCUW hypothesis may also have predictive capabilities, where the effect of climatic changes (i.e. in P, E_p) on E_T and Q can be estimated. This will also allow us to evaluate the implications of the CCUW hypothesis under different hydro-climatic states (P, E_p, E_T).

The derivation of an analytical expression for prediction of streamflow or evapotranspiration given a climatic change signal is straightforward. First consider two long-term average hydro-climate state spaces (P_0, $E_{p,0}$, $E_{T,0}$), (P_1, $E_{p,1}$, $E_{T,1}$) of a given basin. With that the changes in relative excess water ΔW and energy ΔU can be expressed by using Eq. (3.4) as:

$$\Delta W = \frac{E_{T,0}}{P_0} - \frac{E_{T,1}}{P_1}, \quad \Delta U = \frac{E_{T,0}}{E_{p,0}} - \frac{E_{T,1}}{E_{p,1}}. \tag{3.7}$$

Applying the CCUW hypothesis Eq. (3.5) to the definitions of ΔW and ΔU (Eq. 3.7), we find that the sum of E_T/P and E_T/E_p of a given basin is constant and thus invariant for different climatic conditions:

$$\frac{E_{T,0}}{P_0} + \frac{E_{T,0}}{E_{p,0}} = \frac{E_{T,1}}{P_1} + \frac{E_{T,1}}{E_{p,1}} = C_E = \text{const.} \tag{3.8}$$

We name this constant "catchment efficiency" (C_E). C_E is useful as it provides a measure which considers, both the water and energy balance equations, with respect to (a) actual evapotranspiration and (b) its main drivers, water and energy supply. C_E is at maximum, if water and energy supply are equally large (climatic precondition) and if E_T fully utilizes all water and energy supplies (catchment conditions). In this extreme case we would find a value of $C_E = 2$. Contrarily, if $E_T = 0$ then C_E would also be zero. Under extreme arid or humid conditions and assuming that $E_T = \min(P, E_p)$, we would find a value of C_E of about 1.

Finally rearranging Eq. (3.8) yields an expression to compute the evapotranspiration of the new state ($E_{T,1}$):

$$E_{T,1} = E_{T,0} \frac{\frac{1}{P_0} + \frac{1}{E_{p,0}}}{\frac{1}{P_1 + \frac{1}{E_{p,1}}}}. \tag{3.9}$$

By applying the long term water balance equation with $P = E_T + Q$ the expected new state in streamflow Q_1 can also be predicted:

$$Q_1 = \frac{\frac{Q_0}{P_0} - \frac{P_0 - Q_0}{E_{p,0}} + \frac{P_1}{E_{p,1}}}{\frac{1}{P_1} + \frac{1}{E_{p,1}}}. \tag{3.10}$$

So, given a reference long term hydro-climatic state space of a basin (P_0, $E_{p,0}$, $E_{T,0}$) or (P_0, $E_{p,0}$, Q_0) and changes in the climate state (P_1, $E_{p,1}$), the resulting hydrologic states Q_1 or $E_{T,1}$ can be predicted.

3.2.4 DERIVATION OF CLIMATIC SENSITIVITY USING THE CCUW HYPOTHESIS

The elasticity concept of Schaake and Liu (1989) describes that relative changes in streamflow are proportional to the inverse of the runoff ratio (P/Q) multiplied with a term describing how runoff is changing with changes in precipitation $\partial Q/\partial P$:

$$\varepsilon_{Q,P} = \frac{P}{Q}\frac{\partial Q}{\partial P}. \tag{3.11}$$

Thus determination of the unknown term $\frac{\partial Q}{\partial P}$, which can also be written as $1 - \frac{\partial E_T}{\partial P}$ (Roderick and Farquhar, 2011), is the key to predict the sensitivity of streamflow to changes in precipitation $\varepsilon_{Q,P}$.

Next, we derive sensitivity coefficients by applying the CCUW hypothesis. To assess the sensitivity of a basin at a given hydro-climatic state space (P, E_p, E_T) to changes in climate, we derive the first derivatives of W and U. The result is a tangent at a given hydro-climatic state space. First W and U are expressed as functions of E_T, E_p and P, respectively:

$$W = w(P,\ E_T) = 1 - \frac{E_T}{P}, \quad U = u(E_\mathrm{p},\ E_T) = 1 - \frac{E_T}{E_\mathrm{p}}.$$

Then their first total derivatives and solutions of the partial differentials are:

$$\mathrm{d}W = w'(P,\ E_T) = \frac{\partial w}{\partial P}\mathrm{d}P + \frac{\partial w}{\partial E_T}\mathrm{d}E_T \tag{3.12}$$

$$\mathrm{d}U = u'(E_\mathrm{p},\ E_T) = \frac{\partial u}{\partial E_\mathrm{p}}\mathrm{d}E_\mathrm{p} + \frac{\partial u}{\partial E_T}\mathrm{d}E_T \tag{3.13}$$

$$\frac{\partial w}{\partial P} = \frac{E_T}{P^2},\ \frac{\partial w}{\partial E_T} = -\frac{1}{P},\ \frac{\partial u}{\partial E_\mathrm{p}} = \frac{E_T}{E_\mathrm{p}^2},\ \frac{\partial u}{\partial E_T} = -\frac{1}{E_\mathrm{p}}. \tag{3.14}$$

Combining Eqs. (3.12) and (3.13) with the CCUW hypothesis Eq. (3.5) yields an expression for changes in E_T:

$$\mathrm{d}E_T = \frac{-\frac{\partial u}{\partial E_\mathrm{p}}\mathrm{d}E_\mathrm{p} - \frac{\partial w}{\partial P}\mathrm{d}P}{\frac{\partial u}{\partial E_T} + \frac{\partial w}{\partial E_T}}.$$

Finally, dividing by E_T (i.e. the long term average) and term expansions we yield an expression for the relative sensitivity of E_T to relative changes in P and E_p, in which the partial solutions of relative excess water and energy Eq. (3.14) are applied to gain an analytical solution:

$$\frac{\mathrm{d}E_T}{E_T} = \left[\frac{E_\mathrm{p}}{E_T}\frac{-\frac{\partial u}{\partial E_\mathrm{p}}}{\frac{\partial u}{\partial E_T} + \frac{\partial w}{\partial E_T}}\right]\frac{\mathrm{d}E_\mathrm{p}}{E_\mathrm{p}} + \left[\frac{P}{E_T}\frac{-\frac{\partial w}{\partial P}}{\frac{\partial u}{\partial E_T} + \frac{\partial w}{\partial E_T}}\right]\frac{\mathrm{d}P}{P} \tag{3.15}$$

$$\frac{\mathrm{d}E_T}{E_T} = \left[\frac{P}{E_\mathrm{p} + P}\right]\frac{\mathrm{d}E_\mathrm{p}}{E_\mathrm{p}} + \left[\frac{E_\mathrm{p}}{E_\mathrm{p} + P}\right]\frac{\mathrm{d}P}{P}. \tag{3.16}$$

By Eq. (3.16) we derived an analytical expression of the relative sensitivity of basin E_T to changes in climate. The terms in brackets are sensitivity coefficients, also referred to as climate elasticity coefficients (Schaake and Liu, 1989; Roderick and Farquhar, 2011; Yang and Yang, 2011). They express the proportional change in E_T or Q due to changes in climatic variables. Further, it can be seen from Eq. (3.16), that the relative sensitivity of E_T to climatic changes is only dependent on the aridity (E_p/P).

The sensitivities of streamflow to climate can be derived by applying the long term water balance equation $\mathrm{d}Q = \mathrm{d}P - \mathrm{d}E_T$ to Eq. (3.16):

$$\frac{\mathrm{d}Q}{Q} = \left[\frac{P(P - Q)}{Q(E_\mathrm{p} + P)}\right]\frac{\mathrm{d}E_\mathrm{p}}{E_\mathrm{p}} + \left[\frac{P}{Q} - \frac{(P - Q)E_\mathrm{p}}{Q(E_\mathrm{p} + P)}\right]\frac{\mathrm{d}P}{P}. \tag{3.17}$$

So, besides of being dependent on aridity, streamflow sensitivity itself is also dependent on the long term average streamflow. Again the bracketed terms denote elasticity coefficients.

3.2.5 THE BUDYKO HYPOTHESIS AND DERIVED SENSITIVITIES

The relation of climate and streamflow has already been empirically described in the early 20th century. In the long term it has been found that annual average evapotranspiration is a function of P and E_p. This is also known as the Budyko hypothesis. There exist many non-parametric functional forms (e.g. Schreiber, 1904; Ol'Dekop, 1911; Budyko, 1974), which allow to estimate E_T from climate data alone. However, actual E_T is often different from the functional non-parametric Budyko forms. To account for the manifold effects of basin characteristics on E_T, various functional forms have been proposed, which introduce an additional catchment parameter to improve the prediction of E_T. Widely applied is the function established by Bagrov (1953) and Mezentsev (1955)

$$E_T = E_p \cdot P / \left(P^n + E_p^n\right)^{1/n}, \tag{3.18}$$

to which we will refer to as Mezentsev function. Yang et al. (2008) derived Eq. (3.18) from mathematical reasoning and found that the parameter of the function suggested by Fu (1981) has a deterministic relationship with the parameter n in Mezentsev's equation. Choudhury (1999) found that n is about 1.8 for data from river basins. Further, Donohue et al. (2011) showed that for $n = 1.9$ the Mezentsev is quite similar to the Budyko curve, being the geometric mean of the curves of Schreiber and Ol'Dekop.

So more generally, the Budyko functions express E_T as a function of climate and a catchment parameter n: $E_T = f(E_p, P, n)$. Once the functional type of f is known, climate changes causing a change in E_T (dE_T) from its long-term average can be computed (Dooge et al., 1999). Usually, the first total derivative of f is being used (Roderick and Farquhar, 2011):

$$dE_T = \frac{\partial E_T}{\partial P} dP + \frac{\partial E_T}{\partial E_p} dE_p + \frac{\partial E_T}{\partial n} dn. \tag{3.19}$$

Next, by employing the long term water balance equation $dQ = dP - dE_T$ to Eq. (3.19), an expression for the change in streamflow (dQ) is gained (Roderick and Farquhar, 2011):

$$dQ = \left(1 - \frac{\partial E_T}{\partial P}\right) dP - \frac{\partial E_T}{\partial E_p} dE_p - \frac{\partial E_T}{\partial n} dn. \tag{3.20}$$

With Eqs. (3.19), (3.20) and solutions of the respective partial differentials being dependent on the type of Budyko function used, we have analytical solutions for evapotranspiration and streamflow changes due to variations in climate conditions (dP, dE_p) and changes in basin characteristics (dn) (Roderick and Farquhar, 2011). In the case of the non-parametric Budyko functions, the last term in Eqs. (3.19) and (3.20) can be omitted.

Climatic elasticities (dE_T/E_T and dQ/Q) can easily be obtained from Eqs. (3.19) and (3.20) by dividing by E_T or Q and term expansions on the right side (Roderick and Farquhar, 2011):

$$\frac{dE_T}{E_T} = \left[\frac{P}{E_T}\frac{\partial E_T}{\partial P}\right]\frac{dP}{P} + \left[\frac{E_p}{E_T}\frac{\partial E_T}{\partial E_p}\right]\frac{dE_p}{E_p} + \left[\frac{n}{E_T}\frac{\partial E_T}{\partial n}\right]\frac{dn}{n} \tag{3.21}$$

$$\frac{dQ}{Q} = \left[\frac{P}{Q}\left(1 - \frac{\partial E_T}{\partial P}\right)\right]\frac{dP}{P} + \left[\frac{E_p}{Q}\frac{\partial E_T}{\partial E_p}\right]\frac{dE_p}{E_p} + \left[\frac{n}{Q}\frac{\partial E_T}{\partial n}\right]\frac{dn}{n}. \tag{3.22}$$

As in the previous subsection, the bracketed terms denote the elasticity coefficients for P, E_p and n. For the computation of dE_T, dQ and the elasticity coefficients, we only need to enter

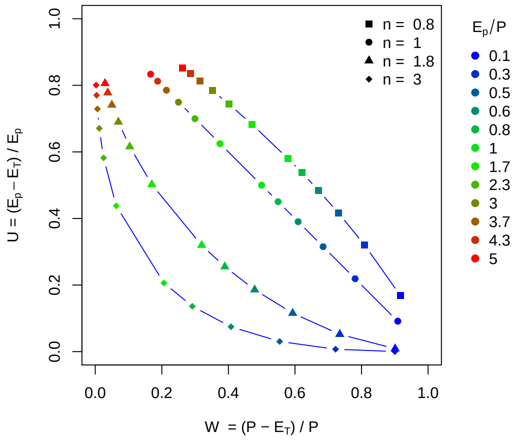

Figure 3.2: Mapping different parameterised Mezentsev functions into UW space using Eq. (3.25). The colours depict certain aridity (E_p/P) values indicated by the legend in the right.

the respective partial differentials. Roderick and Farquhar (2011) report these terms for the Mezentsev function and they are repeated for completeness below:

$$\frac{\partial E_T}{\partial P} = \frac{E_T}{P}\left(\frac{E_\mathrm{p}^n}{P^n + E_\mathrm{p}^n}\right), \quad \frac{\partial E_T}{\partial E_\mathrm{p}} = \frac{E_T}{E_\mathrm{p}}\left(\frac{P^n}{P^n + E_\mathrm{p}^n}\right) \tag{3.23}$$

$$\frac{\partial E_T}{\partial n} = \frac{E_T}{n}\left(\frac{\ln\left(P^n + E_\mathrm{p}^n\right)}{n} - \frac{\left(P^n \ln(P) + E_\mathrm{p}^n \ln\left(E_\mathrm{p}\right)\right)}{P^n + E_\mathrm{p}^n}\right). \tag{3.24}$$

3.3 SENSITIVITY ANALYSIS

In this section the properties and implications of the CCUW hypothesis are evaluated, discussed and compared with the established Budyko streamflow sensitivity approaches.

3.3.1 MAPPING OF THE BUDYKO FUNCTIONS INTO UW SPACE

The variables (P, E_p, E_T) used by the Budyko and the CCUW hypothesis are identical and can be easily related between both diagrams (spaces):

$$W = 1 - f\left(E_\mathrm{p}, P, n\right), \quad U = 1 - \frac{f\left(E_\mathrm{p}, P, n\right) P}{E_\mathrm{p}}. \tag{3.25}$$

Figure 3.2 illustrates the functional behaviour of the Mezentsev function for different catchment parameters n in UW space. The Budyko functions describe curves in the UW space, whereby values of $n > 1$ result in smaller values of both, W and U. Also note that for $n = 1$ the Mezentsev function Eq. (3.18) follows the negative diagonal of the climate change hypothesis, cf. Fig. 3.1.

More important for streamflow change assessment is that the Budyko functions display curves in the UW space. Generally, the derived climate sensitivity is a tangent at some aridity value of a Budyko function. Meaning that there are different climate change directions in UW

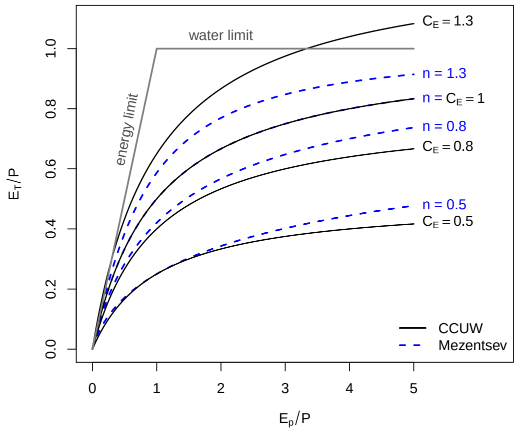

Figure 3.3: Mapping of CCUW hypothesis into Budyko space for different values of catchment efficiency (C_E) using Eq. (3.26). For comparison different parameterisations of the Mezentsev curve are also shown. The grey lines depict the theoretical limits for water and energy.

space (CCD), depending on the aridity of a basin and the respective Budyko curve. So, under humid conditions climatic changes are more sensitive on relative excess water (larger change in runoff ratio than in relative excess energy). Thus the slope of the tangent for $n > 1$ will be larger than -1, but not exceed 0. Under arid conditions changes are more sensitive to relative excess energy and the slope will always be smaller than -1. That means, independent of any given condition (P, E_0, n) and any climatic change ($\Delta\frac{E_p}{P}$), the slope will always be negative and thus $-\infty < \Delta U/\Delta W < 0$, which refers to change directions into the 2nd ($90° < \alpha < 180°$) or the 4th quadrant ($270° < \alpha < 360°$) in Fig. 1. Moreover, it is interesting to note, that if $P = E_p$ the CCD obtained by the Budyko framework using Mezentsev's curve is identical to the one of the CCUW hypothesis. The differences to the CCUW hypothesis are generally increasing the more humid/arid a given basin is. Further, the larger the catchment parameter n, the larger the differences. A mathematical derivation of the climate change direction of the Mezentsev function (α_{mez}) can be found in the Appendix 3.A.

3.3.2 MAPPING CCUW INTO BUDYKO SPACE

For comparison of the CCUW hypothesis with the established Budyko functions we map the CCUW hypothesis into Budyko space and visualise the differences. For the purpose of mapping we come back to Eq. (3.8), where C_E is assumed to be a constant, which is a consequence of the climate change impact hypothesis in UW space. With that we can rearrange Eq. (3.8) to achieve a mapping to Budyko space:

$$\frac{E_T}{P} = C_E \frac{E_p}{P + E_p}. \tag{3.26}$$

Figure 3.3 illustrates the functional form of change predictions of the CCUW hypothesis for different values of C_E. These can be compared with the curves for different parameterisations of Eq. (3.18). The curves of the CCUW hypothesis are strongly determined by C_E, similar to the effect of different values for the catchment parameter n in the parameterised Budyko model of

Mezentsev (1955). By recollecting Eqs. (3.18) and (3.26) we can see, that for $n = 1$ and $C_E = 1$ both functions are identical.

It is, however, important to note, that there is a different asymptotic behaviour of the CCUW hypothesis compared to the Budyko hypothesis. The actual value of the catchment efficiency C_E determines the asymptote for $E_p/P \to \infty$. This makes a distinction from the Budyko hypothesis, which employs the water limit $E_T/P = 1$ as asymptote for $E_p/P \to \infty$. Especially under more arid climatic conditions the differences in climatic sensitivity are apparent. When $C_E > 1$, the slopes of the CCUW function are steeper than those of the Budyko functions and if $C_E < 1$ the slopes are more levelled. For example, let us consider the case of increasing aridity and a basin on the curve for $C_E = 1.3$ as shown in Fig. 3.3. At some point the water limit ($E_T = P$) will be reached, which means that all precipitation is evaporated and there is no runoff anymore. Any points on the curve above the water limit violate the water balance, because actual evapotranspiration can not be larger than the water supply. Thus, for physical reasons, C_E has to decrease when approaching the Budyko envelope. This means that the strong assumption of the CCUW hypothesis with constant C_E can not be valid for all hydro-climatic states and streamflow sensitivity results of basins close to the Budyko water and energy limits are probably not realistic.

3.3.3 CLIMATIC SENSITIVITY OF BASIN EVAPOTRANSPIRATION AND STREAMFLOW

In the theoretical section of this paper we derived analytical equations (i) for predicting the absolute hydrological response for variations in climate and (ii) for estimating the climatic sensitivity, i.e. the proportional change in E_T or Q by a proportional change in climate.

Figure 3.4 illustrates the general behaviour of the CCUW hypothesis under changes in precipitation or potential evapotranspiration, which is expressed by Eqs. (3.9) and (3.10). The left panels of Fig. 3.4 show the relative change of streamflow to P (upper) and E_p (lower panel). From Eq. (3.10) follows that climatic sensitivity of streamflow is regulated by runoff ratio $W = Q/P$ and aridity E_p/P. We find that the smaller the runoff ratio, the larger the climatic effect on streamflow. The slopes of curves depicting the relative change of streamflow are modulated by aridity, with more arid (humid) basins having a smaller (larger) sensitivity. In the right panels of Fig. 3.4 the relative changes in E_T due to relative changes in P (upper panel) and in E_p (lower panel) are shown. The figures highlight that the magnitude of relative change is dependent on the aridity of the given basin. So the more arid the climate, the larger are changes in E_T due to changes in P, while changes in E_p show the opposite behaviour.

In addition, the curves shown in Fig. 3.4 display substantial nonlinear behaviour to changes either in P or E_p. Considering the rainfall-runoff relation, this means that the relative change in streamflow is not proportional to the change in precipitation, but also depends on the magnitude of change in precipitation. In general, positive precipitation changes result in stronger changes in streamflow, than negative precipitation changes. Such features have e.g. been reported by Risbey and Entekhabi (1996), analysing the response of the Sacramento River basin (US) to precipitation changes. While Risbey and Entekhabi (1996) argue that hydrological memory effects are related to this nonlinear behaviour, our analysis suggests that the coupled nature of water and energy balances is the primary cause of the nonlinear response of streamflow to climate.

Next, we discuss and compare climate elasticities derived by the CCUW and the Budyko sensitivity approaches. Kuhnel et al. (1991) showed that $\varepsilon_P + \varepsilon_{E_p} = 1$. Therefore, we only discuss the elasticity to precipitation. Figure 3.5 displays the elasticity of E_T ($\varepsilon_{E_T,P}$) as a function of aridity. In more humid or semi-arid conditions ($E_p/P < 2$), the differences between the Budyko function elasticities and the ones derived by the CCUW hypothesis are small. In each case the sensitivity increases with aridity. In more arid conditions larger differences of the CCUW hypothesis to the Budyko sensitivity functions become apparent. Thereby, the parametric Budyko function with $n > 1$ approaches the upper limit ($\varepsilon_{E_T,P} = 1$) distinctly faster than the CCUW method.

So for example, a precipitation decrease of 10 % in an arid basin with $E_p/P = 4$ results in an estimated change of E_T by 8 %, when the CCUW hypothesis is applied. However, applying the Budyko framework with the Mezentsev function and $n = 1.9$, E_T changes by 9.3 %. Even though this seems to be a small difference, in absolute values such changes are large, when considering the fact that in such arid basins annual E_T is almost as large as annual precipitation.

Regarding the elasticity of streamflow, the picture gets more complicated. First, the sensitivity of streamflow is also dependent on streamflow itself, cf. Eqs. (3.17) and (3.22). Secondly, in arid conditions, streamflow is typically very small compared to all other variables considered here. So even small absolute changes in Q may result in very large elasticity coefficients. In Figure 3.6 we show $\varepsilon_{Q,P}$ as a function of aridity. Because of the dependency to streamflow, or rather to catchment efficiency, we plot $\varepsilon_{Q,P}$ as computed by CCUW for different values of C_E. The effect of C_E on streamflow is shown in the left panel of Fig. 3.6, where we plot the runoff ratio Q/P as a function of aridity. The streamflow elasticities derived by the CCUW method clearly show for arid conditions, that the larger C_E (and thus smaller Q), the larger gets $\varepsilon_{Q,P}$. In contrast the elasticities of the Mezentsev functions converge to a maximal level of $\varepsilon_{Q,P} = n + 1$ for $E_p/P \to \infty$.

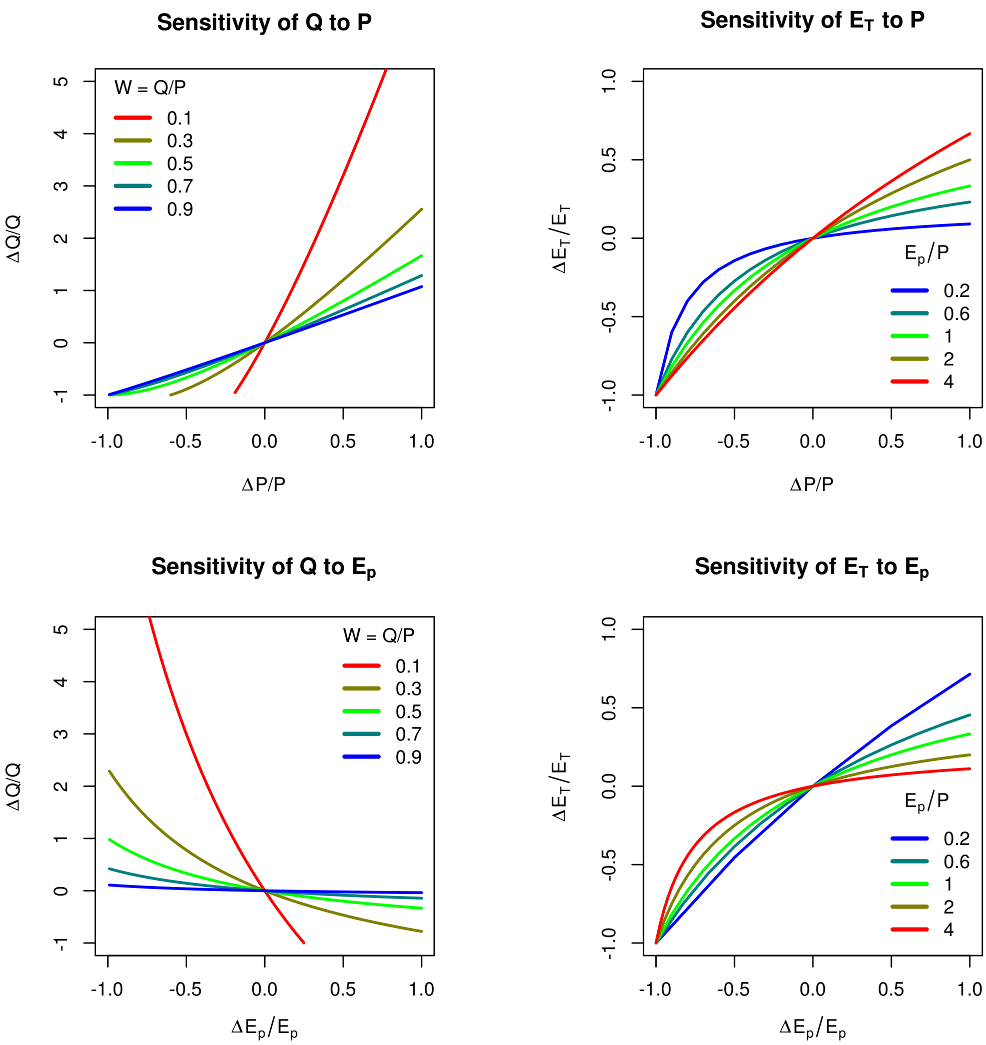

Figure 3.4: Relative change in response to relative changes in P (top panels) and in E_p (bottom panels) of Q (left panels) and E_T (right panels) as predicted by the CCUW hypothesis. Changes in Q are dependent on runoff ratio W and on aridity E_p/P and are coloured with respect to the respective runoff ratio and shown for an aridity index of $E_p/P = 1$. Relative changes in E_T are dependent on aridity only and lines are coloured with respect to different aridity indices. Note that changes of $\Delta Q/Q$ smaller than -1 are not physical.

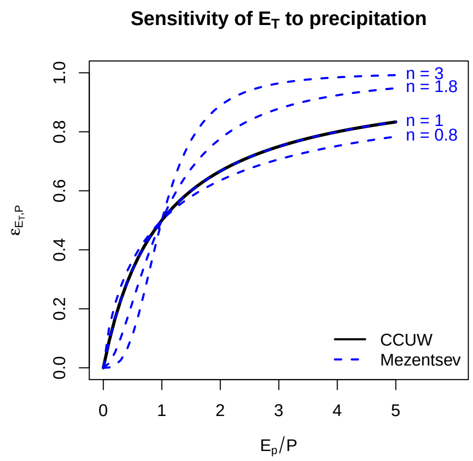

Figure 3.5: Sensitivity (elasticity) of basin evapotranspiration with respect to changes in precipitation ($\varepsilon_{E_T,P}$). The bold black line depicts elasticity of the CCUW, while the dashed line shows different elasticities for the Mezentsev function. The elasticity of the CCUW corresponds with the slope of the curves shown in the top right panel of Fig. 3.4.

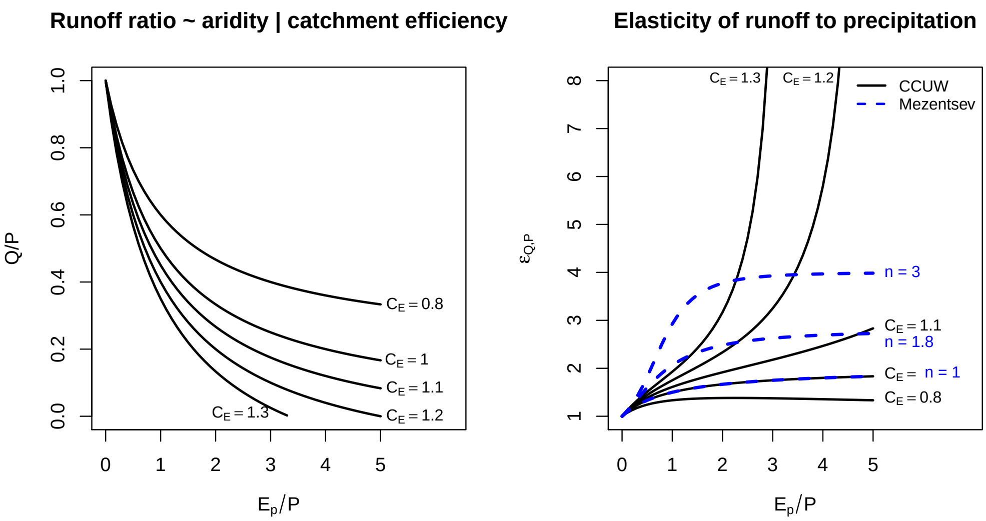

Figure 3.6: Left panel: runoff ratio as a function of aridity for different, but fixed values of catchment efficiency (C_E) using Eq. (3.26). Right panel: elasticity coefficient of streamflow to precipitation $\varepsilon_{Q,P}$ as a function of aridity. Displayed are the elasticities derived from the CCUW hypothesis (black for different values of C_E), and the elasticities derived from different parameterisations of the Mezentsev functions using Eq. (3.22).

3.3.4 Climate-vegetation feedback effects

As detailed in the theory section and illustrated above, both, the Budyko functions and the CCUW hypothesis provide analytical solutions for the problem of how E_T or Q are changing when P or E_p are changing. However, there are very different outcomes with respect to the determined sensitivity. In the following we discuss the origins and implications of these differences in more detail.

The key difference of the parametric Budyko approach is that the sensitivity of the hydrological response (E_T, Q) is also dependent on changes in the catchment parameter n, cf. Eqs. (3.21) and (3.22). In contrast the CCUW approach is only sensitive to changes in P and E_p, cf. Eqs. (3.16) and (3.17). Thus, it is interesting to study the influence of the catchment parameter encoding catchment properties on hydrological response under transient climatic conditions. Further, the elasticity concept of Schaake and Liu (1989), Eq. (3.11), shows that the sensitivity coefficients are composed of two components, which is also apparent in the sensitivity terms within Eqs. (3.21) and (3.22).

For the purpose of illustration we conduct the following experiment: we derive E_T and Q for different aridity indices E_p/P from 0 to 5 using the water balance equation of the Mezentsev function with n set to 1.8. In Fig. 3.7 we plot the two components of the sensitivity coefficients $\varepsilon_{E_T,P}$, $\varepsilon_{E_T,n}$ and $\varepsilon_{Q,P}$, $\varepsilon_{Q,n}$ as functions of the humidity index P/E_p and the aridity index E_p/P, respectively. The purpose of the different x-axes is to highlight the differences in sensitivity, which become apparent for E_T under humid conditions and for Q under arid conditions.

The top panels show the sensitivity of E_T to P and n, which can be decomposed into $\varepsilon_{E_T,P} = P/E_T \cdot \partial E_T/\partial P$ and $\varepsilon_{E_T,n} = n/E_T \cdot \partial E_T/\partial n$, respectively. Panel a displays the first terms of these sensitivity coefficients, which are both increasing with humidity. In panel b solutions of the partial differential terms are displayed for the CCUW hypothesis ($\partial E_T/\partial P = E_T/P \, \frac{E_\mathrm{p}}{E_\mathrm{p}+P}$) and the Mezentsev function Eq. (3.23). The curves of $\partial E_T/\partial P$ of the Budyko and the CCUW approach intersect at a humidity index of $P/E_\mathrm{p} = 1$ and show somewhat larger differences when $P/E_\mathrm{p} > 1.5$, whereby the Budyko curve approaches 0 faster than the CCUW curve. Panel c then displays the resulting sensitivity coefficients, which is the product of both terms shown in panels a and b. While the differences between the two approaches must be similar to the ones shown in panel b, we find that the sensitivity of E_T to the catchment parameter is larger than the sensitivity to P when $P/E_\mathrm{p} > 1.5$. The reason for this behaviour is mainly due to the first term of the coefficients: n/E_T is rising faster than P/E_T (if $n > 1$).

The lower panels of Fig. 3.7 are constructed analogously, but display the sensitivity of streamflow as a function of the aridity index. From panel d we see that the inverse of the runoff ratio is strongly increasing with aridity, but similar to the panel above n/Q is rising faster than P/Q. Panel e is only different from panel b, as P/E_T has been switched. It highlights that there are larger differences between CCUW and the Budyko approach, when $E_\mathrm{p}/P > 1.5$, which we already discussed with respect to Fig. 3.5. From panel f, we can see that these differences in $\partial E_T/\partial P$ have large consequences for the resulting streamflow sensitivities. Whereby, $\varepsilon_{Q,P;\mathrm{ccuw}}$ is proportionally increasing with P/Q and $\varepsilon_{Q,P;\mathrm{mez}}$ approaches its maximal level of $n+1$. Thus, the strong exponential effect of the inverse runoff ratio shown in panel d is heavily reduced. And mirroring the results of E_T above, the sensitivity of Q to changes in the catchment parameter is strongly increasing with aridity and apparently larger than the sensitivity to precipitation in arid basins.

Combining these findings, some important and scientifically interesting conclusions can be

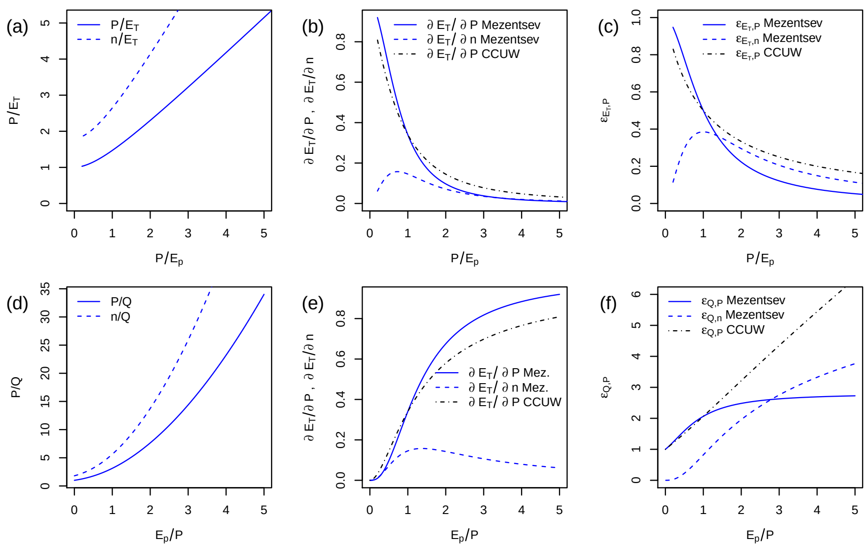

Figure 3.7: Sensitivity coefficients and their components as functions of the humidity and aridity index, respectively. Baseline water balance terms (E_T and $Q = 1 - P$) have been determined with the Mezentsev function and $n = 1.8$. For illustration purposes we set $P = 1$ and $E_p = 0 \ldots 5$. Top panels display components of the sensitivity of actual evapotranspiration E_T to precipitation P and the catchment parameter n as functions of the humidity index P/E_p using terms of Eq. (3.21). The bottom panels display components of the sensitivity coefficient of streamflow Q to P and n as functions of the aridity index E_p/P using terms of Eq. (3.22). The left panels depict the left term of the sensitivity coefficients, the middle panels the right term (solutions of the partial differentials $\frac{\partial E_T}{\partial n}$ and $\frac{\partial E_T}{\partial P}$) and the right panels show the sensitivity coefficients.

made. First, under limiting conditions, either a lack of water or a lack of energy, we find an increasing importance of the catchment properties reflected in the catchment parameter of the parametric Budyko model. Considering the similarities of the Mezentsev function in Eq. (3.18) and the CCUW hypothesis transformed into Budyko space in Eq. (3.26), we conclude that the inclusion of the catchment parameter essentially accounts for these limiting conditions. This agrees with the mathematical derivation of the Mezentsev function by Yang et al. (2008). Secondly, the inclusion of the catchment parameter results in larger sensitivities of streamflow and actual evapotranspiration to changes in catchment properties than to changes in climate. This can explain the levelled climatic sensitivity of streamflow under arid conditions even though P/Q is strongly increasing with aridity.

A direct consequence is that the separation of impacts from climate and land-use (e.g. the concept of Wang and Hejazi, 2011) in water or energy limited basins is likely to be much less certain, because even small basin changes (e.g. in vegetation) can have large effects on the hydrological response. Last, the CCUW hypothesis does not lead to such a determined climate-basin characteristic (vegetation) feedback relation as the Budyko approach. This is most apparent in water limited basins, where the sensitivity of streamflow to changes in aridity derived from the CCUW approach can be much larger than the one derived from the Budyko approach. While the Budyko approach respects the conservation of mass and energy, the CCUW hypothesis may lead to a conflict with the water limit. This indicates that the assumptions of the CCUW hypothesis are not applicable under limiting conditions.

3.4 APPLICATION: THREE CASE STUDIES

To demonstrate the applicability of the newly derived streamflow sensitivity method, we selected data of three different large river basins. We compare the climate sensitivities and absolute streamflow change predictions with the Budyko approaches.

For the case studies we selected the Murray-Darling Basin (MDB) in Australia (Roderick and Farquhar, 2011), the headwaters of the Yellow River basin (HYRB) in China (Zheng et al., 2009), and the Mississippi River Basin (MRB) in North America (Milly and Dunne, 2001). These large basins differ in climate and include arid (MDB), cold and semi-humid (HYRB) and warm, humid (MRB) climates. All basins have already been subject to climate sensitivity studies. Using hydro-climate data from the above references we derived climate sensitivity coefficients and compute the change in streamflow, given the published trends in climate. All data and computations can be found in Table 3.1.

3.4.1 MISSISSIPPI RIVER BASIN (MRB)

The largest observed trend in climate of the three basins is found for the Mississippi River Basin (upstream of Vicksburg). In the period from 1949–1997 we find a marked trend towards a more humid climate with an increasing trend in P and a decreasing trend in evaporative demand (E_p). As one would expect, the observed streamflow increased (26 %) and all predictions are around that magnitude, thus providing evidence that climatic variations explain most of the observed change in runoff. The prediction of the Budyko approach is very close to the observed change in runoff. Also the observed change direction of $\Delta U/\Delta W$ with $\alpha = 304°$ is close to the climate change direction derived from the Mezentsev function, with $\alpha_{mez} = 310°$.

The CCUW method yields somewhat larger sensitivities $\varepsilon_{Q,P}$, and thus predicts a larger change in streamflow (about 7 mm yr^{-1}) given the climatic changes. From Table 3.1 we see that C_E increased by 1 %. This is consistent with the increase in the catchment parameter (Δn), where larger values of n indicate larger E_T. So we can conclude that most of the changes in streamflow in the MRB can be attributed to the increase in humidity, but the increase in both, n and C_E, indicates that changes in basin characteristics may have contributed to increasing E_T. Note, that the numbers given for changes in human water use (e.g. dam management, groundwater harvesting) as given by Milly and Dunne (2001), do not significantly change the magnitude in observed and predicted changes.

3.4.2 HEADWATERS OF THE YELLOW RIVER BASIN (HYRB)

The headwaters of the Yellow River basin are at high altitudes (above 3480 m a.s.l.) and thus relatively cold and receive seasonal monsoon precipitation (Zheng et al., 2009). This basin is also different to the others considered here, as the observed decrease in streamflow (−20 %) comparing the periods 1960–1990 and 1990–2000 cannot be explained by the long term average changes in precipitation and potential evapotranspiration, which almost neutralise each other. As a result, the methods considered here can attribute only 24% of the observed change to climate variations. Further, the change direction in UW space with $\alpha = 210°$ implies, according to the

Table 3.1: Observations and predictions of streamflow change of three case-study river basins, Mississippi River basin (MRB), the headwaters of the Yellow River (HYRB), and the Murray-Darling River Basin (MDB). Data are taken from the respective reference publications. For prediction of streamflow change we compare the CCUW method (ΔQ_{ccuw}) with the sensitivity approach employing the Mezentsev function (ΔQ_{mez}). Change direction in UW space α, corresponding with Fig. 3.1, is computed by Eq. (3.6). The theoretical climate change direction derived for the Mezentsev function (α_{mez}) is computed by Eq. (3.31).

	unit	MRB	HYRB	MDB
area	km^2	3.0e + 06	1.2e + 05	1.1e + 06
P	$\mathrm{mm\,yr}^{-1}$	835.0	511.6	457.0
E_p	$\mathrm{mm\,yr}^{-1}$	1027.0	773.6	1590.8
Q	$\mathrm{mm\,yr}^{-1}$	187.0	179.3	27.3
E_p/P	–	1.2	1.5	3.5
Q/P	–	0.22	0.35	0.06
ΔP	$\mathrm{mm\,yr}^{-1}$	85.4	−21.0	−17.0
ΔE_p	$\mathrm{mm\,yr}^{-1}$	−17.8	−23.0	21.0
ΔQ	$\mathrm{mm\,yr}^{-1}$	48.9	−36.2	−5.6
n	–	2.00	1.13	1.74
Δn	–	0.04	0.18	0.06
C_E	–	1.41	1.08	1.21
ΔC_E	–	0.01	0.09	0.00
$\varepsilon_{Q,P;\mathrm{mez}}$	–	2.38	1.71	2.60
$\varepsilon_{Q,P;\mathrm{ccuw}}$	–	2.55	1.74	4.51
ΔQ_{mez}	$\mathrm{mm\,yr}^{-1}$	50.0	−8.8	−3.2
ΔQ_{ccuw}	$\mathrm{mm\,yr}^{-1}$	56.1	−8.8	−5.7
α	°	304	210	135
α_{mez}	°	310	134	111

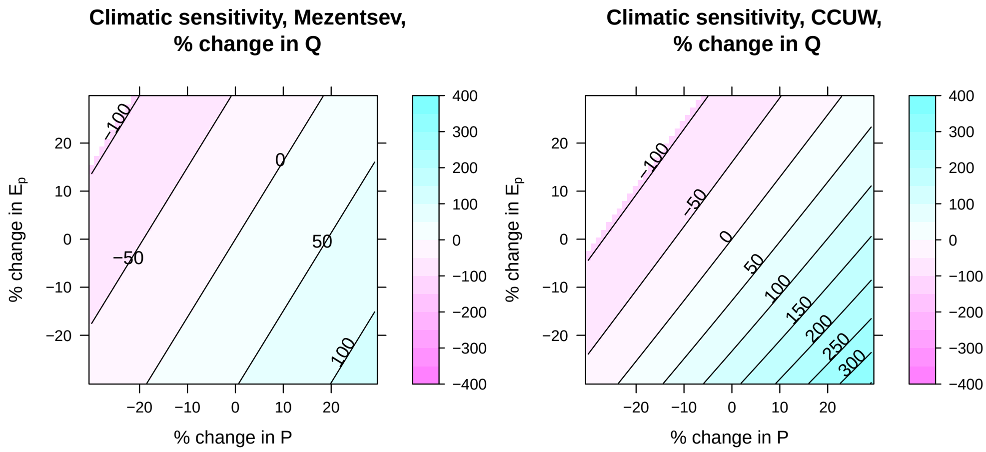

Figure 3.8: Sensitivity plots of streamflow to percent changes of precipitation and E_p, estimated for the long term hydro-climatic states of the Murray-Darling Basin (as given in Table 3.1). Contour lines depict the percent change in streamflow. Note that changes of $\Delta Q/Q$ smaller than -100 % are not physical. Left panel: The Budyko framework using the Mezentsev function and Eq. (3.22) in accordance to Roderick and Farquhar (2011, Fig. 2). Right panel, sensitivity estimation by the CCUW framework Eq. (3.10).

concept of Tomer and Schilling (2009) (Fig. 3.1), that the main direction of the observed change is in basin change direction. Both frameworks indicate that the catchment properties have been changing, with significant increases in n and C_E over time. The data reported on changes in land cover fractions before and after 1990 in Zheng et al. (2009) also implicate land-use change. Especially the increase in cultivated and forested land (above 120 %) at the cost of grassland supports this direction of change towards higher catchment efficiency.

3.4.3 MURRAY-DARLING RIVER BASIN (MDB)

For a more detailed discussion of the case studies, the MDB has been selected. It has the driest climate ($E_\text{p}/P = 3.5$) of all three basins considered. Also the climatic sensitivity coefficients are largest and climate effects on streamflow are expected to be large. We concentrate on the CCUW hypothesis and the parameterised Budyko function approach, a framework which was presented by Roderick and Farquhar (2011), especially for the MDB.

Roderick and Farquhar (2011) report long-term average data for the period 1895–2006 and a period of the last ten years 1997–2006. Comparing these periods, the climate shifted towards increased aridity, with less rain (−3.7 %) and increased potential evapotranspiration (1.3 %). The observed decrease in streamflow is $-5.6\,\text{mm}\,\text{yr}^{-1}$ (−20.5 %).

From Table 3.1 we see that (i) the elasticity coefficients to precipitation and (ii) the predicted changes in streamflow are quite different between the Budyko and the CCUW approach. When using the Budyko approach, following Roderick and Farquhar (2011), the sensitivity of streamflow to a relative change in precipitation is $\varepsilon_{Q,P;\text{mez}} = 2.6$, which is close to the theoretical upper bound of $\varepsilon_{Q,P;\text{mez}} = 1 + n = 2.74$. Employing data of the climatic changes in the second period we predict a change of $-3.2\,\text{mm}\,\text{yr}^{-1}$. Roderick and Farquhar (2011) argue that this underprediction may be due to several reasons such as a change in long term storage. They also argue that a change in catchment characteristics and changes in the spatial distribution of precipitation might explain the difference in observed and predicted streamflow. That means, following the Budyko approach of Roderick and Farquhar (2011), $-3.2\,\mathrm{mm\,yr^{-1}}$ can be attributed to a change in aridity, while the remainder ($-2.4\,\mathrm{mm\,yr^{-1}}$) must be attributed to uncertainties or to changes in catchment properties. This is supported by the observed change of n with $\Delta n = 0.06$. Using the CCUW method, we predict a change of $-5.7\,\mathrm{mm\,yr^{-1}}$, which is very close to the observed value. This means that by only considering climate impacts, the CCUW hypothesis is seemingly able to predict the observed change using the changes of P and E_p only. We also find that $\alpha = 135°$, i.e. the observed change is in climate change direction of the CCUW hypothesis, with increased aridity resulting in increased W and reduced U with quite similar absolute values. In contrast, the Budyko framework predicts $\alpha_\mathrm{mez} = 111°$, i.e. there is a larger relative change in the energy partitioning than in the partitioning of water.

Figure 3.8 illustrates the differences between the parameterised Budyko and the CCUW method on climate sensitivity. A diagram which may be practically considered for the assessment of future hydrological impacts of predicted changes in precipitation and evaporative demand (E_p) (Roderick and Farquhar, 2011). We see that the contour lines of the estimates by the CCUW method are about two times more dense compared to the contours of Roderick and Farquhar's approach. This is due to the fact that the sensitivity to precipitation is almost twice as large, cf. Table 3.1. The CCUW method predicts a larger sensitivity, because the sensitivity is mainly determined by the inverse of the runoff ratio, which is very large for the MDB ($P/Q = 16.7$). However, the result obtained with the CCUW hypothesis should be taken with care, because it is derived by putting the strong assumption that the concept of Tomer and Schilling (2009) and thus the CCUW hypothesis is valid for any given aridity index. Still, with respect to the discussion in Sect. 3.3.4, the resulting difference in streamflow sensitivity illustrates the impact of the inherent assumptions on the role of climate-vegetation feedbacks in arid environments.

3.5 CONCLUSIONS

This paper is based on a conceptual framework published by Tomer and Schilling (2009), which links shifts in ecohydrological states of river basins to shifts in climate and basin characteristics. The original concept is based on the observation that climate impacts on streamflow produce shifts in the ecohydrological states of relative excess water and relative excess energy, which are orthogonal to shifts induced by land-use or land management changes. Particularly interesting is the hypothesis that changes in the supply of water and energy (i.e. changes in the aridity index) lead to distinct changes in the relative partitioning of water and energy fluxes at the surface. According to the climate change hypothesis (CCUW), an increase (decrease) in the ratio of actual evapotranspiration to precipitation balances with the decrease (increase) in the ratio of actual to potential evapotranspiration. A direct consequence of the CCUW hypothesis, is that the sum of both terms, to which we refer to as "catchment efficiency" (C_E), is constant. We then utilise the CCUW hypothesis under the assumption that it is applicable for any aridity index, to derive analytical solutions, (i) to predict the impact of variations of the aridity index on evapotranspiration and streamflow, and (ii) to assess the climatic sensitivity of river basins. Both issues are of great practical and scientific concern.

3.5.1 POTENTIALS AND LIMITATIONS

To understand the properties and implications of the method for estimating climate sensitivity, a thorough discussion of its properties is needed for different climates, expressed by aridity and different possible hydrological responses.

The results of the sensitivity analysis and the case studies of three large river basins show that the sensitivity estimates of the CCUW hypothesis are similar to the results obtained with the Budyko framework, when conditions are far from water or energy limitation, i.e. $2/3 < E_{\mathrm{p}}/P < 3/2$. However, under limiting conditions close to the Budyko envelope large differences between both frameworks are apparent. The transformation of the CCUW hypothesis into Budyko space showed that under such conditions the CCUW hypothesis does not adhere to the water ($E_T \leq P$) and energy limits ($E_T \leq E_{\mathrm{p}}$) proposed by Budyko (1974).

As we show, the effects are largest for the sensitivity of streamflow under arid conditions, where the sensitivity of CCUW tends to increase with the inverse of the runoff ratio, while the sensitivity of the Budyko method approaches a constant value. These findings exclude the use of sensitivity estimates derived by the CCUW hypothesis under hydro-climatic conditions being close to the water limit and limits its use compared to the more general approach of Roderick and Farquhar (2011). In contrast to the CCUW sensitivity framework, their Budyko sensitivity framework respects the conservation of mass and energy even under limiting conditions.

However, our study allows some conclusions on how to use the simple concept of Tomer and Schilling (2009) to separate climate from land-use effects on evapotranspiration and streamflow. First, the concept (Fig. 3.1) with the diagonals representing the change directions, is a special case of sensitivity frameworks using the Mezentsev function under the condition that long-term average precipitation equals evaporative demand. The catchments considered by Tomer and Schilling (2009) have been close to this condition and therefore the Budyko framework estimates similar attributions. If conditions are different, the climate change (and the basin change) directions given in Fig. 3.1 need a case specific correction. As we have shown, if a rotation of the original concept is applied for correction, the result will depend on the aridity index and the catchment parameter n. Generally, when $n > 1$ and under arid conditions, the climate change direction is corrected towards the ordinate in Fig. 3.1, while under humid conditions the arrows are towards the abscissa.

3.5.2 INSIGHTS ON THE CATCHMENT PARAMETER

We compare our results with a parametric Budyko function, which was first proposed by Mezentsev (1955) and recently was also applied for the problem of streamflow sensitivity by Roderick and Farquhar (2011). Yang et al. (2008), who derived the Mezentsev (1955) equation by mathematical reasoning, showed that accounting for the water and energy limits leads to a catchment specific constant. This catchment parameter has a range of effects, which increase in magnitude under the lack of water or energy.

This has several interesting implications. First, the catchment parameter, describing the integral effect of all processes forming the hydrological response of a catchment, influences the sensitivity of catchment E_T to climatic changes. For example the type of vegetation of a basin can significantly affect climatic sensitivity of E_T. This was for example shown for the aerodynamic and canopy resistance parameters in the Penman-Monteith equation (Beven, 1979). Second, the influence of catchment properties is increasing under limiting conditions. As we show, the direct sensitivity of E_T to changes in the catchment parameter can be larger than the sensitivity to changes, e.g. in precipitation, under very wet or very dry conditions. This means that a small change in catchment properties can have large relative effects on evapotranspiration in very humid basins, whereas streamflow would be highly affected in arid basins. On the one hand, this relation will complicate the detection of effects of climatic changes on the water budget in limited environments. On the other hand, we expect that catchment ecosystems adapt to transient climatic changes in order to keep their functionality. Such adaptions are likely to have considerable impact on the resulting hydrological response, however, such climate-vegetation feedback relations are not explicitly considered in any of the two frameworks considered here.

3.5.3 Validation

In this paper we have compared two hypotheses about how streamflow is changing when long-term average precipitation or evaporative demand are changing. Still, both hypotheses need to be tested and validated.

Here, we give only some recommendations. First, there is the necessity to control for catchment property changes, which complicates any attempt of validation. Still, one could try to trace the hydro-climatic states of individual basins through time, hoping for different climatic boundary conditions. Possible test setups are, (i) controlled small scale experiments preferably under more extreme climatic conditions (humid, semi-arid, arid). Examples are the agricultural experiments described by Tomer and Schilling (2009), long-term experimental watershed programs (Moran et al., 2008) or the Long-term Ecological Research project `http://www.lternet.edu/`. Another approach is, (ii) the evaluation of large hydro-climate datasets, where the effect of basin changes can be treated statistically. One example has been presented by Renner and Bernhofer (2012), using a large set of river basins in the United States. In parallel, one could use physically based models, where controlling of basin characteristics is easy, but difficult to prove as the choice of parameters evidently effects the resulting sensitivity coefficients.

Independent of the approach taken, we believe that normalising observations such as relative excess energy and water can reveal interesting relationships of complex data sets.

3.5.4 Perspectives

We are aware that this paper opens a range of further questions and perspectives. Therefore, we would like to discuss a few directions of further research. Most important is to provide empirical evidence of the validity of hypotheses linking climate and hydrological response. Particularly, the role of catchment properties under transient climates needs to be quantified. But also the role of other climatic properties, which are not reflected in the simple water-energy balance models, is of great interest.

Given the significance of vegetation and ecosystems (Donohue et al., 2007) we believe that ecohydrological models and conceptualising such processes at the catchment scale (Klemeš, 1983) is of great importance. Inspiring research introduced the role of soils (Milly, 1994; Porporato et al., 2004), the stochastic role of precipitation (Choudhury, 1999; Gerrits et al., 2009) and the role of self-organising principles of catchment ecosystems (Rodriguez-Iturbe et al., 2011) on the mean annual water balance. However, the remaining challenge is to describe their role under transient climatic conditions.

3.A DERIVATION OF THE CLIMATE CHANGE DIRECTION IN UW SPACE FOR THE MEZENTSEV FUNCTION

Consider a Budyko function which expresses the evaporation ratio as a function of the aridity index $\Phi = E_p/P$ and a catchment parameter n as

$$E_T/P = f(\Phi,\ n). \tag{3.27}$$

With Eq. (3.25) we obtained a mapping of f to the UW space. Using the aridity index as $\Phi = E_p/P$, Eq. (3.25) can be written as:

$$W = 1 - f(\Phi,\ n) \tag{3.28}$$

$$U = 1 - \frac{f(\Phi,\ n)}{\Phi}. \tag{3.29}$$

To estimate the climate change direction in UW space (CCD) of some Budyko function at any given Φ, n, we need to compute the first derivative U' of $U = g(W,\ n)$, whereby W is obtained by Eq. (3.28). Because Eq. (3.29) includes both $f(\Phi,\ n)$ and Φ, we need to derive the inverse of Eq. (3.27). The analytical solution for Mezentsev' function Eq. (3.18) is derived below. First, Eq. (3.18) can be rewritten as a function of $f(\Phi,\ n)$ by $ET/P = 1/(1\ \Phi^{-n})^{1/n}$. Next, we obtain $\Phi = f(W, n)$ through the inverse of the Mezentsev' function:

$$\Phi = \left(\frac{1}{\frac{1}{1-W}^{n} - 1}\right)^{\frac{1}{n}}. \tag{3.30}$$

Then by inserting Eq. (3.30) into Eq. (3.29) and differentiating with respect to W yields a term for CCD of the Mezentsev' equation:

$$\alpha_{\mathrm{mez}} = g'(W,\ n) = \left((1 - W)^{2n} - (1 - W)^{n}\right) \cdot \left(\frac{(1 - (1 - W)^{n})^{1-2n}}{(1 - W)^{n}}\right)^{\frac{1}{n}}. \tag{3.31}$$

Last, by substituting W with Eq. (3.28) in Eq. (3.31) a relation of the CCD as function of Φ, n can be obtained.

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4 APPLYING SIMPLE WATER-ENERGY BALANCE FRAMEWORKS 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. (2012) introduced the climate change impact hypothesis (CCUW), 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 (the Budyko approach of Roderick and Farquhar (2011) and the CCUW) with data of more than 400 basins distributed over the

continental United States. We first estimate the sensitivity of streamflow to changes in precipitation using long-term average data of the period 1949 to 2003. This provides a hydro-climatic status of the respective basins as well as their expected proportional effect to changes in climate. Next, we test the ability of both approaches to predict climate impacts on streamflow by splitting the data into two periods. We (i) analyse the long-term average changes in hydro-climatology and (ii) derive a statistical classification of potential climate and basin change impacts based on the significance of observed changes in runoff, precipitation and potential evapotranspiration. Then we (iii) use the different climate sensitivity methods to predict the change in streamflow given the observed changes in water and energy supply and (iv) evaluate the predictions by (v) using the statistical classification scheme and (vi) a conceptual approach to separate the impacts of changes in climate from basin characteristics change on streamflow. This allows us to evaluate the observed changes in streamflow as well as to assess the impact of basin changes on the validity of climate sensitivity approaches.

The apparent increase of streamflow of the majority of basins in the US is dominated by an increase in precipitation. It is further evident that impacts of changes in basin characteristics appear simultaneously with climate changes. There are coherent spatial patterns with catchments where basin changes compensate for climatic changes being dominant in the western and central parts of the US. A hot spot of basin changes leading to excessive runoff is found within the US Midwest. The impact of basin changes on the prediction is large and can be twice as much as the observed change signal. Although the CCUW and the Budyko approach yield similar predictions for most basins, the data of water-limited basins support the Budyko framework rather than the CCUW approach, which is known to be invalid under limiting climatic conditions.

4.1 INTRODUCTION

4.1.1 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 land 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, 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 (Klemeš, 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. (2012), 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.

4.1.2 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 1970s (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 hydrographical features of many basins, especially land-use changes, dam construction and operation, and irrigation; but also changes in forest and agricultural management practices 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).

4.1.3 AIMS AND RESEARCH QUESTIONS

This paper presents an evaluation of two conceptual hypotheses, the newly developed water-energy balance framework of Renner et al. (2012) and the Budyko framework presented by Roderick and Farquhar (2011) 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 light of the rich literature already existing for the hydro-climatic changes observed over the continental US.

4.2 METHODS

4.2.1 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-dimensional 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., 2012):

$$W = 1 - \frac{E_T}{P} = \frac{Q}{P}, U = 1 - \frac{E_T}{R_n/L} = 1 - \frac{E_T}{E_p}. \tag{4.1}$$

Relative excess water W considers the amount of water that 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 instead of R_n/L to describe the energy supply term. 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. The concept carries three interesting cases relevant for streamflow sensitivity to climate and changes in basin characteristics. First, a change in E_T without any changes in climate must evidently be caused by changes in the basin properties. Thus, both U and W change simultaneously. Second, a change in climate without any changes in the basin properties also leads to changes in U and W, but in opposing directions. Taking this further we assume that

$$\Delta U / \Delta W = -1 \tag{4.2}$$

under the presence of climate changes, which we refer to as the climate change impact hypothesis (abbreviated as CCUW). If, however, both climate and basin characteristics change, we assume that the direction of change as seen in the UW space

$$\omega = \arctan \frac{\Delta U}{\Delta W} \tag{4.3}$$

provides a first-order estimate on the relative importance of past climatic and basin change impacts on the hydrological response of river basins.

4.2.2 STREAMFLOW CHANGE PREDICTION BASED ON A COUPLED WATER-ENERGY BALANCE FRAMEWORK

The simplicity of the climate change impact hypothesis (CCUW) allows to derive sensitivity estimates of streamflow to changes in climate. However, there are strong underlying assumptions which limit the potential use of this method (Renner et al., 2012). Ideally, the CCUW hypothesis is only valid for non-limited conditions, i.e. $P \approx E_\mathrm{p}$ and E_T/P sufficiently smaller than 1. This means that, for any application, we have to assume that the CCUW is invariant to climate as well as to the hydrological response (E_T) of a certain basin. These strong assumptions can theoretically lead to conflicts with the physical laws of water and energy conservation. For example, the CCUW may predict that Budyko's water limit is crossed when the aridity index is increasing (Renner et al., 2012).

Taking these assumptions and limitations into account, the following practical relations can be deduced. First, by using the total derivative of the definitions of W and U in Eq. (4.1) and combining with the CCUW hypothesis (Eq. 4.2), the sensitivity coefficient of streamflow to precipitation can be derived (Renner et al., 2012):

$$\varepsilon_{Q,P} = \frac{P}{Q} - \frac{(P-Q)E_\mathrm{p}}{Q(E_\mathrm{p}+P)}. \tag{4.4}$$

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: $\varepsilon_{Q,P} + \varepsilon_{Q,E_\mathrm{p}} = 1$ (Kuhnel et al., 1991).

The CCUW hypothesis may also be used to predict absolute changes. Therefore, consider two long-term average hydro-climate state spaces ((P_0, $E_{\mathrm{p},0}$, Q_0); (P_1, $E_{\mathrm{p},1}$, 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., 2012):

$$Q_1 = \frac{\frac{Q_0}{P_0} - \frac{P_0-Q_0}{E_{\mathrm{p},0}} + \frac{P_1}{E_{\mathrm{p},1}}}{\frac{1}{P_1} + \frac{1}{E_{\mathrm{p},1}}} \tag{4.5}$$

Last, a direct consequence of the CCUW 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 ($E_\mathrm{T}/E_\mathrm{p}$):

$$C_\mathrm{E} = \frac{E_\mathrm{T}}{P} + \frac{E_\mathrm{T}}{E_\mathrm{p}} \tag{4.6}$$

is constant for a given basin. Any changes in C_E, which we denote as catchment efficiency, would then be assigned to a change in basin characteristics.

4.2.3 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), which we refer to as aridity index (E_p/P). There are various functional forms which describe this relation, e.g. Schreiber (1904); Ol'Dekop (1911); Budyko (1948). In this paper, we use a parametric form first described by Mezentsev (1955):

$$E_\mathrm{T} = \frac{E_\mathrm{p} \cdot P}{(P^n + E_\mathrm{p}^n)^{1/n}}; \tag{4.7}$$

see e.g. Roderick and Farquhar (2011) for the history of this equation. 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 of streamflow to climatic changes (P, E_p) and to changes in the basin properties represented by n (Roderick and Farquhar, 2011; Renner et al., 2012). Thereby, by applying the first total derivative of the respective Budyko function and assuming steady state conditions of the water balance with $P = E_T + Q$, absolute changes in streamflow (dQ) can be predicted (Roderick and Farquhar, 2011):

$$\mathrm{d}Q = \left(1 - \frac{\partial E_T}{\partial P}\right)\mathrm{d}P - \frac{\partial E_T}{\partial E_p}\mathrm{d}E_p - \frac{\partial E_T}{\partial n}\mathrm{d}n. \tag{4.8}$$

To compute the change in streamflow given some change in climate, one generally sets d$n = 0$. Note that using Budyko approaches for predicting the effects of a change in climate will also result in a change in C_E. This change is determined by the functional form and the catchment parameter as well as the aridity index of the basin (Renner et al., 2012).

Last, by dividing by the long-term average Q and term expansions, an expression can be obtained which contains the sensitivity coefficients of streamflow to P, E_p and n, respectively (Roderick and Farquhar, 2011):

$$\frac{\mathrm{d}Q}{Q} = \left[\frac{P}{Q}\left(1 - \frac{\partial E_T}{\partial P}\right)\right]\frac{\mathrm{d}P}{P} + \left[\frac{E_p}{Q}\frac{\partial E_T}{\partial E_p}\right]\frac{\mathrm{d}E_p}{E_p} + \left[\frac{n}{Q}\frac{\partial E_T}{\partial n}\right]\frac{\mathrm{d}n}{n}.$$

The sensitivity coefficients, also referred to as elasticity coefficients (Schaake and Liu, 1989), are given within the brackets. For example, a sensitivity coefficient of $\varepsilon_{Q,P} = 2$ means that a relative change in precipitation of 10 % amounts to a twofold change in Q, i.e. 20 %. The partial differentials for the Mezentsev function are listed in the Appendix 4.A.

Mapping the Mezentsev function into UW space reveals that the CCUW approach can be regarded as a special case of the Budyko approach, because both are identical when $P \approx E_p$. However, the theoretical climate change direction of the Mezentsev function (ω_{Mez}) largely depends on the aridity index and on the catchment parameter n, whereas the CCUW assumes the climate change direction to be constant. A mathematical derivation of ω_{Mez} is given in Renner et al. (2012).

4.2.4 STATISTICAL CLASSIFICATION OF POTENTIAL CLIMATE AND BASIN CHANGE IMPACTS

As we are aiming to test the streamflow sensitivity frameworks with historical data, we also need to take other factors of potential streamflow changes into account (Jones, 2011). Most likely are human alterations such as land-use change, change in agricultural management and other factors that influence the hydrological response of river basins. In the following, we will refer to these type of changes as basin changes.

For the retrospective analysis of past changes on river basin level, we need data of the water and energy balance components. Usually, climatic data (P, E_p) and streamflow data are available. For evaluation of potential impacts, the conceptual model of Tomer and Schilling (2009) can be used to separate climate from basin change impacts. Thereby, simultaneous changes in the water and energy balance reflected by ΔU and ΔW are investigated (Renner et al., 2012, Fig. 1).

However, it is also possible to directly investigate the changes in the hydro-climatic data by using statistical tests, e.g. testing for changes in the mean of two periods. Then, the significance results of climatic variables (P, E_p) and hydrological variables (Q) can be combined to construct a data-based classification of likely impacts on streamflow change. Generally, four different hypotheses for changes in these variables can be formulated: first, the null hypothesis of "no change" in any of these variables. And three alternative hypotheses based on significant changes are possible: "climate only", "runoff only" and "climate & runoff". So, we expect that, if climatic changes directly lead to changes in runoff, these are most likely to be found in the "climate & runoff" group. Contrarily, the other alternative hypotheses suggest that some type of basin changes occurred. Given the background signal of increased humidity, the "climate only" hypothesis suggests that there has been some compensation of climatic changes by changes in the properties of the basin. This could be vegetational responses to past disturbances such as succession, but also (natural or anthropogenic forced) adaptations of vegetation to climate changes (Jones, 2011). In contrast, the "runoff only" hypothesis suggests that factors other than the long-term average change in climate lead to changes in streamflow. A similar grouping of basins has been used by Milliman et al. (2008), who defined "normal rivers" which match with the "climate & runoff" group, "deficit rivers" where the signal in climate is much larger than the signal in runoff, which matches with the "climate only" group. And "excess rivers" where the runoff change cannot be explained by climatic changes, which is similar to the "runoff only" hypothesis.

In this paper, we split a large dataset into two periods and test these hypotheses for each basin by evaluating the combination of two-sample t-tests results for P, E_p and Q. This resulted in 9 different groups, which are set as follows: if none of the three t-tests is rejected at a certain significance level α, we define this as "no change", denoted as "–" in the figures and tables. If, for example, P and E_p changed significantly, while Q did not, we denote this group as "P, E_p".

4.3 DATA

The aforementioned approaches 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. The dataset (available at: `ftp://hydrology.nws.noaa.gov/pub/gcip/mopex/US_Data/`) covers 431 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 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 annual E_p is needed, we considered two temperature-based E_p formulations (Hargreaves and Hamon) and one E_p product (CRU TS 3.1) being a modification of the Penman-Monteith method.

The temperature-based formulations are attractive as these allow a computation of E_p from the available data in the MOPEX dataset. The Hargreaves equation (Hargreaves et al., 1985) can be used to estimate daily E_p:

$$E_{p,\text{Hargreaves}} = a \cdot \text{sd}_{\text{pot}} \left((T_{\max} - T_{\min})/2 + b \right) \cdot \sqrt{T_{\max} - T_{\min}}, \tag{4.9}$$

where sd_{pot} is the maximal possible sunshine duration of a given day at given latitude and two empirical parameters ($a = 0.0023$, $b = 17.8$). It has minimal data requirements ($T_{\min}$ and $T_{\max}$), but yields a good agreement with physically based E_p models (Hargreaves and Allen, 2003; Aguilar and Polo, 2011). Potential evapotranspiration by Hamon equation (Hamon, 1963) depends on daily average temperature (T) and daytime length (L_d) only (Lu et al., 2005):

$$E_{p,\text{Hamon}} = \begin{cases} 1.9812 \cdot L_d \cdot \rho_{\text{sat}}(T) \cdot k & \text{if } T > 0°\text{C} \\ 0 & \text{if } T < 0°\text{C} \end{cases} \tag{4.10}$$

Thereby, the saturated vapour density is $\rho_{\text{sat}}(T) = 216.7 \cdot e_{\text{sat}}/(T + 273.3)$ [g m^{-3}], with the saturated vapour pressure being $e_{\text{sat}} = 6.108 \cdot \exp(17.26939 \cdot T/(T + 237.3))$ [mb]. The calibration parameter k was set to 1.2 in accordance with Lu et al. (2005). Both methods have been tested in previous studies, mostly comparing E_p estimates with Penman estimates for selected weather stations, e.g. Amatya et al. (1995). Lu et al. (2005) found, by comparing E_p formulations at the annual time scale for watersheds in the south-east of the US, that the Hargreaves method yields the most extreme estimates, while the Hamon equation showed the most reasonable results under the temperature-based methods.

Radiation-based formulations are more difficult to derive for the domain and the period considered in this paper. However, the Climatic Research Unit (CRU), University of East Anglia, established a globally available gridded dataset (0.5°) of monthly E_p, which is based on the FAO (Food and Agricultural Organization) grass reference evapotranspiration method (Allen et al., 1994). Essentially, these estimates are based on observed and spatially interpolated data (Mitchell and Jones, 2005) of temperature (mean, minimum, maximum), vapour pressure and cloud cover. Here, we used monthly data from `http://www.cgiar-csi.org/component/k2/item/104-cru-ts-31-climate-database` and extracted basin average values (using the R function raster::extract, Hijmans and van Etten, 2012).

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 to 2003 with 430 basin series.

4.4 RESULTS AND DISCUSSION

4.4.1 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, $E_{p,\text{clim}}$, $E_{p,\text{CRU}}$) in Fig. 4.1. The basins with most precipitation are found in the Northwest, the Southeast and along the east coast. The central part of the US receives considerably 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 show the climatological E_p estimates from the evaporation atlas (Farnsworth and Thompson, 1982) and the long-term averages of the CRU TS 3.1 potential evapotranspiration estimates

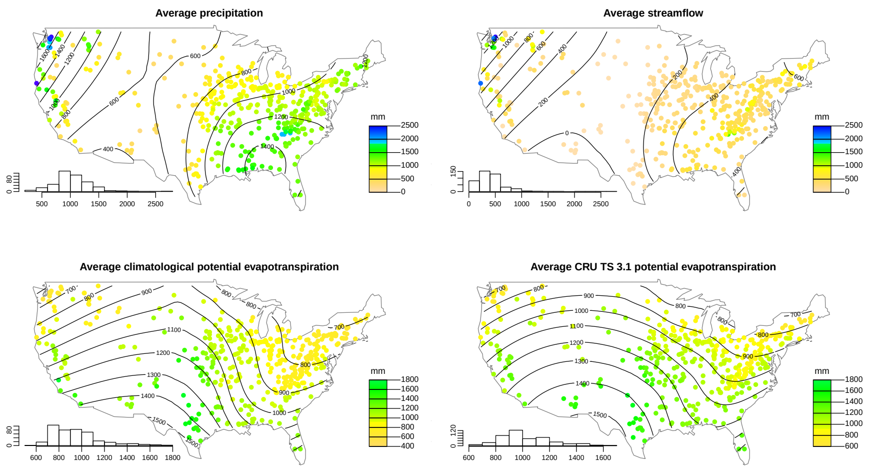

Figure 4.1: Long-term annual average of hydroclimatic variables of the US MOPEX dataset (1949–2003). The contour lines are derived from fitted polynomial surfaces using the R function stats::loess (R Development Core Team, 2011) of the variables using the river gauge locations. The map of the US is taken from the maps package (Becker et al., 2011).

$E_{\mathrm{p,CRU}}$. The long-term basin averages of $E_{\mathrm{p,CRU}}$ show the highest spatial correlation ($r = 0.89$) to $E_{\mathrm{p,clim}}$, while Hamon ($r = 0.57$) and Hargreaves ($r = 0.46$) have lower correlation and somewhat different spatial patterns. Therefore, we selected $E_{\mathrm{p,\,CRU}}$ for further analysis.

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. 4.2, 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 Budyko's function explains 69 % of the variance. The aridity index E_{p}/P of the basins ranges between 0.25 and 5.52, with most basins clustering around 1. The right panel of Fig. 4.2 displays the relationship of the non-dimensional 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. (2012). The hydro-climatic data cover 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 has one negative value (not shown because of the scales used for the axes). This is probably due to an underestimation of $E_{\mathrm{p,\,CRU}}$ for this basin.

4.4.2 CLIMATE SENSITIVITY OF STREAMFLOW

Next, we compare the climate sensitivity coefficients of the CCUW with the Budyko framework using the long-term averages of the MOPEX dataset. In particular, we concentrate on the sensitivity of streamflow to precipitation $\varepsilon_{Q,P}$.

Using the CCUW approach, $\varepsilon_{Q,P;\mathrm{CCUW}}$ is determined by Eq. (4.4), which shows that the coefficient is dependent on the aridity index and the inverse of the runoff ratio. In particular, the correlation of the sensitivity coefficient to the aridity index (correlation $r = 0.53$) is much lower than the correlation to P/Q ($r = 0.99$). This means that, using the CCUW hypothesis, the

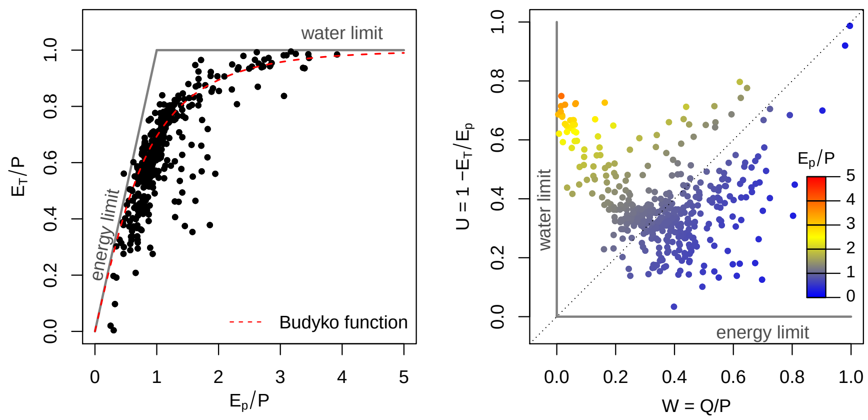

Figure 4.2: Budyko (left) and UW space (right) plots of the period (1949–2003) of the MOPEX dataset. E_p is obtained from the CRU TS 3.1 E_p product. The 1 : 1 line in the UW space diagram separates areas with energy limitation ($E_p/P < 1$) and water limitation ($E_p/P > 1$). Grey lines indicate the water and energy limits.

inverse of the runoff ratio (P/Q) is the main controlling factor in determining runoff sensitivity to climate.

To further illustrate this functional relationship, we plot $\varepsilon_{Q,P}$ in Fig. 4.3 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 in the 55-yr period and display the interquartile range (25 %–75 % percentile range) of all those annual sensitivity coefficients as vertical grey lines. The uncertainty ranges increase with E_T/P. For values of $E_T/P > 0.6$, the ranges get more apparent with about 25 % of $\varepsilon_{Q,P}$, which can be up to the order of $\varepsilon_{Q,P}$ for $E_T/P > 0.8$. 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. 4.3 provides a comparison of the sensitivity estimates of CCUW with the parametric Budyko function approach of Roderick and Farquhar (2011) using the Mezentsev function, with n estimated for each basin separately. The non-parametric Budyko sensitivity approaches are determined by aridity only (Arora, 2002) and have large differences to CCUW, already at medium values of E_T/P (not shown). The parametric Budyko function approach yields similar sensitivities as the CCUW approach for $E_T/P < 0.9$. This is due to the parameter n, which inherently includes some dependency to E_T/P (the correlation of $\varepsilon_{Q,P;\mathrm{Mez}}$ to P/Q is $r = 0.63$). 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 $n = 4$ and the largest sensitivity with $\varepsilon_{Q,P,\mathrm{Mez}} = 4.7$. 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. However, the theoretical assessment of the CCUW hypothesis by Renner et al. (2012) revealed that these large streamflow sensitivity estimates for strongly water-limited basins are probably incorrect, because the CCUW does not obey Budyko's water limit.

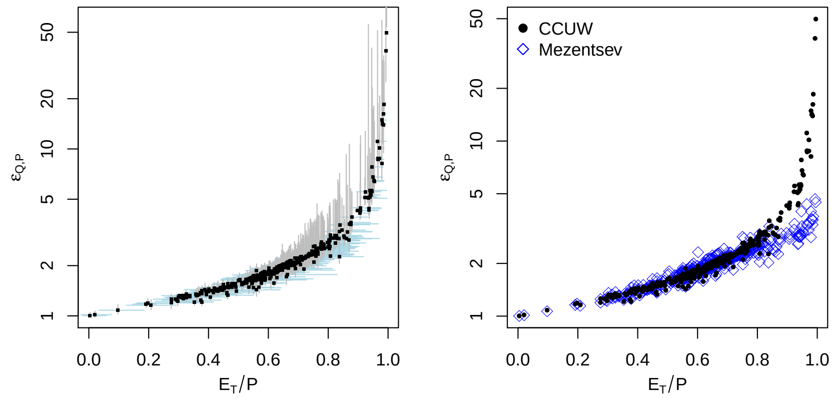

Figure 4.3: Sensitivity coefficients of streamflow to precipitation as function of E_T/P. Left panel: $\varepsilon_{Q,P;\mathrm{CCUW}}$ computed for the CCUW method. Dots represent $\varepsilon_{Q,P;\mathrm{CCUW}}$ using long-term average data of the respective basin. Vertical grey lines depict the interquartile range of $\varepsilon_{Q,P;\mathrm{CCUW}}$ estimated for each year in the record, while light blue 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.

4.4.3 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).

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 32 % of the basins ($\alpha = 0.05$, Welch two-sample t-test with unknown variance, using the function stats::t.test in R (R Development Core Team, 2011)). The top left map in Fig. 4.4 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).

The assessment of changes in potential evapotranspiration necessarily depends on the

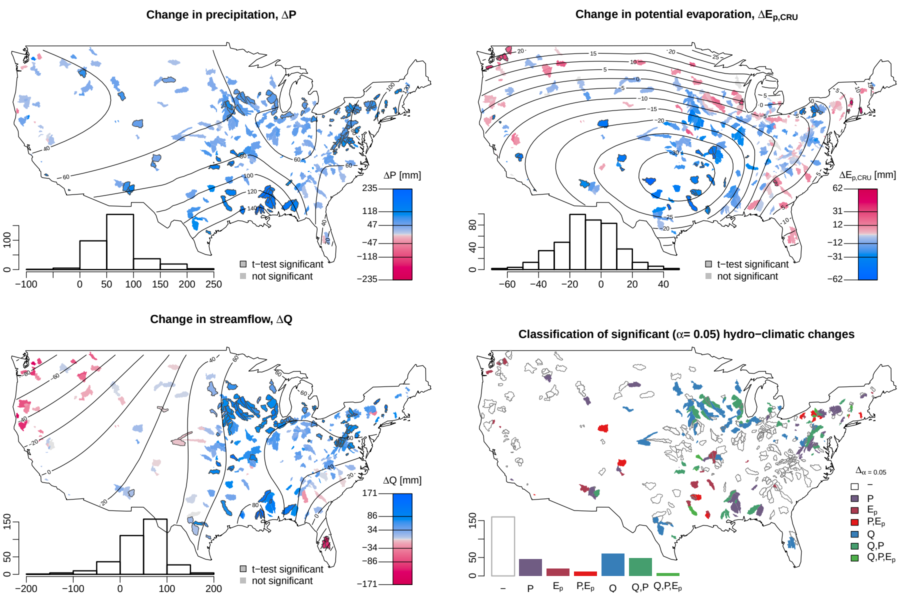

Figure 4.4: Maps of absolute change in hydro-climatic variables of the MOPEX dataset, comparing changes between the periods 1949–1970 and 1971–2003. Annual changes are given in millimeter. Significant changes in the mean of both periods are tested by univariate two-sample t-tests with $\alpha = 0.05$ and are denoted by a grey borderline. For each variable, a histogram of the changes is given in the lower left corner. The map in panel **(d)** provides a classification based on the univariate t-tests.

method of choice and thus the input data, which can influence magnitude and even the sign of these trends (Donohue et al., 2010). A summary of changes for each E_p method is given in Table 4.1. In general, there is a negative trend at more than 50 % of the basins. The Hargreaves method yields the strongest trends and shows a much larger number of basins with a significant change in E_p and, to a lesser degree, also in the change of the aridity index. The correlation matrix of average changes in several variables (given in Table 4.2) shows that this trend in Hargreaves E_p is directly related to a decrease in the diurnal temperature range ΔTR, which has also been reported by Lettenmaier et al. (1994). The E_p changes by the Hamon and CRU product are smaller and less significant. Changes in the Hamon equation are directly and positively related to changes in average temperature ΔT (cf. Table 4.2). Changes in the $E_\mathrm{p,CRU}$ product are positively related to changes, both in average temperature and diurnal temperature range (Table 4.2). This finding further supports the usage of the CRU E_p dataset. The top right map of Fig. 4.4 shows that negative significant changes in average E_p are common in the southern central parts (about -30 mm) and a few patches throughout the US.

Both the increase in precipitation and the decrease in potential evapotranspiration should ideally lead to an increase in annual streamflow. This is supported by the strong positive correlation with precipitation changes and the negative correlation coefficients with the E_p changes (Table 4.2). Further, we find that 32 % of the basins show a significant increase. The map in the bottom left panel of Fig. 4.4 shows that basins with significant increases in streamflow are predominantly found within the Upper Mississippi River basin and the northern

Table 4.1: Statistics of the average change of the three E_p estimates. The first three columns depict quantiles of ΔE_p; the forth and fifth columns denote the relative frequency of basins with significant change ($\alpha = 0.05$, two sample t-test) for E_p and the aridity index (AR).

	10 % [mm]	50 % [mm]	90 % [mm]	$\Delta E_p \leq \alpha$ [%]	$\Delta AR \leq \alpha$ [%]
CRU	−32	−8	13	13	19
Hargreaves	−41	−23	−6	69	40
Hamon	−16	−6	9	6	26

Table 4.2: Pearson correlation coefficients for the average change between the two periods, assessed for all basins with data available. Significance of correlation is denoted with letters ([a] 0.001, [b] 0.01, [c] 0.05), with significant correlations ($\alpha < 0.05$) set in bold for visual aid.

	ΔQ	ΔP	$\Delta E_{P,\mathrm{CRU}}$	$\Delta E_{P,\mathrm{HAR}}$	$\Delta E_{P,\mathrm{HAM}}$	ΔT
ΔQ						
ΔP	**0.57**[a]					
$\Delta E_{P,\mathrm{CRU}}$	−**0.18**[a]	−**0.17**[b]				
$\Delta E_{P,\mathrm{HAR}}$	−**0.13**[c]	−0.01	0.08			
$\Delta E_{P,\mathrm{HAM}}$	−**0.23**[a]	−**0.23**[a]	**0.30**[a]	−0.04		
ΔT	−**0.22**[a]	−**0.19**[a]	**0.24**[a]	0.01	**0.96**[a]	
ΔTR	−**0.18**[a]	−0.01	**0.13**[b]	**0.99**[a]	−0.02	0.02

Appalachian Mountains and a few basins on 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 only use basins for further analysis, which have more than 10 yr of data in any of the two periods and that we removed 2 basins, because the water balance was suspect ($Q > P$). So in total 351 basins are kept for further analysis.

EVALUATION OF STREAMFLOW CHANGE PREDICTIONS

In the previous subsection, we described the changes observed in precipitation, potential evapotranspiration and streamflow by 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. 4.5) and the Budyko approach illustrated by Roderick and Farquhar (2011). For the Budyko approach, we use Eq. (4.8) and the functional form of Mezentsev (1955). In particular, we use the hydro-climatic state of the first period, described by P_0, $E_{p,0}$, Q_0, as well as the climatic states of the second period P_1, $E_{p,1}$ 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_{clim} signals.

A scatterplot of predicted versus observed changes is shown in the left panel of Fig. 4.5, 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

Table 4.3: Group average statistics of hydro-climatic changes for 4 groups of basins ("no change", "climate only", "runoff only" and "climate & runoff"), classified by results of a combination of two-sample t-test results grouping basins with significant change at $\alpha = 0.05$. For the hydro-climatic changes, also the group standard deviation is given. For the normalised basin changes, the first and third quartiles are given. In total 351 basins have been tested.

	Unit	No change	Climate only	Climate & runoff	Runoff only
N	–	160	75	55	61
E_p/P	–	1.15	1.22	1.10	1.28
ΔP	mm	50 ± 24	97 ± 43	113 ± 42	61 ± 21
ΔE_P	mm	-8 ± 14	-13 ± 23	-8 ± 18	-7 ± 12
ΔQ	mm	23 ± 41	44 ± 45	95 ± 31	67 ± 43
ω_{obs}	°	300	286	337	2
ΔC_E	–	0.02	0.05	−0.04	−0.06
Δn	–	0.05	0.15	−0.12	−0.15
$\mathrm{RMSE}_{\mathrm{CCUW}}$	mm	37.98	49.22	31.89	43.94
$\mathrm{RMSE}_{\mathrm{Mez}}$	mm	37.70	48.18	34.57	46.22
$\Delta Q_{\mathrm{basin,CCUW}}/P$	–	−0.03–0.01	−0.04–0	−0.01–0.05	0.02–0.06
$\Delta Q_{\mathrm{basin,Mez}}/P$	–	−0.02–0.01	−0.03–0	0–0.05	0.02–0.06
$\Delta Q_{\mathrm{basin,CCUW}}/Q$	–	−0.08–0.03	−0.12–−0.01	−0.01–0.19	0.05–0.27
$\Delta Q_{\mathrm{basin,\ Mez}}/Q$	–	−0.07–0.05	−0.1–0	0–0.24	0.06–0.29

other. Based on all basins ($N = 351$), both methods yield similar differences compared with the observed change in streamflow ($\mathrm{RMSE}_{\mathrm{CCUW}} = 40.9$ mm, $\mathrm{RMSE}_{\mathrm{Mez}} = 41.3$ mm). A direct comparison is shown as scatterplot in the right panel of Fig. 4.5. The graph indicates that there is a general agreement between both estimates ($r = 0.99$). 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 approach, which was already discussed above. These changes are small in absolute values, but quite large when seen relative to the annual totals of streamflow.

SEPARATING THE INFLUENCE OF CLIMATE AND LAND-USE IMPACTS ON STREAMFLOW

From the maps in Fig. 4.4, it is apparent that basins with significant changes in streamflow do not necessarily match with those having significant changes in the climatic variables (P, E_p). Such inconsistency between climatic and streamflow trends was also reported in previous literature such as in Lettenmaier et al. (1994).

For further analysis, we combined the results of the univariate t-tests ($\alpha = 0.05$), which resulted in 9 different classes. These are further aggregated to the four different hypotheses on streamflow change elaborated in Sect. 4.2.4. In Table 4.3, we provide summary statistics for each class. The map in Fig. 4.4d shows the location of the groups in the US, with a bar plot in the lower left corner showing the counts of each group. For most basins (46 %), we found no significant change in any of the three observed variables. The group of basins where only streamflow changed significantly while climatic variables show insignificant changes is large and consists of 17 % of all basins. These are mostly found in the central north of the US,

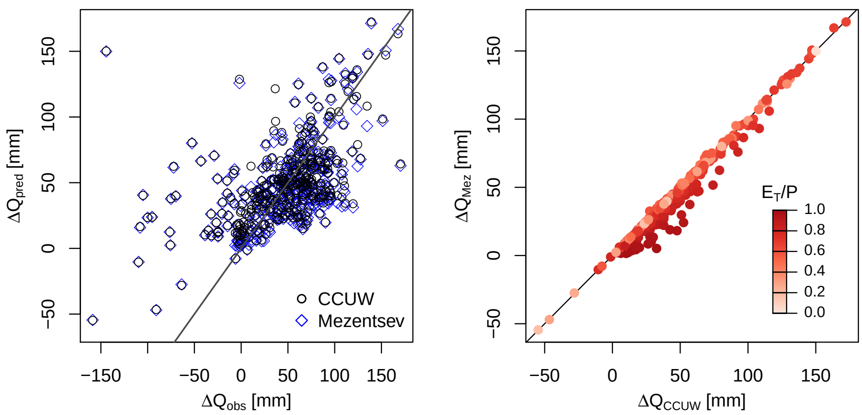

Figure 4.5: Left: Scatterplot of observed vs. predicted annual average changes in streamflow for MOPEX dataset without stations with missing data. The vertical difference to the 1 : 1 line depicts the deviation of the prediction to the observed value. Right: predicted change in streamflow due to climatic changes, comparing the estimates of the Budyko framework with the CCUW estimates. The colour of the dots represents the evaporation ratio E_T/P.

west of the Great Lakes. In the other extreme, there are basins, where significant climatic changes occurred, while streamflow did not change significantly. Combining these classes to the "climate only" group, 21 % of the basins are affected. For this group, reddish colours have been used in the map in Fig. 4.4d. This group is dominant in the west and shows some clusters in the South- and Northeast. Coloured in shades of green, the smallest groups are those where at least Q and P changed significantly. Adding up these groups to the "climate & runoff" change group comprises 16 % of the basins.

The differences between observed and predicted streamflow changes may be due to model deficiencies or input data uncertainty only. In this case, we would expect that the differences are distributed randomly in the set of basins. However, if we take basin changes as alternative hypotheses into account ("climate only", "runoff only"), we would expect that the differences are not random, but carry typical signals of basin change impacts being different from zero.

To investigate this, we analysed the differences normalised by annual average precipitation for the classes of basins determined by the combined t-tests. Results are shown for the four main classes in Table 4.3 and in the boxplot in Fig. 4.6. In the "no change" group, we find a large scatter with the median close to 0 and the interquartile ranges below and above 0, indicating that there is no general trend in the model differences. This behaviour is expected because there are no large and significant changes in the hydro-climate of these basins. The other basins are more interesting. The group of basins where we found significant "climate only" changes shows that most of the basins in this group are below 0. For this group, the Budyko framework has an average difference of −2.1 % and CCUW −2.7 % of the annual water balance. This means that basin changes compensate for the detected climatic changes (with a group average decrease in aridity of −10.2 %). Further analysis shows that E_T strongly increased with 6.2 % of the annual water balance. Also the catchment parameter n and C_E show significant increases (cf. Table 4.3). In contrast, the "runoff only" group shows significant positive basin change impacts (Budyko 3.8 % and CCUW 3.2 % of the annual water balance). In these basins, we find predominant increases in streamflow, along with significantly decreasing catchment parameters. This indicates that changes in the basin properties took place, which led to predominant runoff increases (7.7 % of the water balance) on similar magnitude of the group average precipitation increase (7.3 %). Thus, on average, the increase in precipitation did not increase E_T (-0.4 % of the water balance).

The "climate & runoff" change group reveals smaller errors; however, most of these tend to be influenced by basin changes with positive differences. The map in Fig. 4.4d displaying the location of the groups shows that many of these basins are actually close to the "runoff only" group and so we expect that basin changes are quite likely.

The ecohydrological framework of Tomer and Schilling (2009) is based on analysing changes in the relative partitioning of the surface water and energy fluxes. In Fig. 4.7, we plot the observed changes, i.e. ΔU vs. ΔW, using data of all MOPEX basins. From the figure, it becomes apparent that most of the basins shifted towards the right of the positive diagonal, which is an effect of the general trend of increasing humidity (increasing P and widely decreasing E_p) over the US. The differences of the predicted changes to the observed changes in streamflow are depicted by the size of the dots and the colour palette. Generally, the smallest deviations are found in the lower right quadrant, which represents the climate impact change direction of the ecohydrological concept. Towards the upper right quadrant, we find that basin impacts are increasing, leading to an excess of streamflow, while towards the lower left quadrant basin impacts show compensating effects leading to streamflow deficits.

In the right panel of Fig. 4.7, we use the plotting characters corresponding to the t-test classification groups. Most basins in the "runoff only" group are in the upper right quadrant, while the "climate only" group is concentrated in the lower two quadrants and predominant ΔU increases. So, although the concept of Tomer and Schilling has certain limitations such as the dependency to the aridity index and the hydrological response (Renner et al., 2012), it is generally able to separate the basin and climate impacts on E_T and streamflow.

In summary, the analysis shows that the differences $\Delta Q_{obs} - \Delta Q_{clim}$ are unlikely to be random and due to model deficiencies, but rather reveal distinctive impacts of basin changes under the general trend of increasing humidity. Further, frequency and impacts of basin changes are large and evidently much larger than the differences between both frameworks.

Change direction in UW space

For further analysis, we concentrate on the direction of change in the UW space ω, introduced with Eq. (4.3), which approximately yields a measure of the relative impact of both climatic and basin changes. Graphically, ω represents the angle between the positive x-axis and some point in a ΔU vs. ΔW plot, such as Fig. 4.7. As a reference, we computed the theoretical change direction of climatic changes using the Budyko framework with the Mezentsev curve being dependent on the aridity index and the catchment parameter n. In the scatter plots of Fig. 4.8, we plot the observed change direction ω_{obs} as a function of the theoretical climate change direction ω_{Mez}. If we assume that there are only changes in climate which impact streamflow, we would find all points at the $1:1$ line. We also show the climate change direction of the CCUW hypothesis, which is constant at 315°. Any deviations from these lines indicate the concurrence of basin changes, assuming the models and input data are correct. The size and colour of the dots correspond to the magnitude of normalised difference to the observed change in runoff. We find that there is a clear relation between ω_{obs} and the normalised difference,

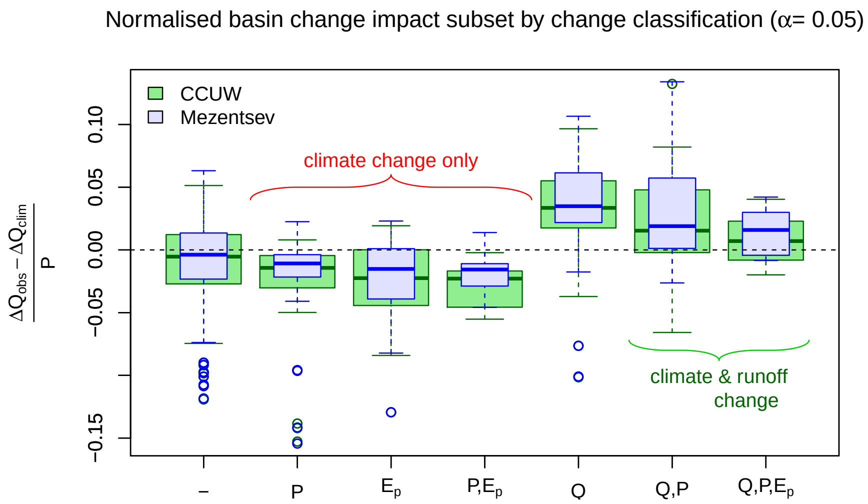

Figure 4.6: Boxplot of the difference of observed and predicted annual average changes in streamflow normalised by annual average of precipitation of the first period. Subsets are in accordance with the t-test classification.

which is positive above the 1 : 1 line and negative below. The largest differences and thus impacts occur at $\omega_{obs} \approx 225°$ when both n and C_E increase strongly, whereas at $\omega_{obs} \approx 45°$, n and C_E decrease strongly. This is confirmed by the scatterplot in the right panel, where the plotting character corresponds to the statistical classification of the basins. Most "climate only" basins are below the 1 : 1 line, while "runoff only" basins are found mainly above. Also note that the "climate & runoff" group has quite a few basins far above the 1 : 1 line.

The combination with the independent classification shows that in general both frameworks seem to be valid for predicting climate change impacts and separating them from basin change impacts. Also the differences between both approaches are generally relatively small. However, very interesting is the performance under limiting conditions, where larger differences must become apparent. Unfortunately, the MOPEX dataset has not too many arid or humid basins and inferences are rather limited. In the left panel of Fig. 4.8, we also depict isolines of the aridity index of the respective basins, where arid basins have a lower ω_{Mez} than more humid ones. We see that arid basins with significant changes follow the 1 : 1 line, rather than the $\omega_{CCUW} = 315°$ line. This supports the validity of the Budyko framework and suggests that the CCUW is not valid under arid conditions.

The theoretical climate change direction reflecting the aridity index and the catchment parameter is mapped in the left panel of Fig. 4.9. This reveals how the actual hydro-climatic conditions in the US modify relative changes in the partitioning of water and energy fluxes at the surface. Most basins have no water or energy limitation (aridity close to 1), and a climate change would equally alter the relative partitioning of water and energy fluxes (i.e. $\Delta U = -\Delta W \rightarrow \omega \approx 135, 315°$), which is the assumption of the concept of Tomer and Schilling (2009) and the CCUW hypothesis. The more arid climate in the central US, however, results in much larger relative changes of the partitioning of energy fluxes than in the water fluxes ($|\Delta U| > |\Delta W|$). This means that an increase in precipitation would decrease the normalised sensible heat flux much

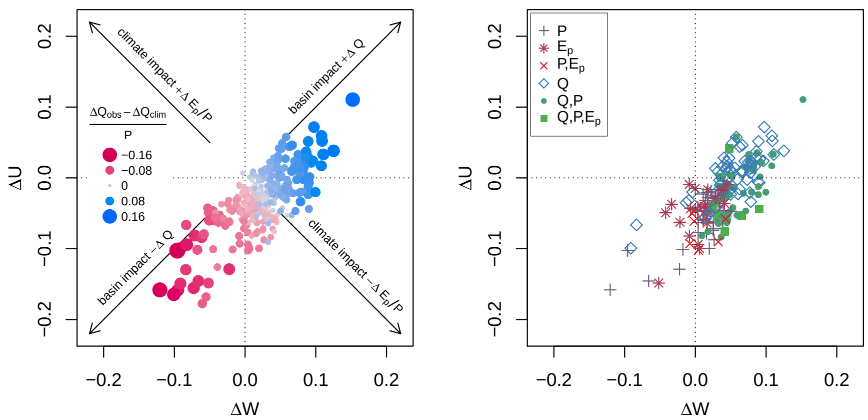

Figure 4.7: Observed changes in UW space between the two periods. The arrows represent the conceptual framework of Tomer and Schilling (2009) to separate climate and basin change impacts. Changes falling approximately below the positive diagonal are related to a decrease in aridity. Under a general trend of decreasing aridity, basin changes leading to an increase of the runoff ratio are approximately above the negative diagonal, while basin changes compensating for climatic impacts are below. In the left panel, the size and colour of the dots depict the normalised difference between observed and climatic related streamflow change. The right panel is restricted to basins with significant changes only, using the t-test classification scheme. Note that for displaying reasons we do not show the change for the Snoqualmie River near Carnation, Wash. In this northwestern river, streamflow dropped strongly, while precipitation increased slightly, which resulted in large changes in $\Delta W = -0.12$ and $\Delta U = -0.38$.

more than the runoff ratio would increase.

The mapping of ω_{obs} in the right panel of Fig. 4.9 provides a quick overview on climatic and basin change impacts. If we consider a segment of 45° centred at ω_{Mez}, this would reflect roughly constant n and valid conditions for the Budyko framework. About 29 % of the basins are actually within this boundary. According to the map in Fig. 4.9, these basins are mainly found in the southern central part of the US, along a band following the Appalachian Mountains, and a few single basins in the west. Basins with distinct climate impacts and compensating basin effects with increasing n and C_{E} ($\omega_{\mathrm{obs}} - \omega_{\mathrm{Mez}} < -22.5°$) are also quite frequent (32 %) 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 the right panel of Fig. 4.9, we see a transition of changes in ω_{obs} over the Mississippi River basin. While the western part shows $\omega_{\mathrm{obs}} < \omega_{\mathrm{Mez}}$, there is a strong transition towards the Midwest, where we find a large cluster of basins with $\omega_{\mathrm{obs}} > \omega_{\mathrm{Mez}}$. This transition may be primarily linked to the precipitation changes, which also show a west to east gradient (cf. map in Fig. 4.4). But agricultural cultivation, especially in basins of the US Midwest, may have

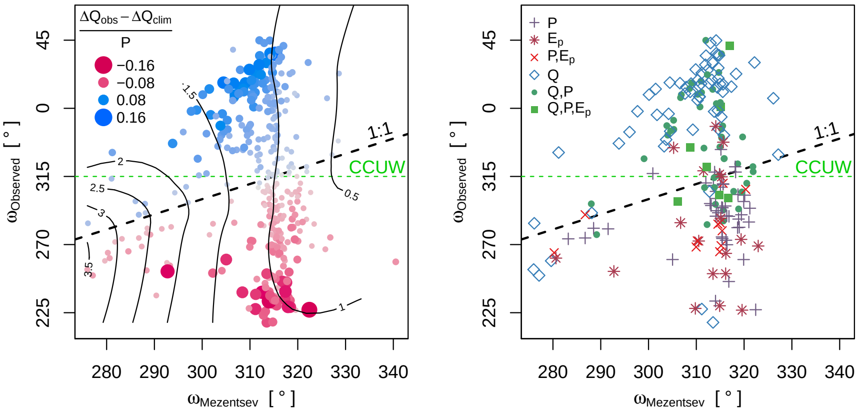

Figure 4.8: Observed change direction in UW space $\omega = \arctan(\Delta U/\Delta W)$ as a function of the theoretical change direction according to the Mezentsev function for all 351 MOPEX basins. In the left panel, the size and colour of the dots depict the normalised difference between observed and climatic related streamflow change. The right panel is restricted to basins with significant changes only, using the t-test classification scheme.

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 increase streamflow on average, but also led 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 ω_{obs} are spatially more heterogeneous. This is probably because topography and land use are more diverse compared to the west. However, it is important to note that the density of river gauge records is much larger. The types of impacts are almost equally frequent, but as the maps of hydro-climatic changes already show, significant changes are rather concentrated in the north and south.

4.4.4 Uncertainty discussion

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; 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 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 two temperature-based E_{p} models: the Hargreaves equation being based on diurnal temperature ranges; and the Hamon equation, which is based on daily average temperatures.

The results show that there are large differences on the long-term average as well as for the detected trends over time. So for example, we found that the changes in E_p derived with the Hargreaves equations are uncorrelated to the changes estimated by the Hamon equation or the E_p time series product of CRU (see Table 4.2). This is in accordance with previous studies on potential evapotranspiration models for hydrological applications. As, e.g., Donohue et al. (2010) note, the reliability of E_p estimates can be improved by adding more relevant input variables. Therefore, we used the E_p time series product of CRU, which includes humidity and cloudiness information. We find that this dataset is more consistent with respect to the long-term average and its spatial distribution as well as the temporal trends.

Still, there are certainly other reasons for the change in evaporative demand which are not reflected in the CRU E_p dataset – 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. This is because the observed changes in the partitioning of water and surface fluxes can be attributed to a much larger part to the change in precipitation.

UNCERTAINTIES DUE TO INHERENT ASSUMPTIONS

While introducing the theoretical framework by Renner et al. (2012) and the Budyko framework, considerable assumptions have been made that lead to uncertainties. 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 identify periods for computing long-term averages. However, this relatively subjective method may also introduce other problems. Secondly, we assume steady state conditions of the water and energy balances. 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 with impacts on evapotranspiration and thus streamflow, which can be larger than impacts of climatic variations. So we found that both catchment parameters (n, C_E) expressing the ability for evaporation have 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 each other. In the companion paper, Renner et al. (2012), we discussed the different assumptions on catchment efficiency and climate changes. While the Budyko functions inherently assume that C_E is changing with the aridity index, the CCUW method assumes C_E 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 characteristic change. But we found that both frameworks yield comparable results under non-limited conditions, and both are generally able to discern climatic and different basin change impacts on streamflow. However, data of the few basins in arid conditions suggest that the CCUW sensitivity framework is unreliable under these generally water- limited conditions.

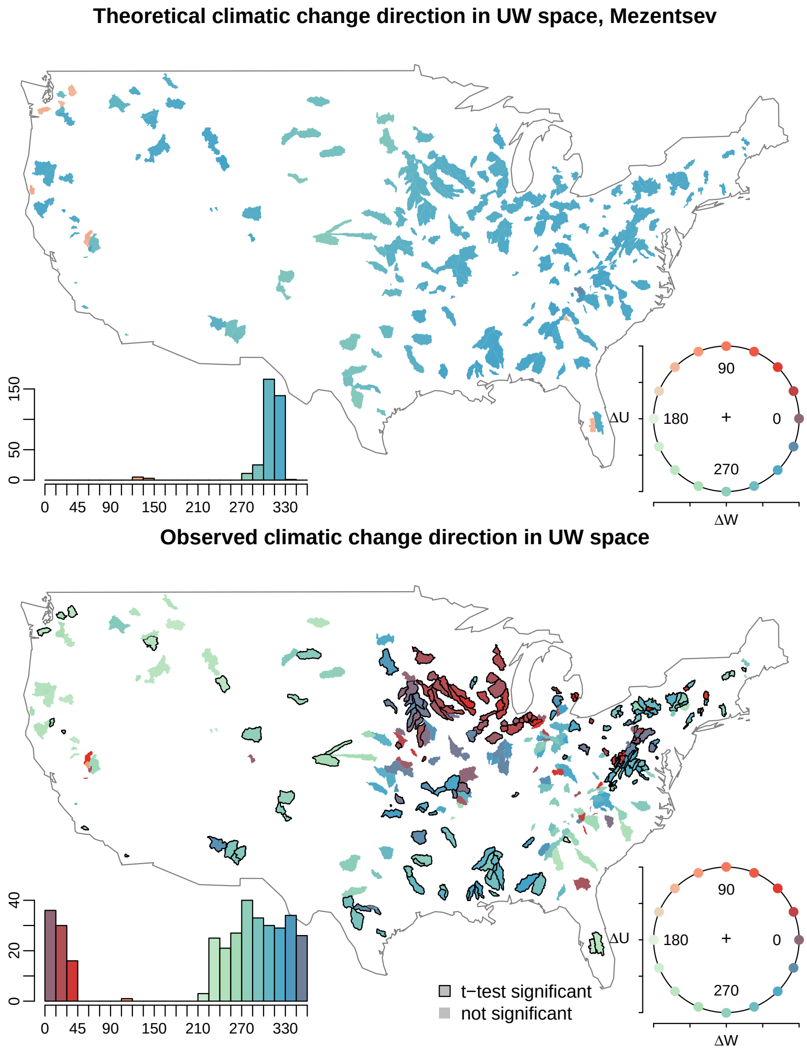

Figure 4.9: Mapping of the change direction ω in UW space. The colour of the polygons indicates the value of ω with the corresponding wheel legend in the bottom right. The upper panel shows the theoretical climatic change direction using the Mezentsev function. The lower panel shows the observed change direction using Eq. (4.3). Polygons with a grey border indicate significant (t-test, α =0.05) changes in any of the observed variables (P, $E_{\mathrm{p,CRU}}$, Q).

4.5 CONCLUSIONS

This paper presents an application and examination of two water-energy balance frameworks for the problem of estimating the sensitivity of streamflow to changes in long-term average precipitation and evaporative demand. In particular, we test and compare the CCUW framework with the Budyko framework by employing 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 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. This fundamental difference results in sizeable 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, on the basis of observed variations in the climate of the second part of the 20th century. The combination with the conceptual framework of Tomer and Schilling (2009) and the statistical classification to discern climate from basin characteristic changes yields comprehensive insights into the hydro-climatic changes in the US. We can reinstate that increased annual precipitation leads 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 ω, 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 increased global greenhouse gas emissions (Thomson, 1995; Barnett et al., 2008) and intensified agricultural land use, especially by artificial drainage and irrigation. The results suggest that the direction and magnitude of human impacts distinctly vary with climate, soil, land-use and hydrographic conditions.

Still, changes in basin characteristics and uncertainties, which are essentially attributed to basin characteristic changes, might have had trends in the past but cannot be extrapolated to the future. However, these impacts play a role and one needs to consider such changes when applying any kind of climate sensitivity framework.

4.A MATHEMATICAL DERIVATIONS FOR THE MEZENTSEV FUNCTION

The first-order perturbation of the Mezentsev function in Eq. (4.8) provides analytical solutions for the problem of streamflow sensitivity. Here, the respective partial differentials are given (Roderick and Farquhar, 2011):

$$\frac{\partial E_{\mathrm{T}}}{\partial P} = \frac{E_{\mathrm{T}}}{P}\left(\frac{E_{\mathrm{p}}^{n}}{P^{n}+E_{\mathrm{p}}^{n}}\right) \tag{4.11}$$

$$\frac{\partial E_{\mathrm{T}}}{\partial E_{\mathrm{p}}} = \frac{E_{\mathrm{T}}}{E_{\mathrm{p}}}\left(\frac{P^{n}}{P^{n}+E_{\mathrm{p}}^{n}}\right) \tag{4.12}$$

$$\frac{\partial E_{\mathrm{T}}}{\partial n} = \frac{E_{\mathrm{T}}}{n}\left(\frac{\ln(P^{n}+E_{\mathrm{p}}^{n})}{n} - \frac{(P^{n}\ \ln(P)+E_{\mathrm{p}}^{n}\ln(E_{\mathrm{p}}))}{P^{n}+E_{\mathrm{p}}^{n}}\right) \tag{4.13}$$

To compute the sensitivity of streamflow to precipitation, we insert Eq. (4.11) into the first bracketed term of Eq. (4.9). Then, by applying the water balance $Q = P - E_{\mathrm{T}}$ and substituting E_{T} by Eq. (4.11), an analytical solution is obtained:

$$\varepsilon_{Q,P;\mathrm{Mez}} = \frac{P}{Q}\left(1-\frac{\partial E_{\mathrm{T}}}{\partial P}\right) \tag{4.14}$$

$$\varepsilon_{Q,P;\mathrm{Mez}} = 1+\frac{E_{\mathrm{p}}\left(1-\frac{E_{\mathrm{p}}^{n}}{P^{n}+E_{\mathrm{p}}^{n}}\right)}{\left(P^{n}+E_{\mathrm{p}}^{n}\right)^{-n}-E_{\mathrm{P}}}. \tag{4.15}$$

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5 SUMMARY AND CONCLUSIONS

5.1 SHIFTING SEASONS IN HYDROLOGY

The seasonal cycle of river regimes is investigated with emphasis on the timing of streamflow and potential links to climatic changes. Thereby, a hydro-climatic data set from 1930-2009 has been prepared for 27 river basins throughout Saxony, Germany. Next, answers to the posed research questions are provided and conclusions are drawn from the study presented in chapter 2.

5.1.1 MAJOR FINDINGS

1. *How to estimate the seasonal timing of river flow?*

The most suitable approach for estimating the timing of runoff was found to be the approach of Stine et al. (2009) which fits a harmonic function for each year of data. This relatively simple approach describes the seasonality of a time series by the phase, to which the author refers to as timing, and the amplitude. By determining the quality of the harmonic fit, this approach also allows to determine the accuracy of the timing measure. However, the method a-priori implies that the underlying seasonal signal has a harmonic nature and different forms would result in poorer fits and estimations. A promising direction of improvement may be more sophisticated time series decomposition methods such as empirical mode decomposition (EMD) which was e.g. applied for temperature by Vecchio et al. (2010) and (Qian et al., 2009, 2011).

2. *Can we assume stationarity for the seasonality of hydrological records?*

While the timing of hydrological records can vary from year to year, stationarity would require that there are no trends or abrupt changes in the data. Stationarity of the timing, being a circular variable, has been tested by cumulative sums of anomalies. Thereby, a new method for circular variables embedded in a generalised fluctuation test framework (Zeileis and Hornik, 2007) has been proposed, cf. section 2.2.4. In most lower altitude basins of Saxony the variability of the timing is large and also no significant changes have been detected. In higher basins with considerable snow storage the amplitude of the seasonal signal increased. Further, nonstationary signals, including trend reversals, have been detected for the highest basins in

Saxony, see also Figures 2.6 and 2.10. Thus, assuming stationarity would bias any seasonal statistic of streamflow and may mislead future water resources management decisions.

3. *What are the physical processes and where to expect changes in the timing of hydrological records?*

Seasonal changes, in particular changes in the timing of river regimes, can be linked to changes in the phase of air temperature. This link is particularly strong in areas where the timing of the freezing point of water is impacted by a seasonal change of temperature. In other words, the timing of the main annual snow melt is affected, which in turn dominates the respective river regime. This link has been found to increase with the altitude of the basin (cf. Table 2.3), resulting in an amplified effect on the timing of regimes in low mountain ranges (cf. Fig. 2.9).

5.1.2 SOCIO-ECONOMIC AND POLITICAL RELEVANCE

The recent sequence of high temperature records in spring time, in connection with spring and early summer time droughts in Central Europe showed to have large impacts on agriculture and vegetation. Thus, assuming that the detected trend towards earlier timing of temperature is likely to persist in the future, would imply a chain of effects with serious consequences, also for water management: higher temperatures during late winter and spring, triggering earlier snow melt, earlier runoff peak flow and less runoff from head water catchments during warm summer months. Thus, this trend increases the vulnerability of water supply if supplies are mainly drawn from fresh water of low mountain ranges. So, this would require larger dams. However, snow storage in a watershed can be of similar or even larger magnitude than existing artificial water storages (Mote et al., 2005).

5.1.3 LIMITATIONS AND POSSIBLE DIRECTIONS FOR FURTHER RESEARCH

Further research is needed to understand the impacts of the advance in timing of the temperature signal on snow, soil moisture, growing season length and runoff dynamics during and after snow melt. Particularly, interesting are the different dynamics with respect to the timing of temperature. So, temperature largely determines snow melt and growing season, while the distinct seasonal course of solar radiation, determining evaporative energy, has a rather constant timing. Hence, changes in the average temperature as well as in the timing of temperature result in quite dynamic consequences. Thus, statistical approaches are limited in this respect and physically based models are recommended.

Although the shift in the timing of temperature is a very robust signal, there is yet no complete physical understanding of the causes of these shifts. Consequently, climate models are currently not able to model the past changes in the timing of global average temperature (Stine et al., 2009). Thus, a better understanding of the variability of timing of the annual cycle of temperature could improve the prediction of impacts (Stine and Huybers, 2012) as well as seasonal forecasts of river flow, given the high correlations to the timing of river flow in low mountain ranges.

5.2 LONG-TERM ANNUAL CHANGES IN E_T AND STREAMFLOW

The papers presented in chapters 3 and 4 establish and test water-energy balance frameworks to separate the impacts of climate from basin changes as well as to predict impacts of climate change on the hydroclimatology of river catchments. In the following the main findings are summarised and put into the scope of the thesis.

5.2.1 MAJOR FINDINGS

4. *How to conceptualise the processes of water and energy partitioning of catchments at the hydroclimatic time scale?*

The derived conceptualisation is based on the hydro-climatic equilibrium being described by the simplified water-energy balance equations and neglecting interannual changes in storage of water and energy. This conceptualisation is in accordance with the Budyko Hypothesis, which states that evapotranspiration is limited by the supply of water P and energy E_p. Further, E_T is modified by catchment properties, such as soil water storage, topography, vegetation and other factors. This feature has led to the inclusion of a catchment parameter n into the classic Budyko curves (Bagrov, 1953; Mezentsev, 1955; Choudhury, 1999), writing $E_T = f(P, E_p, n)$. More generally, the catchment parameter reflects the implicit nature of E_T, which can be written as $E_T = f(P, E_p, E_T)$ (Eagleson, 1978; Yang et al., 2008). This step essentially summarises the complexity of multiple interacting processes at the long term annual average scale into a catchment parameter, or more generally the catchment response E_T.

The functional form allows a straight-forward definition and distinction of the types of externally driven changes on E_T. Thus, climate changes are defined as changes in the climatic forcing variables P, E_p. All other changes are defined as basin changes. For the parametric form of the Budyko curve this reflects a change in the catchment parameter n. Or more generally, this means that the functional form of the implicit nature of E_T is being changed. Thus, this broad definition of basin change includes any types of LULC changes, (soil) hydraulic changes, but also any kind of subscale climate changes. The problem of subgrid (subscale) variability will be addressed in the outlook section.

5. *How to identify and distinguish impacts of climate and land-use change from hydro-climate records?*

The required variables for the model are already given by the functional relationship $E_T = f(P, E_p, E_T)$, where catchment E_T can be estimated through water balance closure. A further step is the normalisation of the water-energy balance equations by P and E_p, respectively. This step can be justified by the Budyko Hypothesis and allows to compare water with energy partitioning. This normalisation is referred to as UW space following Milne et al. (2002) and Tomer and Schilling (2009). Figure 3.2 shows an example of the UW space. The separation of past changes is then based on the analysis of changes in the relative partitioning of water and energy. Tomer and Schilling (2009) proposed that the directions of change in the UW diagram can be used to distinguish climate and LULC changes. In chapter 3 it is shown that their model, depicted in Figure 3.1, is valid only for non-limiting climate conditions, i.e. $P = E_p$ and a correction for limiting conditions is necessary to account for conservation of water and energy, which is elaborated in chapter 3.3.1.

The direction of a basin change in the UW diagram can be directly drawn from the definitions of relative excess water $W = 1 - \frac{E_T}{P}$ and energy $U = 1 - \frac{E_T}{E_p}$. To illustrate an example of a basin change, just consider that E_T changed, while the climatic variables have been constant. The resulting change of U and W will always have the same sign. For the direction of a climate change a further assumption is required. Thus, to predict the effect of climate change the assumption is made that a single basin moves along a predifined Budyko curve, e.g. the Schreiber or the Mezentsev; Choudhury curves. This first order time derivation is based on the works of Dooge (1992); Arora (2002); Roderick and Farquhar (2011). Anyhow, independent of the Budyko function being applied, a climate change will always result in changes in W, U of opposite signs.

6. *What determines the sensitivity of streamflow and evapotranspiration to changes in climate?*

The water-energy frameworks include hypotheses and analytical solutions on changes of the water and energy partitioning under a change in climate. These have been evaluated for any reasonable hydro-climatic state.

A first order control is the climatic supply of water and energy, usually combined to the non-dimensional ratio known as the aridity index or its inverse, the humidity index. Thus, given humid base conditions, a change in climate will alter the relative partitioning of water more than the relative partitioning of energy, see Figure 3.2. This is reversed under arid conditions. Further, the sensitivity of E_T increases with aridity of the climate. The catchment conditions, as described by the catchment parameter, play a secondary role. However, the sensitivity of streamflow to changes in the catchment conditions increases with the aridity index and it can be larger than the sensitivity to changes in climate, cf. Figure 3.7. This finding demonstrates the increasing role of catchment properties under limitation of resources. It further highlights that the prediction of climate effects on the hydrology of arid systems is quite uncertain given the large role of catchment conditions and the problem of reliably estimating them.

7. *How large are basin change impacts compared with changes in climate?*

The large hydro-climatic dataset of the US allowed to test the water-energy frameworks with respect to (i) analytical solutions of the climatic sensitivity of streamflow and (ii) the applicability of the framework to separate climate from basin changes. This was done through a split-sample test and combined with a statistical classification strategy to detect climate and basin changes.

First, the results showed for the case of significant changes both, in streamflow and climate, that the analytical solutions of the parametric Budyko curve, which have been presented by Roderick and Farquhar (2011), provide satisfactory results, see Fig. 4.6. The separation technique provided consistent results and was able to discern impacts of compensating climatic changes from excessive river flow changes, see Fig. 4.6 and the UW change diagram in Fig. 4.7.

The evaluation of observed hydro-climatic changes showed that increases in precipitation increase runoff and evapotranspiration, but with distinctive differences according to physiographic properties. The most relevant property is the index of dryness or its inverse, the humidity index. For example an increase in precipitation as observed in the 20th century over the continental U.S., has larger impacts on the partitioning of energy balance components in dry areas (decreasing sensible heat fluxes), whereas in wet catchments the water balance partitioning is stronger affected with increases in streamflow, cf. Fig. 4.9. Besides the climatic property, it was shown that human induced land cover changes, e.g. through land management, sometimes resulted in larger impacts on streamflow than climate impacts. We observed compensating effects with predominant evapotranspiration increases, possibly through intensive water use for agriculture in drier regions. But also excessive effects where the additional precipitation increased runoff only. It is likely that this was caused by artificial drainage of these predominant humid regions. Severe leaching of nutrients from the soils might occurred in parallel.

5.2.2 SOCIO-ECONOMIC AND POLITICAL RELEVANCE

Tools for assessment of changes in hydroclimatology The presented water-energy framework provides simple and powerful tools for first order estimates of the sensitivity of evapotranspiration and streamflow. These tools show to be applicable for the assessment of water and energy balance changes using future climate change scenario inputs. The framework also provides simple means to identify potential changes of climate and basin change, such as LULC changes from observed hydro-climate records.

The simplicity, the universal applicability to different hydroclimatic conditions and the relatively low data requirement makes the water-energy balance framework an ideal tool for regional to global change assessments. Thus, the framework can be regarded as complementary means to more complex hydrological models.

Compensating effects Changes in precipitation will not only change the partitioning of the water balance, but also the partitioning of the energy balance with wide-ranging effects. For example the increase of precipitation over the central US generally led to an increase of E_T, cf. section 4.4.3 or see Milly and Dunne (2001). At the same time the evaporative demand has been constant or showed decreases. In effect the partitioning of the energy fluxes has also been changed, leading to the phenomena that air temperature did not increase, as observed elsewhere. Thus, regional changes in precipitation or evaporative demand can hide global changes and may lead to wrong conclusions, e.g. with respect to climate change.

This example illustrates the important role of the coupled water-energy balance at regional scales. And more emphasis should be paid for water and energy balance changes in climate change assessment and regional adaption strategies.

5.2.3 LIMITATIONS AND FURTHER RESEARCH

Subscale variability The water-energy balance framework and also any type of Budyko function, generally work at the scale of basins and long term annual averages. Thus, any temporal variability, such as seasonality, intensity of precipitation events, variability changes in the meteorological variables effecting evaporative demand is disregarded by taking the mean. This is also true for spatial variability of climate inputs, such as shifts in areal precipitation within a larger basin (Roderick and Farquhar, 2011; Donohue et al., 2011) or small scale land-use changes in mixed cover basins (van Dijk et al., 2012).

Thus, the broad definition of climate and basin changes is an important limitation especially with respect to the interpretation of the results from the presented water-energy frameworks. So, for the attribution of an observed change in the hydroclimatology of a certain river basin to a more specific physical cause, it is necessary to further check for alternative hypotheses which may have contributed to the observed change. Some guidelines have been presented by Merz et al. (2012) for the attribution of changes in the extremes which can also be adopted here.

Although, a detailed attribution was out of the scope of the work presented here, some guidance can be given. A promising way forward is the explicit treatment of input variability and catchment processes through stochastic models working at the time scale of the relevant processes. Examples are given by derivations of Budyko curves from probability distribution functions of rainfall and evaporative demand by (Choudhury, 1999; Gerrits et al., 2009; Fraedrich, 2010), linking seasonality and catchment water storage (Milly, 1994; Feng et al., 2012; Zanardo et al., 2012), or upscaling effects of evaporation and transpiration (Gerrits et al., 2009). Generally, probabilistic approaches allow an upscaling of subgrid variability to the scale of interest. And it is possible to estimate the potential of subscale processes to effect the scale of interest (Williams, 2005). Further, probabilistic approaches allow for testing procedures of alternative hypotheses on the cause of hydroclimatic changes.

The climate change assumption: Trading space for time While the impact of basin changes can be directly derived from the definitions of the water-energy partitioning, the quantification of climate impacts on E_T requires an additional assumption. Next, some of the weak points of this assumption are discussed. In particular we assume that a single basin moves along a Budyko curve which was estimated by drawing through different basins with different climatic conditions. The critical point is, that it is speculative which particular Budyko curve matches to the conditions of the respective basin. However, this restriction may be relaxed by considering the works of Choudhury (1999) and Fraedrich (2010), showing that particular Budyko curves can be derived from the probability distribution functions of precipitation and evaporative demand. These findings reduce the semi-empirical character of the Budyko curves and backs up the working assumption.

Another possible critique is that climate and basin properties are not independent. Especially at the hydro-climate timescale we do not know the implications of adaption of vegetation to the changes, sometimes referred to as co-evolution of climate and vegetation (Berry et al., 2006). This issue gets even more problematic in reality as we have to consider simultaneous changes of climate and basin conditions; thus, the separation results and the mapping of the change direction provide only rough estimates of the magnitudes of climate and basin changes.

5.3 General conclusions and outlook

5.3.1 Regional and temporal limits and validity

As outlined above the choice of the seasonal timescale (Chapter 2) or the climatic timescale of long-term annual averages (Chapter 3,4) has led to integrative model approaches. Hence, this work does not explitly treat processes and components of the water and energy cycles at other temporal scales. Yet, it is clear that climate and land use changes have direct impacts upon specific (subscale) processes, that may sum up to measurable impact signals on the time scale considered here. This means that although it is possible to distinguish climate from basin change impacts at the catchment scale, the methods presented here do not directly allow to identify the specific cause.

Another aspect which is not touched here but of direct socio-economic importance, is that climate and land use changes are also capable to change the behaviour and magnitude of extreme events (Trenberth, 2012). For example storm water runoff, flooding and inundation is likely be altered by land use changes resulting in altered runoff generation processes (Bronstert et al., 2002). So the land use dynamics of urban sprawl increases the sealing of the surface, leading decreasing infiltration capacities and hence increase the probability of costly inundation events. Similarly, land use changes which alter soil conductivities might alter the partitioning of fast and slow runoff components (Zhang and Schilling, 2006; Kochendorfer and Hubbart, 2010).

The difficulties in assessing land use change impacts also arise from the fact that multiple other processes and surface properties are altered (interception of light and precipitation, reflectance, surface and aerodynamic conductivities, ...). Additionally, the detection of past LULC changes is difficult (Ramankutty and Foley, 1999), but also the creation of future land use scenarios is complicated by competing levels of socio-economic dynamics (Niehoff et al., 2002). The idea of "ecosystem services" puts further emphasis on the role of intact landscapes (Daily et al., 2000; Seppelt et al., 2011). So careful land use management and planning might enable to adapt or even mitigate some environmental change impacts (Tilman et al., 2001; Bonan, 2008; Salazar et al., 2012).

5.3.2 Hydrological records carry signals of climate and land use change

The results of this thesis confirm the expectation that signals of external variations are translated into detectable signatures of hydrological variability. That means, we can deduce climate and land surface changes at economic relevant spatial scales, which makes hydrological observations and archives extremely valuealbe. Given this importance it is alarming to observe the decline in hydrological gauging stations (Vörösmarty et al., 2000).

The extraction of useful signals requires an holistic view of the problem. As demonstrated useful information can be extracted from hydrological data, if these are seen as integral components of the water and energy cycles. The resulting minimalistic model serves as valuable a-priori information for statistical analysis.

5.3.3 Statistical significance of past changes

A major concern of environmental change is that the earth system moves from a presumably stationary behaviour into unpredictable, nonstationary states. The analysis of archives of the last century may provide indications if such changes are indeed happening. In this work two larger hydro-climate datasets have been investigated for nonstationary signals. It was found that most of the basins investigated show no statistical significant changes. One reason for that result is the large natural variability in these records, which reduces the ability of signal detection. Other reasons are that catchment conditions and especially resource limitation determine the sensitivity to external changes. So external changes in non-limited areas must be larger than in resource limited areas to result in detectable signals. Nevertheless, a large number of catchments show significant changes which indicate that external changes lead to nonstationary behaviour. And the geographic coherence of these trends highlight impacts of externally driven changes in the hydroclimatology. Examples from this work are the large number of catchments with high impact of basin changes, as well as climate induced changes in the hydrology (Table 4.3). Further, the study in chapter 2 shows that impacts of increasing temperature get important where the typical timing of physical thresholds such as the freezing point of water is affected.

5.3.4 IMPROVEMENTS IN ASSESSING E_T

In this study actual E_T has been assessed through water balance closure. This introduces a number of drawbacks, such as neglecting long-term changes in catchment water storage (Istanbulluoglu et al., 2012), uncertainties in estimating catchment precipitation (MartinezCob, 1996) and the conversion of water levels to discharge (Di Baldassarre and Montanari, 2009).

Hence, much improvement can be made by including other assessment strategies for E_T. Promising ways are mostly promoted in atmospheric sciences. So, is seems very promising to constrain E_T estimates by energy balance closure methods, see e.g. Brutsaert (1982). This would allow to test if the observed changes in the partitioning of water are consistent with the model based estimates of the energy partitioning changes. However, there is the known but not often articulated problem of lacking suitable networks of radiation balance observations. This lack of information leads to large uncertainties in the surface energy balance. For historical assessment of net radiation often sunshine duration are employed, but these provide rather rough estimates. Improvements in this respect are expected from remote sensing based radiation data (Sommer, 2009; Posselt et al., 2012) and high resolution re-analysis experiments (Troy and Wood, 2009).

Further promising is the use of micrometeorological measurements, such as Eddy-Covariance techniques (Grünwald and Bernhofer, 2007; Aubinet et al., 2012) or Bowen Ratio methods (Malek and Bingham, 1993; Perez et al., 1999) for estimating E_T (Bernhofer, 1992). Recent research demonstrated the value of micrometeorological observations of the FLUXNET network (Baldocchi et al., 2001) for model-based upscaling of E_T (Jung et al., 2009, 2010) and the validity of the Budyko hypothesis at the local scale (Williams et al., 2012).

5.3.5 REMOTE SENSING

Remote sensing data deliver spatial information of various earth system processes. Thus, it allows to assess the roles of spatial variability and structures of landcover, topography, soil moisture or various meteorological variables, such as radiation. Exploitation of these spatially structured information into models is, however, not straightforward and requires interdisciplinary research approaches (Schulz et al., 2006).

Another direct gain from remote sensing data is that we are able to trace land use changes at high resolution over time (DeFries and Eshleman, 2004). This also allows to further verify the separation technique proposed in chapter 3. For example, Renner et al. (2013) show that increases in E_T, being identified as basin changes using the separation method, corresponds to regeneration after massive forest decline in the Ore Mountains as identified by the CORINE land cover classification product. This signal is shown to correspond to observed increases of NDVI as detected from the AVHRR mission.

5.3.6 LEARNING FROM THE PAST TO PREDICT THE FUTURE?

The analysis of past impacts of climate and land-use on the hydroclimatology of river catchments revealed highly diverse signals of change and impacts. On the one side climatic variability controls E_T on the other side LULC can be dominant. These result in spatially varying trends and also temporally changing trends, breakpoints and trend reversals. Thus, there are hardly any monotonic stable trends in hydrological records and any extrapolation of a detected trend into the future is rather speculative and thus not recommended.

For past analysis to be useful for future predictions, it is rather recommended to identify the fundamental laws which are valid even under extreme changes of the boundary conditions. Most relevant for assessments of evapotranspiration and hydro-climatological changes are water and energy conservation. Water and energy conservation should be the most basic requirement for model development. In fact, consideration of the conservation laws actually led to the rejection and correction of the separation framework proposed by Tomer and Schilling (2009), see section 3.5.1.

Exciting ways forward can be expected from constraining more complex models through the optimality hypothesis (Eagleson and Tellers, 1982; Schymanski et al., 2009; Katul et al., 2012). This would allow future predictions as we can expect that even under changed environmental conditions life will optimise for reproduction, changing the living conditions. This requires to understand and model ecological strategies and self-organised structures within the environment (Schaefli et al., 2011). On a more fundamental level the development of earth system processes can be seen as a consequence of the second law of thermodynamics (Paltridge, 1979). Hence, thermodynamic limits of these processes and interactions can be identified and allow theory based constraining of earth system processes (Kleidon, 2012).

With these perspectives we come back to the starting statement that evapotranspiration is an interactive, constantly changing process. And we have to consider large changes simply because of the needs of the increasing human population (Vörösmarty et al., 2000). Thus, we humans are already a significant part of the earth system (Crutzen, 2002) and interact with various processes. Hence, the role of socio-economic dynamics for the earth system increases. This could be worrying, but it also includes possibilities for adapting to a more sustainable future.

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