Hydrological Sciences Journal
Publication details, including instructions for authors and subscription information:
http://www.tandfonline.com/loi/thsj20

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Accepted author version posted online: 15 May 2014.Published online: 11 Nov 2014.

Hydrological Sciences Journal – Journal des Sciences Hydrologiques, 2015
http://dx.doi.org/10.1080/02626667.2013.866710

Understanding hydrological trends by combining the Budyko hypothesis and a stochastic soil moisture model

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Received 26 January 2013; accepted 1 October 2013

Editor D. Koutsoyiannis; **Associate editor** A. Porporato

Abstract The actual evapotranspiration and runoff trends of five major basins in China from 1956 to 2000 are investigated by combining the Budyko hypothesis and a stochastic soil moisture model. Based on the equations of Choudhury and Porporato, the actual evapotranspiration trends and the runoff trends are attributed to changes in precipitation, potential evapotranspiration, rainfall depth and water storage capacity which depends on the soil water holding capacity and the root depth. It was found that the rainfall depth increased significantly in China during the past 50 years, especially in southern basins. Contributions from changes in the water storage capacity were significant in basins where land surface characteristics have changed substantially due to human activities. It was also observed that the actual evapotranspiration trends are more sensitive to precipitation trends in water-limited basins, but more sensitive to potential evapotranspiration trends in energy-limited basins.

Key words hydrological trend; Budyko hypothesis; evapotranspiration; climate change

Comprendre les tendances hydrologiques en combinant l'hypothèse de Budyko et un modèle stochastique de l'humidité du sol

Résumé Les tendances de 1956 à 2000 de l'évapotranspiration réelle et de l'écoulement de cinq grands bassins chinois ont été étudiées en combinant l'hypothèse de Budyko et un modèle stochastique de l'humidité du sol. Sur la base des équations de Choudhury et Porporato, les tendances de l'évapotranspiration réelle et de l'écoulement ont été attribuées à des changements des précipitations, de l'évapotranspiration potentielle, de la hauteur des précipitations et de la capacité de stockage d'eau, qui dépend de la capacité de rétention du sol et de la profondeur de l'enracinement. Il a été constaté que la hauteur de pluie a considérablement augmenté en Chine au cours des 50 dernières années, en particulier dans les bassins du Sud. Les contributions de l'évolution de la capacité de stockage d'eau étaient importants dans les bassins où les caractéristiques de surface ont considérablement changé en raison des activités humaines. On a également observé que les tendances de l'évapotranspiration réelle sont plus sensibles à l'évolution des précipitations dans les bassins pauvres en eau, mais plus sensibles aux tendances de l'évapotranspiration potentielle dans les bassins limités en énergie.

Mots clefs tendance hydrologique ; hypothèse de Budyko ; évapotranspiration ; changement climatique

1 INTRODUCTION

The hydrological cycle has been greatly influenced by climate changes and human activities in the past five to ten decades (Solomon *et al*. 2007). It is important to understand the hydrological trends that have occurred in the past in order to predict future changes (Cong *et al*. 2010). Long-term trends in the terrestrial hydrological fluxes in the context of climate and land-use changes have been extensively investigated (Gleick 1989, Peterson *et al*. 2002, Huntington 2006, Dai *et al*. 2009). There are four major hydrological variables: precipitation, runoff, actual evapotranspiration and water storage change. The changes in precipitation and runoff can be calculated relatively easily from observations at weather stations and river gauges. Regional and global evapotranspiration may be derived from satellite observations (Tang *et al*. 2010), whose temporal coverage is, however, generally too short. The terrestrial water

budget method can also be used to estimate evapotranspiration, but it does not provide further insights into the underlying cause of the changes in evapotranspiration. The water storage change is usually ignored for long-term analysis (Zhang *et al*. 2001, Roderick and Farquhar 2011). The Budyko hypothesis, which describes the relationship between actual evapotranspiration and potential evapotranspiration, has been widely used to examine the terrestrial water budget at the annual, basin scale (Budyko 1958, 1963, Budyko 1974, Milly 1994, Choudhury 1999, Zhang *et al*. 2001, Sankarasubramanian and Vogel 2002, Yang *et al*. 2007, Donohue *et al*. 2010). It can also be used to analyse the inter-annual variations in evapotranspiration or runoff (Donohue *et al*. 2011, Roderick and Farquhar 2011). Several empirical and physically-based equations have been proposed to parameterize the Budyko curve (Budyko 1958, 1963, Pike 1964, Budyko 1974, Fu 1981, Choudhury 1999, Zhang *et al*. 2001, Sankarasubramanian and Vogel 2002, Gerrits *et al*. 2009, Jothityangkoon and Sivapalan 2009). Some of these equations, such as those of Fu (1981) and Choudhury (1999), have one empirical coefficient that needs to be determined *a priori*. The performances of these equations are similar in terms of estimating annual actual evapotranspiration (Zhang *et al*. 2004, Potter and Zhang 2009) and in this study we use the Choudhury (1999) equation.

According to the Budyko hypothesis, changes in actual evapotranspiration might be induced by changes in precipitation, potential evapotranspiration and the shape factor of the Budyko curve. Thus a decreasing trend in potential evapotranspiration (pan evaporation or reference evapotranspiration), which has been observed in the past 50 years over large parts of the global land surfaces (Peterson *et al*. 1995, Roderick and Farquhar 2004, Cong *et al*. 2009b, Fu *et al*. 2009), most likely (but not necessarily) indicates a decreasing trend in actual evapotranspiration due to the complementary relationship between actual evapotranspiration and potential evapotranspiration in non-humid regions (Bouchet 1963, Brutsaert and Parlange 1998, Yang *et al*. 2006). Previous studies mainly focused on the impacts of changing climate factors, i.e. changing precipitation and potential evapotranspiration, on actual evapotranspiration. For example, Ma *et al*. (2008) and Zhao *et al*. (2009) used the Budyko hypothesis to distinguish between the influence of climate changes and human activities on actual evapotranspiration. Roderick and Farquhar (2011) and Donohue *et al*. (2011) presented a simple framework for attributing variations in runoff to changes in climatic conditions and catchment properties based on the differentials of the Choudhury (1999) equation. At that time, the changes in the shape factor of the Budyko curve were not considered quantitatively. Since the shape factor of the Budyko curve also strongly depends on climatic and land surface conditions (Yang *et al*. 2007, Roderick and Farquhar 2011), we assume herein, to understand the hydrological trends, that the Budyko curve changes yearly, which means that the parameter in the equation parameterizing the Budyko curve changes yearly. This assumption is helpful to distinguish the effects on hydrological trends of climate change and land-use change (Tomer and Schilling 2009, Renner *et al*. 2012). In addition, the empirical parameters, such as in the Fu (1981) equation and the Choudhury (1999) equation, still have no clear physical meaning. Porporato *et al*. (2004) used a stochastic soil moisture model to study the average water balance, and their final equation has similar performance to that of other equations describing the Budyko hypothesis. In Porporato *et al*. (2004), the parameter that controls the shape of the Budyko curve is defined as the ratio of water storage capacity to rainfall depth and thus has clear physical meaning; however, the equation is not differentiable. Recently, Donohue *et al*. (2012) combined the equations of Choudhury (1999) and Porporato *et al*. (2004) to quantitatively estimate the shape factor of the Budyko curve and its long-term variation. The sensitivity of runoff in the Murray-Darling Basin was analysed based on this new model.

In this study, the aim has been to understand the hydrological trends in five major basins in China from 1956 to 2000. First, the Budyko framework (the Choudhury 1999 equation and its differentiated form) is applied to determine what the trends in actual evaporation and runoff can be attributed to. Here, the contribution of the trend in the Budyko curve shape (or the parameter in the Choudhury 1999 equation) is considered. Second, we combine the Budyko hypothesis and the stochastic soil moisture model, i.e. the Choudhury (1999) equation and the Porporato *et al*. (2004) equation, with a power-law relationship linking the parameters in the two equations (unlike Donohue *et al*. 2012, who used a linear function relating the two parameters, to obtain an equation with one physical-meaning parameter). Based on the new equation, the trends in actual evapotranspiration and runoff in the five major basins in China are converted into trends in precipitation, potential evapotranspiration, rainfall depth and water

storage capacity. Furthermore, the sensitivities of hydrological trends in each basin are discussed.

2 DATA DESCRIPTION

Figure 1 shows the five major basins in China where the measurements are available. Basins 1, 2 and 3 are in northern China with a semi-arid climate (water-limited basins) and basins 4 and 5 are in southern China with a relatively humid climate (energy-limited basins). The basin hydrological characteristics are given in Table 1. The daily precipitation and pan evaporation data for the period 1956–2005 at 317 weather stations were collected from the China Meteorological Data Sharing Service System and potential evapotranspiration was calculated from pan evaporation data following Fu *et al*. (2004). A kriging method was used to interpolate the weather station data to obtain basin-averaged values of annual precipitation, annual potential evaporation and annual average rainfall depth (Cong *et al*. 2009b). The annual runoff data for the period 1956–2000 at the river outlet hydrological stations of the five basins were obtained from the Water Resources Information Centre, Ministry of Water Resources, China (Cong *et al*. 2010). The results of GRACE (the Gravity Recovery and Climate Experiment) show that the annual water storage changes between 2003 and 2008 in these five basins were less than 20 mm/year (Swenson and Wahr 2002), i.e. <5% of precipitation, so the actual evapotranspiration can be estimated using the water budget method, ignoring water storage change. Since we focus on the hydrological trend, this simplification is acceptable.

In the northern basins (1–3), annual precipitation is far smaller than annual potential evapotranspiration, and annual actual evapotranspiration takes up about 90% of the annual precipitation. In the southern basins (4 and 5), annual precipitation exceeds annual potential evapotranspiration, and annual actual evapotranspiration is about 50% of annual precipitation.

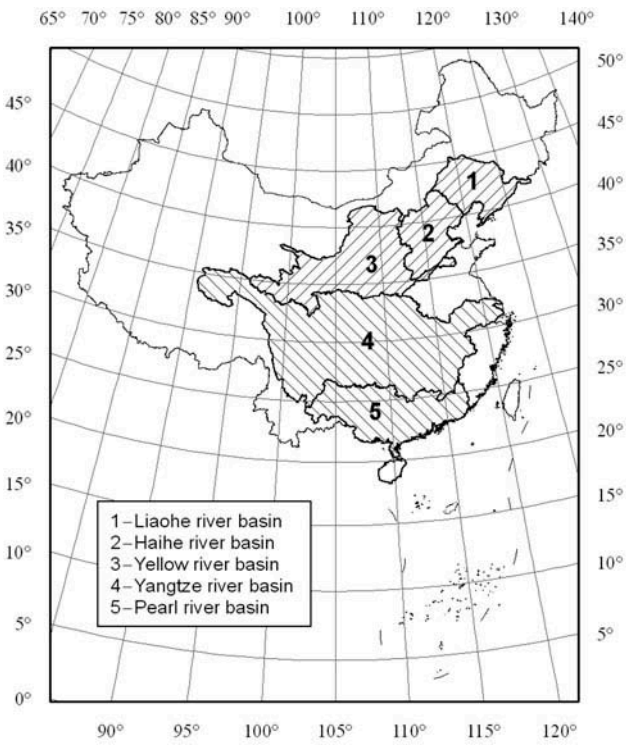

Fig. 1 The five major basins in China.

3 METHODOLOGY

3.1 Budyko hypothesis and Choudhury equation

We here consider the Choudhury (1999) equation, hereafter referred to as the Choudhury equation, to describe the Budyko hypothesis because it combines dimensional analysis with mathematical reasoning (Yang *et al*. 2008).

The Choudhury equation is expressed as:

$$\varepsilon = \frac{\phi}{(1+\phi^n)^{1/n}} \tag{1}$$

Table 1 Hydrological characteristics of five major basins in China for the period 1956–2000.

Basin no.	Name	Area (10^4 km^2)	P (mm/year)	E_0 (mm/year)	R (mm/year)	E (mm/year)	E_0/P	E/P
1	Liaohe	31.5	554	992	60	493	1.79	0.89
2	Haihe	31.9	523	1086	32	491	2.07	0.94
3	Yellow	79.5	444	1074	49	395	2.42	0.89
4	Yangtze	178.1	1031	890	516	515	0.86	0.50
5	Pearl	57.5	1529	964	787	742	0.63	0.49

P: precipitation, E_0: potential evapotranspiration; R: runoff; E: actual evapotranspiration.

where $\varepsilon = E/P$, $\phi = E_0/P$, E is annual actual evapotranspiration, P is annual precipitation, E_0 is annual potential evapotranspiration, and n is a parameter related to climate factors and land characteristics, such as basin size, precipitation intensity, seasonality, slope and soil water storage capacity. (Choudhury 1999, Yang *et al*. 2008, Roderick and Farquhar 2011).

3.2 Stochastic soil moisture model and Porporato equation

The Porportato *et al.* (2004) equation is referred to hereafter as the Porporato equation. The occurrence of rainfall is idealized as a Poisson process of rate λ; the rainfall depth is assumed to be an independent random variable (h), described by an exponential probability density function; and the daily evapotranspiration loss is assumed to increase linearly with relative soil moisture. Based on these assumptions, the stochastic soil moisture model is used to describe the probability of the soil moisture process (Rodriguez-Iturbe *et al*. 1999, Porporato *et al*. 2004, Rodríguez-Iturbe and Porporato 2004):

$$p(s) = \frac{C}{\eta} \mathrm{e}^{-\gamma s} s^{\frac{\lambda}{\eta}-1}$$

$$C = \frac{\eta \gamma^{\frac{\lambda}{\eta}}}{\Gamma\left(\frac{\lambda}{\eta}\right) - \Gamma\left(\frac{\lambda}{\eta}, \gamma\right)} \tag{2}$$

where s is relative soil moisture, $\gamma = n_s Z_r/h$, $\eta = E_{\max}/n_s Z_r$, n_s is the average soil available water, Z_r is the average root depth, h is the average rainfall depth and $E_{\max}$ is the evapotranspiration ability. Furthermore, we define $S = n_s Z_r$ as the average basin water storage capacity and thus $\gamma = S/h$.

The average evapotranspiration can be deduced as (Porporato *et al*. 2004):

$$\begin{aligned} \langle E \rangle &= n Z_r \eta \langle s \rangle \\ &= n Z_r \left[\frac{\lambda}{\gamma} - \frac{\eta \gamma^{\frac{\lambda}{\eta}-1} \mathrm{e}^{-\gamma}}{\Gamma\left(\frac{\lambda}{\eta}\right) - \Gamma\left(\frac{\lambda}{\eta}, \gamma\right)} \right] \end{aligned} \tag{3}$$

with $\langle P \rangle = \lambda h$, $\varepsilon = \langle E \rangle / \langle P \rangle$ and $\phi = E_{\max}/\langle P \rangle$. Equation (3) can be simplified as:

$$\varepsilon = 1 - \frac{\phi \gamma^{\frac{\gamma}{\phi}-1} \mathrm{e}^{-\gamma}}{\Gamma\left(\frac{\gamma}{\phi}\right) - \Gamma\left(\frac{\gamma}{\phi}, \gamma\right)} \tag{4}$$

The Porporato equation (4) can be regarded as another equation to describe the Budyko hypothesis. When γ is taken as 5.5, equation (4) has a very similar performance to the initial Budyko curve (Rodríguez-Iturbe and Porporato 2004). Compared to the Choudhury equation, the only parameter γ in this equation is defined as the ratio of soil water storage capacity to rainfall depth and thus has clear physical meaning. The water storage capacity is defined as the multiple of root depth and soil porosity (Rodríguez-Iturbe and Porporato 2004). According to ecohydrological optimality, the root depth is also controlled by climate conditions and is a balance of carbon cost and water benefit (Guswa 2008, 2010).

3.3 Combining the Budyko hypothesis and the stochastic soil moisture model

When n in the Choudhury equation or γ in the Porporato equation is specified, a relationship between ε and φ can be obtained. These curves are called the Budyko curves. To establish a link between the two parameters, first, γ in the Porporato equation is specified and second, n in the Choudhury equation is optimized to match the Budyko curves from the Porporato equation with the specified γ value. For example, Fig. 2 shows the Budyko curves with γ = 1.0, 2.0, 5.5 and 10.0. The corresponding n values are 0.76, 1.00, 1.95 and 2.73, respectively. The match between the curves is quite good, with the exception of cases when the climate is arid (large ϕ) and the roots are shallow relative to mean storm depth (small γ)—a situation that is unlikely to occur naturally. Hence, a relationship between γ and n is

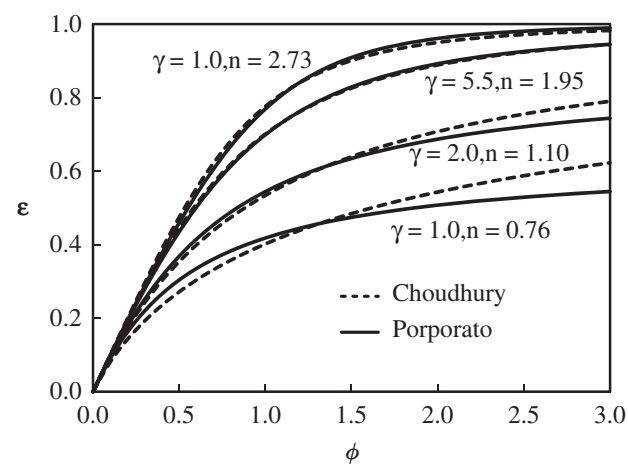

Fig. 2 Budyko curves based on the equations of Choudhury (1999) and Porporato *et al*. (2004), with the parameters n and γ, respectively.

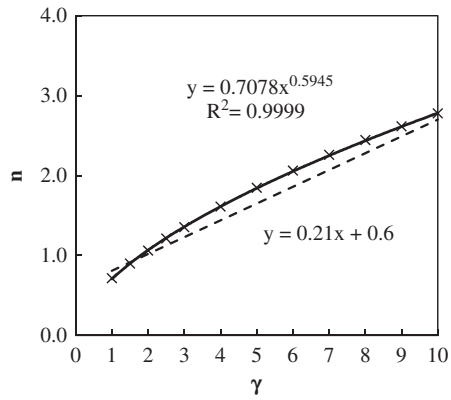

Fig. 3 Relationship between n in the Choudhury (1999) equation and γ in the Porporato *et al*. (2004) equation.

established, as shown in Fig. 3. Note that γ ranges from 1.0 to 10.0 for most basins and n ranges from 0.76 to 2.73, which is consistent with Roderick and Farquhar (2011).

A power-law relationship is further fitted between γ and n (Fig. 3):

$$\begin{aligned} n &= a\gamma^b \\ a &= 0.7078 \\ b &= 0.5945 \end{aligned} \tag{5}$$

We again point out that this equation is different from the linear relationship presented in Donohue *et al*. (2012). This also causes differences in the following derivations and analyses. For instance, the Choudhury equation is rewritten as:

$$\varepsilon = \frac{\phi}{\left(1+\phi^{a\gamma^b}\right)^{\frac{1}{a\gamma^b}}} \tag{6}$$

Since $\gamma = S/h$, the Choudhury equation can be further rewritten as:

$$\varepsilon = \frac{\phi}{\left[1+\phi^{a(S/h)^b}\right]^{\frac{1}{a(S/h)^b}}} \tag{7}$$

Note that h can be estimated from daily precipitation data and, thus, equation (7) has only one unknown, S.

3.4 Attribution of actual evapotranspiration trends

Another form of the Choudhury equation is:

$$E = \frac{PE_0}{\left(P^n + E_0^n\right)^{1/n}} \tag{8}$$

Roderick and Farquhar (2011) present the time derivative of equation (8) as follows:

$$\begin{aligned} \frac{\mathrm{d}E}{\mathrm{d}t} &= \alpha_E \frac{\mathrm{d}P}{\mathrm{d}t} + \beta_E \frac{\mathrm{d}E_0}{\mathrm{d}t} + \gamma_E \frac{\mathrm{d}n}{\mathrm{d}t} \\ &= \mathrm{PE}^* + \mathrm{EE}^* + \mathrm{NE}^* \end{aligned} \tag{9}$$

$$\alpha_E = \frac{\partial E}{\partial P} = \frac{E}{P}\left(\frac{E_0^n}{P^n + E_0^n}\right) \tag{10}$$

$$\beta_E = \frac{\partial E}{\partial E_0} = \frac{E}{E_0}\left(\frac{P^n}{P^n + E_0^n}\right) \tag{11}$$

$$\begin{aligned} \gamma_E &= \frac{\partial E}{\partial n} \\ &= \frac{E}{n}\left[\frac{\ln\left(P^n + E_0^n\right)}{n} - \frac{P^n \ln P + E_0^n \ln E_0}{P^n + E_0^n}\right] \end{aligned} \tag{12}$$

where PE*, EE* and NE* are the contribution to actual evaporation trend of precipitation trend, potential evaporation trend and parameter n trend, respectively; α_E, β_E and γ_E are the sensitivity coefficients; they are also functions of precipitation, potential evapotranspiration and the parameter n. Equation (9) indicates that the changes in actual evapotranspiration can be attributed to the changes in precipitation, potential evapotranspiration and the shape factor of the Budyko curve.

In this study, equation (7) is used since it has one parameter that is physically meaningful. It can be rewritten as:

$$E = \frac{PE_0}{\left(P^{a(S/h)^b} + E_0^{a(S/h)^b}\right)^{\frac{1}{a(S/h)^b}}} \tag{13}$$

The derivative of equation (13) with respect to time is:

$$\begin{aligned} \frac{\mathrm{d}E}{\mathrm{d}t} &= \alpha_E \frac{\mathrm{d}P}{\mathrm{d}t} + \beta_E \frac{\mathrm{d}E_0}{\mathrm{d}t} + \eta_E \frac{\mathrm{d}h}{\mathrm{d}t} + \xi_E \frac{\mathrm{d}S}{\mathrm{d}t} \\ &= \mathrm{PE}^* + \mathrm{EE}^* + \mathrm{HE}^* + \mathrm{SE}^* \end{aligned} \tag{14}$$

$$\eta_E = \frac{\partial E}{\partial h} = -\gamma_E abS^b h^{-b-1} \tag{15}$$

$$\xi_E = \frac{\partial E}{\partial S} = \gamma_E abS^{b-1} h^{-b} \tag{16}$$

where HE* and SE* are the contribution to actual evaporation trend of rainfall depth trend and water storage capacity trend, respectively; η_E and ξ_E are the sensitivity coefficients to rainfall depth and water storage capacity, respectively. Equation (14) indicates that the changes in actual evapotranspiration can be attributed to the changes in precipitation, potential evapotranspiration, average rainfall depth and water storage capacity. Again, due to the power law used in equations (6) and (7), these sensitivity coefficients are different from the ones presented in Donohue *et al*. (2012).

When n is assumed to be constant for a basin over the entire period of record, defined as basin-dependent n, equation (9) becomes (Milly and Dunne 2002, Ma *et al*. 2008, Zhao *et al*. 2009):

$$\frac{\mathrm{d}E}{\mathrm{d}t} = \alpha_E \frac{\mathrm{d}P}{\mathrm{d}t} + \beta_E \frac{\mathrm{d}E_0}{\mathrm{d}t} = \mathrm{PE}^* + \mathrm{EE}^* \tag{17}$$

Consequently, we have three ways to analyse the actual evapotranspiration trends. First, the basin-dependent n is calibrated with the annual precipitation, potential evapotranspiration and actual evapotranspiration data from the period 1956–2000 for each basin according to equation (8), then yearly α_E and β_E are estimated with the basin-dependent n according to equations (10) and (11), and their averages are used to calculate actual evapotranspiration trends based on equation (17). However, this method does not take the changes in the shape factor into consideration. Second, the yearly n and its trends are derived from the annual precipitation, potential evapotranspiration and actual evapotranspiration data for each year according to equation (8); then yearly α_E, β_E and γ_E are estimated with the yearly n according to equations (10)–(12), and their averages are used to calculate the actual evapotranspiration trends based on equation (9). In this way, the parameter n has no clear physical meaning. Third, the yearly S and its trends are derived from the annual precipitation, potential evapotranspiration, average rainfall depth and actual evapotranspiration data for each year according to equation (13); then yearly α_E, β_E, η_E and ξ_E are estimated with the yearly S according to equations (10), (11), (15) and (16), and their averages are used to calculate the actual evapotranspiration trends based on equation (14). In this paper, we focus on the third way to understand the actual evapotranspiration trends in five Chinese basins.

3.5 Attribution of runoff trends

Based on water budget, the annual runoff can be estimated as:

$$\begin{aligned} R &= P - E \\ &= P - \frac{PE_0}{\left(P^{a(S/h)^b} + E_0^{a(S/h)^b}\right)^{\frac{1}{a(S/h)^b}}} \end{aligned} \tag{18}$$

Similar to attribution of actual evapotranspiration trends, we obtain:

$$\begin{aligned} \frac{\mathrm{d}R}{\mathrm{d}t} &= \alpha_R \frac{\mathrm{d}P}{\mathrm{d}t} + \beta_R \frac{\mathrm{d}E_0}{\mathrm{d}t} + \eta_R \frac{\mathrm{d}h}{\mathrm{d}t} + \xi_R \frac{\mathrm{d}S}{\mathrm{d}t} \\ &= \mathrm{PR}^* + \mathrm{ER}^* + \mathrm{HR}^* + \mathrm{SR}^* \end{aligned} \tag{19}$$

$$\alpha_R = \frac{\partial R}{\partial P} = 1 - \frac{E}{P}\left(\frac{E_0^n}{P^n + E_0^n}\right) = 1 - \alpha_E \tag{20}$$

$$\beta_R = \frac{\partial R}{\partial E_0} = -\frac{E}{E_0}\left(\frac{P^n}{P^n + E_0^n}\right) = -\beta_E \tag{21}$$

$$\eta_R = \frac{\partial R}{\partial h} = \gamma_E abS^b h^{-b-1} = -\eta_E \tag{22}$$

$$\xi_R = \frac{\partial R}{\partial S} = -\gamma_E abS^{b-1} h^{-b} = -\xi_E \tag{23}$$

where PR*, ER*, HR* and SR* are the contribution to runoff trend of precipitation trend, potential evaporation trend, rainfall depth trend and water storage capacity trend, respectively; α_R, β_R, η_R and ξ_R are the sensitivity coefficients. Similarly, we have three ways to analyse the runoff trends. In this paper, we focus on the third way that combines the Choudhury and Porporato equations (i.e. equations (18) and (19)) to understand the runoff trends in five Chinese basins.

3.6 Trend analysis

A least squares linear regression model was used to calculate the trends (Roderick *et al*. 2007), and the nonparametric Mann-Kendall test was applied to evaluate the statistical significance of the trends (Maidment 1993).

Table 2 Hydrological trends from 1956 to 2000 of five major basins in China. Underlined values indicate a significance level of >95%.

Basin no.	Name	P (mm/year)^2	E_0 (mm/year)^2	R (mm/year)^2	E (mm/year)^2
1	Liaohe	−1.172	−0.351	−0.418	−0.758
2	Haihe	−1.897	−1.782	−1.387	−0.510
3	Yellow	−1.347	−2.290	−0.784	−0.565
4	Yangtze	1.200	−2.203	1.639	−0.437
5	Pearl	1.548	−1.786	0.800	0.748

4 RESULTS AND DISCUSSION

4.1 Hydrological trends

In the north of China, including the Liaohe River basin, the Haihe River basin and the Yellow River basin, the precipitation, potential evapotranspiration, actual evapotranspiration and runoff decreased between 1956 and 2000. In the south of China, which includes the Yangtze River basin and the Pearl River basin, the precipitation and runoff increased but the potential evapotranspiration decreased between 1956 and 2000. The actual evapotranspiration decreased in the Yangtze River basin but increased in the Pearl River basin. All these trends are shown in Table 2. We try to understand the trends of actual evapotranspiration and runoff according to the trends of precipitation, potential evapotranspiration and the Budyko curve shape.

4.2 Trends of the Budyko curve shape

The basin-dependent parameter n that fits to the 1956–2005 dataset has the values 2.27, 3.00, 1.76, 1.13 and 1.65, respectively, in the five basins. It is larger in northern basins (water-limited) and smaller in southern basins (energy-limited). As mentioned in Section 3.1, n is related to land surface characteristics and climate factors such as soil water storage capacity (influenced by vegetation and reservoirs) and rainfall intensity; thus it is likely that n could vary in different years. In this paper, yearly n is estimated based on annual precipitation, potential evapotranspiration and actual evapotranspiration (equation (8)). Their trends were calculated and are listed in Table 3. Moving averages are also applied to the time series of n but the trends and other results do not show substantial differences. It is shown that n increased in all these basins except for the Yangtze River basin, which corresponds to more actual evapotranspiration under the same precipitation and potential evapotranspiration conditions. The changes in n are dramatic in some basins, such as the Haihe River basin (Basin 2) and the Yellow River basin (Basin 3), where anthropogenic activities are known to be significant (Liu *et al.* 2010, Cong *et al.* 2009a). This suggests that n is closely related to the land surface characteristics, and anthropogenic activities can have significant impacts on the regional hydrological cycle. Since n has large variations in some basins, we believe that the trend of the shape factor, i.e. the n trend, should be taken into account to understand the actual evapotranspiration trends. This is in agreement with the framework proposed by Roderick and Farquhar (2011) and the findings in Donohue *et al.* (2012).

The long-term averaged parameter h is equal to 7.0, 6.9, 5.5, 7.4 and 10.1 mm per event for the five basins, and it is smaller in northern basins (water-limited). The larger rainfall depth corresponds to

Table 3 Mean value and trends in n, h and S between 1956 and 2000 for five major basins in China. n: parameter in the Choudhury (1999) equation; h: average rainfall depth; S: average basin water storage capacity. Underlined values indicate a significance level of >95%.

Basin number	Basin name	Mean value			Trends		
		n	h (mm/event)	S (mm)	n (per decade)	h (mm/event per decade)	S (mm per decade)
1	Liaohe	2.27	7.0	50.2	0.04	0.15	2.32
2	Haihe	3.00	6.9	82.3	0.46	0.18	22.08
3	Yellow	1.76	5.5	26.1	0.10	0.10	2.97
4	Yangtze	1.13	7.4	16.3	−0.0001	0.31	0.63
5	Pearl	1.65	10.1	44.0	0.08	0.33	5.12

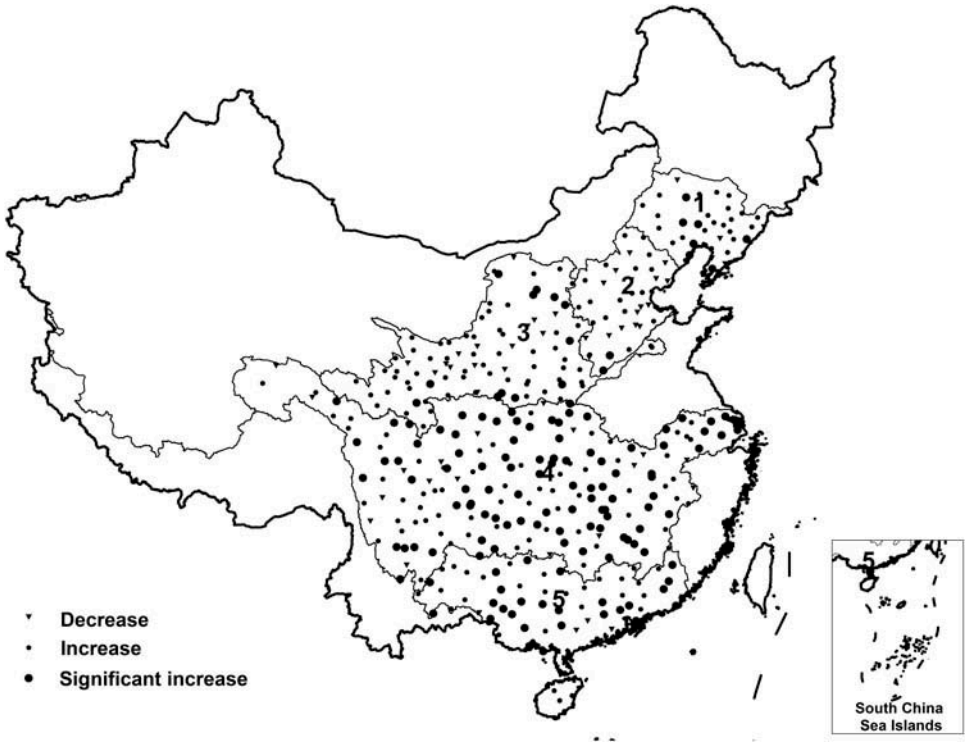

Fig. 4 Rainfall depth trends over China between 1956 and 2005.

smaller n, which means less actual evapotranspiration and more runoff when the total precipitation remains the same. During the past 45 years (from 1956 to 2000), it is found that the average rainfall depth increased significantly at most weather stations in China (Fig. 4). The increasing rainfall depth generates decreasing trend in actual evapotranspiration and increasing trend in runoff, according to equations (14) and (18), which means more flood events. Since the annual precipitation has no significant trend for the same period, more drought events might have occurred at the same time.

The yearly S can be estimated based on equation (7) for each basin, with average S of 50.2, 82.3, 26.1, 16.3 and 44.0 mm for the five basins, respectively. The larger water storage capacity corresponds to a larger n, implying more actual evapotranspiration and less runoff. During the past 45 years (1956–2000), it is found that the water storage capacity increased significantly in all five basins (Table 3). This increasing trend in the water storage capacity would generate more actual evapotranspiration and less runoff, according to equations (12) and (18). Consequently, the increasing trends in h and S will potentially counterbalance each other's impact. It should be noted that all the results and discussion about parameter S depend on the reasonableness of combining the Budyko hypothesis and the stochastic soil moisture model.

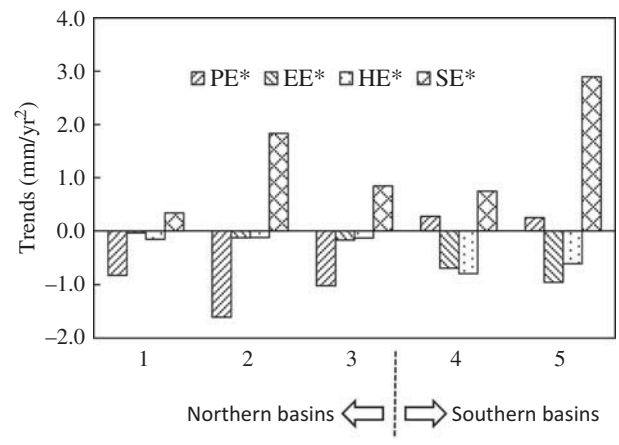

Fig. 5 Contributions of trends in precipitation (PE*), potential evapotranspiration (EE*), rainfall depth (HE*) and water storage capacity (SE*) to actual evapotranspiration trends in five major basins in China.

4.3 Attribution of actual evapotranspiration trends

The actual evapotranspiration trends are attributed to precipitation trend, potential evapotranspiration trend, h trend and S trend. As shown in Fig. 5, the contributions from the four terms are different in magnitude from basin

to basin. For the Liaohe River basin (Basin 1), the decrease in actual evapotranspiration is overwhelmingly caused by the decrease in precipitation, while the contribution of potential evapotranspiration trend, the h trend and the S trend are not pronounced. For the Haihe River basin (Basin 2) and the Yellow River basin (Basin 3), the contribution of S trend is fairly strong but the decrease in precipitation counterbalances and overpowers this contribution, so actual evapotranspiration still decreased. For the Yangtze River basin (Basin 4), the decrease in actual evapotranspiration is mainly caused by decreasing potential evapotranspiration, while the contribution of the precipitation trend is not pronounced, and the contribution of h trend and of S trend counterbalance each other. For the Pearl River basin (Basin 5), the contribution of the S trend is large enough to counterbalance the contributions from the decrease in potential evapotranspiration and the increase in rainfall depth; at the same time the contributions of the precipitation trend is positive, so actual evapotranspiration increased. Based on these observations, we conclude that (a) precipitation decrease plays a dominating role in actual evapotranspiration decrease in northern basins, while potential evapotranspiration decrease plays a dominating role in southern basins; (b) an increase in h can contribute significantly to actual evapotranspiration trend in southern basins (basins 4 and 5); and (c) an increase in S can contribute significantly to actual evapotranspiration trend in some basins (basins 2 and 5).

4.4 Attribution of runoff trends

Runoff decreased in the northern basins due to the negative contributions from both precipitation and S. Runoff increased in the southern basin due to the precipitation increasing and the potential evaporation decreasing. As shown in Fig. 6, the negative contribution from S is pronounced in Basin 5, but it is counteracted by the positive contribution from precipitation and potential evaporation.

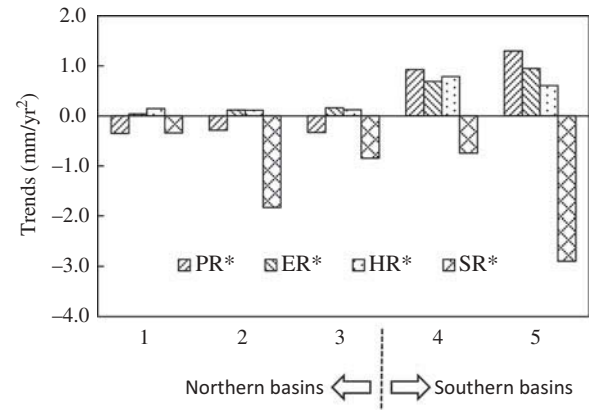

Fig. 6 Contributions of trends in precipitation (PR*), potential evapotranspiration (ER*), rainfall depth (HR*) and water storage capacity (SR*) to runoff trends in five basins in China.

As mentioned in Section 4.2, the increase of rainfall depth induces an increase of runoff, while an increase in water storage capacity reduces the runoff. In the northern basins, the increase in rainfall depth is not significant and the runoff decrease is mainly caused by the increase in water storage capacity. In the southern basins, although the increase in water storage capacity is significant, the increasing precipitation, decreasing potential evaporation and increasing rainfall depth cause the runoff to increase.

4.5 Sensitivity of hydrological trends

The statistics of sensitivity coefficients of actual evapotranspiration to precipitation (α_E), potential evaporation (β_E) and the shape factor n in the Choudhury equation (γ_E) are shown in Fig. 7. The

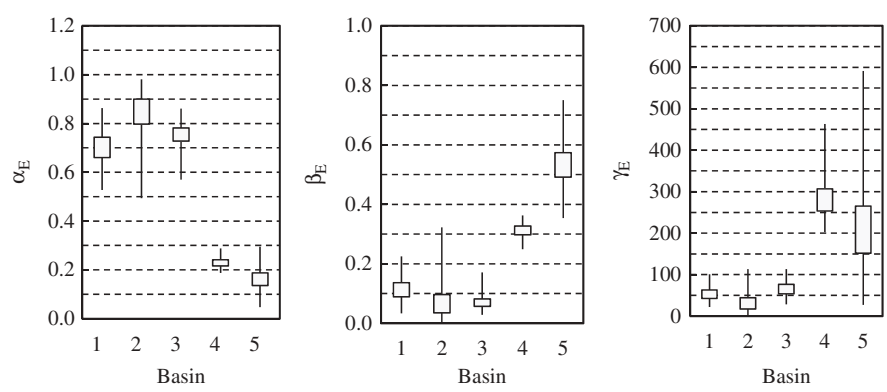

Fig. 7 Statistical characteristics of α_E, β_E and γ_E in different basins. Sensitivity coefficients of actual evapotranspiration to α_E: precipitation; β_E: potential evapotranspiration; and γ_E: the shape factor of the Budyko curve. Boxes are standard deviations centred on averages and vertical lines indicate the maxima and minima.

boxes are the standard deviations centred at the average values and vertical bars indicate the maxima and minima. It is interesting to note that the averages of α_E are significantly larger in the northern basins (about 0.75) than in the southern basins (about 0.20), while the averages of β_E show the opposite, with values in the northern basins (about 0.10) much smaller than those in the southern basins (about 0.40). This implies that actual evapotranspiration trends are more sensitive to precipitation trends in the water-limited basins, while more sensitive to potential evapotranspiration trends in the energy limited basins. In fact, we can generalize from equations (10) and (11) that α_E is always larger than β_E in water-limited basins where potential evapotranspiration is larger than precipitation, and it is the opposite in energy-limited basins. It is also found that the averages of γ_E in the southern basins are larger than those in the northern basins, suggesting the energy-limited basins are relatively more sensitive to variations in n. The fact that the actual evapotranspiration is sensitive to the variations in n in the southern basins is also the basis for the power-law relationship used in equation (5) due to its high accuracy in capturing the relationship between n and γ (see also Fig. 3).

5 CONCLUSIONS

The Budyko curve is used to understand the actual evapotranspiration trends between 1956 and 2000 in five basins in China. Based on the equations of Choudhury (1999) and Porporato *et al*. (2004), actual evapotranspiration trends can be attributed to precipitation trends, potential evapotranspiration trends, rainfall depth trends and water storage capacity trends. The northern basins (1–3) have decreasing precipitation, decreasing potential evapotranspiration and rising Budyko curves (insignificant increasing rainfall depth and significant increasing water storage capacity). The southern basins have increasing precipitation, decreasing potential evapotranspiration, increasing rainfall depth and increasing water storage capacity. The contribution of the four trends to actual evapotranspiration trends and runoff trends varies from basin to basin. The results indicate that the actual evapotranspiration trends are more sensitive to the precipitation trends in water-limited basins, but more sensitive to the potential evapotranspiration trends in energy-limited basins. The contributions of the water storage change trends, which reflect the effects of land surface changes on actual evaporation trend, are pronounced in basins where anthropogenic activities are significant. In the present work, we do not consider the effects of precipitation seasonality, which can sometimes counter the effect of soil water storage, when assessing evapotranspiration using the Budyko framework (Feng *et al*. 2012). This issue is valuable for future research.

Acknowledgements We would like to thank the reviewers, Andrew J. Guswa, Maik Renner and an anonymous reviewer, who gave us good advice which helped to improve this paper.

Funding This study is supported by the National Science Foundation of China, grants 51179083 and 91225302.

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