Potential Evapotranspiration Methods and their Impact on the Assessment of River Basin Runoff Under Climate Change

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Abstract

The current trend in increased amounts of green house gases in the atmosphere will likely affect both precipitation and evapotranspiration, which will in turn affect the runoff response of river basins. These impacts on river basin discharge are discussed in the context of changes in evapotranspiration estimates which are found by coupling a monthly water balance model to account for changes in soil moisture and micrometerological and empirical estimates of potential evapotranspiration. The purpose is to assess the importance of the varying methods for estimating potential evapotranspiration on climate impact assessments of river basin discharge. Four river basins of different size and hydro-climatic variability were selected as case studies.

Introduction

With evidence indicating that atmospheric concentrations of green house gases are increasing, there is growing concern that these changes will have significant impacts on water within the hydrologic cycle in many regions of the world (Skiles and Hanson, 1994). The magnitude of this concern has been expressed by the interagency Committee on Earth and Environmental Sciences (CEES) in the United States which identified the hydrologic cycle as the highest scientific priority for global change research (Rind, et. al., 1992). Concurrently, hydrologists have been busy investigating the response of river basins to possible climatic variability. Naturally hydrologists, primarily concerned with water availability, focus on the hydrologic response of the physical basin and not on the atmospheric components of the evapotranspiration processes that are driving these changes. Often studies of this nature apply a unique hydrologic model of a river catchment and then alter temperature and precipitation to assess basin response (Reibsame, et. al 1994, Gleick 1987, etc.). Yet Rind et. al (1992) point out that different formulae of physical processes as well as different conceptualizations of hydrologic components will likely respond differently under climate change scenarios. This idea motivates the need to look at the myriad of approaches that have been and continue to be used in describing the response of river basins to hydrologic processes. Two main modeling components have been identified in regards to the water balance of a river basin: the modeling of atmospheric processes to remove water vapor from the land surface and the modeling of water transport processes within the soil domain. This paper deals primarily with the first issue, while a companion paper (Yates and Strzepek, 1994) addresses the second.

A key component in the hydrologic cycle (Figure 1) is evapotranspiration (E_V): the conversion of liquid water at the earth-atmosphere boundary to vapor and the subsequent mixing of this vapor with the atmosphere (Hagan, et. al, 1967). The evaluation of evapotranspiration is important in the study of the impact of climate change on water resources, as evapotranspiration can be considered a key "link" between the atmosphere and the soil matrix within the hydrologic cycle (Figure 1). The importance of this link has been observed by Dooge (1992) who states that any estimate of climate change impacts on water resources depends on the ability to relate changes in actual evapotranspiration to predicted changes in precipitation and potential evapotranspiration (E_p). To predict proper *changes* in evapotranspiration it is obviously important to begin with good *estimates* of the driving mechanisms of that change.

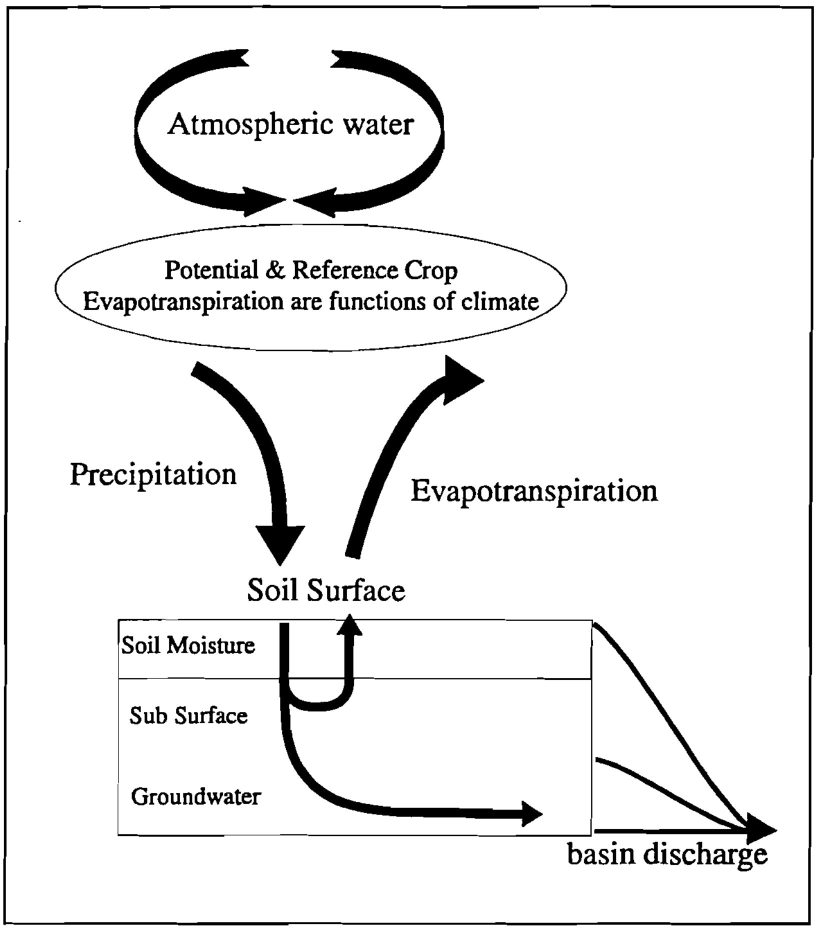


Figure 1. Simplified water balance showing potential and reference crop evapotranspiration (E_p & E_{rc}) as climate variables and actual evapotranspiration (E_v) dependent on soil moisture, plant canopy, and Ep or Erc). E_v = f(available moisture, E_p E_{rc}) and Ep,Erc= f(Climate)

Potential Evapotranspiration: An Imaginary Construct

Runoff from a river basin can be regarded as the by-product (dependent) of two larger processes: precipitation and evapotranspiration (independents). The terms potential and/or reference crop evapotranspiration (E_p and E_{rc} respectively) have often been used as a hypothetical measurement of a climates capacity for the removal of water vapor from the land surface. Ep and Erc can be regarded as "imaginary" concepts that have been used in the estimation of actual evapotranspiration, whose relationship is depicted in Figure 2. Hydrologists, agronomists, climatologists, and agricultural engineers often use the concepts of E_p and E_{rc} as the climatological link between the dynamic processes of vapor diffusion from soil (evaporation) and crop surfaces (transpiration) to the atmosphere. It appears however, that the concepts of actual, potential and reference crop evapotranspiration can be confused because they are often used interchangeably within the literature (Shuttleworth, 1993; Mimikou, et. al. 1991). For this reason, a brief set of definitions are given below (taken from Shuttleworth, 1993).

Evaporation (E): rate of liquid water transformation to vapor from open water, bare soil or vegetation with soil beneath.

Transpiration: part of the total evaporation which enters the atmosphere from the soil through the plants

Evapotranspiration (E_v): combination of evaporation from bare soils and transpiration from plants.

Potential Evapotranspiration (E_p): The idealized quantity of water evaporated per unit area, per unit time from an idealized, extensive free water surface under existing atmospheric conditions.

Reference Crop Evapotranspiration (E_{rc}): rate of evaporation form an idealized grass crop with a fixed crop height of 0.12*m*, an albedo of 0.23 and a surface resistance of 69 sm^{-1}

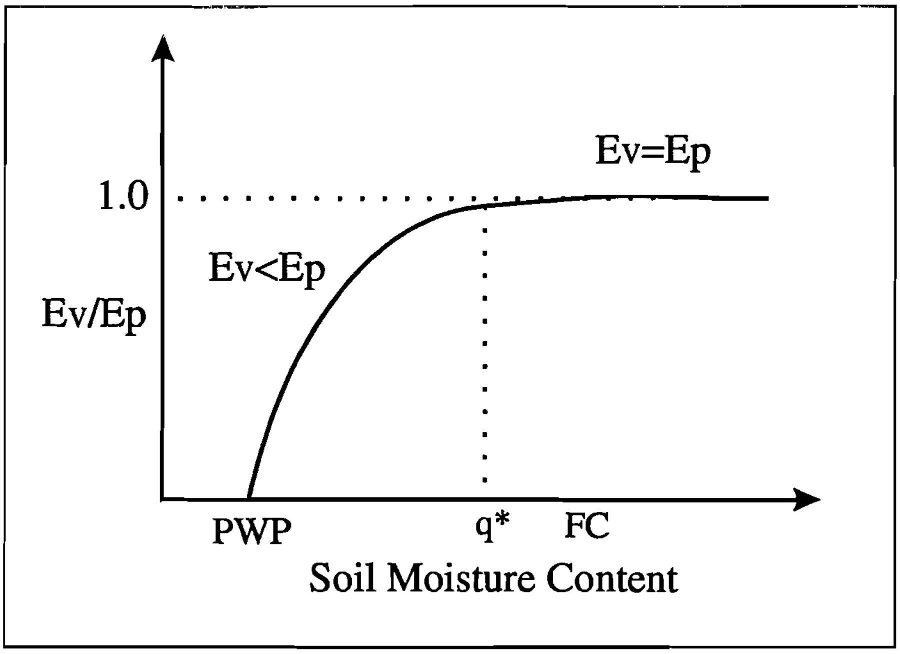


Figure 2. The ratio of E_v to E_p is shown, in an idealized form. It is assumed that E_v achieves the maximum level of evapotranspiration E_p when the soil moisture content achieves the value q* which is smaller than the Field Capacity of the soil (FC). A simpler approximation assumes a linear ratio of E_v/E_p between the Permanent Wilting Point (which is a function of the soil and crop) and the Field Capacity, which is only a function of the type of soil.

It is worthwhile to highlight the difference between potential and reference crop evapotranspiration. Potential evapotranspiration can be idealized as *potential evaporation* using the strict definition, since it uses the the term "extensive free surface" which has nothing to do with the transpiration process of plants. Reference crop evapotranspiration can be thought of as a *potential evapotranspiration* because it incorporates the transpiration process by assuming an idealized plant. Both potential and reference crop evapotranspiration are "potential" estimators of evapotranspiration, but apply different assumptions. For the remainer of this paper, potential and reference crop evapotranspiration will be referred to as E_p when discussing the methods in general.

A number of approaches have been used to assess evapotranspiration and runoff changes within the context of an altered climate.. Gleick (1987) used the Thornthwaite method for estimating E_p and a monthly water balance model in the Sacramento-San Joaquin Basin as did McCabe and Ayers (1989) in the Delaware River Basin. Nemec and Shaake (1982) performed an impact assessment on two basins in the U.S. (one humid and one arid). They used the Budyko radiation method for the computation of E_p combined with a daily hydrologic model to estimate evapotranspiration changes and basin discharge. Reibsame et. al (1994) used a mass balance approach to estimate evapotranspiration (eliminating the need for E_p estimates) for the Zambezi River Basin, while in the same work the Penman method was used to calculate E_p for the Uruguay basin. Mimikou et. al (1991) used the Blaney-Criddle method for estimating potential evapotranspiration in combination with a monthly water balance for three basins in the central mountainous regions in Greece. Schaake (1990) applied uniform changes in potential evapotranspiration of +-10% to river basins in the Eastern US, suggesting that this change was primarily caused by increased temperatures. Rind et. al. (1990) report the use of the Thornthwaite method for computing E_p within the Palmer Drought Severity Index (PDSI) model of the U.S., which was then used to create drought scenarios under climate change.

Methods for Estimating Actual, Potential, and Reference Crop Evapotranspiration

Methods for evaluating evapotranspiration can be divided into three categories: hydrologic or water balance methods, analytical methods based on climate variables, and empirical estimates. The water balance method to determine E_v consists of catchment hydrology, soil water depletion sampling, and lysimetry testing. Because this is primarily a physically based approach, its use in climate change assessment is limited to the laboratory. The second approach, referred to as a micrometeorological method, uses the scientific understanding of the physics of evaporation and transpiration. Mathematical relationships are developed that describe these processes via two key climatological components: energy balance and mass transport. The third method centers around the development of empirical relationships that are often site specific and based on local climatological conditions and often cast in regression analysis. These methods are often "calibrated" by relating the empirical estimates to observed measurements (Hagan, et. al. 1967).

The Micrometerological Approach to evapotranspiration

To estimate evapotranspiration it is necessary to define the components that drive the movement of water away from the earth surface. A brief introduction to the subject of evaporation methods and concepts is presented here (for further discussion see Chow, et. al. 1988 or Shuttelworth, 1993). Evaporation from an open surface is a dynamic process, whose rate depends on three main factors (Chow et al., 1988):

- Ability to provide the *energy* required for evaporation, latent heat, since evaporation absorbs heat from its environment.
- Ability to remove water vapor away from the evaporating surface via *aerodynamic* processes, i.e. wind and the humidity gradient in the air above the evaporating surface.
- Availability of soil moisture to be evaporated when the supply of water is limiting.

Energy Processes in Evaporation (E_r)

In the energy balance method it is assumed that the ability of the system to remove moist air is not limiting to the evaporation process. The energy balance is given by the following equation:

$$\frac{dH}{dt} = R_n - H_s - G \tag{1}$$

where:

dH/dt = rate of change of storage of energy in the water body
R_n = net radiation
H_s = sensible heat lost to air
G = sensible heat lost to ground

and after some manipulations the evaporation estimate E_r may be obtained as:

$$E_r = \frac{1}{\lambda \rho_w}\left[R_n - H_s - G\right] \tag{2}$$

where:

λ = latent heat of vaporization (amount of heat needed, in calories or in MJ K^{-1}, to evaporate 1 gram of water at 1 atmosphere pressure to vapor » 539 cal/gm)

ρ_w = density of water [kg/m^3]

Aerodynamic Processes in Evaporation (E_a)

The second factor that governs evaporation is the ability to transport vapor away from the free surface, i.e.:

- humidity gradient in the air over the surface.
- wind speed across the surface.

Basic equation: *flux = constant * gradient*

and the flux is either momentum or vapor flux. An evaporation estimate which assumes an unlimited availability of energy is the following:

$$E_a = B(e_{as} - e_a) \tag{3}$$

where:

$$B = \frac{0.622k^2 \rho_a u_2}{p\rho_\omega \left[ln\left({z_2}/{z_o} \right) \right]^2} \tag{4}$$

and:

B = Bowen ratio
k = von Karman's constant = 0.4
ρ_a = air density
u_2 = mean wind velocity measured at elevation z_2
p = atmospheric air pressure
ρ_w = water density
z_2 = elevation at which wind measurements are made
z_0 = roughness height

Combining Energy and Aerodynamic Process - Penman (E)

In the above methods is was assumed that either the energy available or the ability to remove saturated air were not limiting when in reality both are in most cases. This gives origin to the combined methods which give a weighted average of the two estimates. The Penman equation is the most widely known combined method of estimating evaporation.

General form:

$$E = \frac{\Delta}{\Delta+\gamma} E_r + \frac{\gamma}{\Delta+\gamma} E_a \tag{5}$$

where:

E = combined evaporation estimate [mm/day]
Δ = slope of the saturated vapor pressure curve
γ = psychometric constant $= C_p\, p\, K_h / (0.622\, \lambda\, K_w)$
where, Cp = specific heat at constant temperature
Kh, Kw = diffusivity [L^2/t]

Evapotranspiration (E_v): Penman-Montieth

The Penman equation has been modified to evaluate actual evapotranspiration (E_v) in the Penman-Monteith equation (Shuttelworth, 1993). The Penman-Monteith equation is regarded as one of the most accurate equations to estimate evapotranspiration and performed very well in a comparison study by Jensen et al (1990). This is an estimate of actual evapotranspiration because of the surface resistance term, r_s. Surface resistance in the Penman Montieth equation is a land cover resistance term which is a measure of available water from the soil surface and the plant canopy. Rosenburg et. al (1989) used this equation to simulate the evapotranspiration response of an irrigated alfafa field under several climate change scenarios. However, this form of the Penman equation is difficult to use in water resources because it does not explicity consider the affects of soil moisture on E_v.

$$E_v = \frac{1}{\lambda}\left[\frac{\Delta A + \rho_a C_p (e_s - e)/r_a}{\Delta + \gamma(1 + r_s/r_a)}\right] \tag{6}$$

where:

A = available energy [MJ m^{-2} day^{-1}] $\cong R_n - G$
$e_s - e$ = vapor pressure deficit [kPa] =D
r_a = aerodynamic resistance [s/m]

r_s = surface resistance [s/m]

At a minimum, the following meteorological data is required:

- Solar Radiation, R_n [ly/day].
- Air Temperature, T_a [°C].
- Saturated Vapor Pressure/Actual Vapor Pressure, e_s and e .
- Wind Speed, u [m/day] [km/day].

The following data may be used to estimate the net radiation (equation 8)

- Specific Humidity, q_v [dimensionless], or Relative Humidity, R_h (%), or Dew-point Temperature, T_d [°C].
- Sunshine Hours , n [hrs] or Cloud Cover f.

Assumptions:

- Steady-State Energy flow prevails (no diurnal cycle, Dt > 1 day)
- Changes in heat storage are not significant (i.e. not adequate for a lake)

All other terms have been defined above.

There are a large number of models and equations that compute evaporation and transpiration as a function of climatological and hydrological data. Unfortunately the availability of this climatological and hydrologic data is not homogeneous in all parts of the world and simpler techniques have to be used to overcome these limitations. Furthermore, scenarios for climate change often provide only information on temperature and precipitation changes. Therefore, investigators are forced to make the assumption that other climatological variables required to estimate evaporation and transpiration either remain at current levels or scale information that is derived from GCMs. Lettenmaier et al. (1994) describe some of these adjustments.

Five methods have been selected for estimating either potential or reference crop evapotranspiration. The first, a modified Penman-Monteith which calculates E_p, is often used when climatological information is available or may be reliably estimated from empirical equations. This equation may be used for daily or longer time steps. A second modified Penman, the Priestly-Taylor equation, makes simplifying assumptions to the Penman equation and computes E_{rc}. Hargreaves, is a temperature based method and may be used when climatological data is limited. It estimates reference crop evapotranspiration on time intervals equal or longer to one month. The fourth method, Thornthwaite, is also a temperature based method which gives estimates of potential evapotranspiration on a monthly basis. Its use has been questioned but because it has seen widespread application throughout the world, it should be investigated as to its applicability to climate impact assessment (Gleick, 1987). The fifth method, Blaney-Criddle, is a well known temperature based method which in its current form contains much empiricism (Shuttelworth, 1993).

Potential Evapotranspiration (E_p): Modified Penman-Montieth

The data requirements for the Penman-Monteith equation are large and the method is primarily designed for small areas in order to compute actual E_v. For these reasons, this specific equation (6) was not used as one of the methods for this work. Reference is made to show the modification of the Penman-Montieth equation to compute potential and reference crop evapotranspiration estimates (Shuttlewoth, 1993). The modified Penman equation for E_p is similar to equation 5. It assumes surface resistance is zero ($r_s = 0$, Equation 6) and the energy supply, A (Equation 6), is replaced by a net energy exchange for the free water surface plus a term for energy advected to the water body, $R_n + A_h$. It is assumed that $A_h=0$ in equation 7 when performing regional hydrologic assessment.

$$E_p = \frac{\Delta}{\Delta+\gamma}(R_n + A_h) + \frac{\gamma}{\Delta+\gamma}\frac{6.43(1+0.536U_2)D}{\lambda} \tag{7}$$

where,
R_n= net radiation exchange (mm/day)
U_2= wind speed at 2m, m/s
D= vapor pressure deficit.

Because net radiation data is often scarce, an equation to derive its value was used. Aside from temperature, the equation uses two additional climate variables; relative humidity and bright sunshine hours per day. These were taken as monthly mean values from the IIASA database, given on a 0.5 x 0.5° basis (Leeman and Cramer, 1993). The value for net radiation can be calculated with the following equation.

$$R_n = \left((1-alb)\left(0.25+0.5\frac{n}{N}\right)R_a\right) - \left(0.9\frac{n}{N}+0.1\right)\left(0.34-0.14\sqrt{e_d}\right)\sigma(T+273.2)^4 \tag{8}$$

R_n = net radiation (MJ $m^{-2}day^{-1}$)
n = bright sunshine hours per day (h)
N = total day length (h)
R_a = extraterrestrial radiation (MJ $m^{-2}day^{-1}$)
σ = Stefan-Boltzmann constant (4.903×10^{-9} MJ m^{-2}°$K^{-4}day^{-1}$)
T = mean air temperature (°C)
e_d = vapor pressure (kPa)
alb = albedo, a measure of surface reflectivity

Equation 8 can be converted to mm/day by dividing by the latent heat of vaporization, λ and assuming constant water density. Actual vapor pressure is estimated using data of mean monthly relative humidity values. Relative humidity, taken from the IIASA data base, is estimated by multiplying the saturated vapor pressure by the relative humidity data. To compute the extraterrestrial radiation and total day length the following equations were used.

$$R_A = 15.392\ d_r\,(w_s \sin f \sin d + \cos f \cos d\ \sin w_s) \tag{9}$$

where;
R_A = extra-terrestrial radiation (mm/day)
N = maximum possible daylight hours, equation (9)

d_r = relative distance earth-sun, equation (10)
w_S = sunset hour angle [radians], equation (11)
f = latitude of site (+ for Northern Hemisphere, - for Southern Hemisphere) [radians]
d = solar declination [radians], equation (12)
J = Julian day

and:

$$N = \frac{24}{\pi}\omega_s \tag{10}$$

$$d_r = 1 + 0.033 \cos(2p\, J / 365) \tag{11}$$

$$w_S = \arc\cos(-\tan f \tan d) \tag{12}$$

$$d = 0.4093 \sin(2p\, J / 365 - 1.405) \tag{13}$$

Reference Crop (E_{rc}): Priestly-Taylor

Priestley and Taylor (1972) found that for very large areas the second term of the Penman equation (5) is approximately thirty percent that of the first. Thus an approximation to the Penman equation that is less data demanding may be written as:

$$E_{rc} = \alpha \frac{\Delta}{\Delta + \gamma}(R_n - G) \quad \text{(mm / day)} \tag{14}$$

where α has been given the value of 1.26 in humid climates (relative humidity greater than 60 percent in the month with the maximum evaporation) and 1.74 for arid climates (relative humidity less than 60 percent in the month with the maximum evaporation). G is the heat conduction to the soil This approximation performed fairly well in the comparison study reported by Jensen et al. (1990). This is a *reference crop* evapotranspiration estimate, which should show lower values than those reported by equation 7, which gives *potential* evapotranspiration. The Priestly and Taylor equation (14) does not require wind speed, in contrast to the Penman-Montieth equation (7). Another advantage to the Priestly-Taylor approximation to E_{rc} for river basin hydrology is evidence supporting its applicability to regional estimates of E_{rc}, (Shuttelworth, 1993).

Emperical (Temperature-Based) Methods

Estimates of $E_p$that are mainly based on temperature have been proposed since the 1920's, their main attractive point being the limited data requirements to produce the estimates. The argument to use temperature is that both the components of evaporation of equation 5 are related to temperature, with the first being substantially larger than the first. Of these methods the most widely used is the Thornthwaite method (1948) which was developed for east-central US and meant only to apply to mid-latitude climates similar to those of east-central US; however it has been used widely throughout the world (sources). There are several studies that show that Thornthwaite usually underestimates Evapotranspiration (see for example Jensen et al., 1990), and so it is often given a multiplying coefficient to increase its relative magnitude. The strength of the method, then, is its simplistic way of generating the seasonal distribution (shape) of E_p.

Hargraves (E_{rc})

The Hargreaves equation is a second temperature based method and although it gives an expression for the reference crop evapotranspiration it is used as a representative expression for potential evapotranspiration (Hargreaves, 1981; Hargreaves et al., 1985). It has a link to solar radiation.

$$E_{rc} = 0.0022 * R_A * \delta'_T{}^{0.5} * (T + 17.8) \tag{15}$$

where:

R_A = mean extra-terrestrial radiation [mm/day], which is a function of the latitude f, (equation 8)

δ'_T = temperature difference = mean monthly maximum temperature - mean monthly minimum temperature for the month of interest [°C].

T = mean air temperature [°C].

This equation gives reasonable estimates of reference crop evapotranspiration because it has a link to solar radiation through R_a and takes into account the impact of radiation warming the surface near the ground by the term, δ'_T.

Thornthwaite (E_p)

The Thornthwaite Method for estimating E_p has been widely used throughout the world (1939). The Thronthwaite computes monthly potential evapotranspiration:

$$E_p = \beta 16 N_m \left(\frac{10\overline{T}_m}{I} \right)^a \quad \text{(mm)} \tag{16}$$

$$I = \sum i_m = \sum \left(\frac{\overline{T}_m}{5} \right)^{1.5} \quad \text{for } m = 1....12 \tag{17}$$

and

$$a = 6.7x10^{-7} I^3 - 7.7x10^{-5} I^2 + 1.8x10^{-2} I + 0.49 \quad \text{(to 2 significant figures)} \tag{18}$$

Blanley Criddle (E_p)

The Blanely Criddle method is an emperical, temperature based approach for calculating potential evapotranspiration. The method uses temperature as well as daily sunshine duration, minimum daily relative humidity, and the 2m daytime wind at. The model is quite sensitive to the wind speed variable and somewhat insensitive to the estimate of relative humidity. Uncalibrated, the Blanley Criddle method gives significantly larger estimates of potential evapotranspiration than the other methods described above for most of the basins. The preferred form of the equation reveals its strong empiricism and is given by (Shuttleworth, 1993)

$$E_p = a + bf \tag{19}$$

where,

$$f = p(0.46T + 8.13) \tag{20}$$

$$a = 0.0043\ \mathrm{RH}_{\min} - (n / N) - 1.41 \tag{21}$$

$$\begin{aligned} b = 0.82 - 0.0041(RH_{\min}) + 1.07(n / N) + 0.066(U_d) \\ - 0.006(RH_{\min})(n / N) - 0.0006(RH_{\min})(U_d) \end{aligned} \tag{22}$$

The variable p in equation 20 is the actual daily daytime hours to annual mean daily daytime hours expresses as a percent, T is the mean air temperature in °C, (n/N) is the ratio of actual to possible sunshine hours, RH_{min} is the percent minimum daily relative humidity, and U_d is the daytime wind at 2m in ms^{-1}.

Calibration of Emperical Methods.

A calibration procedure was used to adjusted the estimates of temperature based Ep estimates. It was based on the idea that the long-term water balance of a large catchment can be simply written as $R_a = P_a - Ev_a$; annual runoff equals annual precipitation minus annual evaporation (Dooge, 1992). If it is assumed that there is no over-year storage when using long-term averages, then a simple monthly runoff model for a basin such as the Blue can be expressed as;

$$\mathrm{R}_i = \begin{cases} BF, & \hat{\mathrm{E}}_{\mathrm{p,i}} \geq \mathrm{P_i} \\ \mathrm{P_i} - \hat{\mathrm{E}}_{\mathrm{p,i}} + BF, & \hat{\mathrm{E}}_{\mathrm{p,i}} < \mathrm{P_i} \end{cases} \tag{23}$$

then summing up the monthly runoff values and setting them equal to the observed values,

$$\sum R_i = \sum Ro_i \tag{24}$$

it is possible to find a coefficient, β, that gives a estimate of the potential evapotranspiration value for the basin based on a given potential evpotranspiration.

$$\hat{\mathrm{E}}_{\mathrm{p,i}} = \beta\ \mathrm{E}_{\mathrm{p,i}} \tag{25}$$

where,

Ro_i = observed runoff in month i

R_i = computed runoff in month i

$\hat{E}_{p,i}$ = adjusted potential or reference crop evapotranspiration estimate in month i

$E_{p,i}$ = potential or reference crop evapotranspiration by Thornwaite and Hargreaves

P_i = Precipitation in month i

β = calibration coefficient for temperature method

BF = baseflow (95% percentile low flow)

Sensitivity of E_p and E_{rc} models to temperature

Computation of both Ep and Erc values depends largely on the climate for which they are applied. For this reason, several basins have been selected to try to span a range of climate variability (Figure 2). Four basins have been selected: The Blue Nile Basin of North East Africa; the Vistula Basin of Central Europe; the Mulberry Basin, a sub-basin of the Arkansas in the South East of USA.; and the East River, a sub-basin of the Colorado River, in the Western USA. A brief description of these basins is given below.

Case Study Basins

Four basins of different scale and climatological characteristics were selected. These included the Blue Nile river basin of Africa, the Vistula river basin in Poland, the East river, a tributary of the Colorado River, in Colorado USA; and the Mulberry river, a tributary of the Arkansas, in Arkansas, USA. These basins were selected because of their range of variability both spatially and climatologically (Figure 2). Selection criteria included basin size, varying climatic and basin characteristics, as well as time series data availability. Two semi-arid basins (Blue Nile and East) were selected and plot to the extreme left and right while the humid Mulberry basin plots the furthest to the top (Figure 2). A brief description of each basin is given below

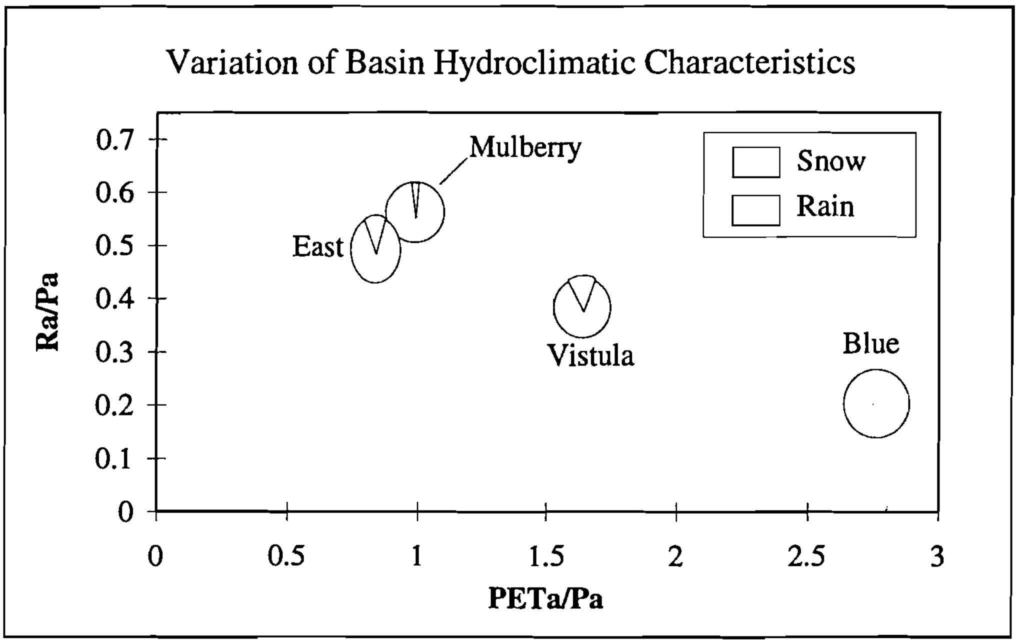


Figure 2. Basin Hydro-climatic Characteristics

Blue Nile

The Blue Nile Basin (12°N 36°) is in a temperate, semi-arid region with little variation in temperature. The average precipitation record reveals that precipitation comes during a three month "rainy season", while the remaining portion of the year is quite dry (Figure 3a). The Blue Nile Basin covers an area of approximately 325,000 km^2, making it one of the largest single basins in the world (Shanin, 1985). Although the annual precipitation is quite high, in some places probably reaching 1500 mm year, the annual runoff for this basin is approximately only 165 mm, giving a runoff coefficient below 0.2. This can attributed to very high evapotranspiration within the basin (Figure 1 and Figure 2).

Vistula

The Vistula basin (52°N, 20°E), covers an area of 194,376, km^2 (87% within the boundaries of Poland). The area can be divided into four diverse climatological area; with the upper, southern portion of the basin residing in a mountainous area. Moving north, the basin is characterized by high and low lands and numerous lake areas (annual precipitation ranging from 500 to 600 mm and mean annual air temperature 7.5°C). The entire Vistula basin has a runoff coefficient of approximately 0.30. The monthly mean discharge values in Figure 3b reveals the rather constant discharge of this basin (Figure 3b).

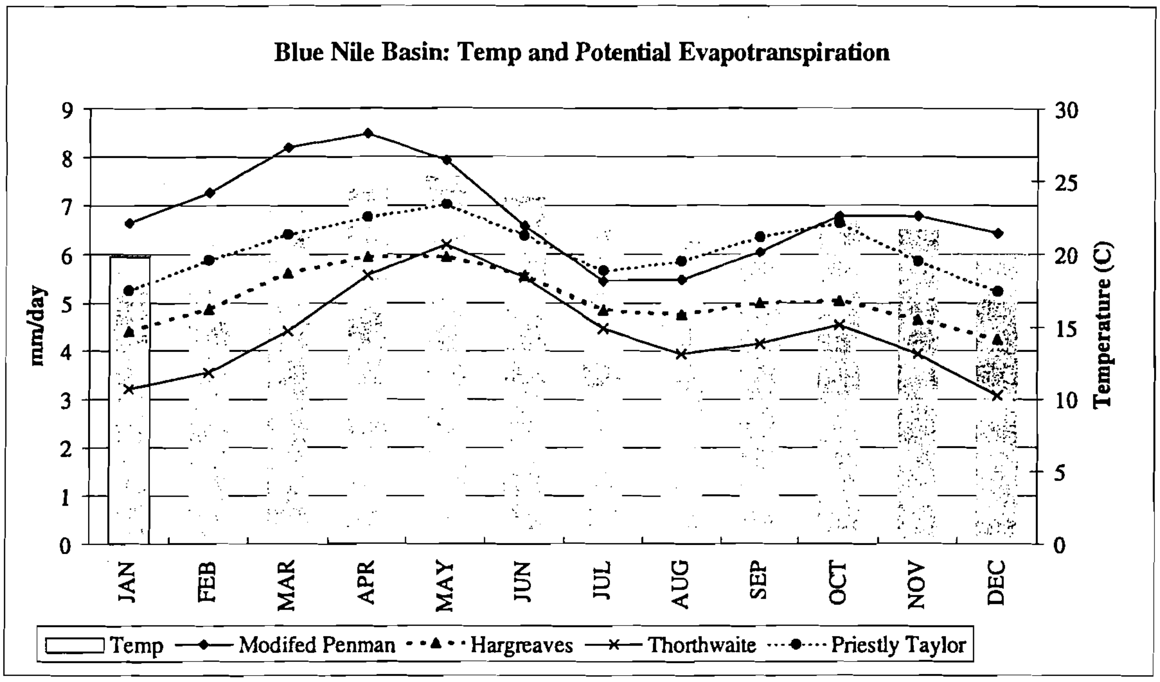
Mulberry

The Mulberry basin in Arkansas U.S.A. is a substantially smaller catchment than those described above and is found at 35N -94W. This is a moderately temperate climate, with a mean annual air temperature of approximately 16°C and only a few incidents of winter mean monthly air temperatures dropping below 0°. The region is characterized by dense ground cover and has little variation in elevation, with the gauging station located at 342 m above sea level. The basin area is a little less than 1000 km^2, making it a relatively "small" catchment. Although Nemec and Shaake (1982) state that modeling such basins should produce minimum error, the climate of this basin produces an interesting seasonal runoff characteristic that can be observed in Figure 3c. Although the overall runoff coefficient is approximately 0.44; the winter season coefficient is as high as 0.70, while the summer season's runoff coefficient drops to below 0.20. This large seasonal change is difficult to model when using models with a limited number of parameters.

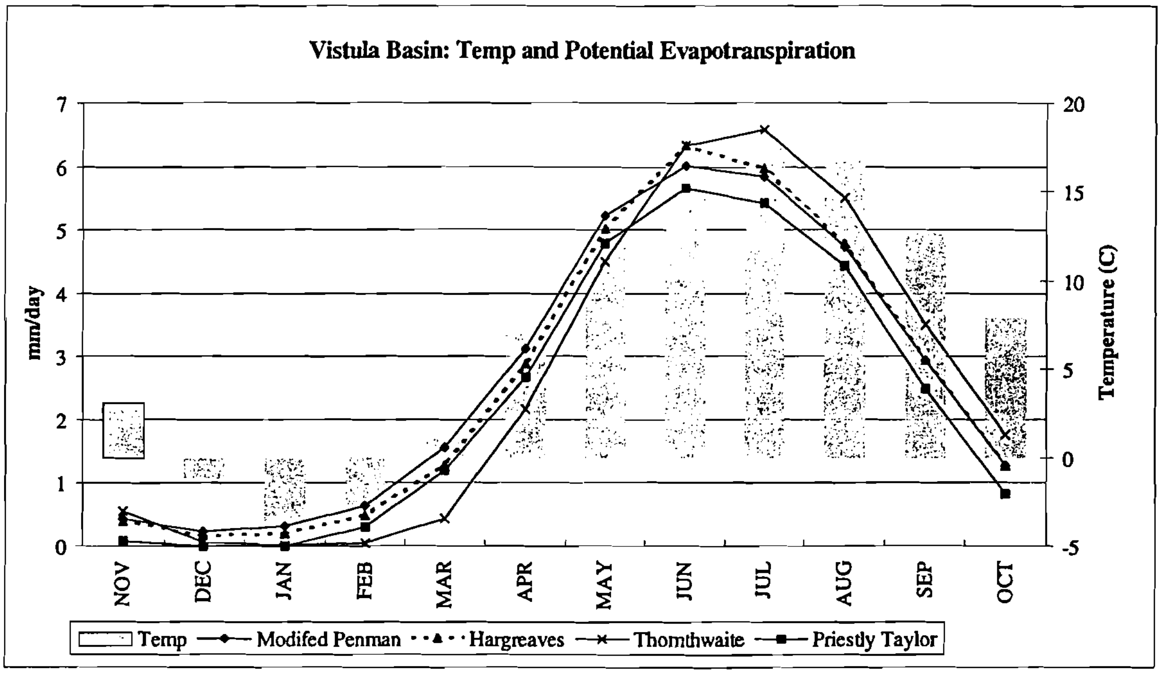
East

The East river in Colorado (40°N -105W) U.S.A. is a tributary of the Gunnison River basin and was the smallest catchment modeled. This basin resides within the Rocky Mountain Range, with most of the basin above 3000m. Although considered a semi-arid region, the runoff coefficient for this basin is highest of those selected because most of the basin runoff comes in the form of spring runoff from snowfall events in the winter. This can be seen from Figure 2, as this basins plots to the extreme left in this figure. The climate station for this basin is located in the Gunnison Valley (elevation 2500m), and so the precipitation records were adjusted to reflect the effect of elevation on precipitation by multiplying the precipitation record by 1.33 in the winter months (November to March). This high mountain basin proved very sensitive to model calibration. A monthly snowmelt model was used to derive an effective precipitation and the specification of the temperature thresholds of freezing and melting proved to be critical to model.

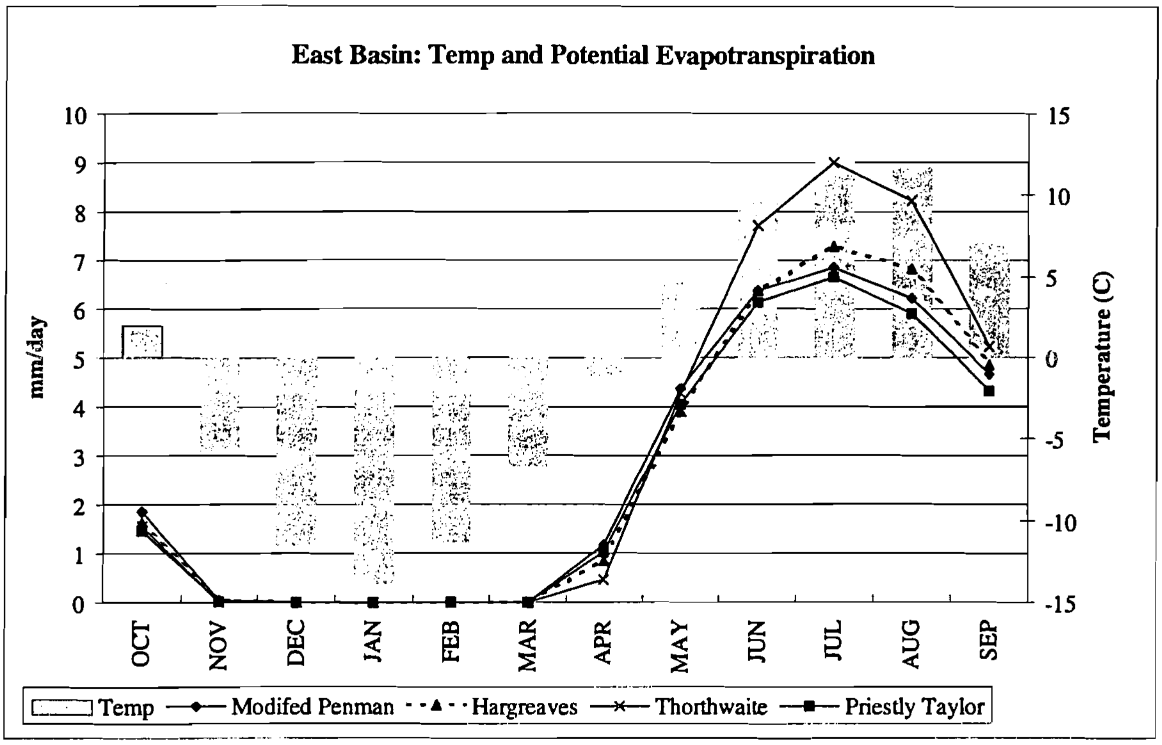
Mean monthly E_p by Methodology for Selected Basins



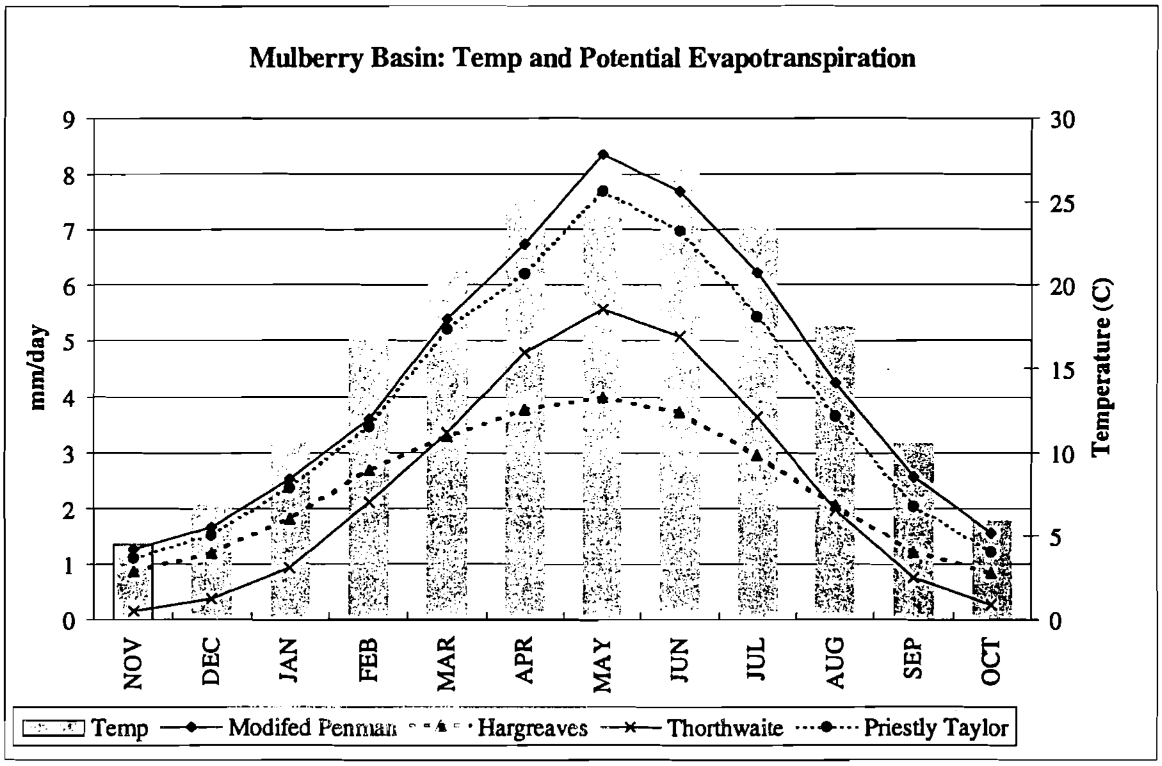
a. Blue Nile



b. Vistula



c. East



d. Mulberry

Figures 3.a-d. Plot of mean monthly temperatures and potential evapotranspiration estimates using four of the methods for the four basins. Hargreaves and Thornthwaite are calibrated values based on the calibration procedure given by equations 23-25.

Sensitivity Results of $E_{p,rc}$ methods

Mean monthly temperature distributions are plotted in Figures 3.a-d. for each of the four basins studied. Three of the basins show high seasal temperature fluctuations. The Blue Nile, at 12°N latitude shows the smallest amount of seasonal variability. A sensitivity analysis of the different estimates of potential evapotranspiration to temperature fluctuations for each basin is shown in Tables 2.a-d. These tables reveal the generally strong linear relationships of the different methods around the margin, with the exception of the Thornthwaite method which becomes non-linear past 4° for higher temperature basins like the Mulberry and Blue Nile. The five methods showed significant difference in the magnitude of the potential evapotranspiration change per degree warming. The Penman based methods showed the strongest linearity and the smallest fluctuations under constant temperature changes (except for the East River). There was little difference in the relative change between the Penman-potential estimates (Modified Penman-Montieth, Equation 7) and the Penman-reference crop (Priestly-Taylor, Equation 14), however it appears that the Penman-reference crop equation gives slightly smaller relative increases. Differences in the two Penman methods with respect to climate change impacts will be more observable when applied to the hydrologic model where the seasonal distribution and monthly relative magnitudes becomes more important. The Blanley-Criddle method also shows a linear response with temperature change, and like the other temperature methods shows larger changes in E_p for each degree Celsius (4.8%/°C), athough the smallest of the emperical relationships.All three emperical methods (Thornthwaite, Hargreaves, and Blaney-Criddle) give higher per degree percent changes. The average 1°C percent change in potential evapotranspiration are given in Table 2.

Nemec and Shaake (1982) report a 4% increase in potential evapotranspiration for a 1°C increase in temperature based on the Budyko method for estimating potential evapotranspiration for an arid and humid basin in the US. Rosenburg et. al (1989) give results of the sensitivity on evapotranspiration by performing a sensitivity analysis with uniform changes in model parameters. Using the Penman equation (6) to estimate evapotranspiration from an irrigated agricultural field in a slightly humid region of the US (Nebraska, USA). They computed an approximate 6%/°C increase in evapotranspiration from this field which implies at least a 6%/°C increase in Ep using a Penman method. This value is higher than those found using the modified Penman-Montieth equation (8) and the Priestly-Taylor equation (14) for similar basins.

a. Blue Nile

	1°C	2°C	3°C	4°C	5°C
Penman	2	4	6	8	11
Priestly-Taylor	2	4	6	8	10
Hargreaves	5	10	15	20	25
Thornthwaite	9	19	32	49	69
Blaney-Criddle	6	12	19	25	31

b. Vistula

	1°C	2°C	3°C	4°C	5°C
Penman	3	6	9	11	14
Priestly-Taylor	3	5	8	10	13
Hargreaves	7	13	20	27	34
Thornthwaite	5	10	16	22	29
Blaney-Criddle	4	9	13	17	21

c. Mulberry

	1°C	2°C	3°C	4°C	T°5
Penman	2	5	7	10	12
Priestly-Taylor	2	4	6	8	11
Hargreaves	5	11	16	22	27
Thornthwaite	8	16	25	36	49
Blaney-Criddle	3	6	9	13	16

d. East

	1°C	2°C	3°C	4°C	T°5
Penman	8	16	25	32	40
Priestly-Taylor	7	15	22	29	36
Hargreaves	13	26	39	53	68
Thornthwaite	5	12	17	24	29
Blaney-Criddle	6	12	17	23	29

Table 2.a-d. Sensitivity of the different Ep and Erc methods to temperature change (% change from $\Delta T=0°C$ and all other variables within the specific method are held constant)

The values in Table 3 are the mean and coefficient of variation of the five methods for the five basins. This table shows that the Penman based methods are on average the least sensitive yet have the greatest amount of variability. The three emperical methods show a range of variability, but tend to be less climate sensitive.

Table 3. Average percent increase change and the coefficient of variation in potential evapotranspiration estimate per degree celcius for the four basins

Method	ΔE_p /°C	
	Mean	CV
Penman	3.75%	0.77
Priestly-Taylor	3.5%	0.68
Hargreaves	5.5%	0.18
Thornthwaite	8.75%	0.38
Blaney-Criddle	4.75%	.031

Affects of Ep and Erc methods on Climate Change Impacts on basin runoff

Four of the E_p and E_{rc} techniques were implemented within a monthly water balance model to assess the impact of these different methods on climate change assessment of basin discharge. The Blaney-Criddle method was not used because of its large degree of empericism. For some of the basins an uncalibrated Blaney-Criddle gave estimates of E_p four to five time that of the Penman method, therefore it was determined to be inappropriate. The four basins have been described above and the hydrologic model is discussed below. Only temperature was altered within the different E_p models. In modified Penman and Prieslty-Taylor; wind speed, relative humidity, and sunshine hours were applied uniformly over the month and held constant. Mean monthly values were taken from the IIASA database (Leemans and Cramer, 1991).

Hydrologic Model

A hydrologic model based on a soil moisture mass balance which incorporates the estimation of evapotranspiration was chosen to test the applicability of the different potential evapotranspiration estimates (Kaczmarek, 1993, Yates, 1994). The water balance uses continuous functions of relative storage to represent surface outflow, sub-surface outflow, and evapotranspiration. The model contains five parameters related to: 1) direct runoff; 2) surface runoff; 3) subsurface runoff; 4) maximum catchment water-holding capacit; and 5) base flow.

Direct runoff (R_d) is given as:

$$R_d = \beta P_{eff} \tag{26}$$

The soil moisture balance is written as:

$$S_{\max}\frac{dz}{dt} = \left(P_{peff}(t)(1-\beta)\right) - R_s(z,t) - R_{ss}(z,t) - E_v(z,t,E_p) - R_b \tag{27}$$

Peff = Effective Preipitation (mm/day)
R_s = Surface runoff (mm/day)
R_{ss} = Sub-Surface runoff (mm/day)
E_v = Evapotranspiration (mm/day)
R_b = Baseflow (mm/day)
S_{max}= Maximum storage capacity (mm)
z = Relative storage ($0 \leq z \leq 1$)
E_p = Potential or reference crop evapotranspiration (mm/day)

The Continuous functional forms that are used in equation 7 are:

1. Evapotranspiration - E_v:

E_v is a function of E_p or E_{rc} and the relative catchment storage state given as

$$Ev(z,t,E_{p,rc}) = E_{p,rc}\left(\frac{5z-2z^2}{3}\right) \tag{28}$$

2. Surface Runoff - R_S:

Surface runoff is described in terms of the storage state, z, the effective precipitation, P_{eff}, and the baseflow. If the precipitation exceeds the predefined baseflow, then surface runoff is zero.

$$R_s(z,P,t) = \begin{cases} z^{\varepsilon}(P_{eff} - R_b) & \text{for } P_{eff} > R_b \\ 0 & \text{for } P_{eff} \le R_b \end{cases} \tag{29}$$

3. Sub-Surface Runoff - R_{SS}:

Sub-surface discharge is a function of the relative storage state times a coefficient, α.

$$R_{ss} = \alpha z^{\gamma} \tag{30}$$

γ is not calibrated and is normally set to a value of 2.0, while α is part of the calibration routine. The 4th model parameter is the maximum catchment holding capacity, S_{max}. The storage variable, Z, is given as the relative storage state: $0 \le Z \le 1$. S_{max} is defined as the maximum storage volume, so when S_{max} is multiplied by z, the current storage volume for the period is given. Total runoff, for each time step, is the sum of the four components:

$$R_t = R_s + R_{ss} + R_b + R_d \tag{31}$$

Inputs to this model include: Effective Precipitation, Ep,Erc, and for calibration purposes - runoff in the units of (length/time). For all basins in this study a monthly time step has been used. A monthly snowmelt model was used to derive an effective precipitation for the for those basins with a portion of their water attributable to snowmelt (Ozga-Zielinska, 1993).

$$Peff_i = \alpha_i(A_{i-1} + Pm_i) \tag{32}$$

where,

$$\alpha_i = \begin{cases} 0 & \text{for } T_i \le T_s \\ 1 & \text{for } T_i \ge T_l \\ \dfrac{(T_i - T_s)}{(T_l - T_s)} & \text{for } T_s < T_i < T_l \end{cases} \tag{33}$$

and snow accumulation is written as,

$$A_i = (1-\alpha_i)(A_{i-1} + Pm_i) \tag{34}$$

Calibration/Validation

The monthly water balance model was calibrated and validated for each basin using four of the $E_{p/rc}$ models (Table 4) A split sample test was used for all basins. Because of the short record for the East river, the first 7 years were used for calibration and the remaining three year were used for validation (calibration: 1979-1985; validation 1986-1988). Two calibration/validation run were performed for the East River: the first with an adjusted temperature (-°2C in all months) to reflect elevation change, the second with an unadjusted temperature. Interestingly, this nominal temperature change produces significantly different results (Table 2). For both the Blue Nile, and Vistula, 26 years of data were selected. The first 13 years of data were used for calibration and the next 13 years were used for validation. For the Mulberry river, 40 years of data were available from 1948 to 1987; the first 20 years were used for calibration and the second 20 for validation. The correlation coefficient and the average monthly error between the observed and modeled discharge are the calibration statistics used.

Table 4. Calibration and Validation Statistics for different E_p and E_{rc} methods for all basins (Std. Err given in mm. The standard error is a measure of the amount of total deviation from the observed series. *Results for East basin with adjusted temperature to reflect decreasing temperature with increasing elevation (-2°C). **Results for East basin with adjusted temperature , showing the sensitivity of this basin to temperature fluctuations. The East basin proved quite sensitive to the snowmelt model and the temperature bounds that were used to derive effective precipitation. Corr = correlation coefficient and S.E. = Standard Error

		Penman (Potential)		Priestly-Taylor (Ref Crop)		Hargreaves (Ref Crop)		Hargreaves Calibrated (Ref Crop)		Thornthwaite (Potential)	
		Calib	Valid	Calib	Valid	Calib	Valid	Calib	Valid	Calib	Valid
Blue Nile	Corr	0.95	0.93	0.95	0.93	0.93	0.93	0.92	0.92	0.92	0.91
Africa	S. E.	5.8	6.2	5.90	6.4	6.7	6.7	7.2	6.8	7.0	6.9
Vistula	Corr	0.70	0.79	0.75	0.80	0.71	0.81	0.72	0.80	0.68	0.75
Poland	S. E.	5.7	5.4	5.24	5.1	5.6	5.2	5.5	5.3	5.8	5.9
East*	Corr	0.97	0.95	0.97	0.95	0.95	0.92	0.97	0.96	0.97	0.96
CO. USA	S. E.	12.4	12.7	12.5	13.0	15.8	16.0	12.7	12.3	12.3	12.2
East**	Corr	0.95	0.97	0.95	0.97	0.95	0.96	0.95	0.97	0.98	0.93
CO. USA	S. E.	17.0	9.5	17.0	9.5	17.4	11.1	16.9	9.3	11.4	15.0
Mulberry	Corr	0.86	0.86	0.86	0.86	0.84	0.84	0.88	0.83	0.91	0.90
AR. USA	S. E.	24.4	25.1	24	25	25.5	26.8	22.9	27.0	19.3	21.3

Table 5. Ranking PET models. Ranking was based on the sum of the standard error of the calibration and validation series; a lower average rank indicates a superior model performance. The lowest ranking ("best model") value was achieved by the Priestly-Taylor method, the highest ("worst model") by the Hargreaves method. *Results for East basin with adjusted temperature to reflect decreasing temperature with increasing elevation (-2°C). **Results for East basin with un-adjusted temperature, showing the sensitivity of this basin to temperature variations.

	Penman	Priestly-Taylor	Hargreaves	Hargreaves C	Thornthwaite
Blue Nile	1	2	3	5	4
Vistula	4	1	2.5	2.5	5
East*	3	4	5	2	1
East**	3.5	3.5	5	1	2
MulberryAR.	3	2	5	4	1
Avg Rank	2.9	2.5	3.5	3.5	2.6

Results using the Hydrologic model with $E_{p,rc}$ methods

Tables 6 amd 7 show that the temperature based methods give consistently different results than the micrometorological methods in all basins. The arid Blue Nile basin is perhaps the most sensitive to the potential evapotranspiration estimate, as the elasticity is approximately 1. Elasticity is defined as

$$\Phi=\%Q/\%E_p \tag{35}$$

The top portion of Tables 7.a-e are the % changes in E_p and E_{rc} for $\Delta T°3$ and $\Delta T°5$. If these values are compared to the T°3 P%0 and T°5 P%0 scenarios it can be seen that the change in runoff is directly correlated to the change in the potential evapotranspiration values. An exception to this is the East River, where the large fraction of runoff from snowmelt complicates the runoff process. Implementing precipitation changes reveals more of the impact of the hydrologic model The relative percent changes in evapotranspiration (E_v) are smaller than runoff because of the significantly larger magnitude of this variable.

Table 6. Elasticities for different methods in different basins for Δ°3C and Δ°5C (no precipitation change). East*(-2°C removed from base); East**(unadjusted)

	Penman	Priestly-Taylor	Hargreaves	Hargreaves-Calib	Thornthwaite
Blue	Ep	Erc	Erc	Erc	Ep
T°3	-1.2	-1.2	-1.0	-1.1	-0.94
T°5	-1.0	-1.1	-0.97	-1.0	-0.71
Vistula	Ep	Erc	Erc	Erc	Ep
T°3	-0.56	-0.50	-0.75	-0.85	-0.89
T°5	-0.50	-0.53	-0.79	-0.74	-0.86
East*	Ep	Erc	Erc	Erc	Ep
T°3	+0.32	+0.40	+0.13	+0.26	+0.33
T°5	+0.10	+0.14	-0.10	-0.07	+0.16
East**	Ep	Erc	Erc	Erc	Ep
T°3	-0.33	-0.27	-0.09	-0.31	+0.07
T°5	-0.26	-0.23	-0.10	-0.25	+0.14
Mulberry	Ep	Erc	Erc	Erc	Ep
T°3	-0.43	-0.50	-0.50	-0.69	-0.52
T°5	-0.41	-0.45	-0.44	-0.63	-0.48

Tables 7.a-d Climate Change scenarios with different E_p and E_{rc} models

a. Blue Nile

Blue Nile	Penman		Priestly-Taylor		Hargreaves		Hargreaves-Calib		Thornthwaite	
%	Ep		Erc		Erc		Erc		Ep	
T°3 *	6		6		15		15		32	
T°5 **	11		10		25		25		69	
CC %	R	Ev	R	Ev	R	Ev	R	Ev	R	Ev
T°3 P%0*	-7	2	-7	2	-15	5	-16	5	-30	9
T°3 P%15	23	13	21	13	13	16	14	15	-4	20
T°3 P%-5	-32	-10	-31	-10	-40	-8	-42	-7	-51	-5
T°5 P%0**	-11	3	-11	3	-24	7	-25	7	-49	14
T°5 P%15	17	14	15	15	2	19	3	18	-29	28
T°5 P%-5	-35	-9	-35	-10	-46	-6	-48	-6	-65	0
giss	66	29	63	30	68	28	74	27	41	14
gfdl	-13	0	-13	0	-26	3	-27	4	-42	8

b. Vistula Basin

Vistula	Penman		Priestly-Taylor		Hargreaves		Hargreaves-Calib		Thornthwaite	
	Ep		Erc		Erc		Erc		Ep	
T°3*	9		8		20		20		16	
T°5**	14		13		34		34		29	
CC	R	Ev	R	Ev	R	Ev	R	Ev	R	Ev
T°3 P%0*	-7	3	-9	4	-15	6	-17	7	-16	7
T°3 P%15	17	14	18	14	7	18	7	18	11	17
T°3 P%-15	-26	-10	-29	-9	-32	-8	-35	-7	-34	-7
T°5 P%0**	-11	5	-13	6	-27	-4	-25	10	-25	10
T°5 P%15	11	17	12	16	-3	22	-4	23	-3	22
T°5 P%-15	-29	-9	-32	-8	-36	-6	-39	-5	-39	-5

c. East Basin, *with 2°C removed from the base temperature record

East*	Penman		Priestly-Taylor		Hargreaves		Hargreaves-Calib		Thornthwaite	
%	E_p		E_{rc}		E_{rc}		E_{rc}		E_p	
T°3*	25		22		39		39		15	
T°5**	40		36		68		68		29	
CC %	R	Ev	R	Ev	R	Ev	R	Ev	R	Ev
T°3 P%0*	8	17	9	15	2	26	4	25	13	9
T°3 P%15	18	24	18	22	18	39	18	36	28	19
T°3 P%-15	-10	2	-10	0	-15	11	-14	8	-7	-5
T°5 P%0**	4	16	5	14	-3	32	-2	28	11	5
T°5 P%15	18	27	20	24	14	48	11	40	31	19
T°5 P%-15	-16	-1	-14	-3	-20	14	-21	9	-10	19

d. East Basin; no adjustment to temperature record

East	Penman		Priestly-Taylor		Hargreaves		Hargreaves-Calib		Thornthwaite	
%	E_p		E_{rc}		E_{rc}		E_{rc}		E_p	
T°3*	21		18		35		35		15	
T°5**	34		30		60		60		28	
CC %	R	Ev	R	Ev	R	Ev	R	Ev	R	Ev
T°3 P%0*	-7	-1	-5	-1	-3	14	-11	5	1	-2
T°3 P%15	9	11	10	9	11	25	7	20	21	13
T°3 P%-5	-24	-14	-18	-11	-19	1	-28	-10	-14	-13
T°5 %0**	-9	-2	-7	-3	-6	17	-15	7	4	-2
T°5 P%15	4	7	10	7	10	31	-3	16	20	9
T°5 P%-5	-23	-13	-21	-14	-22	4	-29	-6	-12	-12

e. Mulberry Basin

Mulberry	Penman		Priestly-Taylor		Hargreaves		Hargreaves-Calib		Thornthwaite	
%	E_p		E_{rc}		E_{rc}		E_{rc}		E_p	
T°3*	7		6		16		16		25	
T°5**	12		11		27		27		49	
CC %	R	Ev	R	Ev	R	Ev	R	Ev	R	Ev
T°3 P%0*	-3	4	-3	3	-8	9	-11	10	-13	10
T°3 P%15	21	14	19	13	17	19	14	18	14	17
T°3 P%-15	-24	-7	-23	-7	-30	-2	-34	1	-36	2
T°5 P%0**	-5	6	-5	5	-12	14	-17	16	-22	17
T°5 P%15	19	16	17	15	11	25	7	24	4	26
T°5 P%-15	-26	-6	-25	-5	-34	2	-40	6	-43	8

Conclusions

The effort here was an attempt to clear some of the confusion regarding the methods and application of the commonly and often misunderstood concept of potential evapotranspiration in estimating the impact of potential climate change on river basin runoff. With so much work being done on impact assessment of water resource systems under climate change over the last several years, it is important to develop physically sound methods that will ensure proper analysis. Several of the more common methods for estimating potential evapotranspiration were detailed and used within a hydrologic model to determine how these various methods impact a climate change assessment. It was generally shown that by simply using a different estimator of E_p, dramatically different impact results can be found.

Not surprising, the magnitude and temporal distribution of the E_p estimates are important. Annual values of different E_p methods might be similar but distribution is critical. Figure 4 is a summary of the 3°C and 5°C scenarios taken as the sum of the impacts of the different E_p methods on the estimation of E_p and runoff (Tables 7.a-d). This figure reveals the generally greater sensitiivty of the emperical methods on impact analysis. Empirical methods, which are often only temperature based, give significantly different marginal changes to temperature fluctuations when compared with the physically (or micromerological) based methods such as the modified Penman equations. Drastically different results are found under the same climate scenarios for a given basin. However, it is difficult to draw definitive conclusion between the empirical and physical methods because different climatological regions show different trends. It appears, however, that physically methods are more soundly based, but are data intensive as well as data sensitive. These methods also require estimates of net radiation, which itself has been estimated by an empirical equation adding an additional element of uncertainty to the physical methods. Of the two physical methods used here, the Priestly-Taylor method would appear to be the method of choice for regional, hydrologic analysis. It is less data intensive and is generally thought of as a "reference crop" estimation of potential evapotranspiration, which is undoubtedly more sound when looking at meso-scale river basins. When coupled with the water balance model, there is little difference in the impact of climate change on runoff when applying either the modified Penman or the Priestly-Taylor method. The monthly distribution of the two methods is only slightly different, with the modified Penman (an E_p estimation) giving slightly higher values than the Priestly-Taylor method (an E_{rc} method). The magnitude and relative change under temperature variations are almost equal which seems to indicate that wind speed is not a significant variable for regional analysis.

The strength of calibrated regional empirical methods is shown in the Mulberry Basin, where the calibration and validation statistics performed best using the Thornthwaite method. However if using a regionally calibrated empirical method for climate change assessment, direct input of temperature is probably not wise especially when applying larger temperature variations (>3°C). Instead, percentage changes of the base should be applied by assuming, for example, a 3-5% increase in E_p for each °C change.

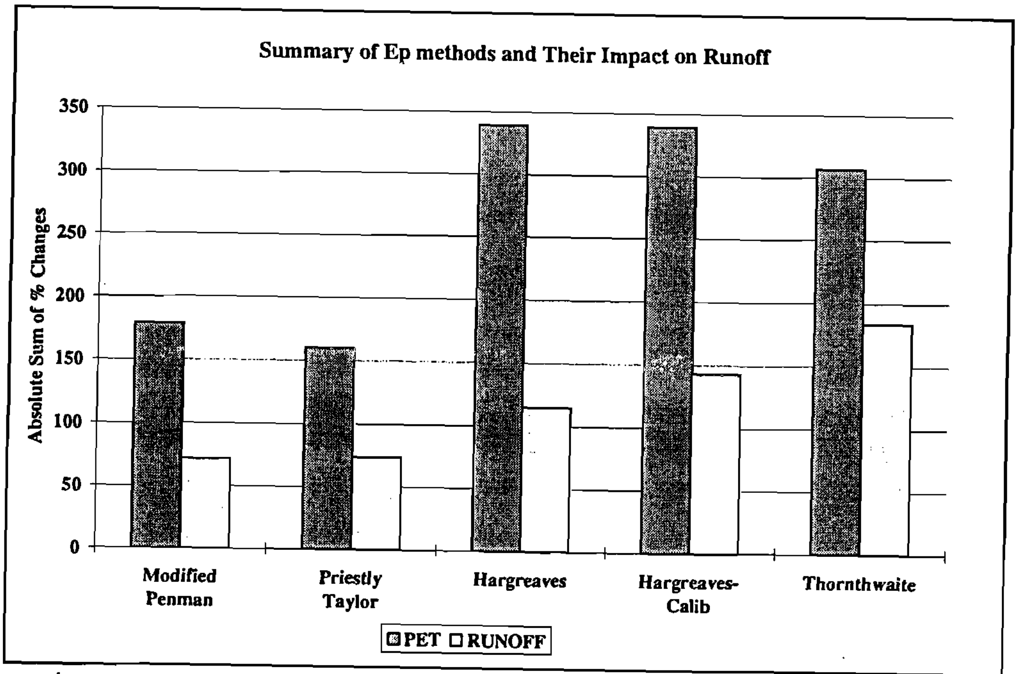


Figure 4. Summary chart of different potential evapotranspiration estimators. The percent change for the 3°C and 5°C scenarios were added from Tables 7.a-d with precipitation held constant.

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