Shift of annual water balance in the Budyko space for a catchment with groundwater dependent evapotranspiration

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Abstract

The Budyko framework represents the general relationship between the evapotranspiration ratio (F) and the aridity index (ϕ) for the mean annual water balanceat catchment scale. It is interesting to investigate if this standard F-ϕ space can be also applied to capture the shift of annual water balance in a catchment with the varying dryness. There are reported cases where the original Budyko framework can't be directly applied for annual water balance due to additional sources of water supply for evapotranspiration besides precipitation.This study investigates how groundwater dependent evapotranspiration causes the shift of annual water balance in the standard Budyko space. A widely used monthly hydrological model, the ABCD model, is modified to incorporate groundwater dependent evapotranspiration in the zone with shallow water table and delayed groundwater recharge in the zone with deep water table. This model is applied in the Hailiutu River catchment in China to estimate the actual annul evapotranspiration, where the depth to water table is less than 2 m in a zone occupying 16% of the catchment area. Results show that the variations in the annual F value with the aridity index do not satisfy the normal Budyko formulas. The shift of the annual water balance in the standard Budyko space is a combination of the Budyko-type response in the deep groundwater zone and the quasi-enegy limited condition in the shallow groundwater zone. Excess evapotranspiration (F>1) could occur in extreme dry years, which is contributed by the significant supply of groundwater for evapotranspiration. Use of groundwater for irrigation can increase the frequency of the F>1 cases.

1 Introduction

Estimating catchment water balance is one of the fundamental tasks in hydrology. Efforts have long been devoted to construct physical, empirical, and statistical models to explain the general relationship among precipitation (P), runoff (Q), potential evapotranspiration (E_0) and actual evapotranspiration (E) in terms of mean annual fluxes at the catchment scale (Budyko, 1948, 1958, 1974; Mezentsev, 1955; Fu, 1981; Porporato et al., 2004; Gerrits et al., 2009). A simple and highly intuitive approach widely used for estimating E at mean annual water balance is the Budyko framework, in which the mean annual evapotranspiration ratio (E/P) was presumed as a function of the climatic dryness as:

$$\frac{E}{P} = F\left(\frac{E_0}{P}\right) = F(\phi), \tag{1}$$

where ϕ is the aridity index defined as E_0/P, and $F(\phi)$ is an empirical function that relates E/P to ϕ based on general water-energy balance ~~behaviorsin catchments~~. The proposed formula by Budyko (1958; 1974) was:

$$F(\phi) = \sqrt{\phi[1-\exp(-\phi)]\tanh(1/\phi)}, \tag{2}$$

which indicates a nonlinear relation between F and ϕ ~~without any parameter~~. This F-ϕ curve has been called the Budyko curve (Zhang et al., 2004; Roderick and Farquhar, 2011) and the F-ϕ space was called Budyko space (Renner et al., 2012).

Instead of using a single curve determined by Eq. (2) in the Budyko space, researchers have introduced a specific catchment parameter in $F(\phi)$ to consider the impacts of catchment properties such as soils and vegetation (Mezentsev, 1955; Fu, 1981). For example, Fu's equation (Fu, 1981) was derived following the idea of Mezentsev (1955) and has been widely used in the last decade (Zhang et al., 2004; Yang et al., 2006; Yang et al., 2007; Zhang et al., 2008;Greve et al., 2015). In particular, Zhang et al. (2001) presented an empirical equation for the Budyko framework in relation to vegetation cover at the catchment scale as:

$$F(\phi) = \frac{w\phi}{1 + w\phi + \phi^{-1}}, \tag{3}$$

where w is called the plant-available water coefficient. Donohue (2007) highlighted the role of vegetation dynamics in application of the Budyko framework. Recently, Wang and Tang

(2014) also developed a one-parameter Budyko model based on the proportionality hypothesis and revealed a complex relationship between the catchment specific parameter and remote sensing vegetation index. These modified formulas suggested a group of Budyko curves instead of the single original Budyko curve, in which a curve represents a specific type of the catchments with similar features controlling the mean annual water balance.

Budyko hypothesis has been directly used to analyzethe interannual change in water balance in catchments (Arora, 2002; Zhang et al., 2008; Potter and Zhang, 2009) even through ~~it~~ ignoring the change in storage (ΔS) under the assumption of steady state water balance.One can plot annually the estimated E/P data in the standard Budyko space to check whether the normal Budyko curves are sufficient or not to represent the interannual variability of evapotranspiration with the varying dryness. In this way, Potter and Zhang (2009) found that the Budyko framework is generally applicable for the catchments in Australia and the optimal curve of annual E/P versus ϕ is highly dependent on the seasonal variations in rainfall. However, this approach should be carefully used when the E/P values are approximated by $(P-Q)/P$ values. Wang et al. (2009) and Istanbulluoglu et al. (2012) reported that the annual data of $(P-Q)/P$ in some basins are negatively related to the aridity index, exhibiting an inverse trend in comparison with the normal Budyko curves. According to long-term groundwater ~~observation~~ in the North Loup River basin, Nebraska, USA, Istanbulluoglu et al. (2012)demonstrated that the annual E/P values estimated by $(P-Q-\Delta G)/P$basically follows the Budyko hypothesis, where $\Delta G$is the change in groundwater storage. However, in some other studies, unexpected high evapotranspiration ratio (E/P>1) was observed (Cheng et al., 2011; Wang, 2012; Chen et al., 2013). Among the 12 watersheds investigated byWang (2012), half of them exhibited such high E/P values in two or more dry years. The physical base of the phenomena is the significant contribution of storage in extremely arid situation by which the high level of evapotranspiration is maintained. Although some of the cases ~~was~~ due to extractinggroundwater for irrigation in farmlands (Cheng et al., 2011; Wang, 2012), it could occur in natural conditions as a result of the temporal redistribution of water from seasonal patterns(Chen et al., 2013). Wang (2012) and Chen et al.(2013) proposed an approach to extend the Budyko framework for annual or even intra-annual water balance by considering the ~~decrease~~ in soil water storage as an potential source of water supply for evapotranspiration. They define $P-\Delta S$ for the selected time scales as the effective rainfall in building the modified Budyko space with $E/(P-\Delta S)$ and $E_0/(P-\Delta S)$, instead of E/P and ϕ, respectively.~~Then, they~~

~~found that the annual water balance of the catchments show the Budyko-type behaviro in the modified Budyko sapce.~~

The excess annual evapotranspiration ~~over the annual precipitation~~ may be originated from both soil water and groundwater. As reported by Wang (2012), during the drought year in 1988, two watersheds in Illinois~~,~~ USA, showed E/P=1.1 with ~100 mm depletion in soil water and ~200 mm decrease in groundwater storage, respectively. It seemed that the contribution of groundwater is more significant (partially enhanced by pumping). Small depth to water table is an advantage to keep a high level of soil water content near ground surface for evapotranspiration (Chen and Hu, 2004). Therefore, it could be argued that the existence of shallow groundwater in a catchment would enhance the occurrence of E/P>1 in dry years. Groundwater dependent evapotranspiration at the regional scale has been noticed in ~~a few of the~~ previous studies (York et al, 2002; Chen and Hu, 2004; Cohen et al., 2006; Yeh and Famiglietti, 2009). Nevertheless, little has been known on the role of groundwater in the interannual variability of the evapotranspiration ratio with the varying dryness.Chen et al.(2013) did not identify the change in groundwater storage to explain the controls of the E/P>1 cases.Wang (2012) mentioned the potential role of groundwater in occurrence of theE/P>1 cases but the individual contribution of groundwater dependentevapotranspiration was not soundly analyzed.

This study aims to investigatehow groundwater dependent evapotranspiration influencesthe annual water balance behavior in the standard Budyko space and develop a modified formula to consider this effect. A monthly hydrological model was developed from the widely used ABCD model (Thomas, 1981) to incorporate the groundwater dependent evapotranspiration as well as the deep infiltration in the vadose zone. The value of E was partitioned into two componentes in accounting for the individual roles of the normal soil water dependent and the specific groundwater dependent evapotranspiration. Then, the modified model was applied toa real ~~world~~ catchment as an example. The calibrated model wasused to produce the annual data of the evapotranspiration components linking with the variable soil water and groundwater storages. With varying climatic dryness, the shift behaviors of the interannual water balance in the standard and modified Budyko space for the catchment were analyzed in detail. The impacts of human activities were also discussed. The study reveals the contribution of groundwater in the interannual variability of catchment water balance under a changing climate.

2 Study Site, Data and Preliminary Analysis

2.1 Study area

The study site isthe Hailiutu River catchment (HRC), with an area of 2,645 km^2, located in the Erdos Plateau in north-central China (Fig.1a). The HRC lies on the southeast edge of the Mu Us Desert and is a sub-catchment of the Wuding River basin, which drains into the Yellow River (Fig.1b). The climate of the Erdos Plateau is typically inland semiarid to arid.The mean annual precipitation in the HRC is ~350 mm/a. More than 60% of the annual precipitation is received in the warm season (June, July,August and September). The main channel of the HRC has a length of approximately 85 km and flows southwards to the Hanjiamao hydrological station, as shown in Fig.1c. Due to the arid climate and desert landscape, the land cover within the catchmentis characterized by desert sand dunes with patches of mostly shrublands. Depression areas and terrace lands with shallow groundwater are covered by meadows and some farmlands. Wind-breaking trees (*Salix matsudana* and *Populus tomentosa*) can be found along the roads and crop areas. Farmlands are mainly located in the southern area and especially in the river valley. Crops cover only ~3% of the total catchment area. Maize is the dominant crop and is irrigated with streamflow and/or groundwater. Several diversion dams have been constructed along the Hailiutu River since the early 1970s for irrigation.

In the study area, groundwater is stored in a~~n~~ thick aquifer system with the sandy sediments and the underlying sandstones. Regional groundwater level distribution has been investigated in Lv et al. (2013) based on a hydrogeological survey carried out in 2010. According to this investigation, depth to water table (DWT) in the area varies in a large range from zeroto 110 m. In more than half of the area, DWT is less than 10 m. The shallow groundwater zone, where DWT is no more than 2 m, occupies 16.0% of the whole catchment area. As investigated in Yin et al. (2015) at the site of the HRC, when DWT is less than 2 m, the transpiration rate of trees is generally ~~higher~~ than 90% of the potential transpiration rate and the soil surface evaporation rate is generally ~~higher than~~ 60% of the potential. As a whole, the evapotranspiration rate would be generally ~~higher than~~ 80% of the potential when DWT is less than 2 m, whereas the evapotranspiration ratio is generally less than 0.4 for the deep groundwater condition (Yin et al., 2015). This investigation confirms that groundwater dependent evapotranspiration is an essential process in the HRC.

2.2 Data

Daily streamflow data since 1957 is availablefrom the Hanjiamao hydrological station. A rainfall gauge was also installed at the hydrological station in 1961, providing daily data of precipitation. In additiontothe Hanjiamao station, rainfall is observed at the city of Uxin Qi,located in the northern half of the basin (Fig.1c), where a meteorological station has been in operation since 1961.

Because of the limitations of only two rainfall gauges in the area and to better account for the variability of monthly rainfall in space and time, we used gridded monthly precipitation data. We developed gridded precipitation data with 1-km resolution between 1957 and 2010 by using rainfall data from 14 national meteorological stations in the Erdos Plateau (Fig.1b). Monthly rainfall data at these 14 stations were downloaded from the China Meteorological Data Sharing Service System (CMDSSS,http://cdc.nmic.cn). We constructed the gridded monthly data using the inverse distance square weighting (IDSW) method due to the moderate topography of the Erdos Plateau in the form of low-relief rolling hills. Figure 1b shows the mean annual precipitation contours of the Erdos Plateauobtained from the gridded data.Within the HRC, precipitation is relatively uniformly distributed because of the flat topography of the region(Yang et al., 2012), but a subtle (~40 mm) increase in precipitation from north to south across the basin can be observed in Fig.1b. In this study, the area-averaged monthly precipitation in the HRC for the period 1963-2010 was estimated by imposing the basin boundaries on the gridded monthly precipitation data and taking the arithmetic average of the grid cells within the HRC boundaries.

The method applied in constructing the gridded precipitation data were further applied in constructing a1-km resolution gridded data setfor monthly pan evaporation between 1957 and 2010 for the Erdos Plateau. The pan evaporation data were based on observations from 200-mm diameter pans that were installed in most stations on the Erdos Plateau and can also be downloaded from CMDSSS (http://cdc.nmic.cn). The average monthly potential evapotranspiration (E_0) in the HRC was estimated from the spatially averaged data of pan evaporation using a local pan coefficient (0.58) for the 200-mm diameter pan.This coefficient was suggested by various investigations of pan coefficients for Chinese meteorological stations (Shi et al., 1986; Fan et al., 2006).

In Fig.2a, the variation patterns of the monthly rainfall and potential evapotranspiration at the catchment scale during 1957-2010 are shown. Both rainfall and evapotranspiration are high in the summer and low in the winter. However, there is a difference in the patterns which may influence the seasonal variation in runoff: the rainfall peak normally arrives in the August but the highest evaporation is exhibited in the June. With respect to these meteorological patterns, the total runoff drops in the Spring and in the early Summer until the heavy rainfall ~~coming~~ in ~~the~~ August, as shown in Fig.2b. In comparison with the rainfall and the potential evapotranspiration, the mean monthly runoff (2.6 mm) and its fluctuation amplitude (0.8-11.9 mm) are quite small. This indicates that most of the precipitation in the HRC returns to the atmosphere by evapotranspiration. During 1957 to 2010, the mean annual P and Q are 350 mm and 32 mm, respectively. The runoff ratio is $Q/P \approx 0.09$. The mean annual potential evaporation in this period is E_0=1248 mm/a, indicating a mean aridity index of $\phi \approx 3.6$. The annual aridity index in this period generally ranged between 2 and 7, covering the semiarid and arid climatic conditions as classified in the scheme recommended by the United Nations Environment Programme (UNEP) (Middleton and Thomas, 1992).

In the HRC,there are interannual fluctuations in E_0, P, and Q.However, no significant trends were detected in theE_0 and$P$data,whereas several regime shifts were found in the streamflow data.Yang et al. (2012) found that the annual regime shifts in streamflow were caused largely by land use policy changes and river water diversions for irrigation. Table 1 shows the mean annual fluxes in four typical periods with different numbers of diversions in the Hailiutu River and major branches during 1957-2010. These diversions influenced the hydrological behavior in the HRC and will be discussed in the following sections. However, before 1967, the Hailiutu River was free ofhydraulic engineering, and the studied area was close to natural conditions.

2.3 Preliminary analysis using (P–Q)/P

In many cases, it is possible to estimate the annual E in a catchment from the annually observed P and $Q$by $P-Q$when the change in storage is sufficiently small. Then it could be treated as the "real" data of the annual E and the shift of annual water balance in the Budyko space could be investigated with the plot of $(P-Q)/P$ versus E_0/P. In this section, we check the validity of this approach in the HRC.

Both the plots of Q/P and $(P-Q)/P$versus E_0/P for the HRC are shown in Figure 3. The annual Q/P value ranges between 0.08 and 0.18, approximately following a linear increasing trend with the aridity index (Figure 3a). If the original Budyko formula is available for annual water balance in the catchment, the annual Q/P value could be calculated as $1-F(\phi)$ and the shift path with the varying aridity index should be a descending curve. However, this trend is contrary to the observed trend of the annual Q/P. Correspondingly, the real annual $(P-Q)/P$ data show a negative trend in the standard Budyko space (Figure 3b). This trend is contrary to the positive E/P trend in the original Budyko framework. In a previous study, Istanbulluoglu et al. (2012) also highlighted this abnormal trend in the North Loup River basin, Nebraska, USA, and they demonstratedthat the trend was due to ignoring the change in storage. They used long-termmonitoring data of groundwater level to estimate the inter-annual change in groundwater storage (ΔG) and replaced the $(P-Q)/P$ data with the $(P-Q-\Delta G)/P$data to reproduce a normal Budyko curve for the basin.

It is a good idea to estimate the change in groundwater storage using groundwater monitoring data. However, this kind oflong-term monitoring was not available in the HRC, China. In addition, the approach of using $(P-Q-\Delta G)/P$ data still has a risk in ignoring the inter-annual change in the soil moisture storage. In this study, we used a hydrological model to estimate the actual annual $E$from monthly modeling steps, in which the groundwater dependent evapotranspiration is incorporated. With the model, both the storage components and the contribution of groundwater for the annual E can be obtained at the catchment scale.

3 Hydroclimatologic models

3.1 The ABCD model

The ABCD model is a conceptual hydrological model with 4 parameters (a, b, c, and$d$) developed by Thomas (1981)to account for the actual evapotranspiration, surface and sub-surface runoff,and storage changes. The ABCD model was originally applied at an annual time step but has been recommended as a monthly hydrological model (Alley, 1984). It was widely appliedas a hydroclimatologic model to investigate the response of catchments ~~on~~ climate change(Vandewiele et al., 1992; Fernandez et al., 2000; Sankarasubramanian and Vogel, 2002; Li and Sankarasubramanian, 2012).

Both the soil water and groundwater storages are considered in the model, as shown in Fig. 4a. At the monthly time step, the change in the soil water storage is determined by

$$W_m - W_{m-1} = P_m - E_m - R_m, \tag{4}$$

where W_{m-1} and W_m are the effective soil water storage at the beginning and the end of the m-th month, respectively; P_m and E_m are the monthly precipitation and evapotranspiration values, respectively; and R_m is the monthly loss of soil water via direct runoff and groundwater recharge. The change in groundwater storage is determined by

$$G_m - G_{m-1} = cR_m - dG_m, \tag{5}$$

where G_{m-1} and G_m represent the groundwater storage at the beginning and the end of the m-th month, respectively; and c and d are two parameters that account for groundwater recharge and discharge from R_m and G_m, respectively. The monthly streamflow is the summation of the monthly direct runoff and groundwater discharge, as follows:

$$Q_m = (1-c)R_m + dG_m. \tag{6}$$

The change in storage in the ABCD model is the summation of the changes in the soil water storage and groundwater storage, which can be expressed as $(W_m - W_{m-1}) + (G_m - G_{m-1})$.

Thomas (1981) proposed a nonlinear function to estimate $(E_m + W_m)$ from $(P_m + W_{m-1})$ as follows:

$$E_m + W_m = \frac{P_m + W_{m-1} + b}{2a} - \sqrt{\left(\frac{P_m + W_{m-1} + b}{2a}\right)^2 - \frac{(P_m + W_{m-1})b}{a}}, \tag{7}$$

where a is a dimensionless parameter, and b is the upper limit of $(E_m + W_m)$. In addition, Thomas (1981) assumed

$$W_m = (E_m + W_m)\exp(-E_{0m}/b), \tag{8}$$

where E_{0m} is the monthly potential evaporation for the m-th month. Substituting Eq. (8) into Eq. (7), the monthly evapotranspiration can be estimated as

$$E_m = \left[\frac{P_m + W_{m-1} + b}{2a} - \sqrt{\left(\frac{P_m + W_{m-1} + b}{2a}\right)^2 - \frac{(P_m + W_{m-1})b}{a}}\right]\left[1 - \exp\left(-\frac{E_{0m}}{b}\right)\right]. \tag{9}$$

Wang and Tang (2014) demonstrated that Eq. (7) can be derived from the generalized proportionalityprinciple and yield an equivalent Budyko-typemodel.

3.2 The ABCD-GE model

To investigate the effects of groundwater dependent runoff and evapotranspiration in basins with both shallow and deep groundwater, the original ABCD model is extended in this study as the ABCD-GE model where ‘GE’ denotes groundwater dependent evapotranspiration. As shown in Fig.4b, a catchment is conceptually divided into two zones where the Zone-1 and Zone-2 represent different areas with deep and shallow groundwater, respectively. Surface water is also included in the Zone-2. The soil water reservoir in the Zone-1 is the same as that in the ABCD model whereas no direct runoff occurs on its surface. In addition, a transition vadose zone is specified between the soil layer and water table to represent the delayed groundwater recharge. In the Zone-2, rainfall and evapotranspiration are the components directly involved in the water balance of groundwater as well as the surface runoff. Thus, three storage components are considered as a chain in the hydrological processes.

Dividing the Zone-1 and Zone-2 in a catchment depends on how groundwater can be accessed by evapotranspiration. It is controlled by the depth of plant roots and the rise of capillary water over groundwater level. In the case study of the HRC, it was observed that some trees have long roots penetrated 2-3 m or more into the earth (Lv et al., 2013), but in general the dominant root zone is less than 2 m below ground surface for shrubs and grasses. When the DWT is larger than 2 m, the contribution of groundwater for evapotranspiration will dramatically ~~decreased in an ignorable level~~(Yin et al., 2015). Thus, it is reasonable to use the contours of 2-m-depth of groundwater as the approximate boundary of the Zone-1 and Zone-2 in the HRC. In the Zone-1, according to the data in Lv et al. (2013), the DWT ranges between 2 m and 110 m. The transition vadose zone is roughly defined as the zone between 2-m-depth below ground surface and 2-m-height above groundwater level. In the assumptions of the ABCD-GE model, this zone could not be influenced by both ~~of~~ the evapotranspiration and groundwater flow processes. Thus, the thickness of the soil layer would be less than 2 m in the model for the HRC. However, one should be aware of that it is not necessary to find the distinct and exact boundaries for the zones, since the ABCD-GE model is a conceptual hydrological model.

In the ABCD-GE model, direct runoff only occurs in the Zone-2 and is assumed to be proportional to the precipitation as $(1-c)P_m$ where c is similar to the dimensionless parameter used in the ABCD model, but now is linked with the precipitation. The total runoff in the catchment is the sum of the direct runoff and groundwater discharge as follows:

$$Q_m = \alpha(1-c)P_m + dG_m, \qquad (10)$$

where α is the ratio of the Zone-2 area to the whole catchmentarea. ~~Using 2 m as the bound value of groundwater depth for the Zone-2, α=0.16 is initially determined in the HRC~~. In comparison with Eq. (6), herein the direct runoff is estimated with the amount of precipitation (αP_m) in the Zone-2, rather than with ~~R_m~~.

Similar to that in the ABCD model, the change in the soil water storage is determined by

$$W_m - W_{m-1} = P_m - E_{1m} - R_m, \qquad (11)$$

where E_{1m} is the monthly evapotranspiration in the Zone-1 determined with Eq. (9), R_m becomes the monthly leakage of soil water, forming the recharge to the transition vadose zone in the Zone-1. The change in storage of this vadose zone is described with

$$V_m - V_{m-1} = R_m - kV_m, \qquad (12)$$

where V_m and V_{m-1}represent the storages in the transition vadose zone at the end and beginning of the m-th month, respectively, and$k$ is the parameter that accounts for groundwater recharge rate as kV_m. In considering of the gain-loss processes of groundwater, the change in the effective groundwater storage is yielded by

$$G_m - G_{m-1} = (1-\alpha)kV_m + \alpha(cP_m - E_{2m}) - dG_m, \qquad (13)$$

where E_{2m} is the monthly evapotranspiration in the Zone-2, which depends on the effective groundwater storage as follows:

$$E_{2m} = gG_mE_{0m}, \qquad (14)$$

where g is a parameter controlling the intensity of groundwater dependent evapotranspiration. Eq. (14) assumes that the evapotranspiration rate in the Zone-2 is simply proportional to both the groundwater storage (which is positively related to groundwater level) and the potential evapotranspiration rate. Thus, the evapotranspiration rate as a whole in the catchment is summarized as

$$E_m = (1-\alpha)E_{1m} + \alpha E_{2m} \text{ .} \tag{15}$$

Eqs. (11)-(13) are solved one by one and finally the value of $G_m$is substituting into Eq. (9) to obtain the runoff. The solutions of the ABCD-GE model are controlled by 7 parameters as: *a*, *b*, *c*, *d*, *g*, *k* and α. The parameter values should be identified with the model calibration process.

4 Model Calibration and Modelling Results

4.1 Model calibration

We applied the ABCD-GE model to estimate the monthly evapotranspiration and the change in storage in the HRC afterthe model parameters were calibrated. The monthly evapotranspiration data were then summed up to estimate the annual evapotranspiration for further analysis. The model calibration was based on the observed monthly streamflow data at the Hanjiamao station and the separated baseflow data.

Because groundwater discharge has been included in the model, a baseflow analysis wasperformed to obtain the expected groundwater discharge for the model calibration. Using the automated hydrograph separation methodHYSEP (Sloto and Crouse, 1996)on the daily streamflow data, such 'observed'groundwater discharge data were obtained for the period 1957-2010. These data were partly reported in Zhou et al. (2013). The baseflow index ranges between 0.80 and 0.95 for the annual streamflow, indicating that groundwater flow is the dominant hydrological process in the HRC. Variation patterns of the monthly groundwater discharge are shown in Fig.2b.

The ordinary least squares (OLS) criterion was applied for parameter estimation. The errors of both log-streamflow and log-baseflow were included in the OLS objective function, as follows:

$$\min U = \sum_{m=1}^{N} \left(e_m^2 + q_m^2\right), \tag{16}$$

where

$$e_m = \ln(\hat{Q}/Q)_m \text{ ,} \quad q_m = \ln(\hat{Q}_b/Q_b)_m \text{ ,} \tag{17}$$

and U is the value of the objective function;N is the number of months; $\hat{Q}$ and Q are the simulated and observed monthly streamflow, respectively; $\hat{Q}_b$ is the simulated monthly groundwater discharge through dG_m in Eq. (12); andQ_b is the 'observed' monthly groundwater discharge obtained from the base flow analysis. The log form errors given in Eq. (17) can be used to obtain homoscedastic residuals (rather than the residual errors) of the normal absolute differences between the observed data and the model outputs(Alley, 1984). The nonlinear optimization algorithmGeneralized Reduced Gradient (GRG) (Lasdonet al., 1978) was used to determine the optimum values of the parameters. The Nash-Sutcliffeefficiency (NSE) (Nash and Sutcliffe, 1970)was also applied to evaluate the performance of the model. The NSE value ranges in ($-\infty$, 1) whereas a higher than zero value is required for a well-perform model.

The parameters in the model were firstly identified using the 1957-1966 data, and this calibrated model was considered to be a 'natural' model due to the minimum impact of human activities during this 10-years period. The initial storage values were also regarded as the unknown parameters to be determined in the calibration process. Changes in the initial conditions generally influenced the simulated results in the first and second years. Therefore, the residual errors in the later years were applied to estimate the parameter values with less influencefrom the initial conditions. A sensitivity analysis was carried out to schematically capture the ranges of the parameter values. The best fitting parameter values obtained through the model calibration are shown in Table 2. The a value approximates to 1.0. In previous studies using the ABCD model, the a value was found generally to be higher than 0.9 (Alley, 1984; Sankarasubramanian and Vogel, 2002; Li and Sankarasubramanian, 2012). The b and d valuesfall into theranges suggested byAlley (1984). The$c$ value is 0.92, indicating that ~~there are~~ 8% of the precipitation in the Zone-2 ~~were~~ transferred to direct runoff during the 1957-1966 period. The fractional area of the shallow groundwater zone, α, is 0.21, which was larger than the current data (16.0%) of the zone with the DWT less than 2 m. Such a difference is reasonablebecause groundwater level before 1967 should be higher than that at present as indicated by the decrease trend of the baseflow began from 1967. The k value controls the rate of groundwater recharge below the transition vadose zone. The transition vadose zone is a necessary component in the HRC as demonstrated by the sensitivity analysis on the k value. When an extremely high value of k is used (k >100), the kV_m value would be almost equal to R_m so that the transition vadose zone does not make sense. However, in this situation the model could never capture the seasonal variation patterns of the runoff and groundwater discharge in the HRC. The best fitting *k* value is significantly less than 1.0, indicating a strong delay effect. Thus, the delayed groundwater recharge is an essential process in this study area.

4.2 Modelling results

For the 1957-1966 period, the mean standard error of the calibrated model is smaller than 15%. The NSE value of the model is 0.51, not very high but significantly larger than zero. It is usuallydifficult to obtain a high NSE value for a catchment with weak seasonal variation in runoff (Mathevet et al., 2006). We used this 'natural' model to estimate the monthly hydrological components during the whole 1957-2010 period. For the runoff estimation, the annual results match the observation much better than the monthly results as indicated by the coefficient of determination (compare Fig. 5a with Fig. 5b). Variations of the monthly total runoff and groundwater discharge estimated by the 'natural' model are shown in Fig. 6a and Fig. 6b, respectively.The model ~~estimated~~ monthly runoff after 1966 are higher than the observed values due to ignoring the impacts of land use changes and increased utilization of water for irrigation. However, the simulated patternsof groundwaterdischarge are similar to the observations: falling in the summer, rising in the winter. This agreement between the simulated and observed patterns demonstrates the ability of the ABCD-GE model in simulating the hydrological behaviors in the studied catchment: significant groundwater-dependent evapotranspiration occurs in the summer, and a strong recovery of storage in the shallow-groundwater zone occurs in the winter due to delayed recharge from the thick vadose zone.

For the periods after 1966, the differences between the model calculated natural annual runoff and the observed values as shown in Fig. 6c could be interpreted as the excess evapotranspiration induced by increasing agricultural water use from river diversion. Enhanced evapotranspiration also occurred in the shallow groundwater zone due to groundwater pumping for irrigation. To evaluate the actual water balance, the following equation

$$E_{ACT} \approx E_{NAT} + (Q_{NAT} - Q_{OBS}), (18)$$

is applied approximately to estimate the actual annual evapotranspiration (E_{ACT}) after 1966 from the 'natural' model result (E_{NAT}) plus the difference of annual runoff between the 'natural' model (Q_{NAT}) and the observation (Q_{OBS}). Thus, the irrigation water use in the catchment is included in E_{ACT}. Results are shown in Fig. 6d. It seems that the difference between E_{NAT} and E_{ACT} is not significantly large in comparison with the mean annual evapotranspiration. The maximum $Q_{NAT} - Q_{OBS}$ value is less than 10% of the mean annual evapotranspiration (~315 mm). Accordingly, the irrigation water use in the HRC did not significantly influence the annual evapotranspiration at the catchment scale. However, it dramatically influenced the streamflow. As shown in Fig. 6a, almost all of the direct runoff was removed from the total runoff after 1987 and groundwater discharge was significantly decreased even though the seasonal patterns ~~were~~ basically remained (Fig. 6b).

4.3 Annual water balance in the standard Budyko space

In Fig. 7a, the E_0/P and E/P data for the annual water balance obtained from the 'natural' model over this 55-years period are plotted in the standard Budyko space. For comparison, both E_{NAT}/P and E_{ACT}/P data are plotted. The E/P values obtained from the 'natural' model(E_{NAT}/P) ~~is~~ a little bit lower than the actual E/P data (E_{ACT}/P). For both data sets, with increase in the aridity index, the evapotranspiration ratio ($F=E/P$) increases almost linearly with ~~the~~ R-square as high as 0.88. When ϕ is larger than 4, the E/P datafall above the line of F=1. Since F=1 is the bound of the mean annual evapotranspiration ratio predicted by the original Budyko hypothesis, the occurrence of such high F values indicates that the normal Budyko formulas, such as Eqs. (2) and (3), cannot be applied in analyzing the annual water balance in the HRC. During extreme dry years when ϕ>4, the annual precipitation is generally less than 290 mm whereas the annual evapotranspiration is generally higher than 300 mm. The excess evapotranspiration is sustained by shallow groundwater.

The effects of groundwater dependent evapotranspiration can be clearly observed when the evapotranspiration ratio is divided into two parts and plotted in the Budyko space separately with respect to the shallow and deep groundwater zones, as shown in Fig.7b. The annual E values in the Zone-1 and Zone-2 are estimated respectively as

$$E_1 = \sum_{m=1}^{12} E_{1m}(W_{m-1}, a, b), \quad \text{and} \qquad E_2 = \sum_{m=1}^{12} E_{2m}(G_m, g), \quad (19)$$

for every year, where E_{1m} was calculated with Eq. (9) whereas E_{2m} was calculated with Eq. (14). The data in Figure 7b were estimated with the parameter values of a, b and g for the 'natural' model. It is obvious that the annual E_1/P values in the Zone-1 (deep groundwater) for the whole range of the aridity index are smaller than 1.0 and fall below the Budyko curve determined by Eq. (2). The low E_1/P value in the Zone-1 is mainly due to the ~~significantly~~ water limited condition. A large portion of precipitation (more than 30%) converts to effective groundwater recharge in the Zone-1 when ϕ is less than 5. The land covers in the deep groundwater zone are dominated by sparse desert grasses which have much lower evapotranspiration rates. The E_1/P trend can be sufficiently fitted by the Budyko curve determined with Eq. (3) for w=0.5. As suggested by Zhang et al. (2001), the plant-available water coefficient, w, ranges between 0.5 and 2.0 where the lower limit refers to short grass or pasture, satisfying the situation in the HRC. However, the relationship between the annual evapotranspiration ratio and the annual aridity index in the shallow groundwater zone definitely could not be explained by any of the normalBudyko formulas, because all the annual F values for the Zone-2 are higher than 1.0. The E_2/P value increases from 1 to 7 when the ϕ value increases from 1.5 to 9.8, approximately following a linear trend. This trend agrees with the relationship betweenE_2 and E_0 ($E_2 \propto E_0$) that described in Eq. (14). When the groundwater storage, G, is relatively stable, the annual E_2/P value would be proportional to the annual E_0/P value and the slope is represented by the annual mean value of gG. In the HRC, the annual mean value of gG is 0.65 according to the 'natural' model. Thus, the annual E_2/P value must be higher than 1.0 when ϕ is higher than 1.5. Such a groundwater dependent evapotranspiration process is the reason for the cases of F>1 occurred at the catchment scale in the HRC. Note that in the original Budyko framework, the $F=\phi$ case denotes an energy limited condition when water supply (only precipitation for mean annual water balance) is sufficient for the evapotranspiration process. The slope of the E_2/P trend (≈0.5) in Figure 7b is less than 1 but is closer to the $F=\phi$ line than the Budyko curve for the ϕ>1 situation. It indicates that in the Zone-2 the evapotranspiration process is in a quasi-energy limited condition, rather than in a water limited condition, because sallow groundwater can effectively serve as an external source of water supply.

5 Discussions

5.1 Controls on *F*>1 cases

It has been demonstrated in Fig. 7a that the annual evapotranspiation ratio, F, would be usually higher than 1.0 when the aridity index, ϕ, is larger than 4.0 in the HRC. In the literature, the F>1 cases were also observed when ϕ is just higher than 1.0 (Cheng et al., 2011; Wang, 2012; Chen et al, 2013). Thus, it is interesting to discuss how the occurrence of the F>1 cases is controlled by the catchment properties when shallow groundwater plays an important role.

The equation for the annual evapotranspiation ratio can be derived from Eqs. (15) and (19) as follows

$$F = (1-\alpha)\frac{E_1}{P} + \alpha g \frac{E_0}{P}\sum_{m=1}^{12}\left(\frac{E_{0m}}{E_0}G_m\right), \qquad (20)$$

where the term E_{0m}/E_0 denotes the proportion of monthly potential evaporation to the annual one with respect to the m-th month. It has been known that the relationship between E_1/P and ϕin the HRCis similar to that predicted by the normal Budyko formulas, as shown in Fig.7b, where E_1/P is less than 1.0. For the groundwater dependent term, defining

$$G_a = \sum_{m=1}^{12}\left(\frac{E_{0m}}{E_0}G_m\right), \qquad (21)$$

as the weighted average of the monthly groundwater storage, Eq. (20) can be replaced by

$$F(\phi) = \frac{(1-\alpha)w\phi}{1+w\phi+\phi^{-1}} + \alpha g \phi G_a, \qquad (22)$$

where E_1/P is represented by Eq. (3). According to Eq. (22), the function $F(\phi)$ is controlled by the parameters, g, w, α and the status of groundwater represented by G_a. As indicated in Eq. (14), $gG_a$is a dimensionless parameter to describe the intensity of groundwater dependent evapotranspiration related to the potentialevaporation. The recommended range of gG_a is 0.5-1.0. In Eq.(22), the term of E_1/P indicates the normal energy-water limited process in Zone-1, whereas the term of groundwater indicates the quasi-energy limited process in Zone-2. The real F value is a mixed result of the different processes.

Typical F- ϕ curves obtained with Eq. (22) are given in Fig.8. It can be seen that the proportion of shallow water table area (α) has large effect on the occurrence of the F>1 case. When the shallow water table area is small (α=0.1), the F>1 case occurs only during extreme dry years. When groundwater dependent evapotranspiration (gG_a) increases, the case F>1 occurs with smaller aridity index. The plant available water coefficient (w) also influences the occurrence of the F>1 case. A larger value of $w$shifts the F-ϕcurves (Fig.8b) to the left side indicating that the F>1 case could occur with smaller aridity index.

5.2 Using effective precipitation and modified Budyko space

The standard Budyko space assumes that the potential water supply for evapotranspiration is only rainfall in a catchment. This is true for the long-term average water balance, but exceptions might exist for the annual or intra-annual behaviors. Wang (2012) and Chen et al. (2013) argued that the reduction of storage in a period should be regarded as one of the water supply components and suggestedan approach to replace the evapotranspiration ratio and the dryness index by $E/(P-\Delta S)$ and $E_0/(P-\Delta S)$, respectively, where ΔS is the storage depletion in a studied period and $P-\Delta S$ regarded as the effective precipitation. In this modified Budyko space, evapotranspiration is always less than the water supply so that the original Budyko hypothesis could be applied for small time-scale problems. In this section we attempt to check the characteristics of the annual water balance data in the HRC using such a modified Budyko space. With the results of the ABCD-GE model, the total change in storage for a year can be estimated as

$$\Delta S=\sum_{m=1}^{12}[(1-\alpha)(W_m+V_m-W_{m-1}-V_{m-1})+(G_m-G_{m-1})], \qquad (23)$$

where m is the number of the months in the year, W_0, V_0 and G_0 for m=0 denoting the respective storagecomponents at the end of the last year. Results are shown in Fig.9. It can be seen in the modified Budyko space that the annual water balance data fall into the zone below the limitation: $E/(P-\Delta S)<1$, even below the modified Budyko curve obtained with Eq. (2) using the newly defined evapotranspiration ratio and dryness index. However, the shift path of the data points can not be captured by a single Budyko curve in the modified Budyko space with a constant value of the specific parameter. Such as shown in Fig. 9, the rising trend of $E/(P-\Delta S)$ with the increasing $E_0/(P-\Delta S)$ seems too weak in comparison with any one of the normal Budyko curves determined by the formula of Zhang et al. (2001). Furthermore, the $E/(P-\Delta S)$ value approaches a stable value around 0.90 with the very high$E_0/(P-\Delta S)$ values. It indicates that at least 10% of $P-\Delta S$ is contributed to the annual runoff, in terms of $Q/(P-\Delta S)$. This portion of the water supply seems to be inaccessible for the annual evapotranspiration process.

The difficultiesin using the effective precipitation defined byWang (2012) and Chen et al. (2013) are the unkown ΔS for an investigated time step and the possible existence of the inaccessible part of ΔS for evapotranspiration. Consequently, the estimation of $E/(P-\Delta S)$valueis not straightforward, but requires a complex iteration process. In the original Budyko framework for steady state water balance, the water supply (only precipitation) does not depend on both evapotranspiration and runoff so that the aridity index is an independent variable in assessing the behaviors of the catchments. However, the water supply represented by the effective precipitation is influenced by the evapotranspiration-runoff processes due to the feedback mechanism. This cross-dependency between the water supply and evapotranspiration significantly reduces the efficiency of using the modified Budyko space in analyzing the shift of annual water balance in a catchment. In contrast, it would be an efficient and straightforward approach to extend formulas for annual water balance in the standard Budyko space, such as Eq. (22), keeping an independent index (ϕ) for the climatic conditions.

5.3 Landscape-driven and human-controlled shifts of annual water balance

As illustrated in Figure 6d, the actual evapotranspiration in the HRC has been enhanced by ~~the~~ human activities. This impact might exist in both the shallow and deep groundwater zones. Crops in the HRC are mainly planted in the depressions and terrace lands with shallow groundwater, especially in the river valley. Crops require much more water than the precipitation for growing. For example, maize could consume more than 3 times of rainfall water in growing seasons (Zhou et al., 2013). Thus, irrigation is necessary to maintain the agricultural production. In the farmlands far away from the rivers, groundwater was abstracted from wells for irrigation. In theriver valley, irrigation wasrealized with diversions and channels. Therefore, ~~increasing~~ in evapotranspiration in the shallow groundwater zone is dominated by irrigation. Along the river, the area of the surface water body was significantly enlarged in reservoirs leading to increase in surface water evapotranspiration loss. It is equivalent to increase in groundwaterdependent evapotranspiration in this study because surface water is also included in the shallow groundwater zone. As a result, the shift of the annual water balance in the Budyko spa~~p~~ce was partly caused by change in land use and controlled by regulation of river water for irrigation.

Recently, Jaramillo and Destouni (2014) developed a method to assess the landscape-driven change in the mean evapotranspiration ratio using the difference between the actual change in the F value and the climate-driven change in the F value following the Budyko framework. In this section, we extend their method to assess the landscape-driven change in annual water balance in the HRC. The years between 1957-1966 is selected from Table 1 as the reference period. Changes are evaluated for the different average values of the annual E/P data in the different periods listed in Table 1. The climate-driven change is estimated with the annual E_{NAT} values obtained from the 'natural' model, using a formula similar to Jaramillo and Destouni (2014), as follows

$$\Delta\left(\frac{E_{LD}}{P}\right)=\Delta\left(\frac{E_{ACT}}{P}\right)-\Delta\left(\frac{E_{NAT}}{P}\right), \quad (24)$$

where $\Delta(E_{\mathrm{LD}}/P)$ denotes the landscape-driven change in comparison with the 1957-1966 period. However, this quantity index includes the landscape changes driven by both climatic force and human activities. To check how this index is correlated with the increasing impacts from the reservoirs and diversions in rivers, following Jaramillo and Destouni (2015), the intra-annual variabilityof the monthly runoff (CV_{Q}) was applied. The $CV_{\mathrm{Q}}/CV_{\mathrm{P}}$ value was estimated to reveal the separate influence of such a human-controlled flow regulation from the mixed human-climate controlling, where CV_{P} is the intra-annual variabilityof the monthly rainfall.

Results of the $\Delta(E_{\mathrm{LD}}/P)$ and $\Delta(CV_Q/CV_P)$ data ~~for the three periods after 1966 are shown~~ in Figure 10. The $\Delta(E_{\mathrm{LD}}/P)$ values ~~of the~~ periods are all positive but not big (less than 6%), indicating a slight increase in the evapotranspiration ratio after 1966 driven by changes in natural landscape and human controlled land use. The $\Delta(CV_Q/CV_P)$ values show a significant fluctuation around zero but also limited in a small range (±5%). Both the $\Delta(E_{\mathrm{LD}}/P)$ and $\Delta(CV_Q/CV_P)$ ~~values reach to the maximum during~~ ~~1968-1987~~. Fluctuations of these data could not be fully explained by the increasing number of diversions in the rivers. The negative $\Delta(CV_Q/CV_P)$ ~~value in the 1957-1967~~ period may be due to construction of the two reservoirs since reservoirs commonly smooth the variation of streamflow. ~~During 1968-1987~~, the $\Delta(CV_Q/CV_P)$ value turned to positive when 5 new diversions was built, indicating the opposite impacts of the reservoirs and diversions. It is possible that the streamflow was disturbed by the regulation of water for irrigation on these diversions with small overflow dams. Decrease in the $\Delta(CV_Q/CV_P)$ value ~~during 1988-2010~~ may be caused by the control of river water use under some government policies to prevent ~~the~~ desertification (Yang et al., 2012; Zhou et al., 2015). The following decrease in $\Delta(E_{LD}/P)$ ~~value for the~~ 1988-2010 period is not significant, seems indicating alternative irrigation practice in farmlands (for example pumping groundwater) so that the real water consumption was reduced but still on a high level. As a result, utilization of surface water and groundwater for irrigation can increase the frequency of the F>1 cases.

5.4 Limitation Remarks

Attention should be paid to the simplifications in the conceptual model extended from the ABCD model, when the equations and formulas are applied in complicated catchments. The ABCD model assumes that the storage-evapotranspiration relationship is controlled by the parameters a and b whereas the physical interpretation of them is difficult (Alley, 1984). Equation (8) in the ABCD model is also hypothesized from a simplified storage-loss process ~~that~~ controlled by the parameter b(Thomas, 1981). Sankarasubramanian and Vogel (2002) suggested that the b value for annual water balance could be approximately represented by the maximum soil moisture field capacity plus maximum E_0 for ϕ<1 or maximum P for ϕ≥1. The a value is generally estimated in a small range between 0.95 and 1.0. In this study, the model output is not sensitive to the a value. The correlation between a and b may exist because both of them are positively related with E_m+W_m in Eq. (7).The ABCD model neglects the possibility of groundwater-dependent evapotranpiration which has been incorporated in the ABCD-GE model. The ABCD-GE model divides the area into shallow and deep groundwater zones, without considering a complicated spatial distribution of groundwater depth. For the shallow groundwater zone, the evapotranspiration is assumed to be proportional to groundwater storage. Nonlinear behavior in groundwater dependent evapotranspiration could be further included if it can besuccessfully parameterized. Linear groundwater storage-discharge relationship is adopted in both of the ABCD and ABCD-GE models. These simplifications could cause systematic errors in modeling a catchment where the nonlinear behaviors in the hydrological processes are significant.

In fact, when the Budyko framework is applied for small time-scale water balance in a catchment, the other additional sources of water supply should be considered, apart from groundwater. Significant changes in soil moisture, snow cover or frozen water in cold regions could also cause 'abnormal' shift of annual water balance for a catchment in the standard Budyko space. The effects of these storage components are ~~ignorable~~ in the HRC but may be essential in other study areas. In particular, the special processes in cold regions are not included in the ABCD-GE model. However, one can refer to Martinez andGupta (2010) who proposed the snow-augmentedABCD model, which is easy to be incorporated into an extension of the ABCD-GE model.

6 Conclusions

The Budyko framework was developed for long-term mean annual water balance in catchments, which estimates the evapotranspiration ratio (F) as a function of the aridity index(ϕ). It can be represented by curves for the F-ϕ relationship in the standard Budyko space that were determined by the original Budyko formula without any parameter or formulas with a catchment specific parameter. It is interesting to investigatewhether the Budyko space can be also applied to capture the annual water balance in a catchment with ~~the~~ varying dryness. However, the shift of annual water balance in the standard Budyko space could be significantly different from that presumed from the normalBudyko curves, in particular, when the cases of F>1 occur as that have been observed in a number of catchments.

In this study, we highlighted the effects of groundwater dependent evapotranspiration in triggering the abnormal shift of annual water balance in the standard Budyko space. A conceptual monthly hydrological model, the ABCD-GE model, wasdeveloped from the widely used ABCD model to incorporate the groundwater-dependent evapotranspiration in the zone with shallow water table and delayed groundwater recharge in the zone with deep water table. The model wassuccessfully applied to analyze the shift of annual water balance in the Hailiutu River Catchment (HRC), China, where 16% of the area is occupied with shallow groundwater (depth to water table is less than 2 m).

The results show that the normalBudyko formulasare not applicable for the interannual variability of catchment water balance in the standard Budyko space when groundwater dependent evapotranspiration is significant. The shift of annual water balance in the F-ϕspace is a combination of the Budyko-type response in the deep groundwater zone and the quasi-energy limited condition in the shallow groundwater zone. Shallow groundwater supplies excess evapotranspiration during extreme dry years, leading to F>1 cases. The occurrence of the F>1 cases depends on the proportion area of the shallow groundwater zone, the intensity of groundwater dependent evapotranspiration and the catchment properties determining the

Budyko-type trend in the deep groundwater zone. Water utilization for irrigation may enhance this excess evapotranspiration phenomenon. The modified Budyko space with the effective precipitation incorporating the change in storage can force *F* values below 1.0. However, the computation is ~~tidious in dealing~~ with the feedback between the water supply and evapotranspiration loss as well as the existence of inaccessible storage for evapotranspiration. The empirical formula proposed in this study for the standard Budyko space provides a straightforward method to predict the trend of annual water balance with the varying dryness.

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Table 1. Mean annual fluxes in the Hailiutu River catchment (HRC) in different periods

Periods	P(mm)	E_0(mm)	Q(mm)	NumberOf Diversions(reservoirs)†
1957-1966	387.0	1230.2	42.3	0(0)
1967-1987	337.0	1269.6	32.6	4(2)
1988-1997	329.9	1240.2	23.4	9(2)
1998-2010	352.8	1234.0	28.0	10(2)

† According to Yang et al. (2012).

Table 2. Best fitting parameters of the ‘natural’ model for the HRC.

Period	a	b	c	d	g	k	α	Error†	NSE
		mm		$\times10^{-2}$	$\times10^{-2}$ m^{-1}	$\times10^{-2}$		(%)	
1957-1966	0.97	33	0.92	4.53	1.00	1.68	0.21	13.9	0.51

†Mean standard errors of the monthly runoff and groundwater discharge

Figure Captions:

Figure 1. Geographic information of the study site: (a) location of the study area in north central China; (b) Distribution of meteorological stations in the Erdos Plateau (green points) and contours of mean annual precipitation plotted from 1-km resolution gridded precipitation data; (c) Characteristics of landscape according to Lv et al. (2013).

Figure 2. The monthly meteorological data (a) and streamflow-baseflow data (b) from 1957 to 2010 in the HRC.

Figure 3. The plots of the annual Q/P data (a) and $(P-Q)/P$ data (b) versus the aridity index in the HRC. The dashed lines are determined with the original Budyko formula, Eq. (2). The solid lines are the correlation curves of the scatter data points.

Figure 4. Schematic representations of the ABCD model (a) and ABCD-GE model (b). W and V are the effective soil water storage and the effective storage in the transition vadose zone, respectively. G is the effective groundwater storage.

Figure 5. Correlation plots of the observed and simulated monthly (a) and annual (b) results for the runoff data in 1957-2010. The simulated results are obtianed with the 'natural' model calibrated with theobservation dataduring 1957-1966 when the impacts of human activities are minimum. Both data of total runoff and groundwater discharge are applied for the correlation analysis without bias.

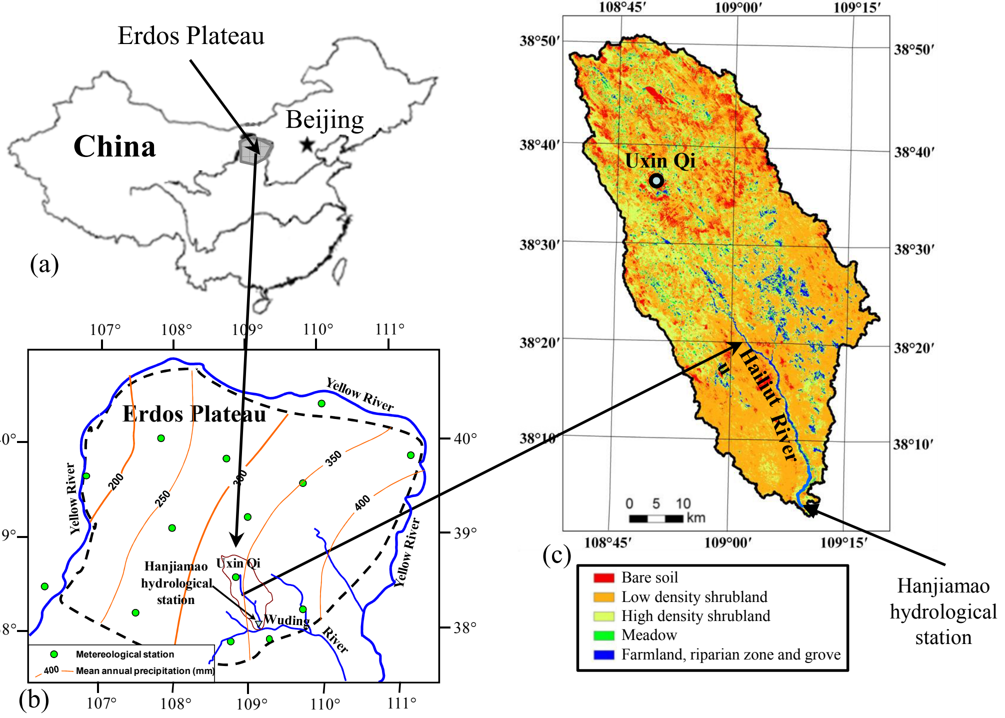
Figure 6. Simulated results of the ‘natural’ ABCD-GE model in comparison with the observation data in the HRC from 1957 to 2010, including: Monthly runoff (a), groundwater discharge (b), annual runoff (c) and annual evapotranspiration (d). The actual evapotranspiration in (d) was estimated with Eq. (18).

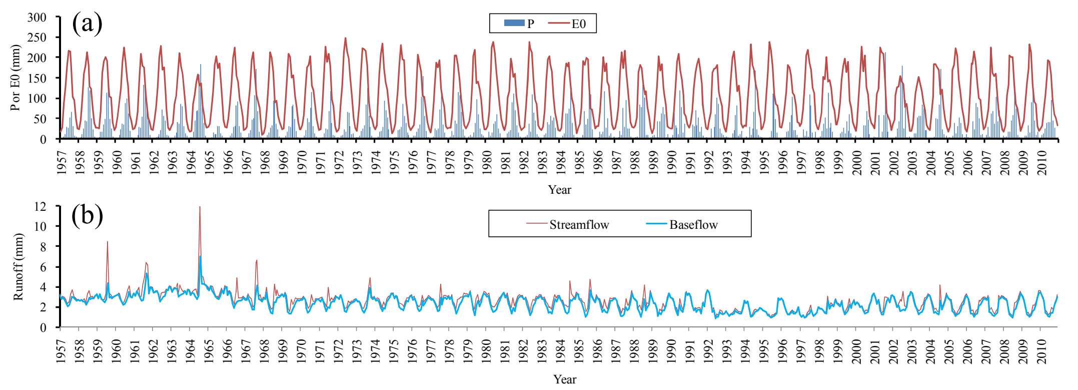
Figure 7. Plots of the annual evapotranpiration ratio in the HRC versus the annual aridity index in the standard Budyko space: (a) using the E_{NAT} data estimated with the 'natural' model and the E_{ACT} data estimated with Eq. (18); (b) using the E_1 data for the Zone-1 and the E_2 data for the Zone-2 that estimated with Eq. (19) on the basis of the 'natural' model .The dashed lines are the linear correlation curves for the 'natural' model data.The F=ϕ line represents the energy-limited condition.

Figure 8. The typical F-ϕ curves for annual water balance in the standard Budyko space determined with Eq. (22) when w=0.5 (a) and w=2.0 (b). The solid and dashed line curves are estimated using gG_a=0.5 and gG_a=1.0, respectively. The gray blocks denote the potential F>1 zones. The actual data of the HRC are shown as the scatter points.

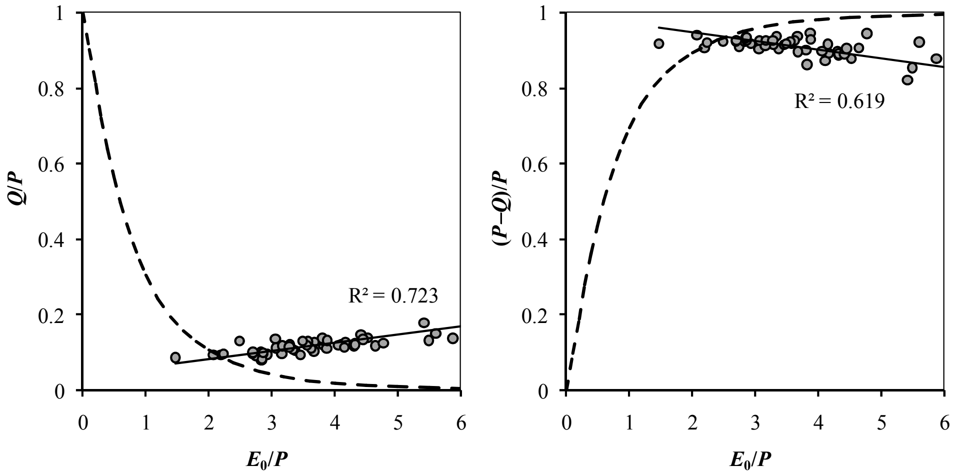
Figure 9. Annual water balance data in the modified Budyko space with the effective precipitation defined by Wang (2012). Dots are the data obtained for the HRC using the 'natural' model. The curve represents the normal Budyko curves determined with Eq.(3)using$E_0/(P-\Delta S)$ and $E/(P-\Delta S)$, respectively, instead of F and ϕ. The dashed line approximately represents the actual bound of the $E/(P-\Delta S)$ data.

Figure 10. Historgram of the $\Delta(E_{\mathrm{LD}}/P)$data determined with Eq. (24) and the $\Delta(CV_Q/CV_P)$ data determined with the intra-annual variabilities of the monthly runoff (CV_Q) and rainfall (CV_P) for the different periods in the HRC. The numbersof diversions(reservoirs) are shown on the top of the blocks according to Table 1.

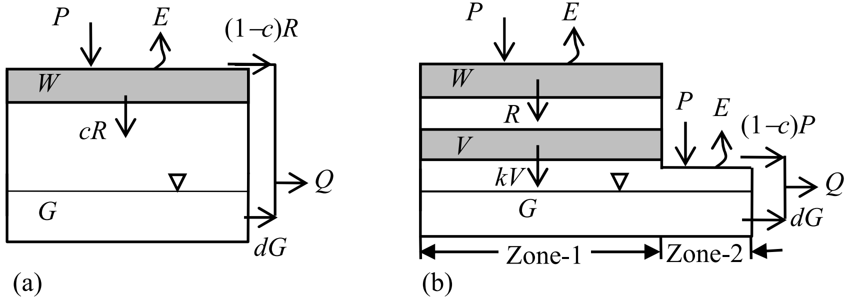

Erdos Plateau
Beijing
China
(a)
107°
108°
109°
110°
111°
40°
39°
38°
Erdos Plateau
Yellow River
Yellow River
Yellow River
200
250
300
350
400
Hanjiamao
hydrological
station
Uxin Qi
Wuding
River
Metereological station
400 Mean annual precipitation (mm)
(b)
108°45′
109°00′
109°15′
38°50′
38°40′
38°30′
38°20′
38°10′
Uxin Qi
Hailiut River
0 5 10
km
(c)
Bare soil
Low density shrubland
High density shrubland
Meadow
Farmland, riparian zone and grove
Hanjiamao
hydrological
station

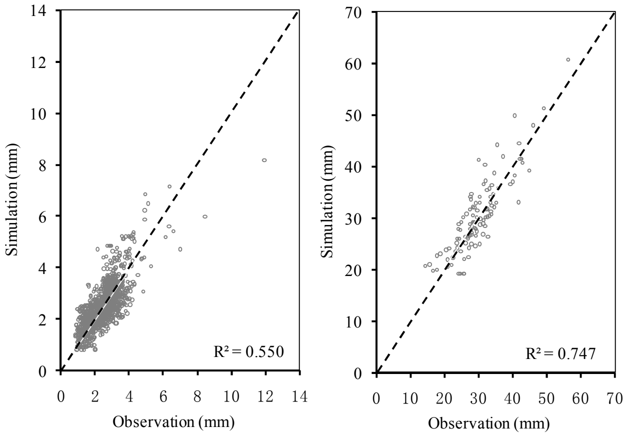


(a)
P
E0
P or E0 (mm)
Year
(b)
Streamflow
Baseflow
Runoff (mm)
Year

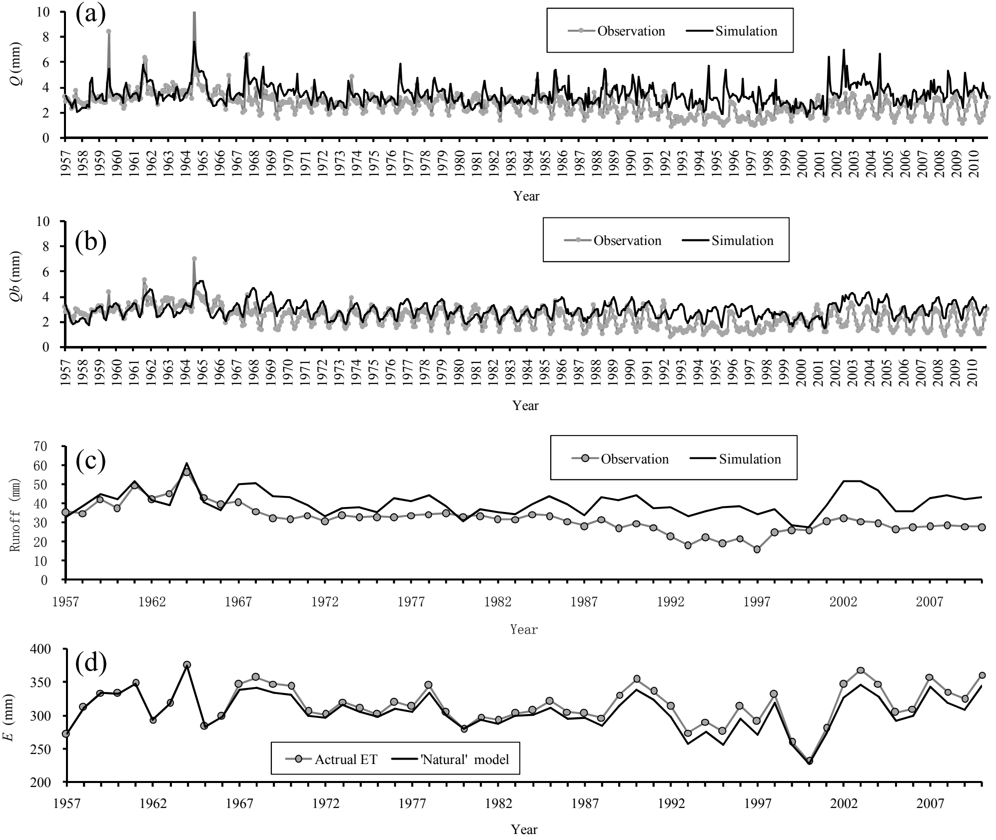


1
0.8
0.6
0.4
0.2
0
Q/P
R² = 0.723
0
1
2
3
4
5
6
E0/P
(P–Q)/P
R² = 0.619
E0/P

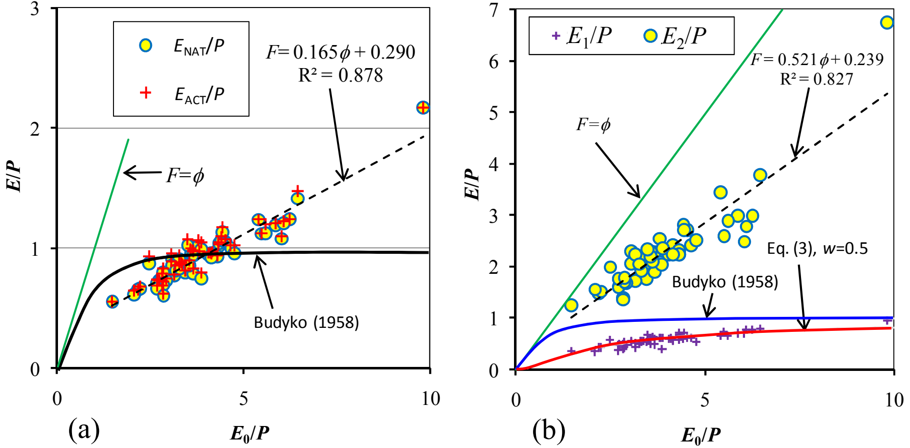

P
E
(1−c)R
W
cR
Q
G
dG
(a)
P
E
W
R
P
E
(1−c)P
V
kV
Q
G
dG
Zone-1
Zone-2
(b)



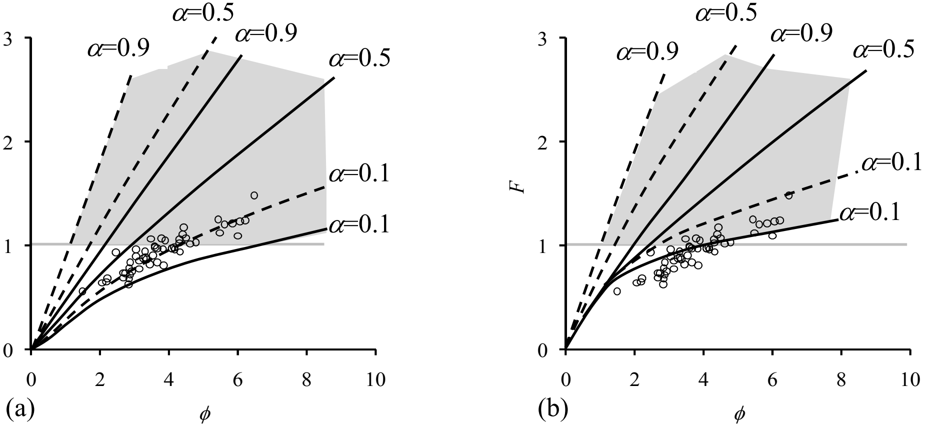
Simulation (mm)
Observation (mm)
R² = 0.550
0
2
4
6
8
10
12
14
Simulation (mm)
Observation (mm)
R² = 0.747
0
10
20
30
40
50
60
70



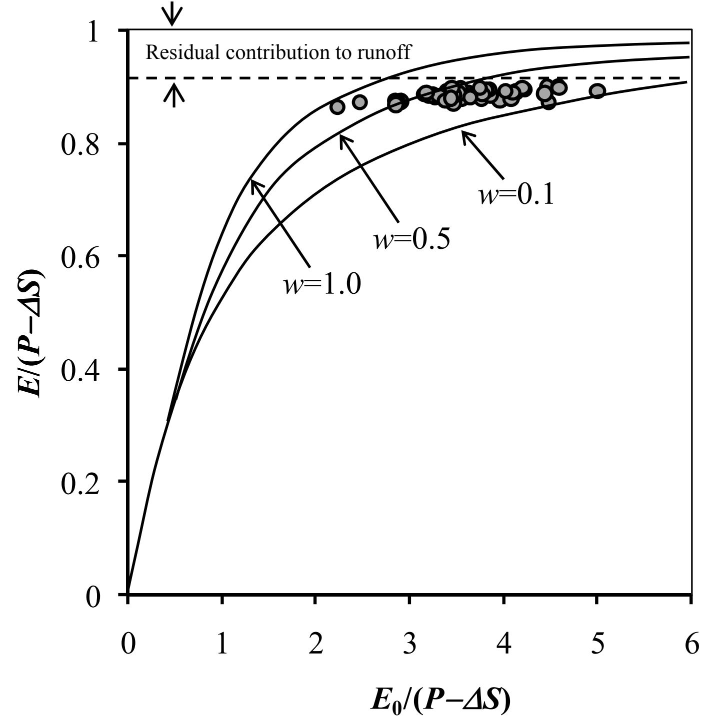
(a)
Observation
Simulation
Q (mm)
Year
(b)
Observation
Simulation
Qb (mm)
Year
(c)
Observation
Simulation
Runoff (mm)
Year
(d)
Actrual ET
'Natural' model
E (mm)
Year

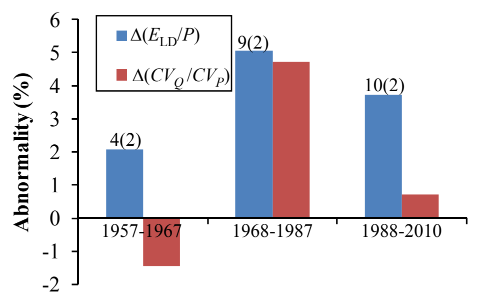


E_{NAT}/P
E_{ACT}/P
$F = 0.165\phi + 0.290$
$R^2 = 0.878$
$F=\phi$
Budyko (1958)
E/P
E_0/P
(a)
E_1/P
E_2/P
$F = 0.521\phi + 0.239$
$R^2 = 0.827$
$F=\phi$
Eq. (3), w=0.5
Budyko (1958)
E/P
E_0/P
(b)



α=0.5
α=0.9
α=0.9
α=0.5
α=0.1
α=0.1
3
2
1
0
F
0
2
4
6
8
10
(a)
ϕ
α=0.5
α=0.9
α=0.9
α=0.5
α=0.1
α=0.1
(b)
ϕ


Residual contribution to runoff
w=1.0
w=0.5
w=0.1
E/(P−ΔS)
E0/(P−ΔS)
0
0.2
0.4
0.6
0.8
1
0
1
2
3
4
5
6



$\Delta(E_{LD}/P)$
$\Delta(CV_Q/CV_P)$
Abnormality (%)
6
5
4
3
2
1
0
-1
-2
4(2)
9(2)
10(2)
1957-1967
1968-1987
1988-2010