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**Title:**

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**Date:**

2004-01-01

**Citation:**

Zhang, L., Hickel, K., Dawes, W. R., Chiew, F. H. S., Western, A. W. & Briggs, P. R. (2004). A rational function approach for estimating mean annual evapotranspiration. Water Resources Research, 40 (2), pp.W025021-W02502114. <https://doi.org/10.1029/2003WR002710>.

**Persistent Link:**

<https://hdl.handle.net/11343/297416>

# A rational function approach for estimating mean annual evapotranspiration

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Received 25 September 2003; accepted 16 December 2003; published 5 February 2004.

[1] Mean annual evapotranspiration from a catchment is determined largely by precipitation and potential evapotranspiration; characteristics of the catchment (e.g., soil, topography, etc.) play only a secondary role. It has been shown that the ratio of mean annual potential evapotranspiration to precipitation (referred as the index of dryness) can be used to estimate mean annual evapotranspiration by using one additional parameter. This study evaluates the effects of climatic and catchment characteristics on the partitioning of mean annual precipitation into evapotranspiration using a rational function approach, which was developed based on phenomenological considerations. Over 470 catchments worldwide with long-term records of precipitation, potential evapotranspiration, and runoff were considered, and results show that model estimates of mean annual evapotranspiration agree well with observed evapotranspiration taken as the difference between precipitation and runoff. The mean absolute error between modeled and observed evapotranspiration was 54 mm, and the model was able to explain 89% of the variance with a slope of 1.00 through the origin. This indicates that the index of dryness is the most significant variable in determining mean annual evapotranspiration. Results also suggest that forested catchments tend to show higher evapotranspiration than grassed catchments and their evapotranspiration ratio (evapotranspiration divided by precipitation) is most sensitive to changes in catchment characteristics for regions with the index of dryness around 1.0. Additionally, a stepwise regression analysis was performed for over 270 Australian catchments where detailed information of vegetation cover, precipitation characteristics, catchment slopes, and plant available water capacity was available. It is shown that apart from the index of dryness, average storm depth, plant available water capacity, and storm arrival rate are also significant. *INDEX TERMS:* 1818 Hydrology: Evapotranspiration; 1836 Hydrology: Hydrologic budget (1655); 1860 Hydrology: Runoff and streamflow; 1878 Hydrology: Water/energy interactions; *KEYWORDS:* water balance, evapotranspiration, vegetation cover

**Citation:** Zhang, L., K. Hickel, W. R. Dawes, F. H. S. Chiew, A. W. Western, and P. R. Briggs (2004), A rational function approach for estimating mean annual evapotranspiration, *Water Resour. Res.*, 40, W02502, doi:10.1029/2003WR002710.

## 1. Introduction

[2] One of the challenges in hydrology is to predict the effect of land-use changes, such as clearing and afforestation, on the partitioning of precipitation into evapotranspiration and runoff. It has been long recognized that evapotranspiration is the result of complex interactions between the atmosphere, soil, and vegetation [Brutsaert, 1982]. Although the processes by which vegetation affects

hydrological response of a catchment have been well documented, it has been difficult to develop models that can be used to make predictions at the catchment scales. This is partly because of the lack of understanding about the interactions and feedbacks among the processes and limited available data. It is almost inevitable that for such a model to be of practical use it has to be simple and based on measurable properties of the catchment. Despite the complex processes and interactions that are associated with evapotranspiration, it has long been assumed that available energy and water are the primary factors determining the rate of evapotranspiration [Budyko, 1958].

[3] A number of simple models for estimating mean annual evapotranspiration have been developed based on the above assumption with various degrees of empiricism [Schreiber, 1904; Budyko, 1958; Pike, 1964]. However, these models do not consider the effect of catchment characteristics on evapotranspiration. Choudhury [1999] generalized these relationships by introducing an adjustable parameter. Recently, Zhang *et al.* [2001] developed a similar model with a parameter assumed to be controlled by soil water storage. A common feature of these models is to assume that evapotranspiration is a known function of precipitation and available energy (as measured by potential evapotranspiration). Although the functional forms of these equations differ, their numerical values are similar. In these models, mean annual evapotranspiration generally approaches precipitation in regions where potential evapotranspiration greatly exceeds the precipitation, and conversely, mean annual evapotranspiration approaches potential evapotranspiration where precipitation is significantly greater than potential evapotranspiration. However, these studies offered no scientific basis as to why a particular function was chosen apart from the fact that it respects the above physical limits.

[4] Milly [1994] developed a mathematical framework for mean annual evapotranspiration that provides insights into these relationships. The model is based on the hypothesis that the long-term evapotranspiration is determined by the local interaction of precipitation and potential evapotranspiration, mediated by soil water storage. With his theoretical framework, he identified several key variables believed to be responsible for the partitioning of precipitation into evapotranspiration and runoff. Although simple in concept, the practical application of this model is limited because of the complex numerical solutions required.

[5] In 1981, B. P. Fu, from Nanjing University, China, published a paper on relationships between long-term evapotranspiration and precipitation [Fu, 1981]. He combined dimensional analysis with mathematical reasoning and developed analytical solutions for mean annual evapotranspiration. The relationships represented by his model are very similar to those known as Budyko's curve. The paper was published in Chinese only, and 20 years later the work is still relatively unknown outside China.

[6] This paper aims to contribute to the understanding of the effects of climatic and catchment characteristics on the partitioning of mean annual precipitation into evapotranspiration. We will revisit the work of Fu [1981] and use this to gain some insight into the key catchment properties controlling the partitioning. The data used in the analysis cover wide ranges of climate, vegetation, and other catchment characteristics. The results therefore are expected to be generally applicable to catchments in most geographic zones.

## 2. Theoretical Framework

[7] The concept of water balance provides a useful framework for evaluating the hydrological response of a catchment under changed land-use conditions. The water balance at a whole-catchment scale (surface plus groundwater) can be written as

$$P = E + Q + \Delta S \quad (1)$$

where  $P$  is precipitation,  $E$  is evapotranspiration, and  $Q$  is total runoff (surface runoff, interflow, and baseflow), and  $\Delta S$  is the change in catchment water storage. When averaged over a long period, the change in catchment water storage ( $\Delta S$ ) can be neglected. The evaluation of the water balance equation requires information about catchment physical characteristics, climatic variables, and further relationships. The uncertainties in determining spatial and temporal distributions of the climatic variables, especially precipitation, constitute a major obstacle to the understanding of hydrological behavior at the catchment scales [Milly and Dunne, 2002].

[8] On the basis of phenomenological considerations, Fu [1981] postulated that over a mean annual timescale for a given potential evapotranspiration rate ( $E_0$ ), the rate of the change in catchment evapotranspiration with respect to precipitation ( $\partial E / \partial P$ ) increases with residual potential evapotranspiration ( $E_0 - E$ ) but decreases with precipitation ( $P$ ). Similarly, for a given precipitation the rate of the change in evapotranspiration with respect to potential evapotranspiration ( $\partial E / \partial E_0$ ) increases with residual precipitation ( $P - E$ ), and decreases with potential evapotranspiration ( $E_0$ ).

[9] Mathematically, these relationships are expressed as

$$\frac{\partial E}{\partial P} = f(E_0 - E, P) \quad (2)$$

$$\frac{\partial E}{\partial E_0} = \varphi(P - E, E_0) \quad (3)$$

where  $f$  and  $\varphi$  are functions to be determined.

[10] On the basis of dimensional analysis, the right-hand side of equation (2) contains only one independent dimensional variable. If we chose precipitation  $P$  as a dimensional variable, then

$$\pi = \frac{\partial E}{\partial P}, \quad \pi_1 = \frac{E_0 - E}{P} \quad (4)$$

Therefore

$$\pi = f(\pi_1) \quad (5)$$

or

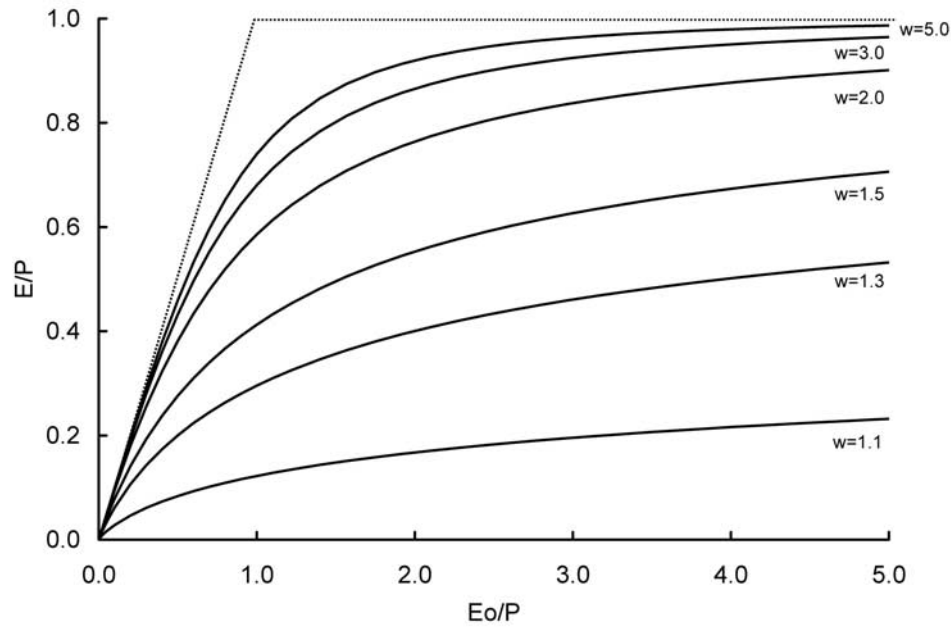
$$\frac{\partial E}{\partial P} = f\left(\frac{E_0 - E}{P}\right) \quad (6)$$

Similarly, equation (3) can be expressed as

$$\frac{\partial E}{\partial E_0} = \varphi\left(\frac{P - E}{E_0}\right) \quad (7)$$

Let

$$\left. \begin{aligned} x &= \frac{E_0 - E}{P} \\ y &= \frac{P - E}{E_0} \end{aligned} \right\} \quad (8)$$



**Figure 1.** Ratio of mean annual evapotranspiration to precipitation ( $E/P$ ) as a function of the index of dryness ( $E_0/P$ ) for different values of parameter  $w$ .

Substituting equation (8) into (6) and (7), one obtains

$$\frac{\partial E}{\partial P} = f(x) \quad (9a)$$

$$\frac{\partial E}{\partial E_0} = \varphi(y) \quad (9b)$$

Under extreme wet conditions,  $E$  approaches  $E_0$  and will not increase with precipitation  $P$  because under this condition  $E$  is limited by potential evapotranspiration  $E_0$ . While under extreme dry conditions,  $E$  approaches precipitation  $P$  and will not increase with potential evapotranspiration. These boundary conditions can be expressed as

$$\left. \begin{aligned} \frac{\partial E}{\partial P} \Big|_{x=0} &= 0 \\ \frac{\partial E}{\partial E_0} \Big|_{y=0} &= 0 \end{aligned} \right\} \quad (10)$$

Solving equations (9a) and (9b) subject to the boundary conditions (10), the following solutions can be obtained:

$$\frac{E}{P} = 1 + \frac{E_0}{P} - \left[ 1 + \left( \frac{E_0}{P} \right)^w \right]^{1/w} \quad (11)$$

$$\frac{E}{E_0} = 1 + \frac{P}{E_0} - \left[ 1 + \left( \frac{P}{E_0} \right)^w \right]^{1/w} \quad (12)$$

where  $w$  is a model parameter. These relationships are shown in Figure 1, and details of the solutions are given in Appendix A. It should be noted that estimation of evapotranspiration using equations (11) and (12) requires

mean annual values of precipitation, potential evapotranspiration, and estimates of the model parameter  $w$ .

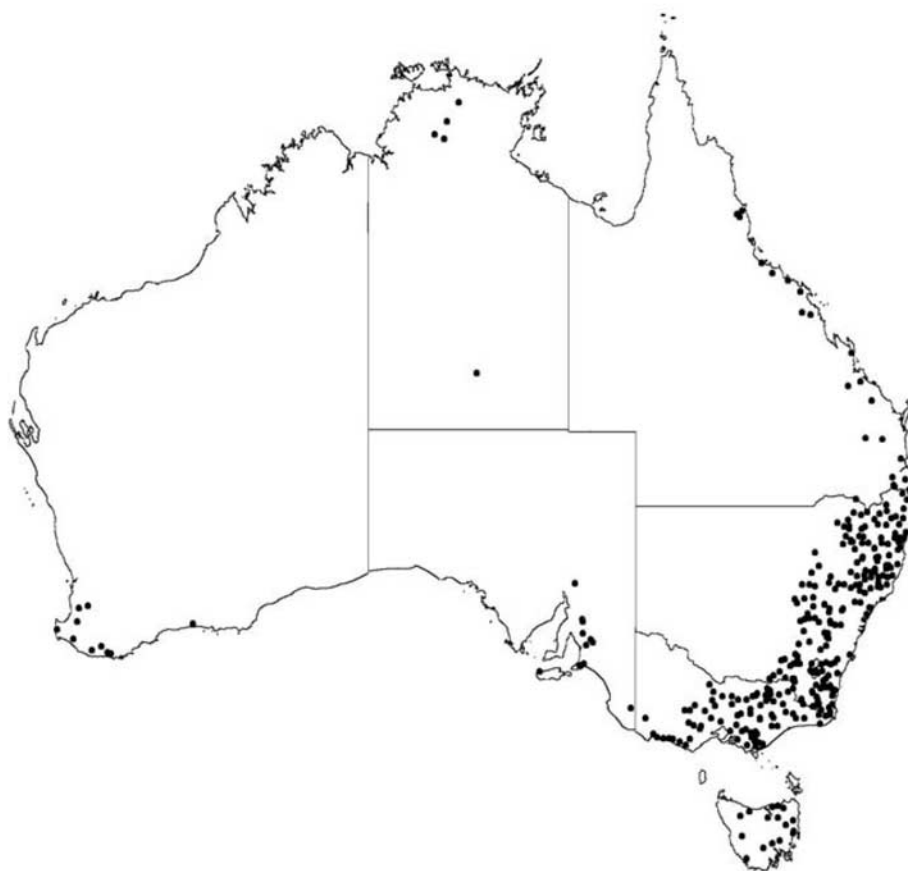
### 3. Data Description

[11] The data used in this study were obtained from two sources: Paired-catchment studies and single-catchment water balance studies. There are some noticeable differences between these two types of studies. The paired-catchment studies generally involve small catchments ( $<10 \text{ km}^2$ ) with detailed precipitation, streamflow, and vegetation data. The main objective of these studies was to estimate changes in catchment water yield. The single-catchment water balance studies focused on relationships among precipitation, runoff, and evapotranspiration. These are generally larger catchments with a long-term record of precipitation and runoff. However, information on vegetation type and history is incomplete. In total there were 470 catchments used in the study including 61 paired catchments, and they represent a range of climates including tropical, dry, and mild midlatitude based on the Köppen classification. It should be noted that rainfall is the dominant form of precipitation in these catchments and average annual precipitation varied from 35 to 2980 mm with different seasonal distributions. The vegetation in these catchments ranges from even-age plantations to native woodlands, open forests, rainforest, to native and managed grasslands and agricultural corps. Details of these catchments are described below.

#### 3.1. Detailed Australian Database

##### 3.1.1. Streamflow

[12] The catchments included in this study have at least 10 years and in most cases 20 years of unimpaired streamflow data and a catchment area between 50 and 2000  $\text{km}^2$ . Unimpaired streamflow is defined as streamflow that is not subject to regulation or diversion. The streamflow data were



**Figure 2.** Location map of 331 Australian catchments used in this study.

assembled by *Peel et al.* [2000]. In total, 308 gauging stations were selected: 6 in the Australian Capital Territory, 178 in New South Wales, 5 in the Northern Territory, 19 in Queensland, 11 in South Australia, 10 in Tasmania, 69 in Victoria, and 10 in Western Australia. The locations of these gauging stations are shown in Figure 2.

### 3.1.2. Precipitation

[13] Annual precipitation was estimated from gridded daily precipitation [*Peel et al.*, 2000]. The spatial resolution of the gridded daily precipitation is  $5 \times 5$  km based on interpolation of over 6000 rainfall stations in Australia. The interpolation uses monthly precipitation data, ordinary kriging with zero nugget, and a variable range. Monthly precipitation for each  $5 \times 5$  km grid cell is converted to daily precipitation using daily precipitation distribution from the station closest to the grid cell. Finally, mean annual precipitation was estimated from the daily precipitation.

### 3.1.3. Potential Evapotranspiration

[14] Monthly potential evapotranspiration was calculated using the Priestley-Taylor equation [*Priestley and Taylor*, 1972]. The data were average over time and space to obtain catchment mean annual potential evapotranspiration. The spatial resolution of the data is  $5 \times 5$  km, and details are given by *Raupach et al.* [2001].

### 3.1.4. Ancillary Data

[15] For each of the catchments shown in Figure 2, data on land use, topographic characteristics, and soil properties

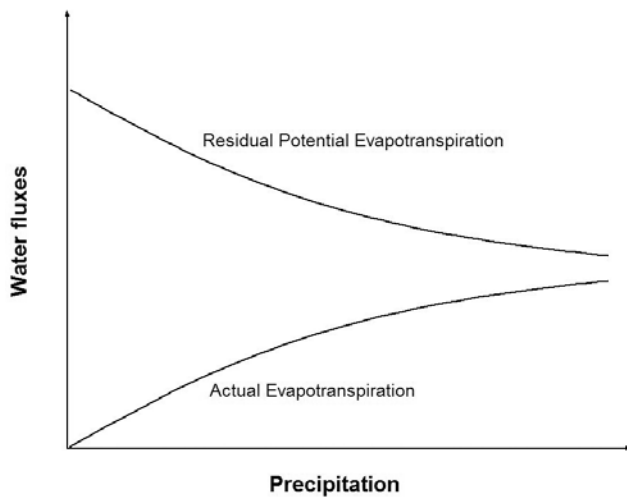
were obtained. The land-use data include mainly percentages of forest cover, cropping areas, and open water. The topographic data (median surface slope, and relief ratio) were derived from 30-m horizontal resolution digital elevation models. The soil properties include upper and lower soil-horizon soil-moisture storage capacities, soil types, and depths of A and B horizon. On the basis of soil properties and the topographic data, plant available water storage capacity (PAWC) was calculated using the method of *McKenzie et al.* [2003].

### 3.2. Other Australian Database

[16] In addition to the database described above, values of long-term average precipitation, streamflow, and potential evapotranspiration are also available for some other Australian catchments. Vegetation cover information was obtained from *Ritman* [1995] for these catchments. *Zhang et al.* [1999] listed the data used in the present study. It should be noted that this database does not include detailed information of precipitation and other catchment characteristics.

### 3.3. International Database

[17] Mean annual precipitation, streamflow, and potential evapotranspiration were also obtained for 162 overseas catchments covering a large range of geographical distribution. These data will allow us to test the applicability of the



**Figure 3.** Schematic representation of the complementary relationship between residual potential evaporation and actual evapotranspiration assuming constant energy supply.

model to various climatic zones. The locations and details of these catchments are given by *Zhang et al.* [1999, 2001].

## 4. Results and Discussion

### 4.1. Model Assumption and Sensitivity Analysis

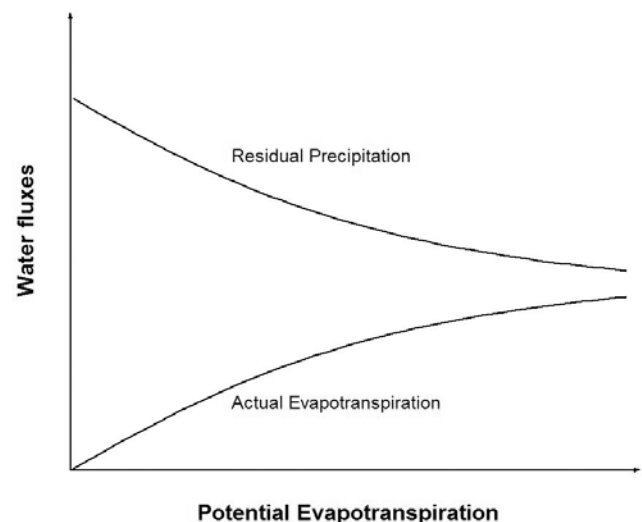
[18] The hypotheses in Fu's model state that for a given potential evapotranspiration, the rate of the change in catchment evapotranspiration with respect to precipitation ( $\partial E/\partial P$ ) increases with residual potential evapotranspiration ( $E_0 - E$ ) but decreases with precipitation ( $P$ ). Similarly, for a given precipitation the rate of the change in evapotranspiration with respect to potential evapotranspiration ( $\partial E/\partial E_0$ ) increases with residual precipitation ( $P - E$ ) but decreases with potential evapotranspiration ( $E_0$ ).

[19] It can be argued that for a given catchment any change in evapotranspiration with respect to precipitation is a function of potential evapotranspiration and precipitation itself. Here it is assumed that precipitation is the only source of water supply for evapotranspiration in a catchment. When there is no precipitation in the catchment, it follows that evapotranspiration is zero, and the air is very dry and hot and that potential evapotranspiration is at its maximum rate. As the precipitation increases, evapotranspiration will increase, and it causes the air above to become cooler and more humid, which in turn causes potential evapotranspiration to decrease. If evapotranspiration rate is reduced below the potential rate due to limited precipitation, then an amount of energy would be released. This is the amount of energy that would otherwise be used for evapotranspiration if water supply were not limiting, and it would equal the residual potential evapotranspiration ( $E_0 - E$ ). Therefore the residual potential evapotranspiration is high when the catchment is dry and it decreases with precipitation.

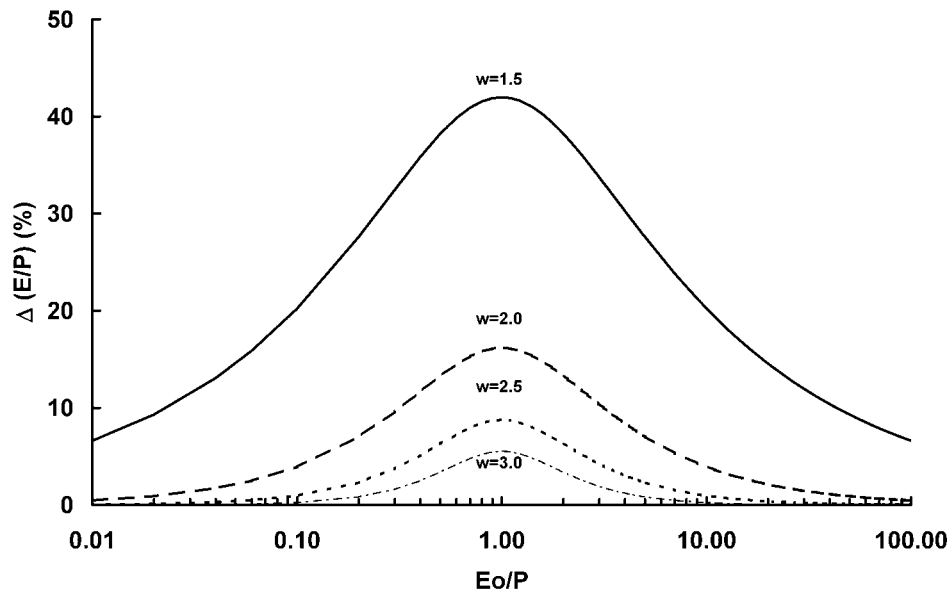
[20] When the catchment is dry, any increase in precipitation will result in an equivalent increase in evapotranspiration since the potential evapotranspiration is not limiting. As a result, the rate of change in evapotranspiration with respect to precipitation will increase with the residual

potential evapotranspiration. However, as precipitation increases evapotranspiration becomes energy limited, and the rate of the change in evapotranspiration due to precipitation will decrease since no additional energy is available for evapotranspiration. The residual potential evapotranspiration can be regarded as a measure of evapotranspiration efficiency in terms of potential evapotranspiration. Fu's assumption can be restated as follows: Actual evapotranspiration from a catchment is in a complementary relationship with the residual potential evapotranspiration, and this can be considered as a complementary relationship in energy. Figure 3 provides a schematic representation of this relationship. This assumption is similar to the hypothesis of the complementary relationship between actual and potential evapotranspiration introduced by *Bouchet* [1963] and further developed by *Morton* [1983]. However, it should be noted that *Bouchet* [1963] assumed the relationship is linear and actual and potential evapotranspiration meet halfway, while *Fu* [1981] only assumed the slope of the relationship and did not impose linearity or the assumption that they meet halfway.

[21] When precipitation is a constant, evapotranspiration varies with potential evapotranspiration, but the rate of the change will decrease with potential evapotranspiration as precipitation is becoming the limiting factor. If evapotranspiration is reduced below the precipitation for a reason independent of water supply, then an amount of water would become available, and it equals the residual precipitation ( $P - E$ ). It is clear that when change in catchment water storage is negligible, the residual precipitation equals runoff. As precipitation is a constant, the residual precipitation decreases with potential evapotranspiration. The rate of the change in evapotranspiration therefore increases with the residual precipitation. This assumption is similar to the first assumption discussed above, and it can be restated as follows: Actual evapotranspiration from a catchment is in a complementary relationship with the residual precipitation, and this can be considered as a complementary relationship



**Figure 4.** Schematic representation of the complementary relationship between residual precipitation and actual evapotranspiration assuming constant water supply.



**Figure 5.** Sensitivity of the evapotranspiration ratio ( $E/P$ ) to the catchment parameter ( $w$ ) as represented by *Fu* [1981]. At extreme values of  $E_0/P$ , there is the least sensitivity to  $w$ , while at the critical point where  $E_0 = P$ , the greatest sensitivity to  $w$ , and for every value of  $w$ , is reached.

in water. Figure 4 provides a schematic representation of the relationship.

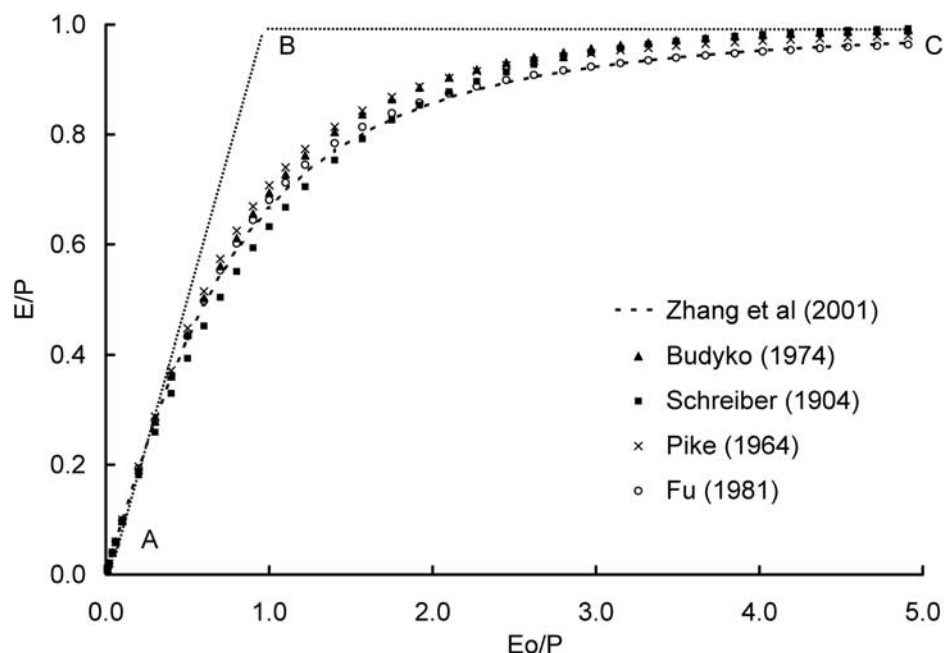
[22] The relationship represented by equation (11) suggests that the evapotranspiration ratio ( $E/P$ ) can be determined by the index of dryness ( $E_0/P$ ) and a catchment parameter ( $w$ ). Similarly, the evapotranspiration efficiency ( $E/E_0$ ) can be determined by the index of wetness ( $P/E_0$ ), which is the reciprocal of the index of dryness and the catchment parameter ( $w$ ) (see equation (12)). It may not be intuitively obvious that the relationships expressed in equations (11) and (12) should have the same form. This is a mathematical feature of *Fu* [1981], and in practice one can use either of the relationships to calculate evapotranspiration. *Dooge et al.* [1999] argued that such a relationship is consistent with the concept of geographical zonality put forward by *Budyko* [1950]. The relationship is also consistent with the theoretical framework developed by *Milly* [1994], which states that the long-term water balance is determined by water supply (precipitation) and demand (potential evapotranspiration), mediated by water storage. The index of dryness (potential evapotranspiration divided by precipitation) in equation (11) represents climatic impact on water balance, while the parameter ( $w$ ) plays a role similar to the water storage term in *Milly's* model. It can be

argued that the storage term in *Milly's* model has a physical definition from a water balance point of view, while the parameter ( $w$ ) in *Fu's* equation is empirical. Nevertheless, one can assume that the parameter ( $w$ ) represents the integrated effects of catchment characteristics such as vegetation cover, soil properties, and catchment topography on water balance. It is also possible to make some generalization as to how the parameter ( $w$ ) will vary with the catchment characteristics. However, it is a challenge to quantify the relationships among them given the feedback processes and interaction are not fully understood.

[23] Figure 5 shows the effect of the catchment parameter ( $w$ ) on evapotranspiration ratio ( $E/P$ ). When the  $w$  parameter increases from 1.5 to 2.0, the evapotranspiration ratio can increase by up to 41%. However, when the  $w$  parameter increases from 2.5 to 3.0, the evapotranspiration ratio only increases by up to 9%. It is clear that the sensitivity of the evapotranspiration ratio to the  $w$  parameter decreases rapidly with increasing  $w$ . The effect of  $w$  on the evapotranspiration ratio is minimal under both very dry and very wet conditions, except for small  $w$  values (i.e.,  $w < 1.5$ ). The reason for this is that under these two extreme conditions evapotranspiration is dominated by precipitation and available energy. When

**Table 1.** Description of Different Relationships for Estimating Annual Evapotranspiration

| Equation                                          | Symbol                                                                                                                                                 | Reference                  |
|---------------------------------------------------|--------------------------------------------------------------------------------------------------------------------------------------------------------|----------------------------|
| $E = P[1 - \exp(-E_0/P)]$                         | $E$ , $P$ , and $E_0$<br>are annual values of actual<br>evapotranspiration, precipitation,<br>and potential evapotranspiration,<br>respectively, in mm | <i>Schreiber</i> [1904]    |
| $E = P/[1 + (P/E_0)^2]^{0.5}$                     | as above                                                                                                                                               | <i>Pike</i> [1964]         |
| $E = [P(1 - \exp(-E_0/P))E_0 \tanh(P/E_0)]^{0.5}$ | as above                                                                                                                                               | <i>Budyko</i> [1974]       |
| $E = P(1 + w(E_0/P))/(1 + w(E_0/P) + P/E_0)$      | as above, $w$ is a coefficient between 0.5 and 2.0                                                                                                     | <i>Zhang et al.</i> [2001] |



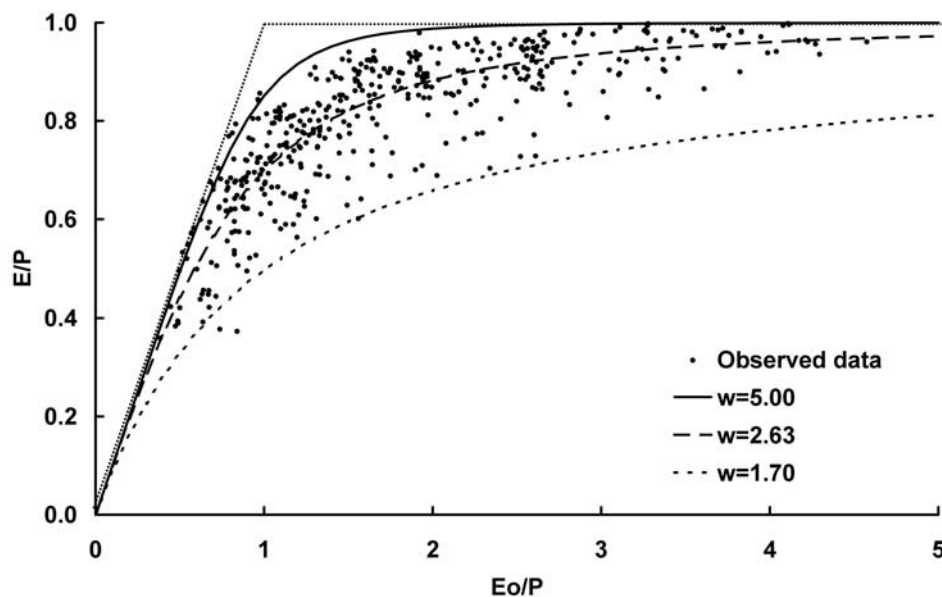
**Figure 6.** Comparison of the relationships developed by *Schreiber* [1904], *Pike* [1964], *Budyko* [1974], *Fu* [1981], and *Zhang et al.* [2001].

the index of dryness is around 1, the evapotranspiration ratio shows maximum sensitivity to the catchment parameter  $w$ . Under this condition, both potential evapotranspiration and precipitation will have the same control over evapotranspiration.

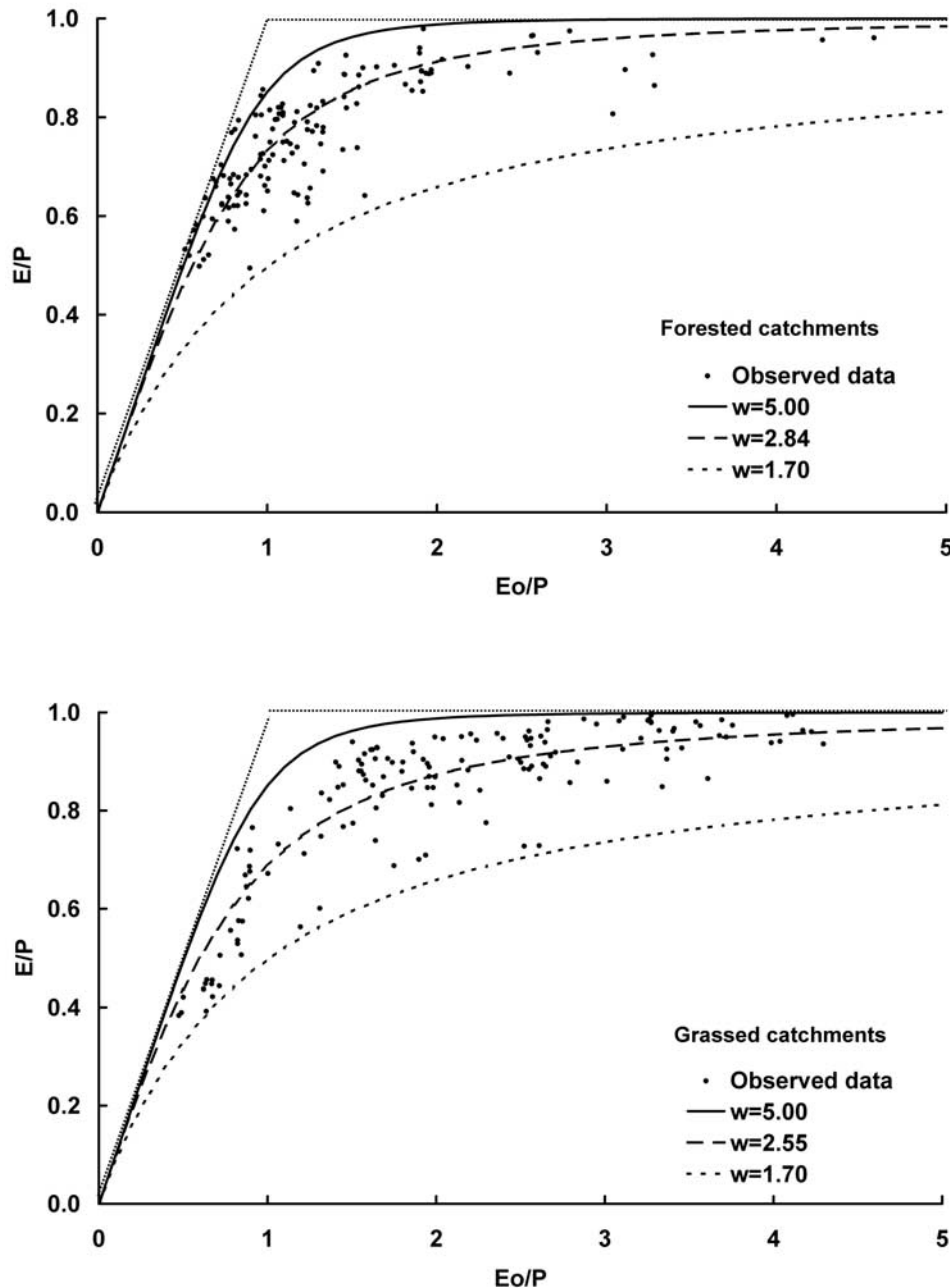
#### 4.2. Comparison With Empirical Equations and the Observed Data

[24] A number of models have been developed for estimating mean annual evapotranspiration. A common

feature of these models is the assumption that evapotranspiration is limited by available water (i.e., precipitation) under very dry conditions and available energy (i.e., potential evapotranspiration) under very wet conditions. A list of these equations is given in Table 1, and it is clear that the functional forms of these models differ. A comparison of these models with *Fu* [1981] shows that there is good agreement between these relationships (Figure 6). In the comparison, the parameter was set to 2.5 by *Fu* [1981] and to 1.0 by *Zhang et al.* [2001].



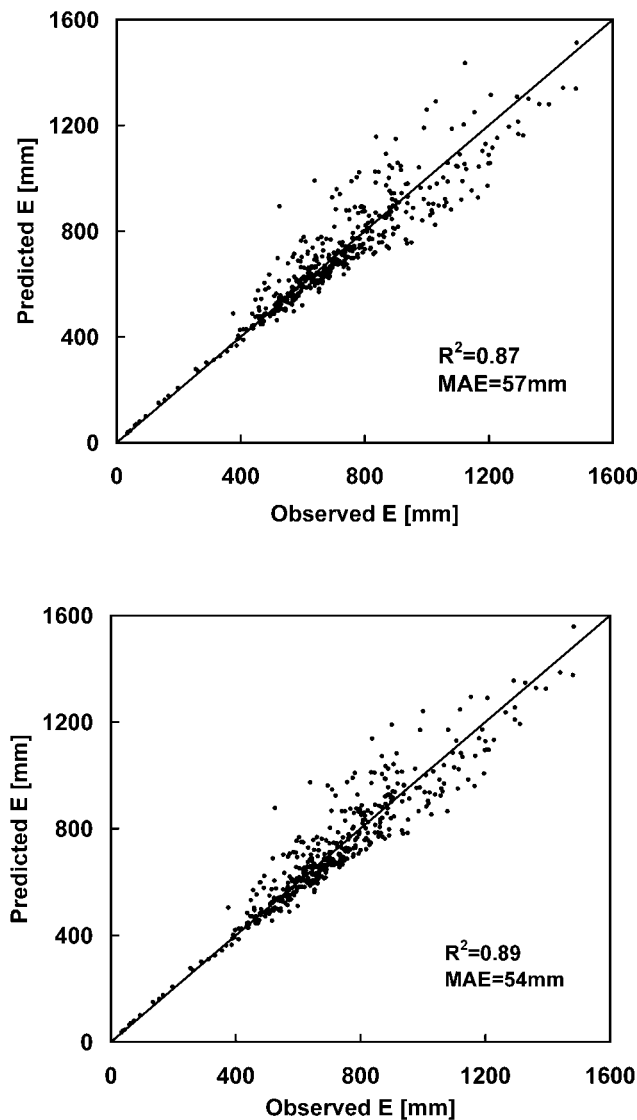
**Figure 7.** Scatterplot of evapotranspiration ratio ( $E/P$ ) against index of dryness ( $E_o/P$ ). Each point represents one catchment, with evapotranspiration taken as the difference between precipitation and runoff. Lines are the relationships represented by *Fu* [1981] with different values of  $w$  parameter.



**Figure 8.** Scatterplot of evapotranspiration ratio ( $E/P$ ) against index of dryness ( $E_0/P$ ). Each point represents one forested catchment, with evapotranspiration taken as the difference between precipitation and runoff. Lines are the relationships represented by *Fu* [1981] with different values of  $w$  parameter. Top panel is for forested catchments with the best fit  $w$  value of 2.84, and bottom panel is for grassed catchments with the best fit  $w$  value of 2.55.

[25] We compared evapotranspiration ratio predicted by *Fu* [1981] with all the observed data in Figure 7. In the calculation, mean annual values of precipitation and potential evapotranspiration were used. The best fit value of  $w$  is 2.63 with a mean absolute error (MAE) of 6.0% and  $r^2$  of 0.78. The top curve practically represents the maximum evapotranspiration ratio for the catchments considered, and the value of  $w$  is 5.0. Catchments following this curve would evaporate at a rate determined by either precipitation or potential evapotranspiration. The

bottom curve is characterized by a  $w$  value of 1.7, and catchments following this curve have lower evapotranspiration ratios. These catchments generally have climatic or catchment properties which are not favorable for evapotranspiration, and these factors include precipitation intensity, seasonality, slope, and soil water storage capacity. As a result, a larger fraction of precipitation becomes runoff in these catchments, resulting in lower evapotranspiration ratios. Smaller values of  $w$  are believed to be associated with steep slopes, high precipitation intensity,



**Figure 9.** Comparison between predicted and observed evapotranspiration taken as the difference between precipitation and runoff for all the catchments. Top panel is for a single  $w$  value of 2.63 for all the catchments, and bottom panel is for  $w$  values of 2.84 (forested), 2.55 (grassed), and 2.53 (mixed) catchments.

and lower plant available water storage capacity. However, it is difficult to represent these factors explicitly in a simple model such as that of *Fu* [1981].

[26] It is clear that there is a scatter in the data and catchments, with an index of dryness ( $E_0/P$ ) around 1.0 tending to show larger scatter, while catchments in dry climatic zones (e.g.,  $E_0/P > 3.0$ ) showed little scatter in the evapotranspiration ratio ( $E/P$ ). It can also be noted that even in very dry climatic zones (e.g.,  $E_0/P > 3.0$ ) evapotranspiration can be less than the mean annual precipitation. This is mainly because the precipitation in these catchments is episodic with high intensity. As a result, surface runoff would occur even though the mean annual potential evapotranspiration can greatly exceed the mean annual precipitation. We selected two types of catchments based on vegetation cover:

Forested catchments (over 75% tree cover) and grassed catchments (over 75% grass cover). The best fit value of  $w$  is 2.84 and 2.55 for the forested and grassed catchments, respectively (Figure 8). These results suggest that forested catchments tend to show higher evapotranspiration compared with grassed catchment. This is mainly because forests and plantations have deeper roots, lower aerodynamic resistance, and higher and more persistent leaf area [*Vertessy and Bessard*, 1999; *Zhang et al.*, 2001].

[27] Equation (11) can be used to calculate actual evapotranspiration when precipitation and potential evapotranspiration are known. A comparison of observed and calculated evapotranspiration is shown in Figure 9. In the calculation, a single value of 2.63 was used for the  $w$  parameter, and it resulted in a MAE of 57 mm, or 8%. The correlation coefficient is 0.87, and the best fit slope through the origin is 1.00. A comparison was also made between observed and calculated evapotranspiration using optimized  $w$  values of 2.84 for forested, 2.55 for grassed, and 2.53 for mixed catchments. The results showed a slight improvement with a mean absolute error of 54 mm. The correlation coefficient is 0.89, and the best fit slope through the origin is 1.00. It can be noted that the optimized  $w$  value for catchments with mixed vegetation is essentially the same the optimized  $w$  value for grassed catchments. However, this does not mean that catchments with mixed vegetation would have the same evapotranspiration as grassed catchments. The models listed in Table 1 may perform equally well given the comparison shown in Figure 6. However, the advantages of *Fu* [1981] over these models are the better conceptual basis and mathematical rigor.

#### 4.3. Model Parameterization

[28] Results in the previous sections showed that the model developed by *Fu* [1981] compared well with observed mean annual evapotranspiration. The model requires precipitation and potential evapotranspiration as input, and it also requires an estimate of the  $w$  parameter. A number of studies showed that mean annual evapotranspiration generally correlated well with mean annual precipitation, and for temperate and dry climate, precipitation is the dominant factor [*Turner*, 1991; *Zhang et al.*, 2001]. Here we evaluate the impact of using a constant potential evapotranspiration term on predicted evapotranspiration for forested and grassed catchments in Australia. In the analysis, equation (11) was optimized using two different objective functions: (1) to minimize the difference between observed and predicted evapotranspiration ratio, and (2) to minimize the difference between observed and predicted evapotranspiration (see Table 2).

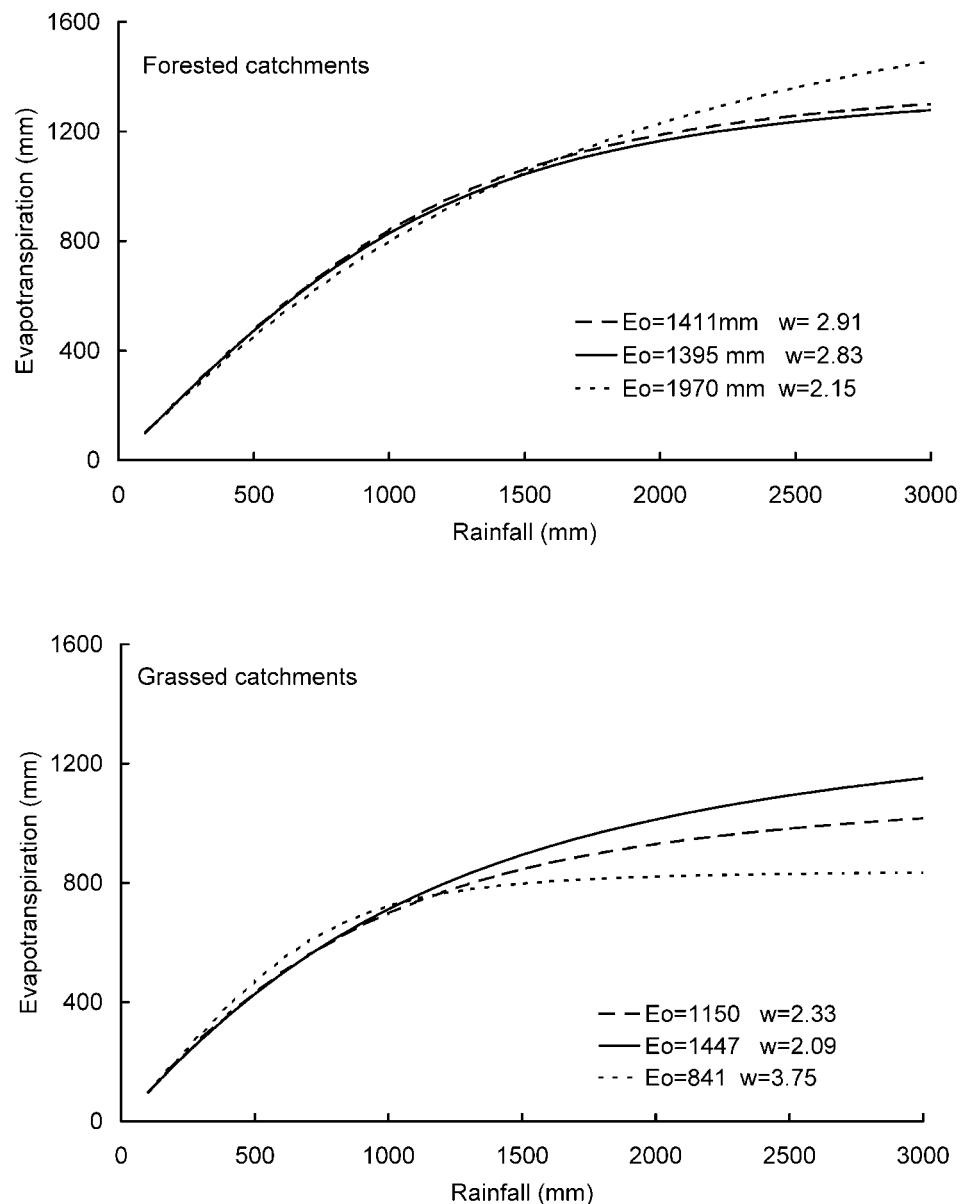
[29] When using observed potential evapotranspiration and optimizing for the  $w$  parameter, the two objective functions yielded slightly different values of  $w$ . However, there is little difference in terms of the mean absolute errors and correlation coefficient (Table 2). For the catchments considered, precipitation is far more variable than potential evapotranspiration and the coefficient of variation is 0.5 for precipitation and 0.25 for potential evapotranspiration. For a fixed value of  $w$ , the optimized value for  $E_0$  is 1411 mm for forested catchments and 1150 mm for grassed catchments. Compared with using observed potential evapotranspiration, this on average increased the MAE by 25%. Obviously, using a constant value of  $E_0$  for all catchments will increase the

**Table 2.** Model Parameterization for Forested and Grassed Australian Catchments With Different Objective Functions

|                                                                                      | $E_0$ , mm   | $w$          | $R^2$ | MAE, mm |
|--------------------------------------------------------------------------------------|--------------|--------------|-------|---------|
| <i>Forest</i>                                                                        |              |              |       |         |
| Variable $E_0$ , optimize for $\alpha$ to minimize objective function 1 <sup>a</sup> | ...          | 2.84         | 0.87  | 65      |
| Variable $E_0$ , optimize for $\alpha$ to minimize objective function 2 <sup>b</sup> | ...          | 2.91         | 0.87  | 65      |
| Fixed $\alpha$ optimize for $E_0$ to minimize objective function 2                   | 1411         | 2.91 (fixed) | 0.82  | 82      |
| Fixed $E_0$ , optimize for $\alpha$ to minimize objective function 2                 | 1395 (fixed) | 2.83         | 0.82  | 80      |
| Optimize both $E_0$ and $\alpha$ to minimize objective function 2                    | 1970         | 2.15         | 0.83  | 76      |
| <i>Grass</i>                                                                         |              |              |       |         |
| Variable $E_0$ , optimize for $\alpha$ to minimize objective function 1              | ...          | 2.55         | 0.90  | 45      |
| Variable $E_0$ , optimize for $\alpha$ to minimize objective function 2              | ...          | 2.33         | 0.90  | 47      |
| Fixed $\alpha$ optimize for $E_0$ to minimize objective function 2                   | 1150         | 2.33 (fixed) | 0.89  | 62      |
| Fixed $E_0$ , optimize for $\alpha$ to minimize objective function 2                 | 1447 (fixed) | 2.09         | 0.86  | 70      |
| Optimize both $E_0$ and $\alpha$ to minimize objective function 2                    | 841          | 3.75         | 0.91  | 43      |

<sup>a</sup>Objective function 1:  $\text{Obj1} = \min \sum \left[ \left( \frac{E}{P} \right)_{\text{obs}} - \left( \frac{E}{P} \right)_{\text{est}} \right]^2$ .

<sup>b</sup>Objective function 2:  $\text{Obj2} = \min \sum [(E)_{\text{obs}} - (E)_{\text{est}}]^2$ .

**Figure 10.** Comparison of the relationship developed by *Fu* [1981] with different parameterization schemes. Top panel is for forested catchments, and bottom panel is for grassed catchments.

**Table 3.** Results of Stepwise Regression Between Evapotranspiration Residuals and the Nominated Variables

| Variables      | Coefficients         | Const  | $R^2$ | Standard Error |
|----------------|----------------------|--------|-------|----------------|
| ASD            | 122.6                | -352.9 | 0.148 | 81.7           |
| ASD, PAWC      | 122.5, -0.2397       | -324.4 | 0.177 | 80.3           |
| ASD, PAWC, SAR | 60.4, -0.2915, 294.0 | -195.0 | 0.198 | 79.3           |

error in evapotranspiration predictions, and it is expected that the impact will be greater for wet catchments than for dry catchments. When average  $E_0$  was used and the model was optimized for  $w$ , the best fit value of  $w$  is 2.83 for the forested catchments and 2.09 for the grassed catchments. Finally, both  $E_0$  and  $w$  were optimized. For the forested catchments, the optimal values are 1970 mm for  $E_0$  and 2.15 for  $w$  and the mean absolute error is 76 mm with a correlation coefficient of 0.83. For the grassed catchments, the optimal values are 841 mm for  $E_0$  and 3.75 for  $w$  and the mean absolute error is 43 mm with a correlation coefficient of 0.91.

[30] The results of the model parameterization are summarized in Figure 10. It is clear that for the forested catchments with annual precipitation less than 2000 mm, there is little difference in predicted evapotranspiration whether  $E_0$  or  $w$  or both are optimized. For the grassed catchments, optimizing  $E_0$  tends to yield larger estimates of evapotranspiration when annual precipitation exceeding 1000 mm, while optimizing both  $E_0$  and  $w$  tends to yield larger evapotranspiration values for low precipitation catchments. The value for  $w$  obtained this way is not consistent with the values obtained using the other methods.

#### 4.4. Factors Controlling Average Annual Water Balance

[31] The model developed by *Fu* [1981] provides a useful framework for evaluating average annual water balance at the catchment scale. However, it is clear that the model has some degree of empiricism, especially when it comes to the parameter estimation. Our understanding of hydrological processes and factors associated with them has sufficiently advanced to allow us to make predictions of catchment water balance. At the same time, hydrologists are attempting to develop fully physically based hydrological models that require no calibration. In practice, however, it is inevitable that any hydrological model that describes the partitioning of annual precipitation into evapotranspiration and runoff will have some degree of empiricism because a large number of complex processes and interactions involved. These processes and interactions are further modified by factors such as climatic and catchment characteristics.

[32] The climatic factors include precipitation, solar radiation, air temperature, humidity, and wind speed. It should be noted that not only their mean annual values, but also their variability over time will affect catchment water balance [*Milly*, 1994]. The catchment factors include physical features of the landscape and vegetation characteristics. Soil properties such as water storage capacity and permeability can affect the partitioning of precipitation into evapotranspiration and runoff. The partitioning can also be affected by catchment topography. For example, in catchments with steep slopes, a greater fraction of precipitation may become runoff and this effect can be just as

important as the other factors. The vegetation factors include leaf area, rooting depth, plant structure, radiative properties, and stomatal functions.

[33] Understanding these individual factors in relation to hydrological processes is an important step in predicting catchment water balance. However, we should be aware that spatial variability, feedbacks, and interaction of these factors add complexity. We also need to be reminded that some of these factors may have coevolved (especially in natural catchments) in such a way that it is the combined effect that affects the average water balance.

[34] Recent work by *Arora* [2002] compared a general circulation model (GCM) to several single-line empirical evapotranspiration and precipitation relationships. He found that in the absence of significant permanent snow and ice within a catchment, the GCM model produced mean annual totals of evapotranspiration and runoff very similar to those of *Budyko* [1974], *Pike* [1964], and *Zhang et al.* [2001]. This result can be viewed in two ways. First, it shows mean annual water balance behavior is preserved reasonably well in the GCM, which incorporates more complex and detailed processes. Second, and more important for this work, it affirms that precipitation and available energy are the primary control of mean annual evapotranspiration. It should be noted that the comparison performed by *Arora* [2002] is valid only for ice- and snow-free catchments. *Milly* [1994] also commented on the implications of frozen precipitation, snowpacks, and snowmelt on water balance. For this reason, none of the simple empirical approaches are consistently applicable through the entire climatic range. In very cold catchments with significant and immobile internal storages of water, in snow and ice, additional parameters may be required in simple approaches or completely new models need to be developed.

[35] The processes and interactions controlling the annual water balance of catchments are many and varied and have been documented previously by authors such as *Milly* [1994] and *Zhang et al.* [2001]. The strength of purpose-derived empirical and theoretical relationships, such as *Fu* [1981], is to bypass the need to describe each of these in detail and solve for them simultaneously. Rather they encapsulate the net effect of these many competing forces within a range of the parameter space that describes catchments. In the case of snow- and ice-covered catchments, identified as not fitted well by the curves previously described, it may be possible to derive a different functional form, from first principles or trial and error, with similarly few parameter to represent their behavior.

#### 4.5. Stepwise Regression Analysis

[36] The relationship represented by equation (11) assumes that average annual precipitation and potential evapotranspiration are the dominant factors controlling mean annual evapotranspiration. This has been supported

by the results shown in Figure 9. An attempt was made here to investigate what other factors may significantly affect evapotranspiration predictions using stepwise regression. Only the detailed Australian data set was used in the analysis, as the required variables are not available for the other data sets. Variables were selected based on our conceptual understanding of their relationships to water balance. These variables include relief ratio ( $R_r$ ), defined as the ratio of the total catchment relief and a representative catchment length; percentage forest cover (PFC); plant available water storage capacity (PAWC); coefficient of variation in daily precipitation ( $CV_p$ ); seasonality index ( $SI$ ), average storm depth (ASD); and storm arrival rate (SAR) as defined by Milly [1994]. The stepwise regression was performed for evapotranspiration residual (e.g., difference between predicted and observed evapotranspiration) and each of the variables described above. The  $F$  tests were used to test hypotheses that individual variables were statistically significant at a level of 95%.

[37] Results of the stepwise regression are summarized in Table 3. The only variables that passed the  $F$  test are average storm depth, the plant available water capacity, and storm arrival rate. The other variables failed to contribute a significant new amount of information to mean annual evapotranspiration prediction. This may be because the variables selected are not independent from each other and there exist cross correlations among them, and as a result there are compensating effects. For example, forested catchments would be expected to have increased evapotranspiration over cleared ones, but we might expect to find natural forests mainly in steeper terrain unsuitable for clearing, where we would expect more runoff due to the steep slopes. Of the three variables that passed the 95%  $F$  test, only the plant available water capacity relates to vegetation and soil characteristics.

[38] Alternately, we examined the prospect for fitting the model parameter  $w$  from individual catchment characteristics, such as relief ratio, forest cover, and plant available water capacity. None of these proved to be significant at the 95% level, with least squares correlation coefficients all less than 0.05. As with Fu [1996], however, we found some correlation of  $w$  with average storm depth and coefficient of variation in daily precipitation. Use of  $w$  values estimated with this empirical relationship reduced standard error in predicted evapotranspiration by 16% compared with predictions made by using average  $w$  value.

## 5. Summary and Conclusions

[39] Mean annual evapotranspiration from a catchment is determined by climatic variables and characteristics of the catchment. The climatic variables include precipitation, solar radiation, temperature, wind speed, and humidity, while the catchment characteristics include vegetation cover, soil depth and permeability, and topography. These factors exhibit large spatial and temporal variability and interact with each other. A rational approach for estimating mean annual evapotranspiration is to integrate the net effect of these interacting variables and capture the key response of the processes involved.

[40] The model developed by Fu [1981] is an example of the rational approach, and it was based on phenomenolog-

ical considerations. It is assumed that the rate of the change in catchment evapotranspiration with respect to precipitation increases with residual potential evapotranspiration but decreases with precipitation. This assumption resembles the hypothesis of the complementary relationship between actual and potential evapotranspiration introduced by Bouchet [1963]. It can be argued that Fu's model was derived from first principles rather than a simple empirical fit to data and that it provides a better framework for estimating mean annual evapotranspiration compared with other similar empirical equations. The model compares well with observed data for over 470 catchments around the world with a mean absolute error of 54 mm and a correlation coefficient of 0.89. The parameter  $w$  can be considered to represent the integrated effects of catchment characteristics on evapotranspiration, and the maximum value for  $w$  was found to be 5.0 and the minimum value was 1.7. Any departure from the maximum curve is associated with a smaller value of  $w$ . Mean annual evapotranspiration responds nonlinearly to changes in  $w$ , and catchments with the index of dryness around 1.0 show the greatest sensitivity to  $w$ . The best fit value of  $w$  is 2.84 and 2.55 for the forested and grassed catchments, respectively, suggesting higher evapotranspiration from forested catchments for a given climatic condition. Although the  $w$  parameter is believed to be a function of catchment characteristics, it is very difficult to estimate values of  $w$  a priori.

[41] This study also evaluated the impact of precipitation, potential evapotranspiration, and the model parameter  $w$  on mean annual evapotranspiration. Precipitation was found to be the most dominant factor in determining evapotranspiration followed by potential evapotranspiration; other catchment characteristics appear to play a secondary role. This is confirmed by the stepwise regression analysis for the Australian catchments.

[42] The rational function approach developed by Fu [1981] only considers first-order factors and provides a simple method for evaluating long-term average relationships between precipitation and evapotranspiration. The use of this method can be considered as a useful first step in understanding the dynamic nature of catchment water balance, and it may lead to estimate of annual evapotranspiration and its intra-annual variability.

## Appendix A: Analytical Solutions of Mean Annual Evapotranspiration

[43] According to Fu [1981], mean annual evapotranspiration can be considered as solutions of the following partial differential equations:

$$\left. \begin{aligned} \frac{\partial E}{\partial P} &= f(x) \\ \frac{\partial E}{\partial E_0} &= \varphi(y) \\ \frac{\partial E}{\partial P} \Big|_{x=0} &= 0 \\ \frac{\partial E}{\partial E_0} \Big|_{y=0} &= 0 \end{aligned} \right\} \quad (A1)$$

where  $x = (E_0 - E)/P$  and  $y = (P - E)/E_0$ .

[44] A necessary condition for a general solution of (A1) is

$$\frac{\partial f(x)}{\partial E_0} + \frac{\partial f(x)}{\partial E} \varphi(y) = \frac{\partial \varphi(y)}{\partial P} + \frac{\partial \varphi(y)}{\partial E} f(x) \quad (\text{A2})$$

From equations (A1) and (A2), one can obtain

$$\left. \begin{aligned} \frac{\partial f(x)}{\partial E_0} &= \frac{\partial f(x)}{\partial x} \frac{\partial x}{\partial E_0} = \frac{1}{P} \left( 1 - \frac{\partial E}{\partial E_0} \right) \frac{\partial f(x)}{\partial x} = \frac{1}{P} [1 - \varphi(y)] \frac{\partial f(x)}{\partial x} \\ \frac{\partial f(x)}{\partial E} &= \frac{\partial f(x)}{\partial x} \frac{\partial x}{\partial E} = \frac{1}{P} \left( \frac{\partial E_0}{\partial E} - 1 \right) \frac{\partial f(x)}{\partial x} = \frac{1}{P} \left[ \frac{1}{\varphi(y)} - 1 \right] \frac{\partial f(x)}{\partial x} \\ \frac{\partial \varphi(y)}{\partial P} &= \frac{\partial \varphi(y)}{\partial y} \frac{\partial y}{\partial P} = \frac{1}{E_0} \left( 1 - \frac{\partial E}{\partial P} \right) \frac{\partial \varphi(y)}{\partial y} = \frac{1}{E_0} [1 - f(x)] \frac{\partial \varphi(y)}{\partial y} \\ \frac{\partial \varphi(y)}{\partial E} &= \frac{\partial \varphi(y)}{\partial y} \frac{\partial y}{\partial E} = \frac{1}{E_0} \left( \frac{\partial P}{\partial E} - 1 \right) \frac{\partial \varphi(y)}{\partial y} = \frac{1}{E_0} \left[ \frac{1}{f(x)} - 1 \right] \frac{\partial \varphi(y)}{\partial y} \end{aligned} \right\} \quad (\text{A3})$$

Substitution of equation (A3) into (A2) leads to following expressions:

$$[1 - \varphi(y)] \frac{\partial f(x)}{\partial x} = \frac{P}{E_0} [1 - f(x)] \frac{\partial \varphi(y)}{\partial y} \quad (\text{A4})$$

Since

$$\frac{P}{E_0} = \frac{\frac{E_0 + P - E}{E_0}}{\frac{E_0 + P - E}{P}} = \frac{1 + \frac{P - E}{E_0}}{1 + \frac{P - E}{E_0}} = \frac{1 + y}{1 + x} \quad (\text{A5})$$

Combining equations (A4) and (A5) yields

$$\frac{(1 + x)}{1 - f(x)} \frac{\partial f(x)}{\partial x} = \frac{(1 + y)}{1 - \varphi(y)} \frac{\partial \varphi(y)}{\partial y} \quad (\text{A6})$$

It can be noted that the left-hand side of (A6) is a function of  $x$ , while the right-hand side is a function  $y$ ; a necessary condition for (A6) to be valid is

$$\frac{(1 + x)}{1 - f(x)} \frac{\partial f(x)}{\partial x} = \alpha \quad (\text{A7})$$

$$\frac{(1 + y)}{1 - \varphi(y)} \frac{\partial \varphi(y)}{\partial y} = \alpha \quad (\text{A8})$$

where  $\alpha$  is a constant.

[45] Integration of equations (A7) and (A8) with boundary conditions in (A1) gives

$$f(x) = 1 - (1 + x)^{-\alpha} = 1 - \left( \frac{P}{E_0 + P - E} \right)^{\alpha} \quad (\text{A9})$$

$$\varphi(y) = 1 - (1 + y)^{-\alpha} = 1 - \left( \frac{E_0}{E_0 + P - E} \right)^{\alpha} \quad (\text{A10})$$

Substitution of equations (A9) and (A10) into (A1) yields

$$\frac{\partial E}{\partial P} = 1 - \left( \frac{P}{E_0 + P - E} \right)^{\alpha} \quad (\text{A11})$$

$$\frac{\partial E}{\partial E_0} = 1 - \left( \frac{E_0}{E_0 + P - E} \right)^{\alpha} \quad (\text{A12})$$

Let

$$u = \frac{E_0 + P - E}{P} \quad (\text{A13})$$

Equation (A11) can then be expressed as

$$\frac{d(u^{\alpha+1} - 1)}{u^{\alpha+1} - 1} = -(\alpha + 1) \frac{dP}{P} \quad (\text{A14})$$

Since  $E_0 \geq E$ ,  $u \geq 1$ , integration of (A13) yields

$$u = \frac{1}{P} (k + P^{\alpha+1})^{\frac{1}{\alpha+1}} \quad (\text{A15})$$

where  $k$  can be considered as a function of  $E_0$ .

[46] Substitution of equation (A13) into (A15) yields

$$E = E_0 + P - (k + P^{\alpha+1})^{\frac{1}{\alpha+1}} \quad (\text{A16})$$

Differentiation of equation (A16) with respect to  $E_0$  gives

$$\frac{\partial E}{\partial E_0} = 1 - \frac{1}{\alpha + 1} (k + P^{\alpha+1})^{-\frac{\alpha}{\alpha+1}} \frac{\partial k}{\partial E_0} \quad (\text{A17})$$

Substitution of equation (A17) into (A12) and considering (A17) yields

$$\frac{\partial k}{\partial E_0} = (1 + \alpha) E_0^{\alpha} \quad (\text{A18})$$

Integration of equation (A18) yields

$$k = E_0^{\alpha+1} + C \quad (\text{A19})$$

where  $C$  is an integration constant.

[47] Substitution of equation (A19) into (A16) gives

$$E = E_0 + P - (E_0^{\alpha+1} + P^{\alpha+1} + C)^{\frac{1}{\alpha+1}} \quad (\text{A20})$$

Since  $E \rightarrow 0$  when  $P \rightarrow 0$ , therefore constant  $C = 0$ .

[48] If we assume  $\alpha + 1 = w$ , the final solution of the mean annual evapotranspiration can be expressed as

$$\frac{E}{P} = 1 + \frac{E_0}{P} - \left[ 1 + \left( \frac{E_0}{P} \right)^w \right]^{1/w} \quad (\text{A21})$$

or

$$\frac{E}{E_0} = 1 + \frac{P}{E_0} - \left[ 1 + \left( \frac{P}{E_0} \right)^w \right]^{1/w} \quad (\text{A22})$$

[49] **Acknowledgments.** This study was supported by the Cooperative Research Centre for Catchment Hydrology under Project 2.3 “Predicting the effects of land-use changes on catchment water yield and stream salinity” and the Murray-Darling Basin Commission funded project “Integrated assessment of the effects of land use changes on water yield and salt loads” (D2013). The senior author would like to thank B. P. Fu for discussion on the model used in this study. We would also like to thank David Evans from D.L.P.E. Natural Resources for providing streamflow data for catchments in Alice Springs. We thank Tim McVicar and John Gallant for their helpful comments. Two anonymous reviewers gave very helpful reviews of the manuscript.

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