Abstract

The potential evapotranspiration (ET) is one of the key inputs to hydrological modeling. There are many evapotranspiration models that have been developed and validated but the efficiency is not high enough for poor estimation results or difficult data collection. While, the Shuttleworth-Wallace (S-W) model can consider two coupled sources in a resistance network and some researchers show that the S-W model can receive a good results. In addition, the S-W model is physically-based and all required data are publicly available and can be applied to data-poor or ungauged basins. As such, the S-W model was used to estimate the potential ET of Poyang lake basin for the period 1982 to 2001. The results show that the averaged ET fluctuate at 900 mm and appear an inflexion in 1992. The variation of averaged ET (also ET in particular land cover type) is inconspicuous in autumn and winter, while prominent in spring and summer. The ET is maximal in July and August and minimal in winter. The interannual distribution form is consistent with the temperature distribution. The spatial distribution of ET is largely uneven with larger values in south than in north and has a close relationship with the distribution of LAI.

1. Introduction

Evapotranspiration (ET) is a transfer process of water vapor from the wet soil-vegetation system to the dry atmosphere, including interception evaporation, soil evaporation, and plant transpiration [1]. In most cases, ET is the greatest consumer of the water budget and the magnitude of ET impacts the available water in a hydrologic system affecting water yield, storage and stream flows [2]. ET is also one of the
hydrologic parameters and a major component of the water cycle and important in water resource development and management. Measurement and estimation of this parameter usually provides different results [2].

Traditional means for point ET estimation include the pan, Bowen ratio, eddy correlation, and aerodynamic techniques [3]. But the application of these point measurement to a large area is in question because the environmental conditions (such as vegetation and soil types, soil moisture, landscape and climate) are so heterogeneous in a basin, very different from the measurement sites [1]. And then, spatial ET is commonly estimated through numerical models to take the heterogeneity of basins into account. In the last two decades, many ET models have been developed and validated, from single climatic variable driven equations to energy balance and aerodynamic principle combination methods [4,5]. However, in each of these models either estimation results are poor or input data are difficult to collect. Shuttleworth and Wallace [6] extended the Penman-Monteith (P-M) method to sparse vegetation to consider two coupled sources in a resistance network: transpiration from vegetation and evaporation from substrate soil. Some researchers [7-9] have compared the P-M model and the Shuttleworth-Wallace (S-W) model, showing that the S-W model outperforms the P-M model. Zhou [4,5] successfully applied the S-W model to estimate potential evapotranspiration and drive a distributed hydrological model for the Mekong River basin. In addition, the S-W model is physically-based and all required data are publicly available and can be applied to data-poor or ungauged basins [4]. As such, the S-W model was selected to estimate the potential evapotranspiration of Poyang lake basin.

2. Study area and model descriptions

2.1. Study area

Poyang Lake is located within 28°22’~29°45’N and 115°47’~116°45’E and connects with the Yangtze River in the north (Fig. 1). The catchment has a subtropical wet climate characterized by a mean annual precipitation of 1680mm and an annual average temperature of 17.5°C. The topography varies from highly mountainous (maximum elevation of about 2200m above the sea level) to alluvial plains in the lower reaches of the primary watercourses. Poyang Lake receives water flows from five rivers: Ganjiang, Fuhe, Xinjiang, Raohe and Xiushui, and exchanges water with the Yangtze River. Lake storage and water level variation is controlled by catchment discharges and interactions with the Yangtze River [10]. Each year, the lake experiences large water level fluctuations (>10 m) in response to annual cycle of precipitation. In the wet season in spring and early summer, the lake level rises and lake coverage expands, covering an area of roughly 170 km from the north to the south and 17 km from the east to the west (about 3800 km²) [11]. The lake shrinks to little more than a river during the dry season from late August through the winter months, exposing extensive floodplains and wetland areas which support migrating waterfowls and a variety of invertebrate species [12].
2.2. Data collection

The digital elevation model (DEM) is one of the most important data sets required for the S-W models. In this study, the DEM was derived from the GLOBE (Global Land One-kilometer Base Elevation) data set with a horizontal resolution of 30 arc seconds (approximately 1km). The land cover, another important input data category, was downloaded from Global Land Cover Characteristics (GLCC) Data Base Version 2 [13], using the IGBP (International Geosphere Biosphere Programme) classification scheme which classifies the global land cover into 17 types with 1 km spatial resolution. In the Poyang lake basin, there are 15 categories as Fig. 2 shows. The NOAA /AVHRR (National Oceanic and Atmospheric Administration / Advanced Very High Resolution Radiometer) NDVI data set for monthly temporal resolutions with 8km spatial resolution is publicly available from Pathfinder AVHRR Land Data. In the S-W model, meteorological data such as mean daily temperature, mean diurnal temperature range, cloud cover, vapor pressure and wind speed are required. The IPCC (Intergovernmental Panel on Climate Change) Data Distribution Centre provides the monthly time series data of mean daily temperature, mean diurnal temperature range, cloud cover and vapor pressure at 0.5°×0.5° grids, and the CRU CL (Climate Research Unit /Climatology) 2.0 dataset provides the wind speed at 10 min grids, which was measured at majority of 10m height [14].
2.3. Evapotranspiration model.

The total evapotranspiration in the S-W model is expressed as [4,5]:

\[
\lambda ET = C_c ET_c + C_s ET_s \tag{1}
\]

where \( ET \) is the evapotranspiration (mm/day); \( \lambda \) is the latent heat of water vapourisation (MJ/kg); \( ET_c \) and \( ET_s \) are equivalent to transpiration and evaporation by applying the P-M equation to closed canopy and bare substrate, respectively; \( C_c \) and \( C_s \) are weighting coefficients.

And the every subentry is estimated as:

\[
ET_c = \frac{\Delta(R_n - G) + [(24 \times 3600) \rho c_p (e_a - e_d) - \Delta r_a^c (R_n^s - G)]/(r_a^c + r_a^c)}{\Delta + \gamma [1 + r_s^c/(r_a^c + r_a^c)]} \tag{2}
\]

\[
ET_s = \frac{\Delta(R_n - G) + [(24 \times 3600) \rho c_p (e_s - e_d) - \Delta r_a^s (R_n^s - R_n^s)]/(r_a^s + r_a^s)}{\Delta + \gamma [1 + r_s^s/(r_a^c + r_a^c)]} \tag{3}
\]

\[
C_c = \frac{1}{1 + (R_c R_a)/[R_s (R_c + R_a)]} \tag{4}
\]

\[
C_s = \frac{1}{1 + (R_s R_a)/[R_c (R_s + R_a)]} \tag{5}
\]

\[
R_a = (\Delta + \gamma) r_a^a, \quad R_c = (\Delta + \gamma) r_a^c + \gamma r_s^c, \quad R_s = (\Delta + \gamma) r_a^s + \gamma r_s^s \tag{6}
\]
where \( R_n \) and \( R_n^S \) are the net radiation at the reference height and substrate soil surface (MJ/(m\(^2\)d)); \( G \) is the substrate soil heat flux (MJ/(m\(^2\)d)); \( e_s \) and \( e_a \) are the saturation and actual vapour pressures (kPa); \( \alpha \) is the slope of saturation vapour pressure curve (kPa/°C); \( \rho \) is the mean air density at constant pressure (kg/m\(^3\)); \( c_p \) is the specific heat of moist air at constant pressure (MJ/(kg°C)); \( \gamma \) is the psychrometric constant (kPa/°C); \( r_s^c \) and \( r_a^c \) are the bulk stomatal and boundary layer resistances of the canopy (s/m); \( r_s^a \) and \( r_a^a \) are the aerodynamic resistances between soil and canopy and between canopy and reference height (s/m); and \( r_s^T \) is the surface resistance of soil (s/m).

3. Results and discussions

3.1. Temporal variation of annual evapotranspiration.

The averaged annual evapotranspiration for the period 1982 to 2001 is shown in Fig. 3. At the same time, the model also outputs the evapotranspiration of land grids which covered by evergreen needleleaf forest, woody savannas, croplands and cropland/natural vegetation mosaic respectively. It is obvious that the averaged ET fluctuates at 900 mm and appears an inflexion in 1992. And there is a trend of decreasing from 1998. The ET of evergreen needleleaf forest is largest (about 1400mm) and the cropland is the smallest (about 700mm). Fig. 4 shows the temporal variation of ET in different season. It can be seen that the variation of averaged ET (also ET in particular land cover type) is inconspicuous in autumn and winter, while prominent in spring and summer. The variation trends are consistent with annual ET (Fig. 3).

![Fig. 3 Temporal variation of averaged annual evapotranspiration](image-url)
3.2. Interannual distribution of evapotranspiration.

The averaged monthly ET is calculated to analyse the interannual distribution as Fig. 5 shows. It is obvious that the ET over whole basin and in particular land cover types have the same interannual distribution. The ET is maximal in July and August and minimal in winter. The land covered by evergreen needleleaf forest has large ET and cropland is small. This distribution form is consistent with the interannual distribution of temperature.

3.3. Spatial variation of evapotranspiration.

The S-W model also outputs the average spatial distribution of seasonal evapotranspiration in 20 years as Fig. 6 shows. It can be seen that the spatial distribution of evapotranspiration is largely uneven in study
area, and the ET in south is larger than ones in north in every season. The forest cover area has the larger ET than that of croplands. In addition, Fig.6 also reflects the variation of ET in every season. The ET is largest in summer (more than 500mm) and smallest in winter (about 40mm) and has a close relationship with LAI.

![Fig. 6 Spatial distribution and variation of evapotranspiration](image)

### 4. Conclusions

The S-W model was used to estimate the potential ET of Poyang lake basin for the period 1982 to 2001. The results show that the averaged ET fluctuates at 900 mm and appears an inflexion in 1992. And there is a trend of decreasing from 1998. The variation of averaged ET (also ET in particular land cover type) is inconspicuous in autumn and winter, while prominent in spring and summer. The ET is maximal in July and August and minimal in winter. The interannual distribution form is consistent with the temperature distribution. The spatial distribution of ET is largely uneven with larger values in south than in north and it demonstrates a close relationship with the distribution of LAI.

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References


