WATER RESOURCES RESEARCH, VOL. 37, NO. 3, PAGES 701–708, MARCH 2001

Response of mean annual evapotranspiration to vegetation changes at catchment scale

L. Zhang, W. R. Dawes, and G. R. Walker

Cooperative Research Centre for Catchment Hydrology and CSIRO Land and Water, Canberra, ACT, Australia

Abstract. It is now well established that forested catchments have higher evapotranspiration than grassed catchments. Thus land use management and rehabilitation strategies will have an impact on catchment water balance and hence water yield and groundwater recharge. The key controls on evapotranspiration are rainfall interception, net radiation, advection, turbulent transport, leaf area, and plant-available water capacity. The relative importance of these factors depends on climate, soil, and vegetation conditions. Results from over 250 catchments worldwide show that for a given forest cover, there is a good relationship between long-term average evapotranspiration and rainfall. From these observations and on the basis of previous theoretical work a simple two-parameter model was developed that relates mean annual evapotranspiration to rainfall, potential evapotranspiration, and plant-available water capacity. The mean absolute error between modeled and measured evapotranspiration was 42 mm or 6.0%; the least squares line through the origin had a slope of 1.00 and a correlation coefficient of 0.96. The model showed potential for a variety of applications including water yield modeling and recharge estimation. The model is a practical tool that can be readily used for assessing the long-term average effect of vegetation changes on catchment evapotranspiration and is scientifically justifiable.

1. Introduction

The massive land use change in Australia associated with agricultural development has caused an imbalance in catchment hydrological regime, leading to increased land and water salinization over large areas. It is estimated that each year the total cost of salinization to the nation is about $270 million including cost of lost production, damaged infrastructure, and degraded environmental assets (Prime Minister's Science, Engineering and Innovation Council, Dryland salinity and its impact on rural industries and the landscape, Canberra, ACT, Australia, 1999, available at http://www.dist.gov.au/science/pmseic/2ndmeeting.html). A number of land rehabilitation programs have been established by the commonwealth and state governments to control the degradation. According to Forest Plantations 2020 Vision, a major initiative of the commonwealth and state governments, the area of tree plantations by the year 2020 will treble [*Department of Primary Industries and Energy*, 1997], with part of this increase justified by environmental benefits. If the 2020 Vision is accurate, the plantation area in Australia will increase to over 3 million ha and will have significant impacts on catchment water yield and salinity. The impacts of such plantations on the trade-offs between economic viability, environmental sustainability, and water resource security will depend on the spatial distribution of the plantations. It is important to be able to predict the water balance–vegetation relationships at regional scales to determine these trade-offs. For the relationships to be useful, they must be dependent only on data that is generally available at those scales.

The research on the hydrological role of vegetation has extended over several decades [*Horton*, 1919; *Wicht*, 1941; *Penman*, 1963; *Bosch and Hewlett*, 1982; *Turner*, 1991]. Sources of information on the water balance associated with vegetation change generally fall into two categories. The first involves "paired-catchment" experimental techniques. *Hibbert* [1967] reviewed results from 39 paired experiments. *Bosch and Hewlett* [1982] updated Hibbert's review to include 55 additional catchments. Results from these experiments showed a large variation in catchment responses to changes in vegetation cover. However, a clear conclusion was that a reduction in forest cover increases water yield by decreasing evapotranspiration. The second source of information on the impact of vegetation comes from "single-catchment" water balance studies. These studies were not designed specifically to examine the effects of vegetation changes on water yield. The fact that they represent catchments with diverse climate, vegetation, and soil can provide useful information about the hydrological role of vegetation in catchment water balance. On the basis of these studies a few empirical relationships have been developed linking evapotranspiration to vegetation types for specific sites [e.g., *Holmes and Sinclair*, 1986; *Turner*, 1991]. The applicability of these empirical equations to other catchments needs to be evaluated.

A number of attempts have been made to calculate annual evapotranspiration using existing climatic data [*Brutsaert*, 1982]. *Budyko* [1958] postulated that long-term average annual evapotranspiration from a catchment is determined by rainfall and net radiation. The relationship proposed showed good agreement with the long-term water balance data for a number of catchments in the former USSR. *Pike* [1964] proposed an equation based on water balance data from Malawi that can explain interannual variation in evapotranspiration. Although the functional forms of the *Budyko* [1958] and *Pike* [1964] equations differ, their numerical values are similar [*Dooge*,

Paper number 2000WR900325.
0043-1397/01/2000WR900325$09.00

1992]. *Milly* [1994] developed a mathematical framework for mean annual evapotranspiration that provides a theoretical background for Budyko's equation. Milly's work recognized the importance of storage capacity of the root zone in controlling evapotranspiration and has the potential for assessing the catchment-scale response of vegetation changes. However, the practical application of this model is limited because of the complex numerical solutions required.

The purpose of this paper is to quantify the long-term impact of vegetation changes on mean annual evapotranspiration at catchment scales based on data and parameters that are easily measurable at a regional scale. This study uses a "top-down" approach that links the catchment response to our understanding of processes at finer scales. By reviewing and collating many water balance studies from around the world, we seek to develop generic relationships for assessing the impact of vegetation changes on evapotranspiration.

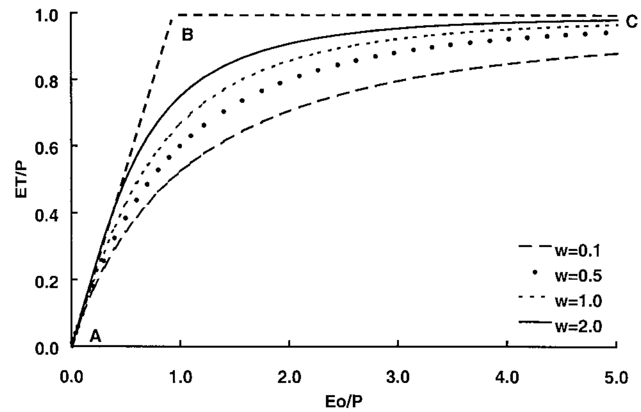


Figure 1. Ratio of mean annual evapotranspiration to rainfall as a function of the index of dryness (E_0/P) for different values of plant-available water coefficient (w).

2. General Framework

2.1. Catchment Water Balance

The concept of water balance provides a framework for studying the hydrological behavior of a catchment. It is useful for assessing how changes in catchment conditions can alter the partitioning of rainfall into different components. The water balance for a catchment can be written as

$$P = \mathrm{ET} + R + D + \Delta S, \tag{1}$$

where P is precipitation, ET is evapotranspiration, R is surface runoff measured as streamflow, D is recharge to groundwater, and ΔS is the change in soil water storage.

Precipitation is the largest term in the water balance equation, and it varies both temporally and spatially. For most of the hydrological applications it is appropriate to assume that precipitation is independent of vegetation type [*Calder*, 1998]. However, on a continental scale some studies using general circulation models suggest that vegetation types may affect precipitation [*Rowntree*, 1988; *Gash et al.*, 1994; *Xue*, 1997]. Evapotranspiration is the second or third largest term in the water balance equation, and it is closely linked with vegetation characteristics. In arid and semiarid regions, evapotranspiration is often nearly equal to precipitation, while in humid areas, it is limited by available energy. Surface runoff is also an important component of the water balance and is affected by the structure of vegetation and through rainfall interception and transpiration. On an annual basis, surface runoff will generally show good correlation with annual rainfall, particularly in areas where potential evaporation and rainfall are out of phase, such as winter dominant rainfall zones [*Budyko*, 1974]. Recharge is generally the smallest term in the water balance equation and is usually inferred from precipitation and evapotranspiration measurements. The last term in the water balance equation is the change in soil water storage. Over a long period of time (i.e., 5–10 years) it is reasonable to assume that changes in soil water storage are zero.

2.2. Climatic Effects on Annual Evapotranspiration

Catchment evapotranspiration is a complex process that is affected by rainfall interception, net radiation, advection, turbulent transport, canopy resistance, leaf area, and plant-available water [*McNaughton and Jarvis*, 1983; *Zhang et al.*, 1999]. Under dry conditions the principal controls on evapotranspiration are plant-available water and canopy resistance. Under wet conditions the dominant controls are advection, net radiation, leaf area, and turbulent transport. Under intermediate conditions the relative importance of these factors varies depending on climate, soil, and vegetation. The challenge in modeling catchment-scale evapotranspiration is to be able to represent these processes and factors in a simple fashion in order to use generally available data yet allow practical prediction of the effect of vegetation changes.

It is a common practice to combine these factors by considering their net effects. One way of approaching catchment evapotranspiration is to assume that evapotranspiration from land surfaces is controlled by water availability and atmospheric demand. The water availability can be approximated by precipitation; the atmospheric demand represents the maximum possible evapotranspiration and is often considered as potential evapotranspiration. Under very dry conditions, potential evapotranspiration exceeds precipitation, and actual evapotranspiration equals precipitation. Under very wet conditions, water availability exceeds potential evapotranspiration, and actual evapotranspiration will asymptotically approach the potential evapotranspiration. On the basis of these considerations, *Budyko* [1958] postulated that the following relationships are valid under very dry conditions:

$$R/P \rightarrow 0 \qquad \mathrm{ET}/P \rightarrow 1 \qquad R_n/P \rightarrow \infty, \tag{2}$$

where R is surface runoff, P is precipitation, ET is evapotranspiration, R_n is net radiation, and under very moist conditions,

$$\mathrm{ET} \rightarrow R_n \qquad R_n/P \rightarrow 0. \tag{3}$$

The dry and wet limits are represented by BC and AB in Figure 1, respectively. *Budyko* [1958] used net radiation (R_n) as a surrogate for potential evapotranspiration; we use potential evapotranspiration (E_0) calculated by the method of *Priestley and Taylor* [1972]:

$$E_0 = \alpha \frac{\Delta}{\Delta + \gamma} R_n, \tag{4}$$

where α is a constant equal to 1.28, Δ is the slope of the saturation vapor pressure curve, and γ is the psychrometric constant.

The dimensionless function (F) that satisfies conditions (2) and (3) must take the following form:

$$\mathrm{ET}/P = F(E_0/P). \tag{5}$$

This formulation is based principally on the influence of climatic conditions, and it assumes that the only effect of vegetation on evapotranspiration is through the influence of surface albedo on net radiation.

2.3. Vegetation Effects on Annual Evapotranspiration

Our belief is that other vegetation effects come primarily through the plant-available water capacity. *Nepstad et al.* [1994] estimated that half of the closed forests of Brazilian Amazonia depend on deep root systems to maintain green leaf areas and evapotranspiration during the dry seasons. In a similar study, *Hodnett et al.* [1995] also showed that during wet seasons evapotranspiration of a terra firma type forest was very similar to that of pasture (*Brachiaria decumbens*) in central Amazonia. The soil moisture under the two vegetation types showed little difference. However, in the dry seasons the forest sustained a higher evapotranspiration rate than the pasture, and the difference was attributed to the ability of the trees to access soil moisture from greater depth. *Calder* [1998] suggested that evapotranspiration in semiarid areas is limited principally by plant-available water, whereas in the wet uplands of the United Kingdom, evapotranspiration is limited principally by radiation and advection. These studies indicate that deep roots play an important hydrological role in plant systems, especially under dry conditions.

Rooting depth determines the soil volume from which plants are able to draw water, and together with soil hydraulic properties, it defines the plant-available water capacity. Trees generally have much larger available water capacity than herbaceous plants. During wet seasons, plants extract most water from shallow layers where the root density is the highest. As the soil progressively dries, more water is extracted from deeper layers to keep stomata open. As a result, trees are able to maintain a relatively constant evapotranspiration rate over time, even when soil moisture in the upper part of the soil is limited. Under such conditions, shallow-rooted plants tend to close their stomata and have a reduced evapotranspiration rate. In regions with dry climates, plant-available water capacity is expected to be a main reason for differences in annual evapotranspiration between trees and shallow-rooted plants.

The depth and distribution of plant roots is affected by a number of factors such as physical barriers, chemical barriers, and nutrient distribution. When soil physical properties such as porosity, pore sizes, strength, and root channels are unfavorable to water and oxygen supply, plant growth can be severely limited. *Tennant* [1976] showed that the available water for wheat in five different soils depended more on the rooting depth than it did on the soil hydraulic properties. *Canadell et al.* [1996] reviewed 290 studies around the world and showed that average maximum rooting depth was about 7 m for trees and 2.6 m for herbaceous plants. Such a difference in average maximum rooting depth will translate into a 540 mm difference in plant-available water for sandy soils and up to 3 times this amount for loamy and clayey soils. Therefore it is expected that rooting depth will contribute to differences in evapotranspiration between forests and herbaceous plants.

Greacen and Williams [1983] reported the plant-available water for some important Australian soils. For example, in a deep red earth under eucalypt woodland, the plant-available water was about 360 mm, although its water-holding capacity was relatively low (Figure 2). However, for a grey clay under

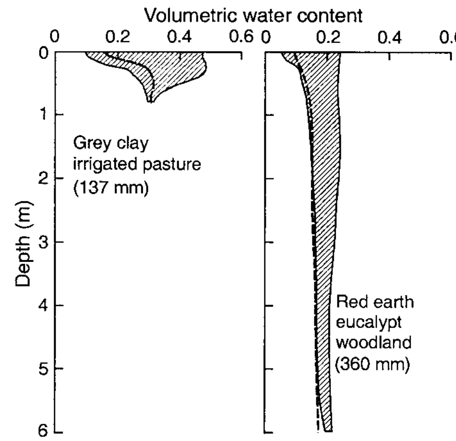


Figure 2. Typical soil moisture profiles and plant-available water capacity for two soil types under different plants (adapted from *Greacen and Williams* [1983]). Solid lines represent upper and lower limits of the soil water store. Numbers on Figure 2 refer to stored soil water.

irrigated pasture the profile was relatively shallow but with high water-holding capacity; the plant-available water was only 137 mm. As shown in Figure 2, deep-rooted plants (i.e., trees) generally have larger storage capacity than shallow-rooted plants (i.e., shortgrass and crops). The differences in both magnitude of the plant-available water and its profile water store will affect plant transpiration. It is clear that the plant-available water capacity is primarily responsible for greater evapotranspiration from forests than from pasture and crops [*Turner*, 1991; *Nepstad et al.*, 1994; *Hodnett et al.*, 1995].

2.4. Rational Function Approach

Two approaches can be used to formulate the relationship embedded in (1). The first is to use a process-based model dependent on a number of variables such as plant-available water content, seasonality of rainfall, soils hydraulic properties, etc. We refer to this as a "bottom-up" approach. An alternative is to use simple interpolators of an appropriate form between the two limits (2) and (3) based on observed data. The interpolators are, in turn, related to physical variables such as potential evapotranspiration and plant-available water content. We refer to this as a "top-down" approach.

The following simple rational function satisfies conditions (2) and (3):

$$\frac{\mathrm{ET}}{P} = \frac{1 + w\,\dfrac{E_0}{P}}{1 + w\,\dfrac{E_0}{P} + \left(\dfrac{E_0}{P}\right)^{-1}}, \tag{6}$$

where w is the plant-available water coefficient and it represents the relative difference in the way plants use soil water for transpiration. We interpret this as mainly owing to differences in root zone depth. The sensitivity of (6) to the plant-available water coefficient w is shown in Figure 1. The area to the left of the line AB defines a region where long-term average evapotranspiration would exceed long-term average rainfall; this is impossible in the dryland situations where rainfall is the only source of available water.

The effect of w on evapotranspiration is minimal under both very dry and very wet conditions (see Figure 1). The maximum

Table 1. Description of Different Relationships for Estimating Annual Evapotranspiration

Equation[a]	Reference
$ET = P[1 - \exp(-E_0/P)]$	*Schreiber* [1904]
$ET = P/[1 + (P/E_0)^2]^{0.5}$	*Pike* [1964]
$ET = \{P[1 - \exp(-E_0/P)]E_0 \tanh(P/E_0)\}^{0.5}$	*Budyko* [1974]

[a]ET is annual evapotranspiration (mm). P is annual rainfall (mm). E_0 is potential evaporation (mm).

difference in the ratio of evapotranspiration to rainfall between trees and herbaceous plants occurs when annual rainfall equals the atmospheric demand (i.e., $E_0/P = 1.0$). Under this condition the ability of trees to exploit a greater depth in soils allows them to use water that has been stored during the times they are least active, while shallower rooted herbaceous plants may allow that water to escape their root zone.

2.5. Comparison With Empirical Equations

A number of relationships have been developed based on the assumption that evapotranspiration is limited by available water (i.e., rainfall) under very dry conditions and available energy (i.e., potential evaporation) under very wet conditions (see Table 1). A comparison of these relationships with (6) is shown in Figure 3. It is clear that (6) is in good agreement with these empirical relationships. The plant-available water coefficient of 1.0 provided better agreement with these empirical relationships than smaller or larger values of w.

3. Testing and Generalization of the Model

3.1. Data Description

As stated in section 1, the data used in this paper 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 involved small catchments (<100 km^2), and the main objective was to detect changes in catchment water yield (i.e., precipitation minus evapotranspiration and recharge) after afforestation or deforestation. Detailed information on vegetation type and cover is available from these studies. The single-catchment water balance studies focused on relationships among rainfall, runoff, and evapotranspiration. These are generally large catchments with good quality rainfall and runoff data over a long period. However, information on vegetation type and cover is not complete. To draw some general conclusions about the impact of vegetation on catchment water balance from these studies, we selected catchments with the following characteristics: (1) Rainfall is the dominant form of precipitation. (2) Slopes of the catchments are gentle. (3) Soil depth is relatively thick (>2 m). Given that detailed information on vegetation is not available for all the catchments concerned, especially for large catchments, we will use the following terms to describe vegetation types: herbaceous plants, mixture of herbaceous plants and trees, and forest ($>70\%$ of canopy cover). Most of the catchments used in this study have long records of annual rainfall and streamflow data, from which we were able to obtain average annual evapotranspiration by assuming zero soil water storage change. In a few catchments, evapotranspiration was measured directly. The size of the catchments varied from less than 1 km^2 to 6×10^5 km^2. These catchments span a variety of climates including tropical, dry, and warm temperate. Mean annual rainfall in these catchments varied from 35 to 2980 mm, and the seasonal distribution varied.

The vegetation ranges from same-age plantation trees to native woodlands, open forests, rainforest, eucalyptus, various species of pine trees and conifers through to native and managed grassland and agricultural cropping. Soil descriptions were not routinely included in the reviewed papers. The sheer variation in geographical location and climatic regime in the data, however, is expected to cover most of the spectrum of soil types, from sand through loams to clays. The location of these catchments is shown in Figure 4, and details are given by *Zhang et al.* [1999].

3.2. Comparison With Observational Data

The relationship represented by (6) is determined by the plant-available water coefficient (w), and larger values of w tend to promote evapotranspiration. For the data listed by *Zhang et al*. [1999], it was found that (6) provides upper and lower limits with w equal to 2.0 and 0.1 (see Figure 5). On the basis of data shown in Figure 5, we assert that w varies between 0.5 and 2.0 for the range of plants. For forests the best fit value was 2.0, while for shortgrass and crops the best fit value was 0.5. It is expected that the value of w for bare soil will be less than 0.5 because the rate of soil evaporation becomes water-limited before plant transpiration does [e.g., *Zhang and Dawes*, 1998]. For bare soils the parameter w simply represents the relative water stored in the soil that can be directly evaporated.

Despite its semiempirical nature the functional form of (6) was found to be in good agreement with the data listed in Appendix A of *Zhang et al*. [1999] combined with E_0 estimated by *Priestley and Taylor* [1972]; all data are shown in Figure 5. The mean absolute error (MAE) in the ratio of evapotranspiration to rainfall (ET/P) between observation and (6) is 5%, and the root mean square error (RMSE) is 6%. In this comparison the plant-available water coefficient (w) was set to 2.0 for forest and 0.5 for pasture. For catchments with mixed vegetation we arbitrarily assigned a value of 1.0 because data were not available to accurately separate herbaceous and forest cover for individual catchments. The potential evapotranspiration (E_0) was calculated using the equation of *Priestley and Taylor* [1972] with average values of temperature and net radiation data. *Milly* [1994] developed a theoretical model that incorporates soil water storage, rainfall seasonality, and other factors. For a midlatitude location and assuming an exponen-

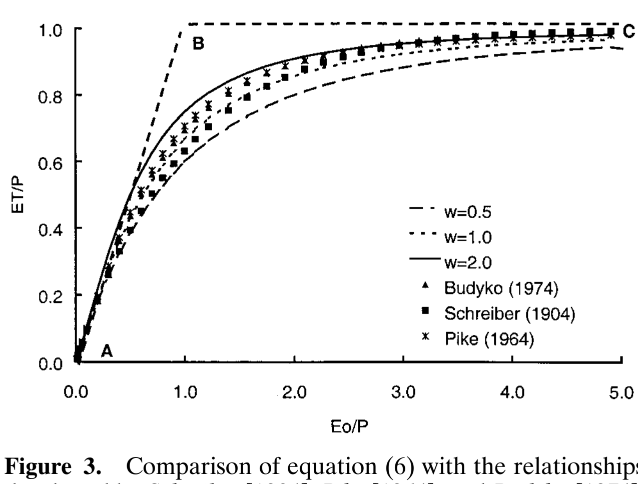


Figure 3. Comparison of equation (6) with the relationships developed by *Schreiber* [1904], *Pike* [1964], and *Budyko* [1974].

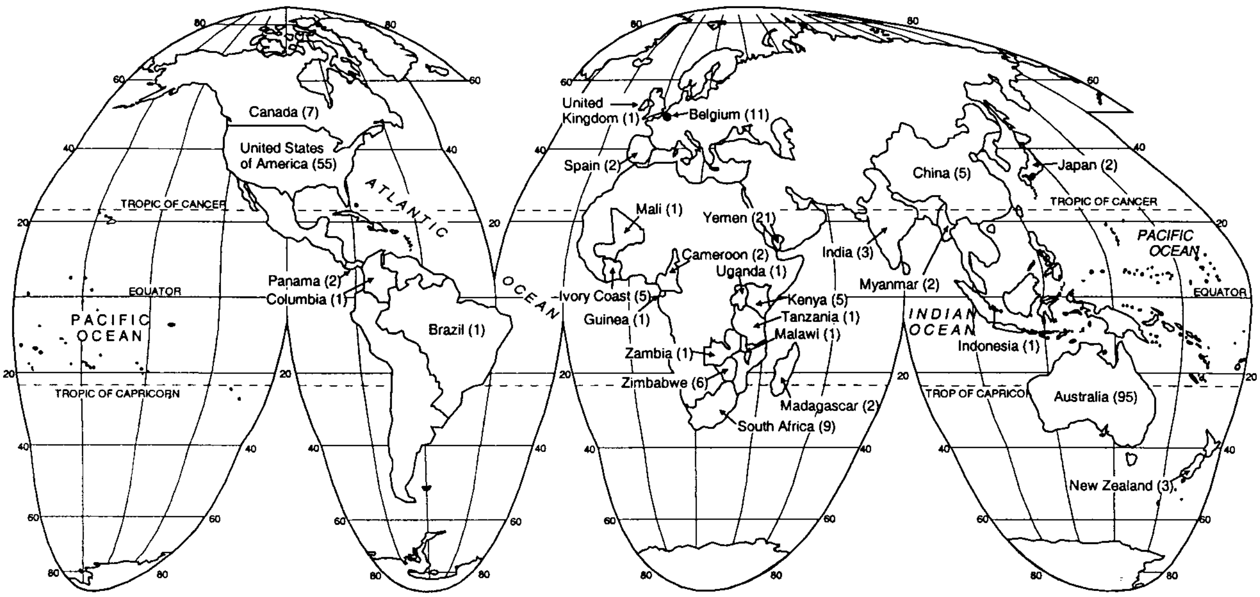


Figure 4. Location map of the catchments used in the study. Details are given by *Zhang et al.* [1999].

tial distribution of soil water storage, his model yielded similar results (Figure 5).

Equation (6) can be used to calculate actual long-term evapotranspiration when both rainfall and potential evapotranspiration are known. A comparison of observed and calculated evapotranspiration from (6) is shown in Figure 6. The MAE between the model estimates and measurements is 42 mm or 6.0%. The correlation coefficient is 0.96, and the best fit slope through the origin is 1.00.

3.3. Model Generalization

To extend the above method to catchments with varying proportions of forest and agricultural land use, a simple catchment scale model is proposed. Following *Eagleson* [1982], we assumed that annual evapotranspiration from a catchment is the sum of the annual evapotranspiration from herbaceous vegetation (including soil evaporation) and that from forest, weighted linearly according to their areas. The general equation can be expressed as

$$\mathrm{ET} = f\mathrm{ET}_f + (1 - f)\mathrm{ET}_h, \tag{7}$$

where ET is the total annual evapotranspiration in millimeters, f is the fractional forest cover, ET_f is the annual evapotranspiration from forests in millimeters, and ET_h is the annual evapotranspiration from herbaceous plants in millimeters.

As demonstrated earlier, (6) is a useful framework for esti-

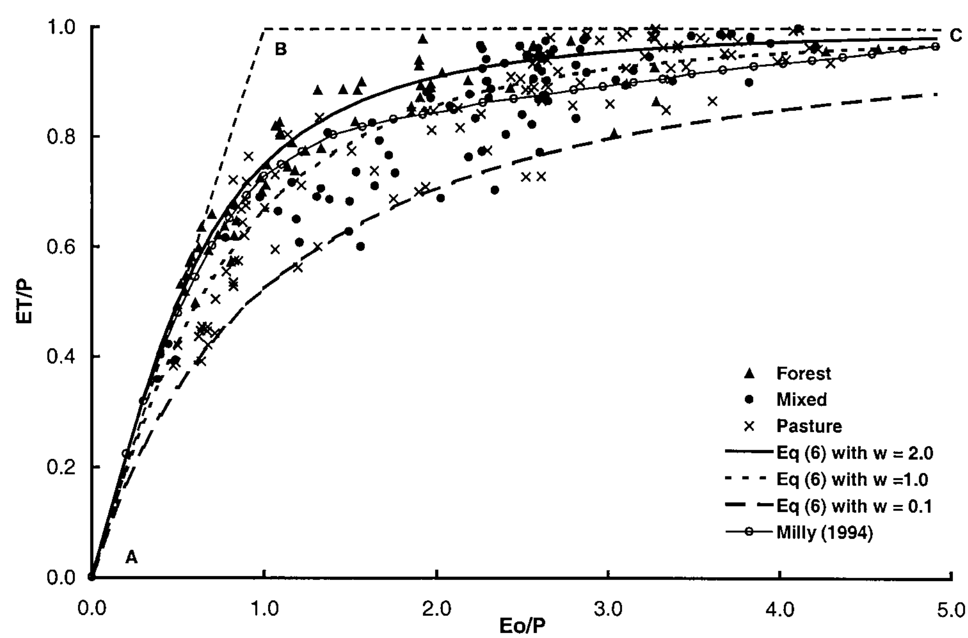


Figure 5. Comparison of equation (6) with measurements for catchments with different vegetation covers. Also shown is curve of *Milly* [1994].

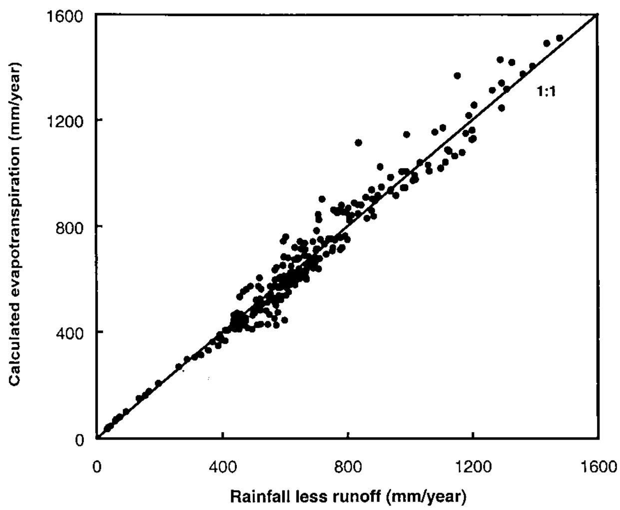


Figure 6. Scatterplot of the observed and calculated evapotranspiration using equation (6). The correlation coefficient between the model estimates and measurements is 0.96, and the slope of the best fit through the origin is 1.00.

mating annual evapotranspiration. However, it requires estimates of potential evaporation (E_0) and plant-available water coefficient (w) for each catchment. *Holmes and Sinclair* [1986] studied 103 catchments within the state of Victoria, Australia, with varying mixtures of grass and native eucalypt forest cover. They found clear differences between evapotranspiration rates for forested and grassland catchments along a rainfall gradient. *Turner* [1991] reported similar relationships based on a study of 68 catchments in California, United States of America.

Inspired by these workers, the parameters of (6) were established for forested and grassland catchments listed by *Zhang et al.* [1999] so that average evapotranspiration could be estimated from average annual rainfall. We assumed that E_0 in (6) is a constant (E_z), which was obtained by a least squares fit based on the data listed by *Zhang et al.* [1999]. Figure 7a shows the fitted function for trees, which has $r^2 = 0.93$, RMSE = 93 mm, $E_z = 1410$ mm, and $w = 2.0$. Figure 7b shows the fitted function for herbaceous plants, which has $r^2 = 0.90$, RMSE = 75 mm, $E_z = 1100$ mm, and $w = 0.5$. Thus the generalized form of (7) can be expressed as

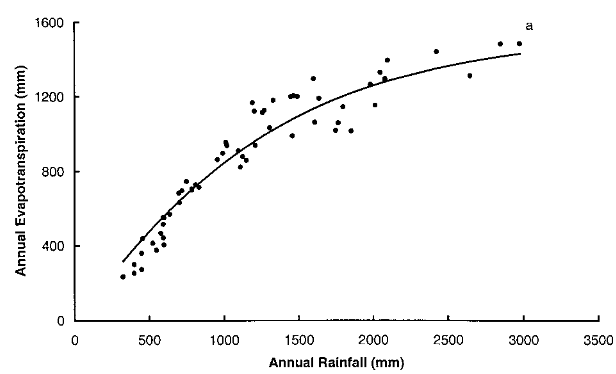
$$\mathrm{ET} = \left(f \frac{1 + 2\frac{1410}{P}}{1 + 2\frac{1410}{P} + \frac{P}{1410}} + (1 - f)\frac{1 + 0.5\frac{1100}{P}}{1 + 0.5\frac{1100}{P} + \frac{P}{1100}} \right) P. \tag{8}$$

A comparison of (8) with the curves described by *Holmes and Sinclair* [1986] and *Turner* [1991] is shown in Figure 8. Over the range 500 to 1500 mm of annual rainfall, curves are very similar. For higher annual rainfall, (8) tends to overestimate forest evapotranspiration compared to the curve of *Holmes and Sinclair* [1986]. As stated earlier, the data listed by *Zhang et al.* [1999] represent varying proportions of forest and grass or crop covers. A scatterplot of these data against (8) is shown in Figure 9. It is clear that most of the forested catchments plotted around the upper curve, and grassed catchments plotted around the lower curve with mixed vegetation catchments in the middle.

4. Discussion

In spite of the complexity of the soil-vegetation-atmosphere system the most important factors controlling mean annual evapotranspiration appear to be annual rainfall, potential evapotranspiration, and vegetation type. A simple model framework has been proposed for estimating long-term mean annual evapotranspiration based on rainfall, potential evaporation, and a plant-available water coefficient (6). The resulting relationship has the same general form as the relationship developed by *Milly* [1994]. Further, a generalized version for direct application has been developed where only annual rainfall and two vegetation-based constants are required for (8). The relationship suggests that long-term average annual evapotranspiration under the same climatic conditions is mainly determined by vegetation characteristics, and the difference may be attributed to the way different kinds of vegetation use soil water. The model is not designed for exploring interannual or intra-annual variability. However, it may be possible to incorporate these features into the model. *Milly* [1994] showed that the spatial distribution of soil water storage capacity and temporal rainfall pattern can affect catchment evapotranspiration, but on a long-term average basis these effects appear to be secondary.

It should be noted that in the formulation of the relationship for long-term average annual evapotranspiration, *Budyko* [1974] used net radiation instead of potential evapotranspiration. The use of net radiation in this context is to represent maximum evapotranspiration that would occur under given climatic conditions. He examined different methods for calculating potential evapotranspiration and concluded that the best way to calculate potential evapotranspiration is to use a combination equation (e.g., Penman's equation). *Budyko* [1974]



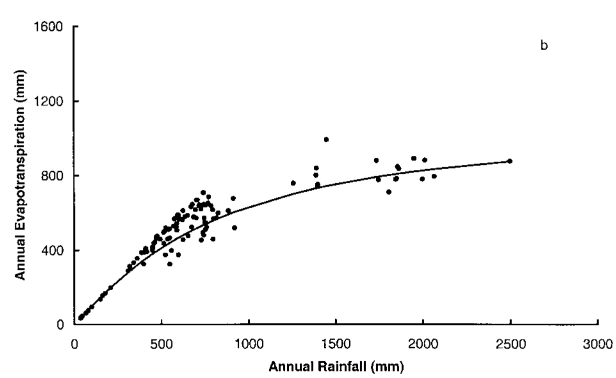


Figure 7. Scatterplots of the least squares fit for (a) forested and (b) herbaceous plant catchments.

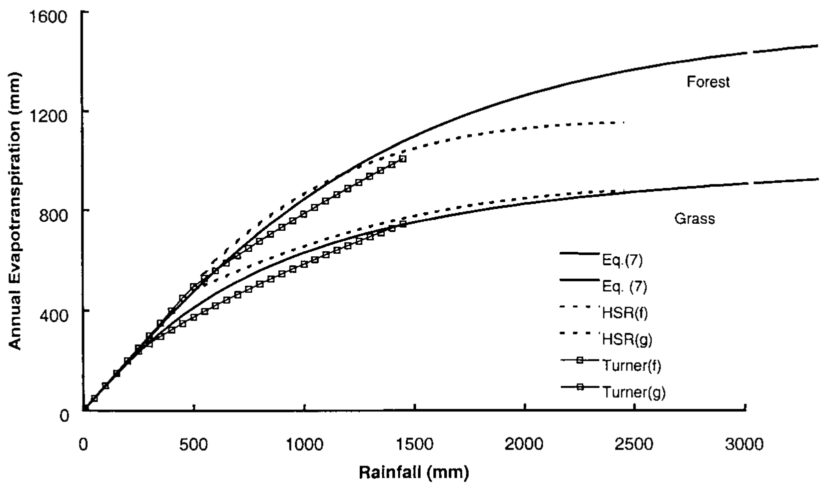


Figure 8. Comparison of equation (8) with the empirical relationships developed by *Holmes and Sinclair* [1986] and *Turner* [1991] for forested and grassed catchments.

argued that on an annual basis, the upper limit of evapotranspiration is equal to net radiation because the annual sums of the sensible heat flux cannot provide significant input of energy to the land surface; hence the latent heat flux must be provided by net radiation. In this study, potential evapotranspiration was calculated using the method of *Priestley and Taylor* [1972] with average values of temperature and net radiation data. It is likely that estimation of potential evapotranspiration using other methods will yield different results. However, evaluation of these methods for estimating potential evapotranspiration is outside the scope of this study.

For the generalized model (8) both the potential evaporation and plant-available water coefficient are set to constant values for trees and grass. A constant E_z in (8) is used to simplify the model and cannot be interpreted as potential evaporation in the traditional sense because actual evapotranspiration can exceed E_z as rainfall increases. The value of E_z is 1410 mm for forests and 1100 mm for shortgrass. Differences in albedo and aerodynamic resistance between these vegetation types may explain these differences in E_z.

Annual evapotranspiration is generally greater for forested than for nonforested catchments. The difference is larger in high-rainfall areas and diminishes in areas with annual rainfall less than 500 mm. From Figure 9 it is clear that catchments with mixed cover have annual evapotranspiration between that observed for fully forested and fully cleared catchments. Therefore we can use the two curves as an envelope; that is, the response of the mixed catchment must lie between these two curves. It was assumed that mean annual evapotranspiration is a linear function of tree cover [*Liu and Zhong*, 1978; *Vertessy and Bessard*, 1999], and this may introduce errors in catchments with mixed cover type in high-rainfall zones. However, *Sahin and Hall* [1996] argue that the effect of tree cover is likely to be a nonlinear function, and thresholds exist below which no changes in evapotranspiration could be observed. It is clear from Figure 5 that a single curve cannot explain all of the

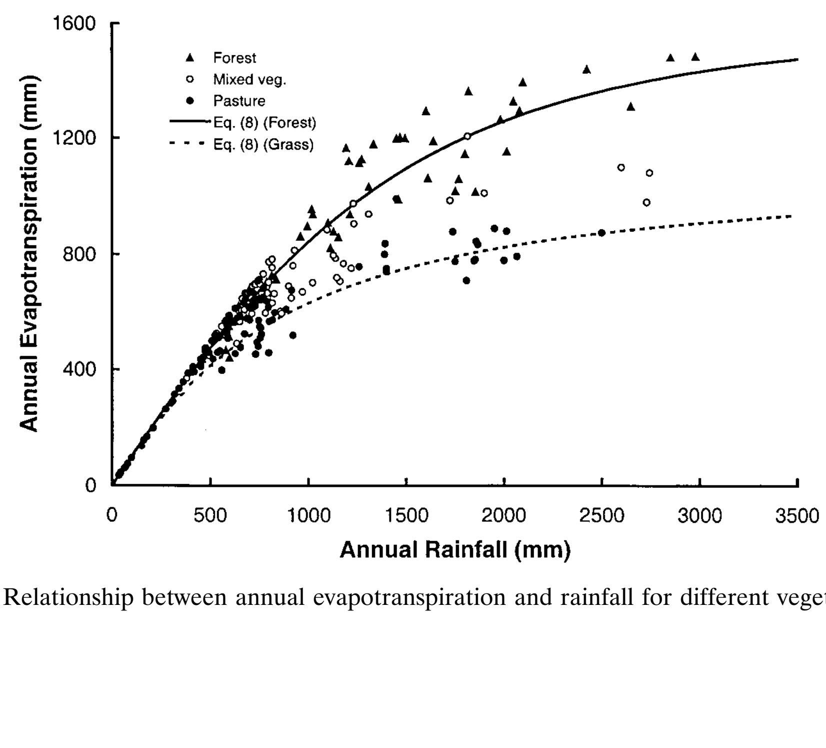


Figure 9. Relationship between annual evapotranspiration and rainfall for different vegetation types.

variability among the data. The uncertainties associated with rainfall, potential evapotranspiration estimates, and estimates of fractional vegetation cover must contribute to the scatter.

5. Conclusions

A two-parameter model has been developed to estimate mean annual evapotranspiration at catchment scales. The model is based on, and constrained by, observations, and the relationship should be both robust and scientifically justifiable. The model has advantages over more traditional process-based models, requiring data generally available at regional scales and being easy to apply either to an individual catchment or in a spatial modeling framework. The model is consistent with previous theoretical work and shows good agreement with over 250 catchment-scale measurements from around the world. The generalized model (8) provides a catchment approach for estimating the order of magnitude of the changes in mean annual evapotranspiration that result from changes in catchment vegetation.

The model is a practical tool that can be readily used to predict the long-term consequences of reforestation and has potential uses in catchment-scale studies of land use change. Using fixed parameters rather than allowing them to vary by catchment reduces the data requirements and facilitates automated implementation of the model within geographic information systems and other frameworks [e.g., *Zhang et al.*, 1997; *Vertessy and Bessard*, 1999]. The approach presented seeks to establish long-term average relationships for annual evapotranspiration, and it can be considered as a preliminary step toward the study of climate-soil-vegetation dynamics.

Acknowledgments. This study was partially supported by an ACIAR grant "Regional Water and Soil Assessment for Managing Sustainable Agriculture" and an NRMS-LWWRDC grant "Improving Dryland Salinity Management through Integrated Catchment Scale Management" (NRMS grant D6026). We thank Bryson Bates and Richard Silberstein for their valuable comments on a draft of this paper. We also acknowledge two reviewers, including Edwin Weeks, for their helpful comments.

References

Bosch, J. M., and J. D. Hewlett, A review of catchment experiments to determine the effect of vegetation changes on water yield and evapotranspiration, *J. Hydrol.*, *55*, 3–23, 1982.

Brutsaert, W. F., *Evaporation Into the Atmosphere*, 299 pp., D. Reidel, Norwell, Mass., 1982.

Budyko, M. I., *The Heat Balance of the Earth's Surface*, translated from Russian by N. A. Stepanova, 259 pp., U.S. Dep. of Commer., Washington, D. C., 1958.

Budyko, M. I., *Climate and Life*, 508 pp., Academic, San Diego, Calif., 1974.

Calder, I. R., Water-resource and land-use issues, *SWIM Pap. 3*, Int. Water Manage. Inst., Colombo, Sri Lanka, 1998.

Canadell, J., R. B. Jackson, J. R. Ehleringer, H. A. Mooney, O. E. Sala, and E. D. Schulze, Maximum rooting depth of vegetation types at the global scale, *Oecologia*, *108*, 583–595, 1996.

Department of Primary Industries and Energy, The 2020 Vision Statement, report, Canberra, ACT, 1997.

Dooge, J. C. I., Sensitivity of runoff to climate change: A Hortonian approach, *Bull. Am. Meteorol. Soc.*, *73*, 2013–2024, 1992.

Eagleson, P. S., Ecological optimality in water-limited natural soil-vegetation systems, 1, Theory and hypothesis, *Water Resour. Res.*, *18*, 325–340, 1982.

Gash, J. H. C., C. A. Nobre, J. M. Roberts, and R. L. Victoria (Eds.), *Amazonian Rainforest and Climate*, 576 pp., John Wiley, New York, 1994.

Greacen, E. L., and J. Williams, Physical properties and water relations, in *Soil—An Australian Viewpoint*, pp. 499–530, Academic, San Diego, Calif., 1983.

Hibbert, A. R., Forest treatment effects on water yield, in *Forest Hydrology*, edited by W. E. Sopper and H. W. Lull, 813 pp., Pergamon, Tarrytown, N. Y., 1967.

Hodnett, M. G., L. Pimentel da Silva, H. R. da Rocha, and R. Cruz Senna, Seasonal soil water storage changes beneath central Amazonian rainforest and pasture, *J. Hydrol.*, *170*, 233–254, 1995.

Holmes, J. W., and J. A. Sinclair, Water yield from some afforested catchments in Victoria, Paper presented at Hydrology and Water Resources Symposium, Inst. of Eng., Brisbane, Queensland, Australia, Nov. 25–27, 1986.

Horton, R. E., Rainfall interception, *Mon. Weather Rev.*, *47*, 603–623, 1919.

Liu, C. M., and J. R. Zhong, The effect of forest on stream flow in the Loess Plateau, Acta *Geogr. Sinica*, *33*, 112–126, 1978.

McNaughton, K. G., and P. G. Jarvis, Predicting the effects of vegetation changes on transpiration and evaporation, in *Water Deficits and Plant Growth*, vol. VII, edited by T. T. Kozowski, pp. 1–47, Academic, San Diego, Calif., 1983.

Milly, P. C. D., Climate, soil water storage, and the average annual water balance, *Water Resour. Res.*, *30*, 2143–2156, 1994.

Nepstad, D. C., C. R. de Carvalho, E. A. Davidson, P. H. Jipp, P. A. Lefebvre, G. H. Negreiros, E. D. de Silva, T. A. Stone, S. E. Trumbore, and S. Vieira, The role of deep roots in the hydrological and carbon cycles of Amazonian forest and pasture, *Nature*, *372*, 666–669, 1994.

Penman, H. L., Vegetation and hydrology, *Tech. Commun. 53*, 124 pp., Commonw. Bur. of Soils, Harpenden, England, U. K., 1963.

Pike, J. G., The estimation of annual runoff from meteorological data in a tropical climate, *J. Hydrol.*, *2*, 116–123, 1964.

Priestley, C. H. B., and R. J. Taylor, On the assessment of the surface heat flux and evaporation using large-scale parameters, *Mon. Weather Rev.*, *100*, 81–92, 1972.

Rowntree, P. R., Review of general circulation models as a basis for predicting the effects of vegetation changes on climate, in *Forests, Climate and Hydrology: Regional Impacts*, edited by E. R. C. Reynolds and F. B. Thompson, pp. 162–192, Kefford, Singapore, 1988.

Sahin, V., and M. J. Hall, The effects of afforestation and deforestation on water yields, *J. Hydrol.*, *178*, 293–309, 1996.

Schreiber, P., Über die Beziehungen zwischen dem Niederschlag und der Wasserführung der Flüße in Mitteleuropa, *Z. Meteorol.*, *21*(10), 441–452, 1904.

Tennant, D., Wheat crop penetration and total available water on a range of soil types, *Aust. J. Exp. Agric. Anim. Husb.*, *16*, 570–577, 1976.

Turner, K. M., Annual evapotranspiration of native vegetation in a Mediterranean-type climate, *Water Resour. Bull.*, *27*, 1–6, 1991.

Vertessy, R. A., and Y. Bessard, Anticipating the negative hydrologic effects of plantation expansion: Results from a GIS-based analysis on the Murrumbidgee Basin, *Rep. 99/6*, Coop. Res. Cent. for Catch. Hydrol., Canberra, ACT, 1999.

Wicht, C. L., Diurnal fluctuation in Jonkershoek streams due to evaporation and transpiration, *J. S. Afr. For. Assoc.*, *7*, 34–49, 1941.

Xue, Y., Biosphere feedback on regional climate in tropical north Africa, *Q. J. R. Meteorol. Soc.*, *123*, 1483–1515, 1997.

Zhang, L., and W. R. Dawes, WAVES—An integrated energy and water balance model, *Tech. Rep. 31/98*, CSIRO Land and Water, Canberra, ACT, 1998.

Zhang, L., M. Stauffacher, G. R. Walker, and P. A. Dyce, Recharge estimation in the Liverpool Plains (NSW) for input to groundwater models, *Tech. Rep. 10/97*, CSIRO Land and Water, Canberra, ACT, 1997.

Zhang, L., W. R. Dawes, and G. R. Walker, Predicting the effect of vegetation changes on catchment average water balance, *Tech. Rep. 99/12*, Coop. Res. Cent. for Catch. Hydrol., Canberra, ACT, 1999.

W. R. Dawes, G. R. Walker, and L. Zhang, CSIRO Land and Water, Canberra Laboratory, GPO Box 1666, Canberra, ACT 2601, Australia. (dawes@cbr.clw.csiro.au; walker@cbr.clw.csiro.au; lu.zhang@cbr.clw.csiro.au)

(Received April 28, 2000; revised September 25, 2000; accepted October 13, 2000.)