Trends of Calculated and Simulated Actual Evaporation in the Yangtze River Basin

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ABSTRACT

Actual evaporation in the Yangtze River basin is calculated by the complementary relationship approach—that is, the advection–aridity (AA) model with parameter validation from 1961 to 2007—and simulated by the general circulation model (GCM) ECHAM5–Max Planck Institute Ocean Model (MPI-OM) from 1961 to 2000. Trends of annual and seasonal estimated actual evaporation and air temperature, net radiation, saturation vapor pressure deficit, wind speed, and precipitation are examined by the linear regression method and nonparametric Mann–Kendall test. The stepwise regression method is used to analyze the significance to reference evapotranspiration of independent variables.

Results show that a significant decreasing trend in annual reference evaporation is caused by a significant decline in wind speed. The annual actual evaporation decreases in the upper and midlower Yangtze reaches; more significantly in the AA model [−9.3 mm (10 yr)\(^{-1}\)] than in the GCM [−3.6 mm (10 yr)\(^{-1}\)]. Significant negative trends are found in spring and autumn, but they show reverse trends in summer and winter within the two methods, which is caused by the different contributors to the seasonal actual evaporation in the two methods.

Decreasing net radiation is the main contributor to annual and spring actual evaporation in the two methods. Decreasing precipitation and net radiation are the main contributors to decreasing autumn actual evaporation in the AA model and the GCM. Increasing net radiation and decreasing precipitation are the main contributors to summer and winter actual evaporation in the GCM. Decreasing net radiation and increasing precipitation are the main contributors to decreasing summer and increasing winter actual evaporation in the AA model.

1. Introduction

Evaporation is expected to increase under global warming. However, many studies show that pan evaporation and potential evapotranspiration have decreased in the past 50 years (for the United States, see Hobbins et al. 2004; Peterson et al. 1995, former Soviet Union, see Peterson et al. 1995, India, see Chattopadhyay and Hulme 1997, China, see Thomas 2000; Liu and Zeng 2004; Chen et al. 2005; Gao et al. 2006; Ren and Guo 2006; Xu et al. 2006a,b; Wang et al. 2007; Y. Zhang et al. 2007; Yin et al. 2009; Cong et al. 2009, and Australia and New Zealand see Roderick and Farquhar 2004, 2005). The contrast between expected and observed evaporation is called the “evaporation paradox” (Brutsaert and Parlange 1998; Roderick and Farquhar 2002; Hobbins et al. 2004). There are three interpretations for the “evaporation paradox” (van Heerwaarden et al. 2010; Cong et al. 2009). First, it is an indication of decreasing actual evaporation (Peterson et al. 1995), decreasing pan evaporation related to increasing cloud cover, and/or to the observed decrease in sunlight resulting from increasing cloud cover and aerosol concentration (Roderick and Farquhar 2002).
However, the decreasing trends in sunlight or solar radiation changed to the reverse direction in the late 1980s (Wild et al. 2005; Pinker et al. 2005). This reversal may substantially affect surface climate and the hydrological cycle (Wild et al. 2005). Second, there is a complementary relationship between actual and potential evapotranspiration (Brutsaert and Parlange 1998; Hobbins et al. 2004; Ramírez et al. 2005): A decrease in potential evapotranspiration hints to an increase in actual evaporation, at least in regions that are not characterized as wet, that is, with limited land surface moisture. Third, the decreasing trends in pan evaporation coincide with decreasing trends in surface winds, especially in water-limited regions (Roderick et al. 2007, 2009). However, the implications for the actual evapotranspiration are not yet sure (van Heerwaarden et al. 2010).

As theorized by Brutsaert and Parlange (1998), the solution to the evaporation paradox turns to the relation between pan evaporation and actual evaporation. Long-term and large-scale actual evaporation datasets do not exist. Therefore, little is known on the relationship between pan evaporation and actual evaporation. Ohmura and Wild (2002) pointed out the importance of trends in actual evaporation; pan evaporation matters insofar as it can indicate the directions in actual evaporation.

Until now, studies on changes in actual evaporation have been few and limited. Using large weighing lysimeters, Golubev et al. (2001) found that the actual evaporation during the warm seasons of 1950–90 tended to increase in some relatively dry parts of southern Russia and Ohio, while there is a decreasing trend of actual evaporation in two wetter places of the taiga. From a regional perspective, Teuling et al. (2009) analyzed the correlation between actual evaporation and its main drivers (radiation and precipitation) in central Europe and central North America. They found the relation between the actual evaporation and radiation was strong in more humid Europe, and the relation between the actual evaporation and precipitation was strong in more arid North America. Van Heerwaarden et al. (2010) use a model to show the regulation of the near-surface temperature and humidity by land–atmosphere feedbacks results in a strong connection between pan evaporation, actual evaporation, and vapor pressure deficit depending on the climate forcing. They raise the main conclusion that an increase in soil moisture leads to more actual evaporation and less pan evaporation under all conditions.

Zhan et al. (2005) calculated monthly and annual actual evaporation using a water balance model. The result shows that the actual evaporation increased from 1991 to 2000 in most parts of China, especially in the arid and semiarid regions. Gao et al. (2007) used an improved water balance model to estimate actual evaporation over China from 1962 to 2002. The estimated annual actual evaporation showed a decreasing trend in most areas east of 100°E and an increasing trend in the west and the north parts of northeast China. Liu et al. (2010) used the complementary relationship approach, the GG model, to estimate actual evaporation in the Poyang Lake basin (South of the Yangtze River in central-east China) from 1955 to 2001 and observed a decreasing trend in actual evaporation.

The basin of the Yangtze River (Fig. 1a), the third longest river in the world, is densely populated and undergoes rapid economic development. It also suffers from frequent flood and drought. Trends of pan evaporation and potential evapotranspiration in the Yangtze River basin since 1960 have been studied recently (Xu et al. 2006a,b; Wang et al. 2007). The trend of actual evaporation during the past 50 years, however, has not been analyzed yet. This work could provide a reference for integrated river basin management and water resource planning.

In this paper, changes of actual evaporation are estimated by two methods: 1) the advection–aridity (AA) model with parameter validation and 2) output data of the general circulation model [ECHAM5–Max Planck Institute Ocean Model (MPI-OM)]. The results are analyzed and compared for the Yangtze River basin.

2. Data and methods

a. Data

Data of 147 National Meteorological Observatory (NMO) stations with daily observations of maximum, minimum, and mean (near surface) air temperature, wind speed, sunshine hours, vapor pressure, relative humidity, and air pressure for the period from 1961 to 2000 are used in this study, provided by the National Climate Center (NCC) of the China Meteorological Administration (CMA). On average, only 0.22% of data entries are missing for all 147 stations and indicators. There are 0.11%, 0.13%, 0.10%, 0.18%, 0.32%, 0.26%, 0.26%, and 0.39% of data missing for maximum, minimum, and mean (near surface) air temperature, wind speed, sunshine hours, vapor pressure, relative humidity, and air pressure, respectively. The homogeneity of the datasets of all stations was analyzed by calculating the von Neumann ratio and the cumulative deviations and the Bayesian procedures (for details see Buishand 1982; Maniak 1997). The datasets of all stations prove to be homogeneous with significance beyond the 95% confidence level.

The runoff data from January 1961 to December 2005 for the Yichang (upper reaches), Hukou (middle reaches),
and Datong (lower reaches) hydrological station of the Yangtze River were provided by the Bureau of Hydrology (BOH) of the Changjiang (Yangtze) Water Resources Commission (CWRC). Other hydrological information for the Beibei hydrological station of Jialingjiang River basin (upper reaches tributary in Sichuan), Pingshan hydrological station of Jinshajiang River basin (upper reaches of the Yangtze River), and Wulong hydrological station of Wujiang River basin (upper reaches in Yunnan province) are extracted from the manual of flood prevention and water regime of Changjiang (Bureau of Hydrology of the Changjiang Water Resources Commission 2000) (Table 1). The Yangtze River basin is divided into two main parts, taking the Yichang hydrological station as the boundary: (i) the upper-Yangtze reaches (average altitude of about 2250 m) with 68 stations and (ii) the middle and lower Yangtze reaches (average altitude of about 270 m) with 79 stations (Fig. 1b).

b. Methods

1) AA MODEL SETUP AND ACTUAL EVAPORATION ESTIMATION

Several methods can be used to evaluate actual evaporation. For example, a weighing lysimeter can provide detailed information about the water balance; however, it is practically and economically impossible to measure evapotranspiration over widespread areas for a considerably long time period (Xu and Chen 2005). Therefore, actual evaporation is usually estimated through less complex physically based or empirical approaches (Gao et al. 2007). Among these methods, the complementary relationship approach, that is, the advection–aridity (AA) model (Brutsaert and Stricker 1979), the complementary relationship areal evaporation (CRAE) model (Morton 1983), and the GG model (Granger and Gray 1989) have been discussed by many researchers and show reasonable
result in different regions (Crago and Brutsaert 1992; Liu et al. 2004; Xu and Singh 2005).

Actual evaporation from lysimeter measurement and other meteorological data from meteorological station in Nanchang County in the middle reach of the Yangtze River basin are used to evaluate the adaptability of the AA, CRAE, and GG models in estimating actual evaporation for the Yangtze River basin. In previous studies, the AA showed highest accuracy to estimate actual evaporation in the Yangtze River basin when comparing the result of the models AA, CRAE, and GG (Liu 2009). In this study, data from lysimeter measurements in Nanchang feed into the calibration of the AA model, which is then used for the extrapolation in the entire catchment.

In the AA model (Brutsaert and Stricker 1979), the $E$ is calculated by combining information from the energy budget and water vapor transfer in the Penman (1948) equation:

$$E = \frac{(2a - 1) \Delta}{\Delta + \gamma} (R_n - G) - \frac{\gamma}{\Delta + \gamma} E_a,$$

where $E$ is the actual evaporation (mm day$^{-1}$), $a$ is a Priestley–Taylor coefficient, $R_n$ is the net radiation near the surface (mm day$^{-1}$), $\Delta$ is the slope of the saturation vapor pressure curve at the air temperature (kPa), $\gamma$ is the psychrometric constant (kPa), $G$ is heat flux into the ground (mm day$^{-1}$), and $E_a$ is the drying power of the air, which in general can be written as

$$E_a = f(U_2)(e^* - e_a) = 0.35(1 + 0.54U_2)(e^* - e_a),$$

where $U_2$ is the wind speed (m s$^{-1}$) at 2 m above surface, $e^*$ and $e_a$ are saturation vapor pressure and actual vapor pressure (mm Hg) of the air, respectively, yields $E_a$ (in mm day$^{-1}$).

Therefore, the required input data for the AA model are air temperature, relative humidity, wind speed, and net radiation. Since the observed net radiation $R_n$ is scarcely available, it is estimated by

$$R_n = 0.77 \times \left( a + \frac{R_n}{N} \right) R_a - \sigma \left( \frac{T_{max}^4 + T_{min}^4}{2} \right) \times (0.34 - 0.14 \sqrt{e_a^*}) \left( 1.35 \frac{a + bn/N}{a + b} - 0.35 \right),$$

where $\sigma$ is the Stefan–Boltzmann constant ($4.903 \times 10^{-9}$ MJ K$^{-4}$ m$^{-2}$ day$^{-1}$), $n$ is actual sunshine duration (h), $N$ is the maximum possible duration of sunshine (h), $R_a$ is the extraterrestrial radiation (MJ m$^{-2}$ day$^{-1}$), $e_a$ is actual vapor pressure (kPa), $T_{max}$ and $T_{min}$ are maximum and minimum temperature (K), and $a$ and $b$ are empirical constants; the United Nations Food and Agriculture Organization suggests that $a = 0.25$ and $b = 0.50$ (Allen et al. 1998).

The Priestley–Taylor coefficient $a$ is the parameter that needs adjustment in this study. Priestley and Taylor (1972) first proposed this parameter to be $\sim 1.26$ for well-watered surfaces, and Brutsaert and Stricker (1979) introduced $a = 1.26$ (or 1.28) in their original AA model equation. Yet, these values lead to strong overestimation of actual evaporation in the Yangtze River basin, which may be caused by an unsuitable selection of $a$ according to the field experiments taken by Priestley and Taylor (1972) and Stewart and Rouse (1976), who concluded that $a = 1.26$ is more representative for saturated land surface, oceans, and sedge meadow, which are under potential evaporation conditions. As many previous researches have revealed (Hobbins et al. 2001), the coefficient $a$, which is directly impacted by soil moisture and has a highly nonlinear relationship with it (Crago 1996), may vary for different land cover types and at a range of temporal scales. Flint and Childs (1991) summarized studies of $a$ under various surface conditions and pointed out that it depends on surface vegetation and microclimatic conditions and ranges from 0.72 for forest conditions to 1.57 for conditions of strong advection.

In the Yangtze River basin, the landform can be divided threefold: the first step is the areas of altitude

Table 1. Hydrological attributes and calculated actual evaporation for the Yangtze River basin and subbasins.

<table>
<thead>
<tr>
<th>Subbasin</th>
<th>Hydrological station</th>
<th>Area ($10^4$ km$^2$)</th>
<th>Runoff (m$^3$ s$^{-1}$)</th>
<th>Runoff depth (mm)</th>
<th>Precipitation (mm)</th>
<th>Evaporation By water balance (mm)</th>
<th>Evaporation By AA model (mm)</th>
<th>Close error of water balance (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jialinjiang</td>
<td>Beibei</td>
<td>16.00</td>
<td>2100</td>
<td>413.9</td>
<td>955.7</td>
<td>541.8</td>
<td>540.1</td>
<td>-0.18</td>
</tr>
<tr>
<td>Jinhajiang</td>
<td>Pingshan</td>
<td>48.51</td>
<td>4530</td>
<td>294.5</td>
<td>725.3</td>
<td>430.8</td>
<td>446.5</td>
<td>2.16</td>
</tr>
<tr>
<td>Wujiang</td>
<td>Wulong</td>
<td>8.79</td>
<td>1590</td>
<td>570.4</td>
<td>1122.0</td>
<td>551.6</td>
<td>569.6</td>
<td>1.60</td>
</tr>
<tr>
<td>Upper Yangtze</td>
<td>Yichang</td>
<td>100.55</td>
<td>14 200</td>
<td>439.6</td>
<td>938.2</td>
<td>498.6</td>
<td>503.0</td>
<td>0.47</td>
</tr>
<tr>
<td>Poyang lake</td>
<td>Hukou</td>
<td>16.22</td>
<td>4720</td>
<td>913.3</td>
<td>1646.0</td>
<td>732.7</td>
<td>695.8</td>
<td>-2.24</td>
</tr>
<tr>
<td>Midlower Yangtze</td>
<td>Yichang–Datong</td>
<td>69.99</td>
<td>14 178</td>
<td>655.9</td>
<td>1277.1</td>
<td>621.2</td>
<td>644.9</td>
<td>1.86</td>
</tr>
<tr>
<td>Entire Yangtze</td>
<td>Datong</td>
<td>170.54</td>
<td>28 378</td>
<td>523.1</td>
<td>1106.9</td>
<td>583.8</td>
<td>579.2</td>
<td>-0.41</td>
</tr>
</tbody>
</table>

$\Delta, \gamma, R_a, T_{max}, T_{min}, b$ (K), $a$ (K), $\sigma$ (mm Hg)
above 3500 m with altiplano and alpine valleys; the second step is the area of altitude between 500 and 2000 m that includes the Qinba Mountains, Sichuan basin, and Guizhou altiplano; and the third step is the areas of altitude below 500 m with hills and large plains as in East China. Climatically, the third step belongs to a warm and wet subtropical monsoon climate; the first and second steps are characterized by a semihumid and semiarid climate.

Based on the above facts and also owing to the limited experiments that describe the relationship between vegetation type and the parameters in more detail, the parameter is selected as $\alpha = 1.01$ for areas of altitude above 500 m and $\alpha = 1.105$ for the lower altitude areas in the Yangtze River basin.

Daily actual evaporation from 1961 to 2007 is calculated for 147 meteorological stations by using the AA model with the above parameters. The mean annual runoff at the Yichang hydrological station from 1961 to 2005 is 523.1 mm. The mean annual precipitation over the whole of the Yangtze River basin in the same period of time is 1106.9 mm. According to the water balance method (precipitation minus runoff), the calculated regional mean annual evaporation is 583.8 mm. The mean annual actual evaporation estimated by the AA model is 579.2 mm. The difference in water balance between the AA model and the water balance method is $-0.41\%$, which is relatively small.

The Yangtze River basin covers a large area, shows significant differences in the physical geographical condition, and data on river runoff are sparse. Therefore, the validation of the estimated actual evaporation in the AA model is only carried out in the Jialingjiang River basin, the Jinshajiang River basin, the Wujiang River basin, the upper–Yangtze River basin, the Poyang Lake basin, and the midlower Yangtze River basin. The test results are shown in Table 1. Overall, the estimated mean annual actual evaporation in the Yangtze River basin shows good results. The estimated data is better in areas with less precipitation (upper Yangtze River basin) than in those areas with higher precipitation (midlower Yangtze River basin).

For example, in the Jialingjiang River basin, the mean annual precipitation is 955.7 mm and the mean annual runoff is 2100 m$^3$ s$^{-1}$ at the Beibei hydrological station, which equals to 413.9 mm. The estimated mean annual actual evaporation by the AA model is 540.1 mm and the calculated mean annual actual evaporation by the water balance method is 541.8 mm. The difference between the approaches is 1.7 mm, which is a minus of 0.18% between the AA model and water balance method. For the Poyang Lake basin, the mean annual precipitation is 1646.0 mm and the mean annual actual evaporation estimated by the AA model is 36.9 mm lower than that of the water balance method. This difference of 2.24% resulting from the models is the highest of all subcatchments. Therefore, the AA model is suitable to estimate actual evaporation in the Yangtze River basin and it produces good results after parameters validation.

2) ECHAM5–MPI-OM AND ACTUAL EVAPORATION SIMULATION

Ensemble simulations with the ECHAM5–MPI-OM for the twentieth century are performed. Three model realizations are initialized by a preindustrial control run at different states; their respective forcings are the observed anthropogenic greenhouse gases (CO$_2$, CH$_4$, N$_2$O, CFCs, O$_3$, and sulfate). The model consists of the European Centre–Hamburg Model (ECHAM5) (Roeckner et al. 2003) and the ocean model, MPI-OM (Marsland et al. 2003). ECHAM5 runs with horizontal T63 (about 1.8°) and vertical 31-layer resolution. The upper and midlower reaches of the Yangtze River basin are represented by 41 (23°–36.5°N, 90°–108.75°E) and 37 model grid points (23°–36.5°N, 108.75°–122°E). The model diagnostics provides the water cycle with actual evaporation, precipitation, and runoff, which is shown to be comparable with the observed components (Hagemann et al. 2006).

3) ANALYSIS METHOD

The trend tests applied are both the linear regression method and the nonparametric Mann–Kendall test (MK test) (Kendall and Gibbons 1981). Linear regression is used to estimate the magnitude of changes of climate indicators in terms of a linear trend. The MK test is used to examine the nonlinear trend of climatic data; this test is performed on all stations to detect mean annual and seasonal trends. Confidence levels of 90%, 95%, and 99% are taken as thresholds to classify the significance of positive and negative trends. Trends at significance levels below 90% are not considered. The stepwise regression method is used to analyze the significance to reference evapotranspiration of independent variables. The seasons are spring (March–May), summer (June–August), autumn (September–November), and winter (December–February).

3. Trends in actual evaporation: Analysis and results

Trends in actual evaporation calculated by the AA model from 1961 to 2007 are presented in this section. In addition, trends of the twentieth century actual
evaporation simulated by the general circulation model ECHAM5–MPI-OM are analyzed.

a. Trends in actual evaporation calculated by the AA model

The MK test is performed for 147 meteorological stations to identify trends of the annual and seasonal actual evaporation during the recent 47 years in the Yangtze River basin. The linear regression is applied to annual actual evaporation in different regions of the Yangtze River basin to detect changes in actual evaporation (Fig. 2a).

Figure 3 shows the number of stations with significant positive or negative trends in actual evaporation beyond the 90% confidence level. The following results are noted.

(i) There are more stations with significant negative (86) than positive (20) actual evaporation trends in the annual totals.

(ii) The actual evaporation shows increasing (positive) trends in spring (30 stations), summer (7 stations), autumn (21 stations), and winter (69 stations). Decreasing (negative) trends are estimated in winter (62 stations), summer (112 stations), autumn (64 stations), and winter (16 stations). The most distinct trends of actual evaporation are estimated to be the negative trends in summer.

(iii) The annual mean actual evaporation decreases by 9.3 mm (10 yr)$^{-1}$ in the whole of the Yangtze basin, 6.0 mm (10 yr)$^{-1}$ in the upper reaches of the Yangtze river, and 11.8 mm (10 yr)$^{-1}$ in the mid-lower Yangtze reaches.

The spatial distribution of the trends in mean annual actual evaporation is shown in Fig. 4. The annual mean actual evaporation decreases in most regions except for the source area of the upper reaches of the Yangtze River, which shows a significant increasing trend.

The results of the MK test and significance levels (1961–2007) for the seasons and basins are shown in Fig. 4. In the whole of the Yangtze River basin, significant negative trends are found in annual, summer actual evaporation (>99% confidence level), and in spring, autumn actual evaporation (>95% confidence level); the winter actual evaporation shows slightly positive trends (>90% confidence level). In the upper Yangtze reaches, significant decreasing trends are found in summer actual evaporation (>99% confidence level), in annual, autumn actual evaporation (>95% confidence level), and in spring actual evaporation (>90% confidence level). The winter actual evaporation shows increasing trend (>90% confidence level). In the midlower Yangtze reaches, marked downward trends are found in annual, spring, and summer actual evaporation (>99% confidence level) and in autumn (>95% confidence level), but there is an upward trend in winter (>90% confidence level).
The decreasing (or negative) trend in summer contributes most to the changes in annual mean actual evaporation. The significance of the negative trends (in annual actual evaporation) is higher in the midlower than in the upper Yangtze reaches.

b. Trends in actual evaporation simulated by the model ECHAM5–MPI-OM

The ensemble mean of the three ECHAM5–MPI-OM simulations of the late twentieth-century climate (1961–2000) is examined in this section. Annual and seasonal time series and trends of the whole Yangtze River basin with its upper reaches and its midlower reaches are analyzed. Figure 4 presents the results of the Mann–Kendall trend test for the ensemble mean evaporation. The linear regression is also applied for annual actual evaporation in different regions of the Yangtze River basin (Fig. 2b). The following results are noted:

(i) The annual mean actual evaporation decreases by 3.6 mm (10 yr)$^{-1}$ in the whole of the Yangtze River basin, 1.6 mm (10 yr)$^{-1}$ in the upper Yangtze reaches, and 5.8 mm (10 yr)$^{-1}$ in the midlower Yangtze reaches.

(ii) In the whole of the Yangtze River basin, significant negative trends are found in annual, autumn actual evaporation (>95% confidence level), and in spring actual evaporation (>90% confidence level). The summer actual evaporation shows slightly positive trend and the winter actual evaporation shows slightly negative trends.

(iii) In the upper Yangtze reaches, significant decreasing trends are found in autumn actual evaporation (>95% confidence level), a significant increasing trend is found in summer actual evaporation (>90% confidence level), and no obvious trends are found in annual, spring and winter actual evaporation. In the midlower Yangtze reaches, downward trends are found in annual, spring actual evaporation (>95% confidence level) and in autumn (>90% confidence level), but there are no trends in summer and winter.

The decreasing (or negative) trend in autumn and spring contributes most to the annual mean actual evaporation. The significance of the negative trends (in annual actual evaporation) is higher in the midlower than in the upper Yangtze reaches. The increasing trend in summer in upper Yangtze reaches leads to a slight increasing trend in the whole of the basin in summer.

4. Discussion and conclusions

From 1961 to 2007 the mean annual air temperature shows significant increasing trends (>99% confidence level) in the entire Yangtze River basin, the upper reaches, and the midlower reaches of the Yangtze River (Table 2). Over the same period of time, the mean annual reference evapotranspiration (calculated by the FAO56–Penman–Monteith model, Allen et al. 1998) shows a decreasing trend (>90% confidence level) in the entire basin and different reaches (Table 2). It may be concluded that

![FIG. 4. The spatial distribution of the trends of annual mean actual evaporation calculated by the AA model for 1961–2007.](https://example.com/figure4.png)

<table>
<thead>
<tr>
<th>Yangtze River basin</th>
<th>Entire basin</th>
<th>Upper reaches</th>
<th>Midlower reaches</th>
</tr>
</thead>
<tbody>
<tr>
<td>Period</td>
<td>ETr</td>
<td>T</td>
<td>ETr</td>
</tr>
<tr>
<td>Year</td>
<td>–1.90$^{a}$</td>
<td>3.60$^{b}$</td>
<td>–1.33$^{c}$</td>
</tr>
<tr>
<td>Spring</td>
<td>0.38</td>
<td>2.63$^{b}$</td>
<td>–0.65</td>
</tr>
<tr>
<td>Summer</td>
<td>–3.37$^{b}$</td>
<td>0.98</td>
<td>–2.19$^{a}$</td>
</tr>
<tr>
<td>Autumn</td>
<td>–0.03</td>
<td>2.43$^{b}$</td>
<td>0.14</td>
</tr>
<tr>
<td>Winter</td>
<td>–1.16</td>
<td>3.57$^{b}$</td>
<td>0.28</td>
</tr>
</tbody>
</table>

$^{a}$ Denotes beyond 95% confidence level.

$^{b}$ Denotes beyond 99% confidence level.

$^{c}$ Denotes beyond 90% confidence level.
**Table 3.** Stepwise regression analysis with $T$ (air temperature), $R_n$ (net radiation), SVPD (saturation vapor pressure deficit), $U$ (wind speed), and $P$ (precipitation) as predictors and $E_{Tr}$ (reference evapotranspiration) as the dependent variable for the year and each season, based on the period of 1961–2007. (Criteria: Probability-of-F-to-enter ≤ 0.05, Probability-of-F-to-remove ≥ 0.05.)

<table>
<thead>
<tr>
<th>Entire Yangtze River basin</th>
<th>Upper reaches of the Yangtze River</th>
<th>Midlower reaches of the Yangtze River</th>
</tr>
</thead>
<tbody>
<tr>
<td>Variable</td>
<td>Standardized Coefficients (%)</td>
<td>Changes in ETr (%)</td>
</tr>
<tr>
<td>$R_n$</td>
<td>0.41</td>
<td>−5.48</td>
</tr>
<tr>
<td>SVPD</td>
<td>0.70</td>
<td>17.96</td>
</tr>
<tr>
<td>$U$</td>
<td>0.34</td>
<td>−23.57</td>
</tr>
<tr>
<td>$R_n$</td>
<td>0.30</td>
<td>−1.89</td>
</tr>
<tr>
<td>$T$</td>
<td>0.13</td>
<td>5.96</td>
</tr>
<tr>
<td>Spring</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$R_n$</td>
<td>0.68</td>
<td>−10.86</td>
</tr>
<tr>
<td>SVPD</td>
<td>0.33</td>
<td>3.38</td>
</tr>
<tr>
<td>$U$</td>
<td>0.13</td>
<td>−15.63</td>
</tr>
<tr>
<td>$T$</td>
<td>0.06</td>
<td>0.85</td>
</tr>
<tr>
<td>Summer</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$R_n$</td>
<td>0.45</td>
<td>−2.82</td>
</tr>
<tr>
<td>SVPD</td>
<td>0.61</td>
<td>16.42</td>
</tr>
<tr>
<td>$U$</td>
<td>0.37</td>
<td>−26.17</td>
</tr>
<tr>
<td>$T$</td>
<td>0.13</td>
<td>5.45</td>
</tr>
<tr>
<td>$P$</td>
<td>−0.12</td>
<td>−22.10</td>
</tr>
<tr>
<td>Autumn</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$R_n$</td>
<td>0.45</td>
<td>−2.82</td>
</tr>
<tr>
<td>SVPD</td>
<td>0.61</td>
<td>16.42</td>
</tr>
<tr>
<td>$U$</td>
<td>0.37</td>
<td>−26.17</td>
</tr>
<tr>
<td>$T$</td>
<td>0.13</td>
<td>5.45</td>
</tr>
<tr>
<td>Winter</td>
<td></td>
<td></td>
</tr>
<tr>
<td>SVPD</td>
<td>0.88</td>
<td>6.43</td>
</tr>
<tr>
<td>$U$</td>
<td>0.30</td>
<td>−29.27</td>
</tr>
<tr>
<td>$R_n$</td>
<td>0.15</td>
<td>−3.11</td>
</tr>
<tr>
<td>$T$</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*RC (relative change) is indicated by the percentage of change in 47 years to absolute average value (Yin et al. 2009).

**Indicates the changes in ETr induced by each climatic variable.

the evaporation paradox existed in the Yangtze River basin from 1961 to 2007. The evaporation paradox, however, is not universal in temporal and spatial dimensions. In the entire Yangtze River basin, only the summer reference evapotranspiration decreases significantly (>99% confidence level), while the change of summer air temperature is not obvious. It may be questioned whether there is the evaporation paradox in the entire Yangtze River basin in summer. In the upper reaches of the river, the evaporation paradox prevails only in summer (reference evapotranspiration decreases significantly while the air temperature increases). In the midlower reaches of the river, the evaporation paradox exists only in winter (reference evapotranspiration decreases while the air temperature increases) (Table 2).

Reference evapotranspiration ($E_{Tr}$) is mostly influenced by net radiation ($R_n$), air temperature ($T$), wind speed ($U$), saturation vapor pressure deficit (SVPD), precipitation ($P$), and so on. To detect the relative contributions of the meteorological variables to the change of reference evapotranspiration, a stepwise regression method is used to regress annual and seasonal $E_{Tr}$ against $T$, $R_n$, $U$, SVPD, and $P$. The significance level for a predictor to be added into the model is set as 0.05, and the significance level for a predictor to be removed from the model is 0.051. The results can be found in Table 3.

Comparing the standardized coefficients in Table 3, it is found that there is no universal explanation for the trend of reference evapotranspiration. The major controlling factors vary from region to region and from season to season. For the entire Yangtze River basin, SVPD is the most important controlling factor for annual and seasonal $E_{Tr}$ except summer $E_{Tr}$, which is mostly controlled by $R_n$. This is in accordance with Gong et al. (2006) and Yin et al. (2009), who reported relative
humidity to be the most sensitive variable in the Yangtze River basin and in China, respectively. For annual ETr, although SVPD is the most important controlling factor, changes in SVPD induced an increment of 7.36% in ETr. The wind speed U is a less important controlling factor, but it has declined significantly with the relative change (RC) of 23.57%. Therefore, it is found that U is the primary contributor that causes ETr to decrease by −8.05% in the past 47 years. Net radiation \( R_n \), with a significant decreasing trend is the most important controlling factor for summer ETr, so it is the main contributor to decreasing ETr in summer in the entire Yangtze River basin. For the upper reaches and midlower reaches of the river, the primary contributor for decreasing annual ETr is \( U \) and for decreasing ETr in summer it is \( R_n \) (Table 3).

In general, for the Yangtze River basin as a whole, decreasing \( U \) is the key contributor to declined annual ETr, and \( R_n \) is the leading factor to decreased ETr in summer. Similar results have been found in China (Yin et al. 2009) and the Tibetan Plateau (Y. Zhang et al. 2007, 2009).

In this paper, the annual actual evaporation estimated by the two methods decreases in the whole Yangtze River basin except the source area of the upper reaches, which is consistent with the research of Gao et al. (2007) who draw a conclusion that the annual actual evaporation increased in some parts of the source area of the Yangtze River basin and decreased in the middle part of the Yangtze River basin from 1960 to 2002. Liu et al. (2010) calculated actual evaporation by the complementary relationship method (GG model) and achieved similar results of a decreasing annual actual evaporation trend in the Poyang Lake basin of the middle reaches of the Yangtze River from 1955 to 2001.

Actual evaporation is driven essentially by meteorological variables, mediated by vegetation and soil characteristics, and constrained by the amount of available water (Gao et al. 2007). Because soil moisture is often difficult to obtain, precipitation is usually a substitute (Cohen et al. 2002). To identify the reason for the change of actual evaporation, correlation between actual evaporation and other meteorological variables on annual and seasonal scales is performed. Tables 4 and 5 show the correlation coefficients between actual evaporation simulated by the ECHAM5–MPI-OM and calculated by the AA model and other observed meteorological variables, respectively. The trends of meteorological variables simulated by the ECHAM5–MPI-OM and observed are detected and the results are shown (Fig. 6).

For the entire basin, combining Tables 4 and 5, both in the AA model and in the ECHAM5–MPI-OM, \( R_n \) is the most important controlling variable for annual, spring, and summer actual evaporation; \( P \) is the one for winter actual evaporation. However, the main controlling factor for autumn actual evaporation is \( P \) in the AA model and \( R_n \) in the ECHAM5–MPI-OM, respectively. Therefore, a decrease in annual and spring \( R_n \) causes the decrease in annual and spring actual evaporation in
two methods; see Figs. 5 and 6. The decrease in $R_n$ in summer and increase in $P$ in winter lead to a decrease in summer actual evaporation and an increase in winter actual evaporation in the AA model, respectively (Figs. 5 and 6). In the ECHAM5–MPI-OM, a slight increase in summer $R_n$ and a slight decrease in winter $P$ cause an increase in summer actual evaporation and a decrease in winter actual evaporation, respectively (Figs. 5 and 6). The decreasing trend in summer actual evaporation is the most significant, which contributes most to the decrease in annual actual evaporation in the AA model (Figs. 5 and 6). As the decreasing trend in annual $R_n$ is more significant in the AA model than in the ECHAM5–MPI-OM (Fig. 6), the significance of the negative annual actual evaporation is higher in the AA model than that simulated by the ECHAM5–MPI-OM (Fig. 5). The annual actual evaporation decreases by about 9.3 mm (10 yr)$^{-1}$ in the AA model as compared to 3.6 mm (10 yr)$^{-1}$ in ECHAM5–MPI-OM. In the AA model, the decrease is caused by significant negative trends in summer actual evaporation. In ECHAM5–MPI-OM, the decrease is caused by significant trends in autumn and spring.

For the upper reaches of the Yangtze River basin, combining Tables 4 and 5, both in the AA model and in the ECHAM5–MPI-OM, it is found that $R_n$ is the most important controlling variable for summer and autumn actual evaporation; $P$ is the one for spring and winter actual evaporation. In the AA model, although $R_n$ is less important controlling factor for spring actual evaporation, it shows significant decrease (Fig. 6), which contributes most to the decrease trend in spring actual evaporation. A decrease in $R_n$ causes the decrease in summer and autumn actual evaporation (Figs. 5 and 6). The increase in $P$ causes an increase in winter actual evaporation. In the ECHAM5–MPI-OM, the decrease in $P$ causes a decrease in spring and winter actual evaporation, and the decrease in $R_n$ causes a decrease in autumn actual evaporation. Increases in $R_n$, SVPD, and $T$ cause an increase in summer actual evaporation (Figs. 5 and 6). Although the controlling factors for seasonal actual evaporation are the same in the two methods, the changes of seasonal meteorological variables vary in the two models, which leads to a negative (positive) trend in the summer (winter) actual evaporation in the AA model and a positive (negative) trend in ECHAM5–MPI-OM.

For the midlower reaches of the Yangtze River basin, combining Tables 4 and 5, both in the AA model and in the ECHAM5–MPI-OM, $R_n$ is the most important controlling variable for spring and summer actual evaporation; $P$ is the one for autumn and winter actual evaporation. In the AA model, although $R_n$ is less important controlling factor for spring actual evaporation, it shows significant decrease (Fig. 6), which contributes most to the decrease trend in spring actual evaporation. A decrease in $R_n$ causes the decrease in summer and autumn actual evaporation (Figs. 5 and 6). The increase in $P$ causes an increase in winter actual evaporation. In the ECHAM5–MPI-OM, the decrease in $P$ causes a decrease in spring and winter actual evaporation, and the decrease in $R_n$ causes a decrease in autumn actual evaporation. Increases in $R_n$, SVPD, and $T$ cause an increase in summer actual evaporation (Figs. 5 and 6). Although the controlling factors for seasonal actual evaporation are the same in the two methods, the changes of seasonal meteorological variables vary in the two models, which leads to a negative (positive) trend in the summer (winter) actual evaporation in the AA model and a positive (negative) trend in ECHAM5–MPI-OM.

### Table 5. Correlation coefficients between actual evaporation and other meteorological variables simulated by the ECHAM5–MPI-OM.

<table>
<thead>
<tr>
<th>Period</th>
<th>Entire Yangtze River basin</th>
<th>Upper-Yangtze reaches</th>
<th>Midlower Yangtze reaches</th>
</tr>
</thead>
<tbody>
<tr>
<td>Year</td>
<td>$T$ $R_n$ $U$ SVPD $P$</td>
<td>$T$ $R_n$ $U$ SVPD $P$</td>
<td>$T$ $R_n$ $U$ SVPD $P$</td>
</tr>
<tr>
<td>Year</td>
<td>0.11 $0.65^*$ $-0.10$ 0.02 $-0.02$</td>
<td>$-0.10$ 0.23 $-0.38^<em>$ $-0.45^</em>$ 0.09</td>
<td>0.24 0.81* 0.18 0.15 $-0.12$</td>
</tr>
<tr>
<td>Spring</td>
<td>0.09 $0.69^<em>$ $-0.33^</em>$ $-0.18$ 0.09</td>
<td>$-0.43^<em>$ $-0.15$ $-0.64^</em>$ $-0.57^<em>$ $0.46^</em>$</td>
<td>0.40* 0.86* 0.09 0.29 $-0.15$</td>
</tr>
<tr>
<td>Summer</td>
<td>0.52* 0.96* 0.32* 0.66* $-0.23$</td>
<td>0.51* 0.90* 0.09 0.52* $-0.55^*$</td>
<td>0.55* 0.96* 0.27 0.63* 0.01</td>
</tr>
<tr>
<td>Autumn</td>
<td>$-0.11$ 0.83* $-0.05$ 0.43* $-0.49^*$</td>
<td>$-0.06$ 0.70* 0.41* 0.17 $-0.24$</td>
<td>$-0.19$ 0.85* $-0.23$ 0.42* $-0.56^*$</td>
</tr>
<tr>
<td>Winter</td>
<td>0.27 0.03 $-0.16$ 0.28 0.28</td>
<td>0.20* $-0.34^<em>$ $-0.50^</em>$ $-0.27$ 0.35*</td>
<td>0.09 0.15 0.13 $-0.40^<em>$ 0.32</em></td>
</tr>
</tbody>
</table>

* Denotes beyond 0.05 significance level.
$P$ is the one for winter actual evaporation. However, the main controlling factor for autumn actual evaporation is $P$ in the AA model and $R_n$ in the ECHAM5–MPI-OM. The decrease in $R_n$ causes the decrease in spring and summer actual evaporation, and the increase in $P$ causes the increase in winter actual evaporation in the two methods. The main reason for the decrease in autumn actual evaporation is the decrease in $P$ in the AA model and the decrease in $R_n$ in the ECHAM5–MPI-OM (Figs. 5 and 6). Although the controlling factors of seasonal actual evaporation are not the same in the two models, the total contributions of meteorological variables to the change of seasonal actual evaporation are similar. Therefore, the change of seasonal actual evaporation shows similar directions in the two models—that is, decreasing trends of actual evaporation in spring, summer, autumn, and increasing trends of actual evaporation in winter.

In general, the decreasing trend in annual, spring, and summer actual evaporation is mostly caused by the decrease in $R_n$, and the change of $P$ is the main contributor in the winter. The decrease in autumn actual evaporation is mainly caused by the decrease in $P$ in the AA model and the decrease in $R_n$ in the ECHAM5–MPI-OM, respectively.

Although the simulation capacity of meteorological variables in the ECHAM5–MPI–OM is strong in China (Zhai et al. 2009), the Yangtze River basin (Zeng et al. 2007; Z. Zhang et al. 2007; Liu et al. 2008), the Zhujiang River basin (Liu et al. 2009), and the Huaihe River basin (Gao et al. 2010), there are differences in the simulations and observations in some meteorological variables. Figure 6 indicates that the trends of $T$ and SVPD as simulated are more similar to observations than that of $R_n$, $U$, and $P$. While $R_n$ and $P$ are the main factors for the change of actual evaporation in the Yangtze River basin, it would increase the uncertainty of the change of actual evaporation. In the following analysis of the relationship between actual evaporation and ETr, actual evaporation calculated by the AA model is used.

Since the annual actual evaporation and reference evapotranspiration in the Yangtze River basin show the similar decreasing trend, it may be concluded that they have a proportional relationship. Further analysis of the correlation between actual evaporation and reference evapotranspiration on annual and seasonal scales is performed. Correlation coefficients are shown in the third column of Table 4.

Actual evaporation and reference evapotranspiration have a strong positive correlation (significance level < 0.01) in summer in the entire basin, the upper reaches, and the midlower reaches of the Yangtze River, and a strong negative correlation (significance level < 0.1) in winter in the entire basin, in spring and winter in the upper reaches, and in autumn and winter in the midlower reaches of the Yangtze River basin (Table 4). This result supports the argument that actual evaporation and ETr have a strong proportional relationship in summer in different reaches of the river where the decrease in $R_n$ contributes most to the decrease in actual evaporation, and they have a strong complementary relationship in winter in the entire basin, in spring and winter in the upper reaches, and in autumn and winter in the midlower reaches of the Yangtze River basin where the change of actual evaporation is mainly caused by the change of precipitation.

To further analyze the climate condition, the radiative index of dryness $R$ [ratio of potential evaporation to precipitation introduced by Budyko (1956)] is calculated spatially and seasonally. The index $R$ is shown in the last column of Table 4. It indicates that a complementary relationship between actual evaporation and ETr occurs in areas with $R > 0.9$ where there are water-limited environments, while for regions with $R < 0.9$, where there are energy-limited environments, a proportional relationship exists between actual evaporation and ETr. Golubev et al. (2001) obtained similar results that colinear trends in pan and actual evaporation occur only in areas with $R < 0.7$, while an inverse relationship exists between pan and actual evaporation in regions with $R \geq 0.8$ over the contiguous United States and the former USSR. Consistent results also have been found in Hobbins et al. (2008), Teuling et al. (2009), and Roderick et al. (2009), who pointed out that, in energy-limited conditions, declining pan evaporation usually indicated declining actual evaporation and, in water-limited regions, declining pan evaporation with increasing actual evaporation occurred. Roderick and Farquhar (2004) and Yang et al. (2006, 2007) pointed out that in energy-limited regions, a decrease in pan evaporation, at constant rainfall, implies a decrease in actual evaporation and an increase in runoff and/or soil moisture. Observations from the midlower reaches of the Yangtze River show a significant decrease in annual actual evaporation (calculated by water balance method, i.e., precipitation minus runoff) with significant decreasing pan evaporation (Wang et al. 2007) and significant increasing runoff (Qin et al. 2005).

Further investigations, including station-based analysis of actual evaporation, pan evaporation, reference evapotranspiration, solar radiation, and precipitation, may provide additional insights into the interplay among them.

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