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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.

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, 2011), we will address the application of this hydro-climatic framework on a multitude of catchments throughout the contiguous United States.

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.

2.1 Coupled water and energy balance

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

$$P = E_T + Q + \Delta S_\mathrm{w} \quad \text{and} \tag{1}$$

$$R_\mathrm{n} = E_T\, L + H + \Delta S_\mathrm{e}. \tag{2}$$

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:

$$E_\mathrm{p} = E_T + H/L. \tag{3}$$

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:

$$W = 1 - \frac{E_T}{P} = \frac{Q}{P}, \quad U = 1 - \frac{E_T}{E_\mathrm{p}} = \frac{H/L}{E_\mathrm{p}}. \tag{4}$$

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. 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. 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 E_T (Gedney et al., 2006), can not be attributed to a climate change direction in Fig. 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.

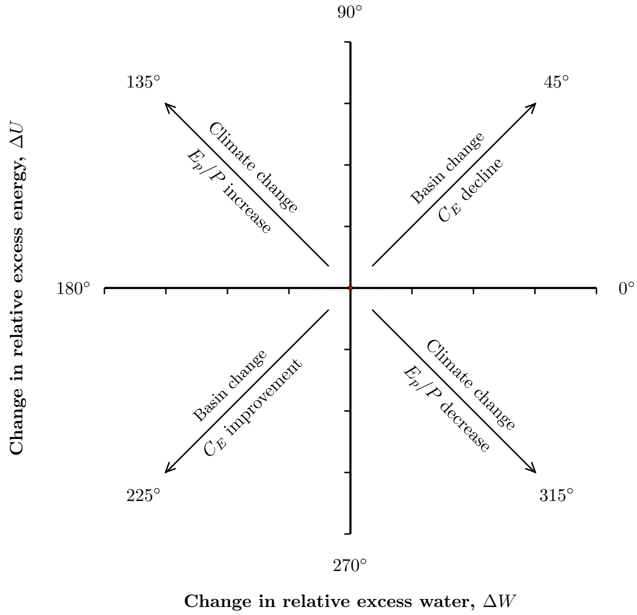

Fig. 1. Illustration of the change attribution framework established by Tomer and Schilling (2009, after their Fig. 2). Considering climatic change effects, a change in either precipitation or potential evapotranspiration, will result in a change of both, relative excess water and energy but in opposite 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. 8), which results in a deviation from the negative diagonal.

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:

$$\Delta U/\Delta W = -1. \tag{5}$$

We refer to Eq. (5) as the climate change impact hypothesis (abbreviated as CCUW).

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

$$\alpha = \arctan \frac{\Delta U}{\Delta W} \tag{6}$$

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

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, $E_{\mathrm{p},0}$, $E_{T,0}$), (P_1, $E_{\mathrm{p},1}$, $E_{T,1}$) of a given basin. With that the changes in relative excess water ΔW and energy ΔU can be expressed by using Eq. (4) as:

$$\Delta W = \frac{E_{T,0}}{P_0} - \frac{E_{T,1}}{P_1}, \quad \Delta U = \frac{E_{T,0}}{E_{\mathrm{p},0}} - \frac{E_{T,1}}{E_{\mathrm{p},1}}. \tag{7}$$

Applying the CCUW hypothesis Eq. (5) to the definitions of ΔW and ΔU (Eq. 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:

$$\frac{E_{T,0}}{P_0} + \frac{E_{T,0}}{E_{\mathrm{p},0}} = \frac{E_{T,1}}{P_1} + \frac{E_{T,1}}{E_{\mathrm{p},1}} = C_E = \text{const.} \tag{8}$$

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 $E_T = \min(P, E_\mathrm{p})$, we would find a value of C_E of about 1.

Finally rearranging Eq. (8) yields an expression to compute the evapotranspiration of the new state ($E_{T,1}$):

$$E_{T,1} = E_{T,0} \frac{\frac{1}{P_0} + \frac{1}{E_{\mathrm{p},0}}}{\frac{1}{P_1 + \frac{1}{E_{\mathrm{p},1}}}}. \tag{9}$$

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:

$$Q_1 = \frac{\frac{Q_0}{P_0} - \frac{P_0 - Q_0}{E_{\mathrm{p},0}} + \frac{P_1}{E_{\mathrm{p},1}}}{\frac{1}{P_1} + \frac{1}{E_{\mathrm{p},1}}}. \tag{10}$$

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

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 $\partial Q/\partial P$:

$$\varepsilon_{Q,P} = \frac{P}{Q}\frac{\partial Q}{\partial P}. \tag{11}$$

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:

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

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

$$\mathrm{d}W = w'(P, E_T) = \frac{\partial w}{\partial P}\mathrm{d}P + \frac{\partial w}{\partial E_T}\mathrm{d}E_T \tag{12}$$

$$\mathrm{d}U = u'\left(E_\mathrm{p}, E_T\right) = \frac{\partial u}{\partial E_\mathrm{p}}\mathrm{d}E_\mathrm{p} + \frac{\partial u}{\partial E_T}\mathrm{d}E_T \tag{13}$$

$$\frac{\partial w}{\partial P} = \frac{E_T}{P^2}, \ \frac{\partial w}{\partial E_T} = -\frac{1}{P}, \ \frac{\partial u}{\partial E_\mathrm{p}} = \frac{E_T}{E_\mathrm{p}^2}, \ \frac{\partial u}{\partial E_T} = -\frac{1}{E_\mathrm{p}}. \tag{14}$$

Combining Eqs. (12) and (13) with the CCUW hypothesis Eq. (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. (14) are applied to gain an analytical solution:

$$\frac{\mathrm{d}E_T}{E_T} = \left[\frac{E_\mathrm{p}}{E_T}\frac{-\frac{\partial u}{\partial E_\mathrm{p}}}{\frac{\partial u}{\partial E_T} + \frac{\partial w}{\partial E_T}}\right]\frac{\mathrm{d}E_\mathrm{p}}{E_\mathrm{p}} + \left[\frac{P}{E_T}\frac{-\frac{\partial w}{\partial P}}{\frac{\partial u}{\partial E_T} + \frac{\partial w}{\partial E_T}}\right]\frac{\mathrm{d}P}{P} \tag{15}$$

$$\frac{\mathrm{d}E_T}{E_T} = \left[\frac{P}{E_\mathrm{p} + P}\right]\frac{\mathrm{d}E_\mathrm{p}}{E_\mathrm{p}} + \left[\frac{E_\mathrm{p}}{E_\mathrm{p} + P}\right]\frac{\mathrm{d}P}{P}. \tag{16}$$

By Eq. (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. (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. (16):

$$\frac{\mathrm{d}Q}{Q} = \left[\frac{P(P-Q)}{Q\left(E_\mathrm{p}+P\right)}\right]\frac{\mathrm{d}E_\mathrm{p}}{E_\mathrm{p}} + \left[\frac{P}{Q} - \frac{(P-Q)E_\mathrm{p}}{Q\left(E_\mathrm{p}+P\right)}\right]\frac{\mathrm{d}P}{P}. \quad (17)$$

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.

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)

$$E_T = E_\mathrm{p} \cdot P / \left(P^n + E_\mathrm{p}^n\right)^{1/n}, \quad (18)$$

to which we will refer to as Mezentsev function. Yang et al. (2008) derived Eq. (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: $E_T = f(E_\mathrm{p}, P, n)$. Once the functional type of f is known, climate changes causing a change in E_T ($\mathrm{d}E_T$) 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):

$$\mathrm{d}E_T = \frac{\partial E_T}{\partial P}\mathrm{d}P + \frac{\partial E_T}{\partial E_\mathrm{p}}\mathrm{d}E_\mathrm{p} + \frac{\partial E_T}{\partial n}\mathrm{d}n. \quad (19)$$

Next, by employing the long term water balance equation $\mathrm{d}Q = \mathrm{d}P - \mathrm{d}E_T$ to Eq. (19), an expression for the change in streamflow ($\mathrm{d}Q$) is gained (Roderick and Farquhar, 2011):

$$\mathrm{d}Q = \left(1 - \frac{\partial E_T}{\partial P}\right)\mathrm{d}P - \frac{\partial E_T}{\partial E_\mathrm{p}}\mathrm{d}E_\mathrm{p} - \frac{\partial E_T}{\partial n}\mathrm{d}n. \quad (20)$$

With Eqs. (19), (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 ($\mathrm{d}P$, $\mathrm{d}E_\mathrm{p}$) and changes in basin characteristics ($\mathrm{d}n$) (Roderick and Farquhar, 2011). In the case of the non-parametric Budyko functions, the last term in Eqs. (19) and (20) can be omitted.

Climatic elasticities ($\mathrm{d}E_T/E_T$ and $\mathrm{d}Q/Q$) can easily be obtained from Eqs. (19) and (20) by dividing by E_T or Q and term expansions on the right side (Roderick and Farquhar, 2011):

$$\frac{\mathrm{d}E_T}{E_T} = \left[\frac{P}{E_T}\frac{\partial E_T}{\partial P}\right]\frac{\mathrm{d}P}{P} + \left[\frac{E_\mathrm{p}}{E_T}\frac{\partial E_T}{\partial E_\mathrm{p}}\right]\frac{\mathrm{d}E_\mathrm{p}}{E_\mathrm{p}} + \left[\frac{n}{E_T}\frac{\partial E_T}{\partial n}\right]\frac{\mathrm{d}n}{n} \quad (21)$$

$$\frac{\mathrm{d}Q}{Q} = \left[\frac{P}{Q}\left(1 - \frac{\partial E_T}{\partial P}\right)\right]\frac{\mathrm{d}P}{P} + \left[\frac{E_\mathrm{p}}{Q}\frac{\partial E_T}{\partial E_\mathrm{p}}\right]\frac{\mathrm{d}E_\mathrm{p}}{E_\mathrm{p}} + \left[\frac{n}{Q}\frac{\partial E_T}{\partial n}\right]\frac{\mathrm{d}n}{n}. \quad (22)$$

As in the previous subsection, the bracketed terms denote the elasticity coefficients for P, E_p and n. For the computation of $\mathrm{d}E_T$, $\mathrm{d}Q$ and the elasticity coefficients, we only need to enter the respective partial differentials. Roderick and Farquhar (2011) report these terms for the Mezentsev function and they are repeated for completeness below:

$$\frac{\partial E_T}{\partial P} = \frac{E_T}{P}\left(\frac{E_\mathrm{p}^n}{P^n + E_\mathrm{p}^n}\right), \quad \frac{\partial E_T}{\partial E_\mathrm{p}} = \frac{E_T}{E_\mathrm{p}}\left(\frac{P^n}{P^n + E_\mathrm{p}^n}\right) \quad (23)$$

$$\frac{\partial E_T}{\partial n} = \frac{E_T}{n}\left(\frac{\ln\left(P^n + E_\mathrm{p}^n\right)}{n} - \frac{\left(P^n \ln(P) + E_\mathrm{p}^n \ln\left(E_\mathrm{p}\right)\right)}{P^n + E_\mathrm{p}^n}\right). \quad (24)$$

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.

3.1 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):

$$W = 1 - f\left(E_\mathrm{p}, P, n\right), \quad U = 1 - \frac{f\left(E_\mathrm{p}, P, n\right)P}{E_\mathrm{p}}. \quad (25)$$

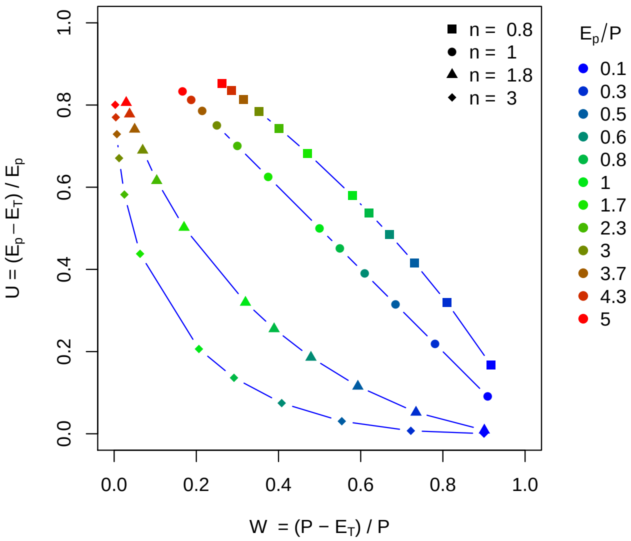

Fig. 2. Mapping different parameterised Mezentsev functions into UW space using Eq. (25). The colours depict certain aridity (E_p/P) values indicated by the legend in the right.

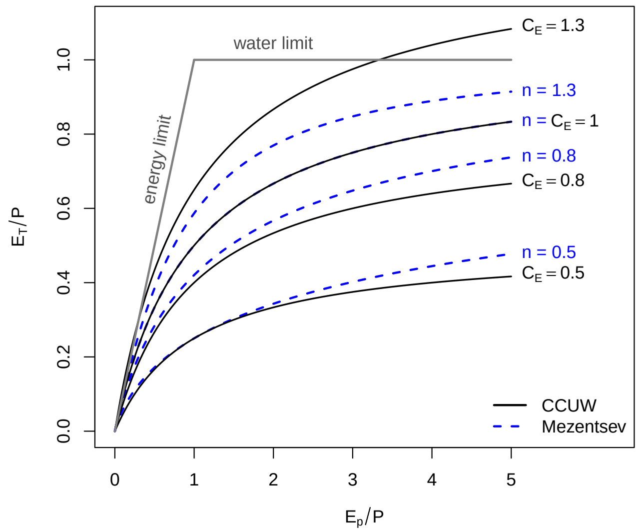

Fig. 3. Mapping of CCUW hypothesis into Budyko space for different values of catchment efficiency (C_E) using Eq. (26). For comparison different parameterisations of the Mezentsev curve are also shown. The grey lines depict the theoretical limits for water and energy.

Figure 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. (18) follows the negative diagonal of the climate change hypothesis, cf. Fig. 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 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 ($\Delta \frac{E_\mathrm{p}}{P}$), 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 $P = E_\mathrm{p}$ 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 A.

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. (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. (8) to achieve a mapping to Budyko space:

$$\frac{E_T}{P} = C_E \frac{E_\mathrm{p}}{P + E_\mathrm{p}}. \quad (26)$$

Figure 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. (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. (18) and (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 $E_\mathrm{p}/P \to \infty$. This makes a distinction from the Budyko hypothesis, which employs the water limit $E_T/P = 1$ as asymptote for $E_\mathrm{p}/P \to \infty$. 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

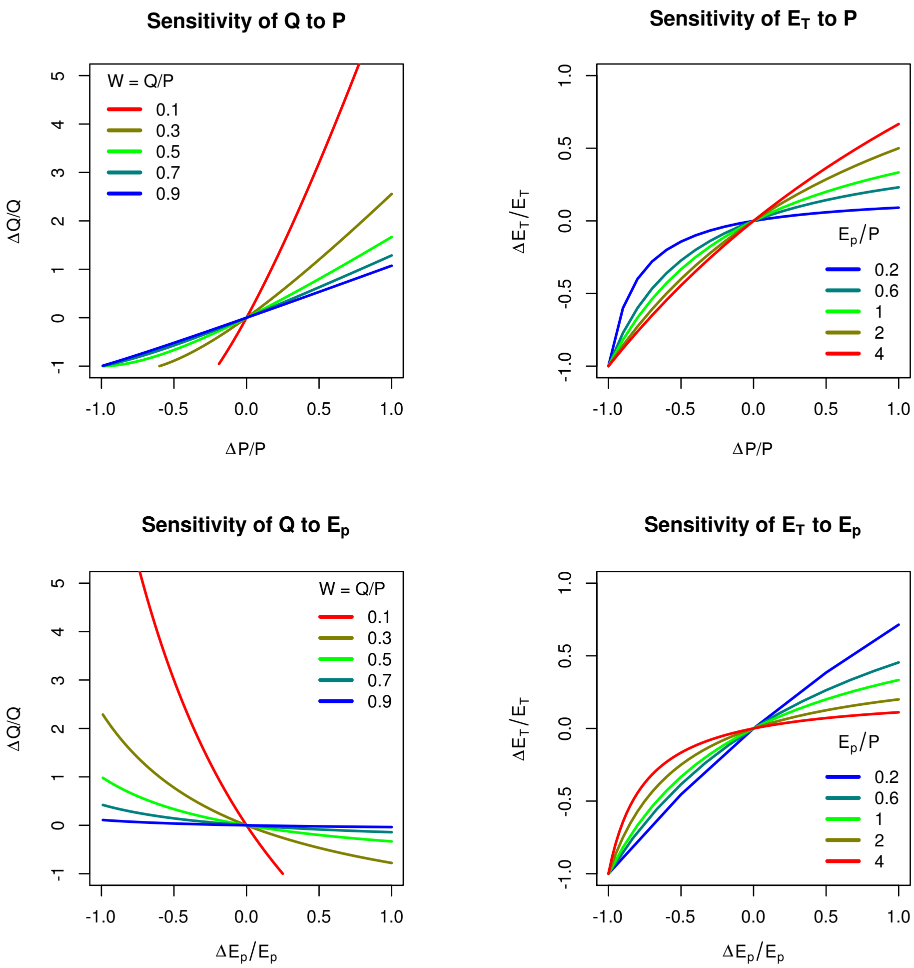

Fig. 4. Relative change in response to relative changes in P (top panels) and in E_p (bottom panels) of Q (left panels) and E_T (right panels) as predicted by the CCUW hypothesis. Changes in Q are dependent on runoff ratio W and on aridity E_p/P and are coloured with respect to the respective runoff ratio and shown for an aridity index of $E_\mathrm{p}/P=1$. 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.

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. 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 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 4 illustrates the general behaviour of the CCUW hypothesis under changes in precipitation or potential evapotranspiration, which is expressed by Eqs. (9) and (10). The left panels of Fig. 4 show the relative change of streamflow to P (upper) and E_p (lower panel). From Eq. (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. 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. 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 $\varepsilon_\mathrm{p} + \varepsilon_{E_\mathrm{p}} = 1$. Therefore, we only discuss the elasticity to precipitation. Figure 5 displays the elasticity of E_T ($\varepsilon_{E_T,P}$) as a function of aridity. In more humid or semi-arid conditions ($E_\mathrm{p}/P < 2$), 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 $E_\mathrm{p}/P = 4$ 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. (17) and (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 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. 6, where we plot the runoff ratio Q/P as a function of aridity. The streamflow elasticities

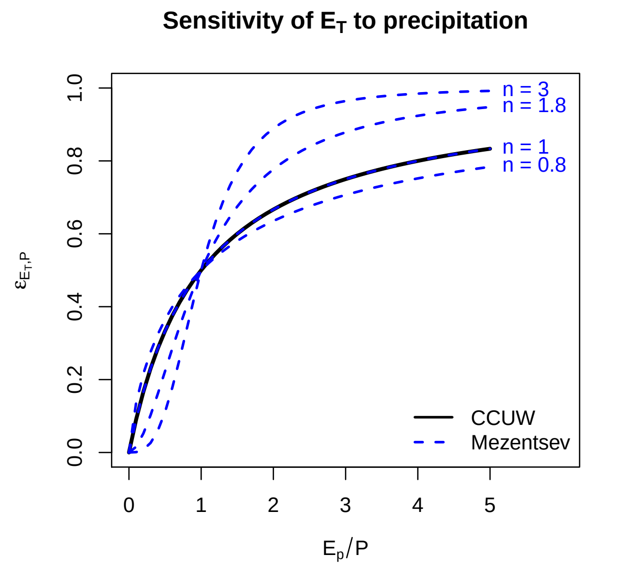

Fig. 5. Sensitivity (elasticity) of basin evapotranspiration with respect to changes in precipitation ($\varepsilon_{E_T,P}$). The bold black line depicts elasticity of the CCUW, while the dashed line shows different elasticities for the Mezentsev function. The elasticity of the CCUW corresponds with the slope of the curves shown in the top right panel of Fig. 4.

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 $E_\mathrm{p}/P \rightarrow \infty$.

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. (21) and (22). In contrast the CCUW approach is only sensitive to changes in P and E_p, cf. Eqs. (16) and (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. (11), shows that the sensitivity coefficients are composed of two components, which is also apparent in the sensitivity terms within Eqs. (21) and (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

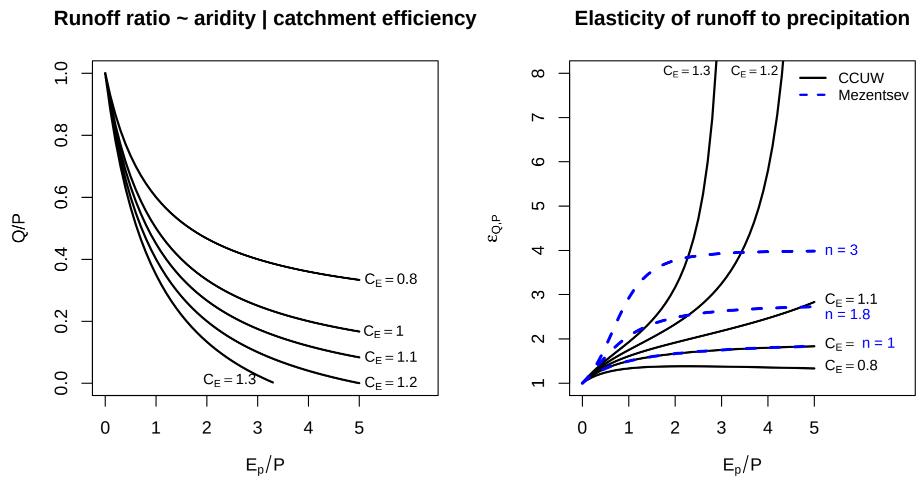

Fig. 6. Left panel: runoff ratio as a function of aridity for different, but fixed values of catchment efficiency (C_E) using Eq. (26). Right panel: elasticity coefficient of streamflow to precipitation $\varepsilon_{Q,P}$ as a function of aridity. Displayed are the elasticities derived from the CCUW hypothesis (black for different values of C_E), and the elasticities derived from different parameterisations of the Mezentsev functions using Eq. (22).

Mezentsev function with n set to 1.8. In Fig. 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. (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. 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. 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 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. (18) and the CCUW hypothesis transformed into Budyko space in Eq. (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

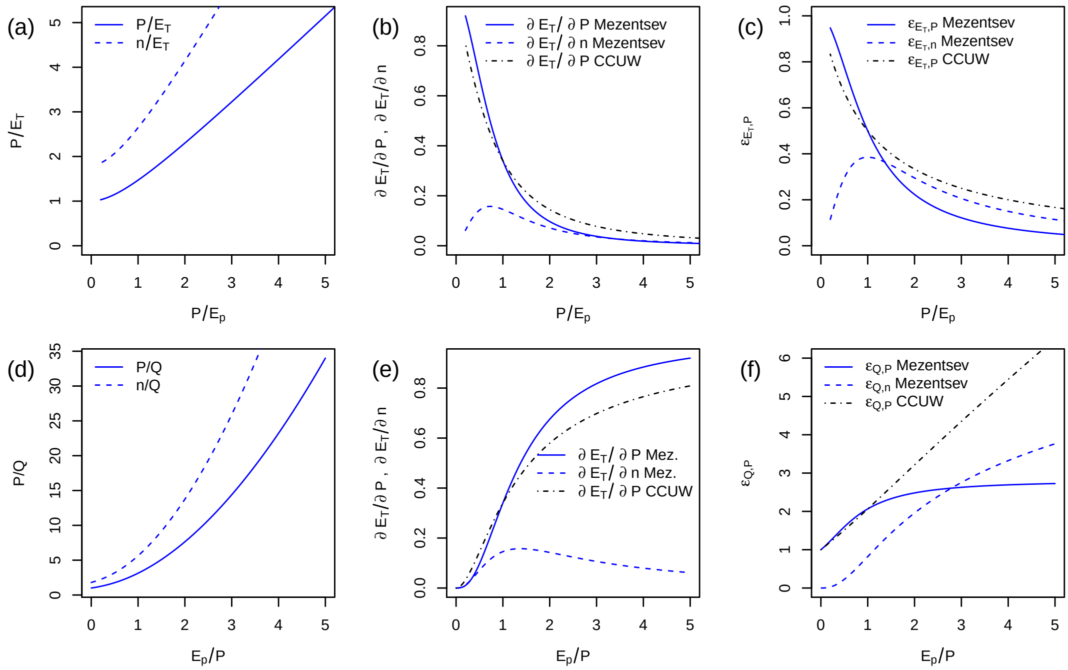

Fig. 7. Sensitivity coefficients and their components as functions of the humidity and aridity index, respectively. Baseline water balance terms (E_T and $Q = 1 - P$) have been determined with the Mezentsev function and $n = 1.8$. For illustration purposes we set $P = 1$ and $E_p = 0 \ldots 5$. Top panels 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. (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. (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.

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.

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 1.

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 1 we see

Table 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. 1, is computed by Eq. (6). The theoretical climate change direction derived for the Mezentsev function (α_{mez}) is computed by Eq. (A5).

	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

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.

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 concept of Tomer and Schilling (2009) (Fig. 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.

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 ($E_\mathrm{p}/P = 3.5$) 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 $-5.6\,\mathrm{mm\,yr}^{-1}$ (−20.5 %).

From Table 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 $\varepsilon_{Q,P;\mathrm{mez}} = 2.6$, which is close to the theoretical upper bound of $\varepsilon_{Q,P;\mathrm{mez}} = 1 + n = 2.74$. Employing data of the climatic changes in the second period we predict a change of $-3.2\,\mathrm{mm\,yr}^{-1}$. 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

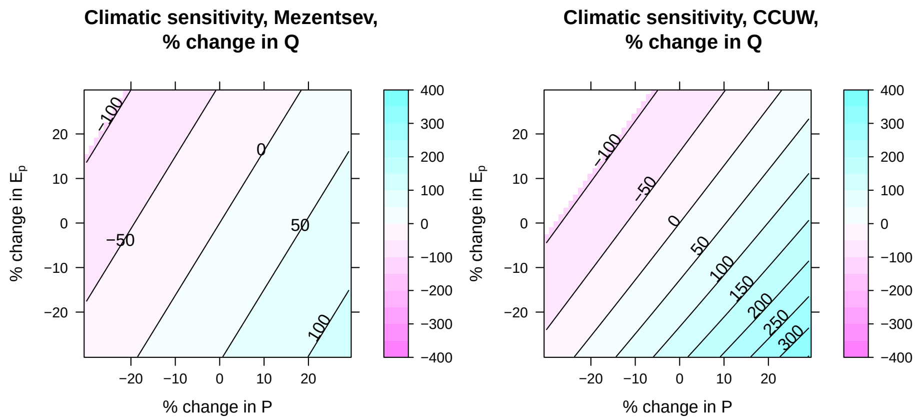

Fig. 8. Sensitivity plots of streamflow to percent changes of precipitation and E_p, estimated for the long term hydro-climatic states of the Murray-Darling Basin (as given in Table 1). Contour lines depict the percent change in streamflow. Note that changes of $\Delta Q/Q$ smaller than $-100\,\%$ are not physical. Left panel: The Budyko framework using the Mezentsev function and Eq. (22) in accordance to Roderick and Farquhar (2011, Fig. 2). Right panel, sensitivity estimation by the CCUW framework Eq. (10).

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 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 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. 3.4, the resulting difference in streamflow sensitivity illustrates the impact of the inherent assumptions on the role of climate-vegetation feedbacks in arid environments.

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.

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 ($E_T \leq E_\text{p}$) 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. 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. 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. 1, while under humid conditions the arrows are towards the abscissa.

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.

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 (2011), 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.

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 (Klemes, 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.

Appendix 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 $\Phi = E_\mathrm{p}/P$ and a catchment parameter n as

$$E_T/P = f(\Phi,\ n). \tag{A1}$$

With Eq. (25) we obtained a mapping of f to the UW space. Using the aridity index as $\Phi = E_\mathrm{p}/P$, Eq. (25) can be written as:

$$W = 1 - f(\Phi,\ n) \tag{A2}$$

$$U = 1 - \frac{f(\Phi,\ n)}{\Phi}. \tag{A3}$$

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. (A2). Because Eq. (A3) includes both $f(\Phi,\ n)$ and Φ, we need to derive the inverse of Eq. (A1). The analytical solution for Mezentsev' function Eq. (18) is derived below. First, Eq. (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:

$$\Phi = \left(\frac{1}{{\frac{1}{1-W}}^{n} - 1}\right)^{\frac{1}{n}}. \tag{A4}$$

Then by inserting Eq. (A4) into Eq. (A3) and differentiating with respect to W yields a term for CCD of the Mezentsev' equation:

$$\alpha_\mathrm{mez} = g'(W,\ n) = \Big((1-W)^{2n} - (1-W)^{n}\Big) \cdot \left(\frac{(1-(1-W)^n)^{1-2n}}{(1-W)^n}\right)^{\frac{1}{n}}. \tag{A5}$$

Last, by substituting W with Eq. (A2) in Eq. (A5) a relation of the CCD as function of Φ, n can be obtained.

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