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Estimating potential evapotranspiration using Shuttleworth–Wallace model and NOAA-AVHRR NDVI data to feed a distributed hydrological model over the Mekong River basin

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Summary One of key inputs to hydrological modeling is the potential evapotranspiration, either from the interception (PET_0) or from the soil water of root zone (PET). The Shuttleworth–Wallace (S–W) model is developed for their estimation. In this parameterization, neither experimental measurement nor calibration is introduced. Based on IGBP land cover classification, the typical thresholds of vegetation parameters are drawn from the literature. The spatial and temporal variation of vegetation LAI is derived from the composite NOAA-AVHRR NDVI using the SiB2 method. The CRU database supplies with the required meteorological data. They are all publicly available. The developed S–W model is applicable at the global scale, particularly to the data-poor or ungauged large basins.

Using the century monthly time series of CRU TS 2.0 and the monthly composite NOAA-AVHRR NDVI from 1981 to 2000, annual PET is estimated 1354 mm over the Mekong River basin, spatially distributed strikingly non-uniformly from 300 to 2040 mm, and seasonally changed significantly with LAI. By replacing the monthly with the 10-day composite NDVI and the albedo of 0.10 with 0.15 for substrate soil surface, annual PET relatively decreases less than 4% and 1.7%, respectively over the whole basin. The correlation with pan evaporation (E_{pan}) is quite scattered but grouped with the vegetation types and consistent with a rough ratio as reported

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in the literature. In contrast, the PET and the reference evapotranspiration (RET) are vegetation-type-dependently correlated very well. The PET_0 is estimated 1.63 times of PET in average over the whole basin. The application of BTOPMC model shows that the derived LAI, PET_0 and PET behave very well in the distributed hydrological modeling.

Introduction

One of key inputs to hydrological modeling is the potential evapotranspiration, which refers to maximum meteorologically evaporative power on land surface. Two kinds of potential evapotranspiration are necessary to be defined: either from the interception, denoted by PET_0, or from the root zone when the interception is exhausted but soil water is freely available, specifically at field capacity (Federer et al., 1996; Vorosmarty et al., 1998), denoted by PET. The actual evapotranspiration is distinguished from the potential through the limitations imposed by the water deficit.

Evapotranspiration can be directly measured by lysimeters or eddy correlation method but expensively and practically only in research over a plot for a short time. The pan evaporation has long records with dense measurement sites. To apply it in hydrological models, however, first, a pan coefficient, K_p, then a crop coefficient, K_c, must be multiplied. Due to the difference on sitting and weather conditions, K_p is often expressed as a function of local environmental variables such as wind speed, humidity, upwind fetch, etc. A global equation of K_p is still lack. The values of K_c from the literature are empirical, most for agricultural crops, and subjectively selected. On the other hand, a great number of evaporation models have been developed and validated, from the single climatic variable driven equations (e.g. Thornthwaite, 1948) to the energy balance and aerodynamic principle combination methods (e.g. Penman, 1948). Among them, probably the Penman equation is physically soundest and most rigorous. Monteith (1965) generalized the Penman equation for water-stressed crops by introducing a canopy resistance. Shuttleworth and Wallace (1985) extended the Penman-Monteith method to the sparse vegetation to consider two coupled sources in a resistance network: the transpiration from vegetation and the evaporation from substrate soil. Now the Penman–Monteith (P–M) model and the Shuttleworth–Wallace (S–W) model are widely employed. The former was even standardized as FAO-24 (Doorenbos and Pruitt, 1992) and FAO-56 (Allen et al., 1998) for the reference evapotranspiration of a hypothetical crop.

The P–M model treats the crop canopy as a single uniform cover, or ''big-leaf'', but neglects the evaporation from soil surface. Over a large basin, however, the big leaf assumption is rarely valid. There are often many vegetation types co-existent, and always some parts or periods where or when the vegetation is not ''closed''. Both the soil surface and the vegetation leaves evaporate or transpire moisture to the atmosphere and their relative importance changes significantly as the vegetation develops. The ideal approach is that applicable at all times and places and able to reflect the changes of surface conditions. The S–W model meets this criterion. Stannard (1993) and Federer et al. (1996) compared a number of models, including the P–M and the S–W, and found that they give very different prediction. The research of Stannard (1993) and Vorosmarty et al. (1998) shows that hydrological modeling is sensitive to the PET methods, higher in humid regions, and the S–W model performs best. Furthermore, the interception plays an important role in water cycle. Only the S–W model is applicable to the evaporation from interception (Federer et al., 1996). Therefore the S–W model is selected for this research.

The S–W is highly complex with many parameters and demands a great deal of data on the meteorology and the land surface characteristics. Most previous work has been focusing on the model validation and comparison with some specific cover types over small experimental catchments in a short time (e.g. Iritz et al., 1999 among others) or in the water balance model at a continent (Vorosmarty et al., 1998). Its application to a large basin in a long term with a physically-based distributed hydrological model is still lack. In this research, first, the S–W model is developed only using parameter values from the literature. Neither experimental measurement nor calibration is introduced. Second, all input data are publicly available, so that it can be applied to the data-poor or ungauged basins, particularly to the large basins. Third, using this method, the spatial distribution of potential evapotranspiration is estimated for a long term over the Mekong River basin and the output is used to drive a distributed hydrological model.

Evapotranspiration model

Shuttleworth–Wallace (S–W) model

Fig. 1 shows the sensible and latent heat transfer structure of S–W model. The total evapotranspiration and each term are expressed (Shuttleworth and Wallace, 1985) as:

$$\lambda \mathrm{ET} = C_c \mathrm{ET}_c + C_s \mathrm{ET}_s \tag{1}$$

$$\mathrm{ET}_c = \frac{\Delta(R_n - G) + [(24 \times 3600)\rho c_p(e_s - e_a) - \Delta r_a^c(R_n^s - G)]/(r_a^a + r_a^c)}{\Delta + \gamma[1 + r_s^c/(r_a^a + r_a^c)]} \tag{2}$$

$$\mathrm{ET}_s = \frac{\Delta(R_n - G) + [(24 \times 3600)\rho c_p(e_s - e_a) - \Delta r_a^s(R_n - R_n^s)]/(r_a^a + r_a^s)}{\Delta + \gamma[1 + r_s^s/(r_a^a + r_a^c)]} \tag{3}$$

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

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

$$R_a = (\Delta + \gamma) r_a^a \tag{6}$$

$$R_c = (\Delta + \gamma) r_a^c + \gamma r_s^c \tag{7}$$

$$R_s = (\Delta + \gamma) r_a^s + \gamma r_s^s \tag{8}$$