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Water Balance Estimates of Evapotranspiration Rates in Areas with Varying Land Use

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1. Introduction

In the continental United States, approximately 2/3 of all rainfall delivered is lost to evapotranspiration (ET; US Water Resource Council, 1978). It follows that the ET rate, representing the combined processes of physical evaporation and biological transpiration, is essential for predicting water yields, designing irrigation and supply projects, managing water quality, quantity, and associated environmental concerns, and negotiating disputes, contracts, or treaties involving water. Water fluxes in catchments are controlled by these physical and biological processes as well as by hydrogeologic properties that are complex, heterogeneous, and poorly characterized by field and laboratory measurements. As a result, practical theories of ET rates and their impact on runoff generation and catchment hydrology remain elusive.

Many methods have been used to determine ET rates in watersheds. Since atmospheric vapor flux is difficult to measure directly, most methods monitor the change of water in the system. Potential ET (PET), the amount of ET that would occur if unlimited water were available, can be measured using an evaporation pan or ET gauge. Pan data can also be used to estimate the actual ET, representing the ET that occurs when water is limited, for the vegetation of interest using relationships presented by Jensen et al. (1990). Lysimeters, soil water depletion, and the energy balance method have also been used to estimate ET (e.g., van Bavel, 1961), though measurements are difficult. Another approach, the water balance method, provides simple, effective estimates of ET rates if accurate stream gauging and precipitation data are available. This method is generally used for large watersheds, and compares water inputs (e.g., precipitation) and outflows (e.g., stream flow) for a given basin over long periods of time.

This paper uses the water balance approach to calculate ET rates by comparing precipitation and runoff data for watersheds. We expand on this traditional method to show how ET can be deconvolved into physical and biological components whose magnitudes vary both

seasonally and with variations in basin properties, such as land use or hydrogeology. We also demonstrate how results for different basins can be directly compared to quantify ET disparities and out-of-basin gains or losses of water.

1.1. Previous work on water balance

The water balance approach entails determining the ET from the following equation:

$$ET \;=\; P \;-\; R \;\pm \Delta W \;\pm \Delta S \qquad (1)$$

where P is precipitation, R is runoff, ΔW represents withdraws, diversions, or interbasin transfers, and ΔS is the change in groundwater and soil water storage (Ward & Trimble, 2004). The main assumptions of the model are: (1) the net groundwater flow across the watershed boundaries is zero and, hence, other than ΔW, the only inflow of water is P and the only outflows are ET and R; (2) the saturated and unsaturated storages are lumped in a single term (S); and (3) the entire water storage is accessible by plant roots in the watershed (Palmroth et al., 2010). For short intervals, ΔS is important and should be measured, but over an annual period it is normally small and to first order can be neglected, so:

$$ET \;=\; P \;-\; R \;\pm \Delta W \qquad (2)$$

Previous workers assert that different land use practices have different and sizeable effects on ET and runoff rates (Dunn & Mackay, 1995; Gerten et al., 2004). In urban settings, impervious surfaces such as buildings, roads, parking lots, and other structures prevent rainfall from infiltrating; consequently, abnormally large fractions of runoff are directed into stream channels (Schilling & Libra, 2003). In contrast, in forest settings more water is evaporated directly off of tree leaves during throughfall, and once rainwater reaches the soil it is absorbed and transpired. Thus, in forested areas, ET is enhanced and runoff is reduced (Murakami et al., 2000). Crops and grasses can also intercept and transpire rainwater, but at significantly lower rates than in forests.

The following discussion tests and expands upon these concepts by comparing water balance calculations for several watersheds. Basins selected for analysis have long-term meteorological and discharge records as well as other special characteristics that simplify calculations or exemplify special processes and effects. Discharge and meteorological data are correlated with GIS data to determine the effects of land use on ET-runoff relationships.

1.2. Vegetation and transpiration

Vegetation and the water cycle, including the relationship between the ET rate and runoff generation, are intrinsically linked (Hutjes et al., 1998; Arora, 2002; Gerten et al., 2004). Basin water balance is a fundamental constraint on the productivity (Clark et al., 2001) and distribution (Stephenson, 1990) of terrestrial vegetation. Similarly, the plant community structure and geographic location are of primary importance for ET and runoff generation dynamics

(Dunn & Mackay, 1995). Transpiration accounts for the movement of water within a plant and the subsequent loss of water as vapor through stomata in its leaves, and the type and abundance of vegetation significantly affects the overall ET rate. Plant communities influence runoff processes in numerous ways, including phenology (Peel et al., 2001), plant maturity (Neilson, 1995), leaf area (Kergoat, 1998), stomatal behavior (Skiles & Hanson, 1994), and rooting strategy (Milly, 1997). In turn, processes such as albedo, interception (Eckhardt et al., 2003), percentage of soil cover, solar radiation, humidity, temperature, and wind (Swank & Douglass, 1974) affect plants transpiration rates. Herbaceous plants generally transpire less than woody plants because they typically have less extensive foliage. Conifer forests tend to have higher ET rates than deciduous forests, which is primarily due to the enhanced amount of precipitation intercepted, evaporated, and transpired by conifer foliage during the winter and early spring seasons.

It is well established that reduced forest cover decreases the ET rate and subsequently increases basin runoff, whereas reforestation typically lowers runoff (Bosch & Hewlett, 1982). For a large part of the southeastern Unites States, forested areas, over a period of decades, have transpired about 33 cm/y (area-depth) more than other land covers (Trimble et al., 1987). Vegetation also effects ET on the global scale; for instance, the variability of annual runoff between continents is controlled not only by differences in precipitation but also by the geographical distribution of various types of vegetation (e.g., evergreen vs. deciduous; Peel et al., 2001). In areas that are not irrigated, actual ET is usually no greater than precipitation, with some buffer in time depending on the soil's ability to hold water. Actual ET will usually be less than precipitation because some water will be lost due to percolation or surface runoff. An exception is areas with high water tables, where capillary action can cause water from the groundwater to rise through the soil matrix to the surface. If PET is significantly greater than precipitation, then soils will dry out if not irrigated.

The role of vegetation in the hydrologic cycle has been extensively studied (Horton, 1919; Wicht, 1941; Penman, 1963; Bosch & Hewlett, 1982; Turner, 1991), and these investigations have generally been split into two categories. The first involves "paired-catchment" experiments, and comparisons among > 90 catchments revealed large variations in runoff response attributable to differences in vegetative cover, with the catchments that have lower forest cover showing increased water yield due to lower ET (Hibbert, 1967; Bosch & Hewlett 1982). The second category, i.e., "single-catchment" water balance studies, is also used to determine the impact of vegetation on runoff. These studies were not designed to study the specific effects of land use on ET rates and water balance, but encompass a diverse group of catchments with different climates, vegetation, and soil types, and thus provide useful information about the role of vegetation in catchment water balance.

1.3. Hydrogeology and runoff

In addition to type and density of vegetation in a watershed, runoff volumes in the stream channel are controlled by the geography and hydrogeology of the basin. The size (Criss & Winston, 2008), shape (Hodge & Tasker, 1995), and orientation relative to the storm path (Ward & Trimble, 2004) of the watershed affect the rate of runoff delivery. Lithology and soil type

significantly influence runoff volumes because permeability varies enormously for different geologic materials, with karstic rocks, gravels, and sand being most permeable, and igneous and metamorphic rocks, shales, and clay being least permeable (Bureau of Reclamation, 1977). If recent rainfall cannot penetrate into the subsurface, it will not be stored and subsequently transpired. In addition, basin topography, especially slope, exerts great influence on the transmittal of water to stream channels. Further, closed depressions can direct water to the groundwater system in karst areas or can cause ponding that enhances ET. Rainfall intensity and rainfall duration also influence runoff percentages and rates, and if they are high, are major causes of flash floods.

Anthropogenic activities also modify the hydrology of basins. Urban watersheds are particularly vulnerable to flash flooding due to the high percentage of impervious surface (low permeability), such as roads, roofs, sidewalks, and parking lots (Konrad, 2003). However, rainfall-runoff relationships in developed areas are highly complicated because of storm sewers and detention basins. Interbasin transfers are also possible, especially where storm and sanitary sewer systems are combined.

1.4. Data sources

Meaningful water balance calculations require accurate, long-term discharge and meteorological records. The US Geological Survey (USGS) currently maintains nearly 8,000 real-time gauging stations that monitor stage and/or discharge of streams and rivers in the United States (Wahl et al., 1995). The monitored watersheds vary greatly in size, climate, lithology, land use, engineering modifications, and other anthropogenic impacts, and thus a huge and diverse database is available for analysis. Long-term records of annual and monthly discharge are available online for many of these sites (USGS, 2012a). To avoid confusion and to simplify correlations with rainfall records, in the following discussion and diagrams we always use calendar years, not the USGS "water year."

Annual and monthly precipitation data were obtained from several National Oceanic and Atmospheric Administration weather stations (NOAA, 2012). The closest weather station to a given basin that had essentially complete records was used to calculate the average precipitation for the catchment. To evaluate the influence of land use and vegetative cover on ET, land use/land cover GIS data from the 2006 National Land Cover Database (USGS, 2012b) were used. The catchment area above each discharge gauging station was calculated and percentages of each type of land use/land cover were generated using ArcGIS 10 software.

1.5. Hydrologic setting of the meramec river basin

The Meramec River, which drains a 10,300 km^2 area of east-central Missouri, USA (Fig. 1), has many special characteristics that render it optimal for the study of ET rates and runoff generation processes, so special reference is made to it below. First, the Meramec River is one of the few remaining large, unimpounded rivers in the United States, as it has been spared from the engineering works and flood management practices found on practically all other waterways in the United States (Jackson, 1984; Ruddy, 1992). Second, water balance relation-

ships are further simplified because the basin has a low population density and negligible withdrawals. Third, very long (46 to 90 years) discharge records are available for eight gauging stations in the basin, and additional sites were monitored for a shorter interval or intermittently. Finally, the high accuracy of the available discharge data can be quantitatively established, justifying their use in making reliable assessments of ET rates in the different subbasins. In particular, the sum of annual flows for the three major subbasins matches that for the downstream station, when a minor adjustment for the evident difference in areas is made; the error is less than 2% (Fig. 2). For all the sites, the average annual runoff is around 32%, indicating that 68% of the precipitation has been removed from the system. This is similar to the average annual ET rate for the continental United States, which is close to 66% of the annual rainfall (US Water Resource Council, 1978).
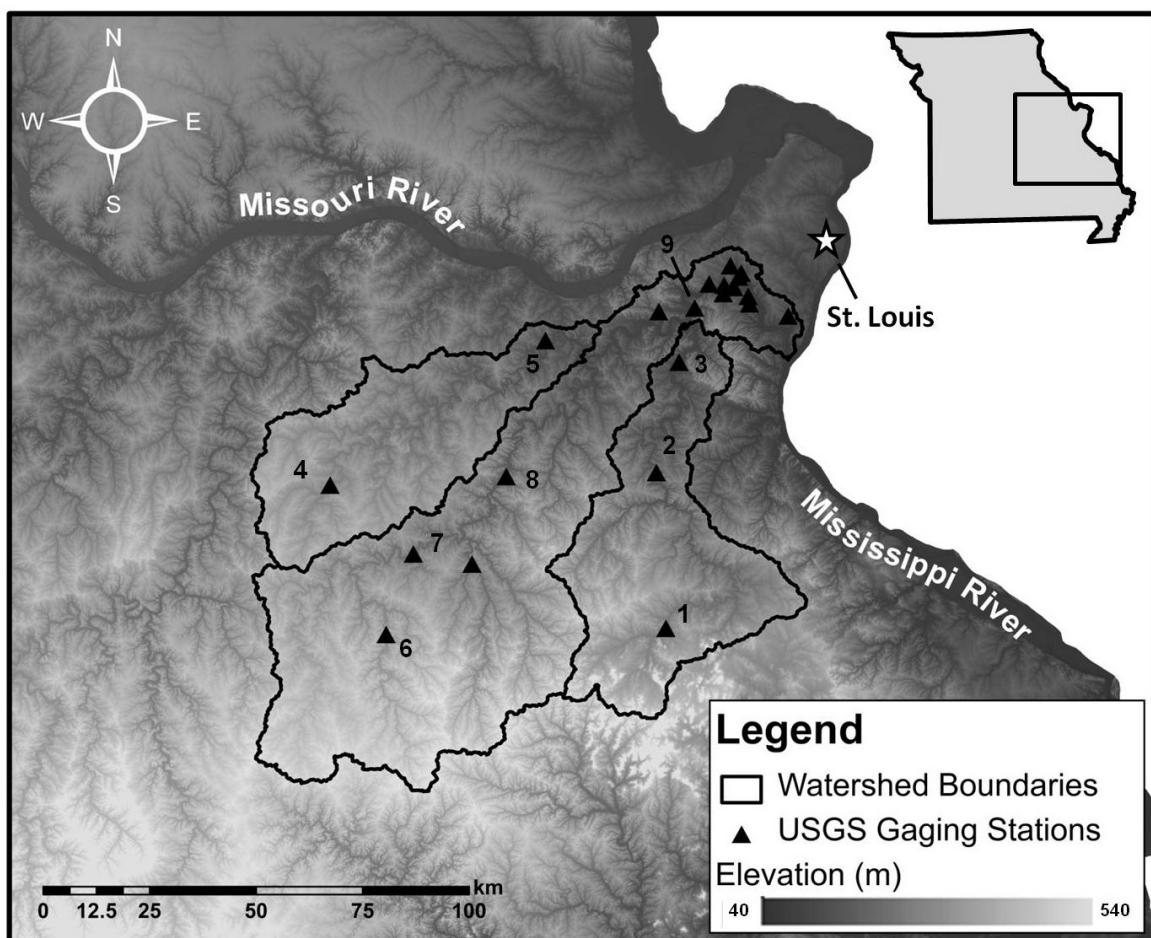

Figure 1. Shaded relief map of east-central Missouri, USA showing the whole Meramec River basin (10,300 km^2) and the Bourbeuse River (2,183 km^2) and Big River (2,513 km^2) subbasins; the City of St. Louis (open star) is shown for reference and the closed triangles are USGS gauging stations in the basin. Selected gauging stations are labeled (1 = Big River at Irondale; 2 = Big River near Richwoods; 3 = Big River at Byrnesville; 4 = Bourbeuse River near High Gate; 5 = Bourbeuse River at Union; 6 = Meramec River at Cook Station; 7 = Meramec River near Steelville; 8 = Meramec River near Sullivan; and 9 = Meramec River at Eureka). The elevation in the map area ranges from 390 m in the southwest to 110 m along the Mississippi River in the southeast; digital elevation model (DEM) base map data generated by the USGS are provided by MSDIS (2012).

The Meramec River and its two main tributaries, the Bourbeuse and Big Rivers, flow generally north and northeast until joining the Mississippi River south of St. Louis (Fig. 1). The flow pattern is asymmetrical as the basin lies on the northeastern flank of the Salem Plateau in east-central Missouri (Fenneman, 1938), and includes the foothills of the Ozark Mountains. Relief is greatest in the south and gradually decreases northward into the rolling hills of the Bourbeuse River subbasin. The unconfined Ozark aquifer crops out throughout the area and predominantly consists of lower Paleozoic dolostone and limestone units that underlie thin soils (Imes & Emmett, 1994). The basin features diverse karst topography including many springs, losing and gaining streams, 'swallow holes,' and sinkholes, which allow rapid connection between surface water and groundwater reservoirs (Vandike, 1995). Recharge occurs exclusively through infiltration of rainwater, with annual precipitation averaging ~ 100 cm and monthly totals for April and May usually exceeding 10 cm.

In addition to its proximity to the unimpounded Meramec River, the St. Louis region is optimal for the study of ET phenomena because discharge data are also available for a diverse suite of small streams. In particular, the USGS currently maintains 39 gauging stations in the City of St. Louis and St. Louis County (USGS, 2012a) that quantify discharge in watersheds that vary in area from 0.65 to 215 km², and include urban, industrial, commercial, residential, agricultural, and rural forested land use. Discharge records for most of these sites span 5 to 15 years.
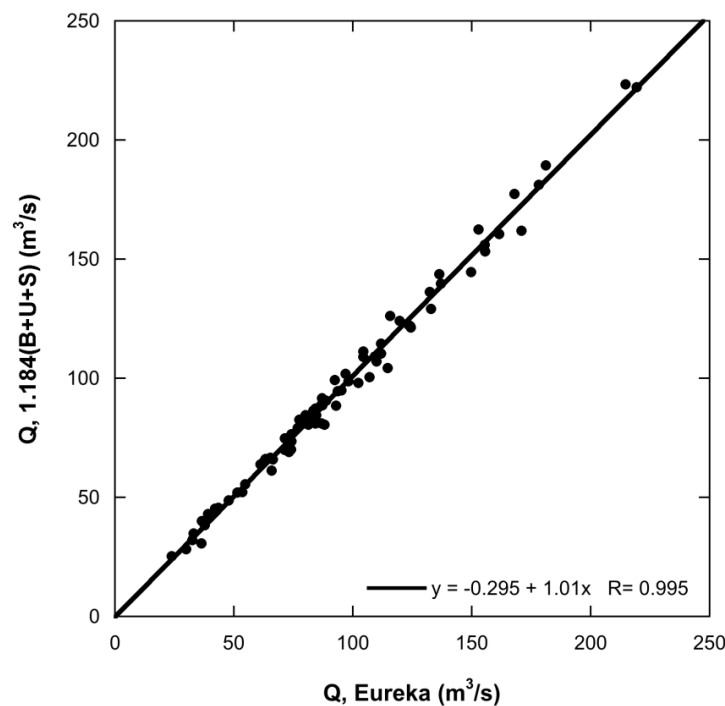

Figure 2. When multiplied by 1.184, the combined annual discharge for the Big River at Byrnesville (B; 2,375 km²), Bourbeuse River at Union (U; 2,093 km²), and upper Meramec River near Sullivan (S; 3,820 km²) compares very closely with the discharge of the lower Meramec River measured at Eureka (9,811 km², see Fig. 1). The factor of 1.184 represents the quotient of the relevant respective basin areas, 9,811 km²/(2,375 km² + 2,093 km² + 3,820 km²). The unit slope, small y-intercept, and high correlation coefficient of the regression line attest to the high quality of these discharge data, which are continuous at all four sites since 1923 except for Sullivan from 1933 to 1943.

2. Results

The observed difference between rainfall delivered to the watershed and the resultant stream discharge can be used to determine the ET rate. For a river with negligible withdrawals or out-of-basin gains or losses, and over a sufficiently long interval, eq. 2 applies and can be rewritten as:

$$R = P - ET \qquad (3)$$

This equation provides an important means of determining the average ET by simply subtracting the long-term mean value of runoff from that of precipitation. The equation also suggests a straightforward graphical procedure for determining ET, namely plotting observed discharge vs. observed precipitation for different months or years and determining ET from the y-intercept. Multiple complications and sources of confusion interfere with the latter approach.

2.1. Units of measure

Use of eq. 3 requires attention to the relevant units of measure. Because precipitation in the basin is measured as meters delivered over a specified interval of time, then both runoff and ET must be expressed in the same units for the same area and over the same time interval. Thus, the relevant runoff quantity, R, can be determined from the total discharge volume (in m^3) flowing out of the watershed over the specified time interval, divided by the contributing basin area (in m^2). The resultant unit, for example m/s, appears to be a "rate," but is actually a volumetric flux of water, $m^3/m^2/s$. All quantities in eq. 3 will therefore reduce to units of "rates," yet they actually all represent volumetric fluxes. Thus, the parameters in eq. 3 must all be expressed as cm/s, m/d, or m/y, etc., as is internally consistent and appropriate for any given case. This elementary matter has caused much confusion over the years, particularly because precipitation is normally reported and casually understood in length units, such as "cm" or "m" of rain. Less commonly precipitation is reported as a rate, (e.g., "cm/d," "m/y," etc.), which at least conveys acceptable units, yet precipitation is almost never considered as a volumetric flux, which is, in fact, what it actually represents.

2.2. Runoff vs. precipitation plots

Eq. 3 suggests that a graph of runoff vs. precipitation data will conform to a simple linear relationship with a negative y-intercept. A unit slope is expected for that relationship, because taken at face value, eq. 3 suggests that $\partial R/\partial P = 1$. In other words, in elementary algebraic parlance, where linear equations are understood to have the form $y = mx + b$, it appears that m in eq. 3 is equal to unity, giving $y = x + b$. Indeed, Fredrickson (1998) demonstrated that plots of discharge vs. precipitation for different parts of the Meramec basin define linear relationships with negative y-intercepts. However, when discharge is reported in appropriate flux units, the dimensionless slopes are only about 0.5 ± 0.2, depending on the site. Annual data for the Meramec River at Eureka and for a tributary near High Gate exemplify this result (Fig. 3).
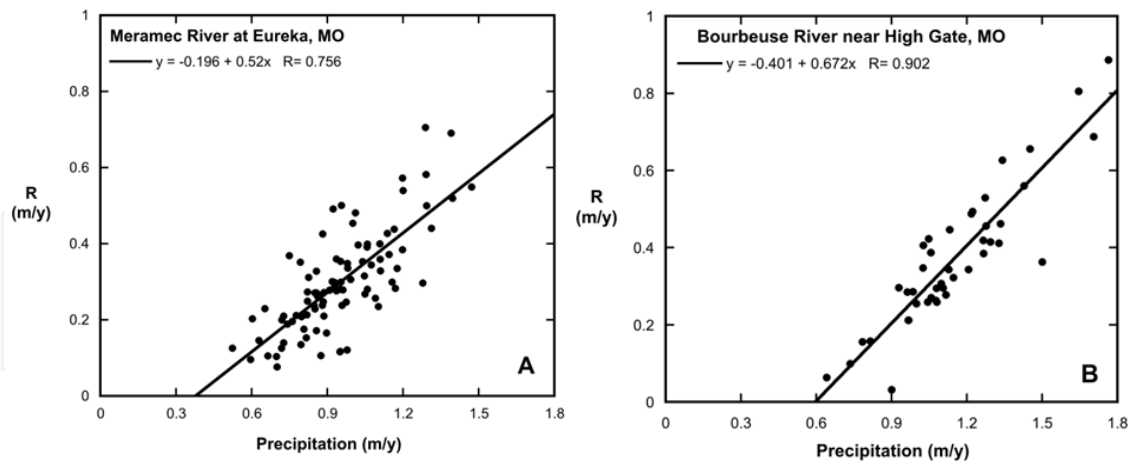
Figure 3. Graph of (A) runoff for the Meramec River at Eureka, Missouri vs. annual rainfall at St. Louis, Missouri for the years 1922 to 2011 and (B) runoff for the Bourbeuse River near High Gate, Missouri vs. annual rainfall at Rolla, Missouri for the years 1966 to 2010. Note that the slope and x-intercept are higher at the High Gate station than at the Eureka station.

In fact, for all basins we have examined, a linear correlation with a negative y-intercept is obtained when basin runoff (R) is plotted against precipitation (P), but unit slopes are never observed. Eq. 3 must be modified to:

$$R = mP + b \tag{4}$$

where m is the observed slope. Note that the dimensionless quantity m represents the fraction of the rainfall in excess of the x-intercept that becomes runoff. Moreover, because $m \neq 1$, b cannot represent negative ET.

Graphs of runoff vs. precipitation for numerous temperate-zone basins display a range of slopes (mostly 0.6 ± 0.25), and have positive x-intercepts that equal $-b/m$. The positive x-intercepts indicate that a certain amount of precipitation is "lost" every year before runoff is generated, equaling approximately 35 ± 20 cm in Missouri. Possible reasons for this "lost" rainfall are evaluated below and include ET processes and groundwater recharge. Once this fixed amount of water is removed from the basin, the river channel then receives a fixed fraction of the remaining rainfall budget, which is represented by the slope of the line. However, even this "excess" rainfall is still subject to additional ET losses, as there is not a 1:1 slope between R and P. This result shows that there are at least two components to ET.

2.2.1. Physical meaning of b

The fundamental reason that $m < 1$ is that ET is a function of P. For this reason, eqs. 1 – 3, although clearly correct and almost universally invoked in hydrologic literature, are highly misleading. Despite all appearances, eq. 3 does not indicate that $y = (1 \times x) + b$, rather it states that $y = x + f(x)$, where $f(x)$, representing the ET, is an unknown function of precipitation. Fortunately, this function is readily determined from available data.

Comparison of eqs. 3 and 4 shows that:

$$ET \;=\; (1-m)P \;-\; b \tag{5}$$

In effect, ET is seen to consist of two components, one that depends on the amount of precipitation delivered, equal to *(1 − m)P*, and the second (*–b*) that is independent of precipitation. In other words, the second component represents the base ET that exists even when the precipitation delivered is zero. It is logical, though not perfectly accurate, to conceptualize the first ET component as physical evaporation effects on standing water, and the second as plant transpiration. Practically no physical evaporation can occur when wet surfaces are absent, yet even in times of little or no rainfall, plants would wilt and die if they were unable to extract large quantities of moisture from soils. Thus, a somewhat simplistic but conceptually useful interpretation of eq. 5 is:

$$ET \;=\; (1-m)P \;+\; ET_0 \tag{6}$$

where *(1 − m)P* mostly represents physical evaporation but includes "excess" transpiration, and ET_0 is the minimum allowable transpiration, numerically equal to *–b*.

In the general case where eq. 1 applies, the right hand side of eq. 6 would need to include the terms $\pm \Delta W$ and $\pm \Delta S$, and the latter quantities could both be functions of *P*, which would greatly complicate interpretations. In what follows, we find it convenient to refer to two equations that are only slightly more complicated than eq. 6, namely:

$$
\begin{aligned}
ET \;&=\; (1-m)P \;+\; (ET_0 \pm \Delta W) \quad \text{(a)} \\
&\text{and} \\
ET \;&=\; (1-m)P \;+\; (ET_0 \pm \Delta S) \quad \text{(b)}
\end{aligned}
\tag{7}
$$

where $\pm \Delta W$ and $\pm \Delta S$ are not neglected but are viewed as constants. We will show that the form of *–b* in eq. 7a is helpful in analyzing annual runoff and precipitation data, while the form of *–b* in eq. 7b is helpful in interpreting monthly data.

Eqs. 5 – 7 suggest another procedure for estimating ET, which would be to graph the quantity *P – R*, representing the estimated ET, directly against *P*. Ideally, the slope *m** on this graph would equal the quantity *1 – m*, and the y-intercept *b** would equal *–b*. Although this construct has certain advantages for visualizing ET, it masks an induced correlation between the x- and y-axes. This defect is considerable because the plotted variables on such a graph are not linearly independent. Among other problems, random errors in the measurements would lead to a systematic overestimation of the slope *m**. In contrast, when *R* is plotted directly against *P* as in Fig. 3, the y- and x-axes represent entirely different, independent variables.

2.2.2. Mean annual ET

Many direct measurements quantify evaporation rates from pans (Farnsworth & Thompson, 1982), and far fewer measurements quantify ET rates using lysimeters (e.g., van Bavel, 1961). Some germane examples are shown in Fig. 4; the data suggest that the annual ET rate in the eastern USA is ~ 0.8 m/y, while the pan rate is ~ 1.3 m/y for the indicated sites. Of course, these rates depend on location and they vary year to year, but as a rule of thumb, ET appears to be about 63% of the annual pan rate.

It is useful to compare the total, mean ET from lysimeters (Fig. 4) with the long-term average for the Meramec basin, using eq. 5 and the regressions given in Fig. 3. For Eureka, the mean ET is 0.65 m/y given the average rainfall of 0.95 m/y for the relevant interval, and at High Gate, ET is 0.78 m/y given the mean precipitation at Rolla of 1.15 m/y for the relevant interval. Thus, over many decades in the Meramec basin, ET is about 68% of total precipitation while the complementary runoff is about 32%.

It is reassuring that the mean ET values secured for the Meramec basin are in reasonable agreement with the mean annual ET data provided by van Bavel (1961; Fig. 4) for several sites in the eastern USA. More importantly, eq. 5 provides a means of showing how ET depends on the annual precipitation. In particular, in years with the lowest observed precipitation, ET is observed to be > 90% of P, while in years having the most rainfall, ET can be < 60% of P.
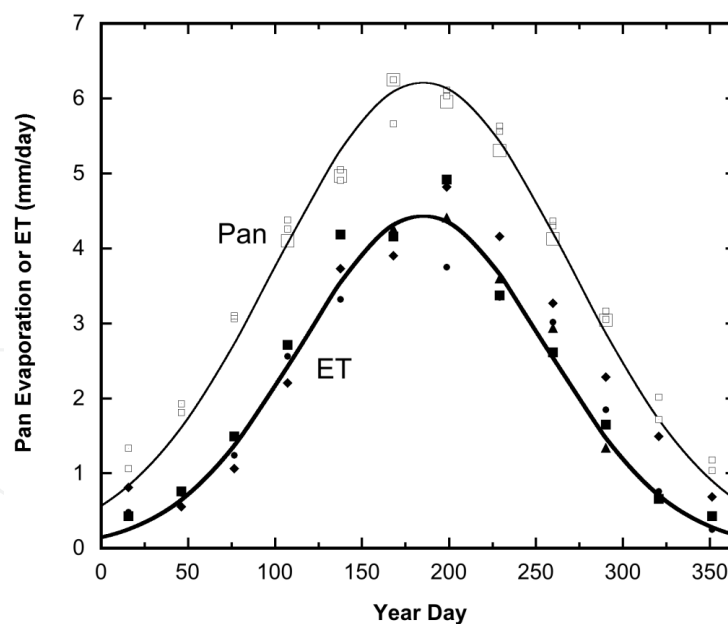

Figure 4. Pan evaporation (open symbols) from Weldon Spring, Missouri (large square) and from Russellville and Stuttgart, Arkansas (small squares), compiled in Farnsworth & Thompson (1982). Weldon Springs pan data were not collected during winter and early spring, likely due to freezing, but data for the available months is similar to that measured at the selected Arkansas sites, which offer complete annual coverage. ET data (solid symbols) are from van Bavel (1961) for Coshocton, Ohio (square); Seabrook, New Jersey (diamonds); Raleigh, North Carolina (triangles); and Waynesville, North Carolina (circles). The data approximate simple bell curves (solid lines), notably $E_{pan} \sim 6.22$ $Exp\{-0.00007(Year\text{-}Day - 185)^2\}$, and $ET \sim 4.44\ Exp\{-0.0001(Year\text{-}Day - 185)^2\}$; modeled after Criss & Winston (2008).

2.2.3. Seasonal behavior of ET

Theoretically, the seasonal behavior of ET can be defined from graphs of runoff vs. precipitation constructed for each month; however, several matters interfere with this approach. The ET determinations require that the change in groundwater storage over the interval of interest is small, a condition much more likely to be realized over an annual cycle than over a short interval. Consequently, large basins are not well suited for monthly analysis. For example, the lower Meramec basin would be a poor choice, as it features considerable seasonal variations in groundwater storage, indicated by its hydrologic residence time of ~ 3 months determined from oxygen isotope data (Frederickson & Criss, 1999). Small basins are much more likely to have short storage constants, but few have long records and they tend to not be gauged as accurately as large basins; a result of their flashy nature. Due to such problems and short-term weather vagaries, the monthly regressions at individual sites were found to have rather low correlation coefficients, causing uncertainties in the slopes and y-intercepts.

Nevertheless, in an attempt to define monthly relationships between runoff and precipitation, we searched for small basins in Missouri and Illinois that have long-term records and proximal meteorological stations. Those selected for examination in Missouri include the Bourbeuse River near St. James (60 km^2; USGS #07015000) and at High Gate (350 km^2; USGS #07015720), Little Beaver Creek (16.6 km^2; USGS #06931500) near Rolla, and in Illinois include Indian Creek (95 km^2; USGS #05588000) at Wanda, Asa Creek (20.8 km^2; USGS #05591500) at Sullivan, and Farm Creek (71 km^2; USGS #05560500) at Farmdale.

Fig. 5 shows the monthly variations in slope and y-intercept for these particular sites. In each case, the variations are "noisy" over an annual cycle, but taken as a group, the variations show systematic behavior. The slopes (Fig. 5A) are steepest in the cold months, consistent with low physical evaporation and enhanced runoff due to frozen ground, and are much smaller during the hot, sunny, dry months when physical evaporation is high. When the data are inverted or graphed as $1 - m$, they feature a "humped" pattern, but one that is skewed compared to the pan and ET data plotted in Fig. 4. This shape probably reveals a key characteristic of physical evaporation disclosed by this analysis. In particular, due to leaf growth, the effective surface area of a basin late in the year is much greater than that early in the year. This effect could produce the asymmetry in the curve because physical evaporation rates depend on surface area, which changes over an annual cycle in natural settings, but not in an evaporation pan. Similar seasonal asymmetry has been calculated by Hoskins (2012).

Fig. 5B shows the monthly variations in y-intercept for these particular sites. The y-intercepts are near zero during winter, when plants are inactive, confirming the expectation that this term is related to ET_0 as implied by eqs. 6 and 7b. This expectation is also consistent with the strongly negative y-intercepts during spring, when plant growth is rapid. Surprisingly, the y-intercepts resume low values during the hottest months, followed by a second minimum in fall, then recover to low values with winter's approach, defining a "w-shaped" pattern over an annual cycle. In the latter half of the year, it is useful to interpret $-b$ using eq. 7b. During the hot summer months when large fractions of precipitation are lost to physical evaporation, plants strive to conserve water, and much of the transpired water is derived from soils, which subsequently dry out. In fact, it appears that ET_0 and ΔS approximately offset each other during summer,

so the intercept representing their sum is small (eq. 7b). In fall when plant activity is greatly reduced and ET_0 is small, rainfall replenishes dry soils before runoff is generated, so this second minimum represents the "repayment" of soil moisture (ΔS) losses incurred during summer.
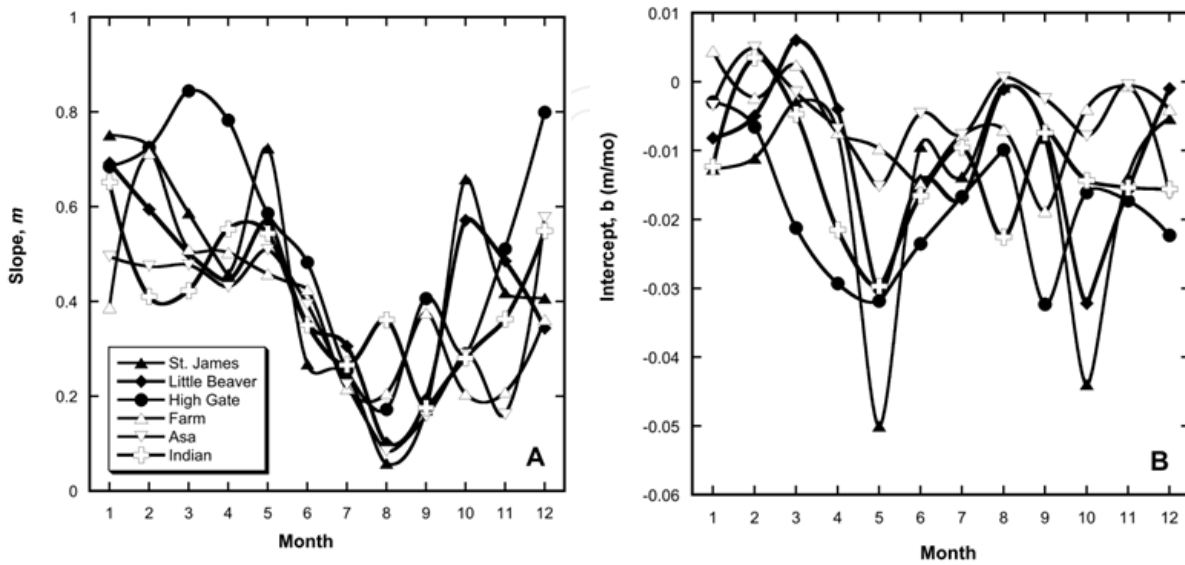

Figure 5. Seasonal variations of (A) slope m and (B) intercept b determined from monthly regressions using eq. 4, for several small watersheds in Missouri (solid symbols) and Illinois (open symbols) with extensive records..

3. Interbasin comparisons

On an annual basis, the total rainfall delivered to geographically proximal basins is similar, and changes in storage are rather small. This feature can be exploited to directly compare the runoff of proximal river basins to define differences attributable to land use disparities or due to out-of-basin transfers (eq. 7a). Consider that two proximal basins, denoted by subscripts 1 and 2, have annual runoff vs. precipitation regressions that can be expressed in the form of eq. 4:

$$R_1 = m_1 P_1 + b_1 \text{ and } R_2 = m_2 P_2 + b_2 \tag{8}$$

If P_1 and P_2 are similar for these basins, a direct consequence is that:

$$R_2 = \left(m_2 / m_1\right) R_1 + \left\{\left(m_2 / m_1\right) b_1 - b_2\right\} \quad \text{(a)}$$

So:

$$\partial R_2 / \partial R_1 = m_2 / m_1 \quad \text{(b)} \tag{9}$$

Eqs. 9a and 9b provide a means to directly compare basin runoff and to gather insights about their relative amounts of ET (Table 1). Importantly, note that the slopes and y-intercepts on such diagrams are *not* the same as *b* and *m* derived from runoff vs. precipitation plots (e.g., eqs. 9a and 9b).

Site	USGS Site Number	Area (km²)	Altitude‡ (m)	Intercept† (m/y)	Slope†	r*(m,b)	Years of Data
Irondale	07017200	453	229.6	-0.004	1.170	0.919	46
Richwoods	07018100	1,904	159.4	+0.002	1.079	0.967	62
Byrnesville	07018500	2,375	132.2	+0.002	1.079	0.970	89
High Gate	07015720	350	244.5	-0.016	1.248	0.938	46
Union	07016500	2,093	148.9	-0.009	1.057	0.961	90
Cook Station	07010350	515	263.5	-0.019	0.881	0.963	18
Steelville	07013000	2,023	207.8	-0.001	0.896	0.958	89
Sullivan	07014500	3,820	177.3	+0.003	0.951	0.978	79
Eureka	07019000	9,811	123.2	0.000	1.000	1.000	92

‡ Altitude of gauging station.

† Note that these slopes and y-intercepts are not the same as m and b (e.g., see eq. 9b).

r* is the correlation coefficient.

Table 1. Regression lines for runoff in numerous subbasins of the Meramec River compared to runoff from the lower basin measured at Eureka.

3.1. Interpretation of slope on Runoff vs. Runoff plots

In theory, a runoff vs. runoff plot would provide a complete characterization of ET effects in basin 2 if those effects in basin 1 were fully characterized, i.e., if m_1 and b_1 are known. This supposition requires that *P* is highly similar in the basins being compared. For example, any systematic, proportional differences in *P* would directly factor into the slope on this diagram. Interpretations are best for small, proximal watersheds, and even this restriction may be insufficient in mountainous areas where *P* varies strongly with altitude, rain shadow effects, etc.

Given this caveat, on runoff vs. runoff plots, approximately unit slopes indicate that the physical ET losses in basin 2 are similar to those of basin 1, whereas low slopes (< 1) indicate that the losses for basin 2 exceed those of basin 1, and high (> 1) slopes indicate the opposite. An example is shown for two subbasins in the Meramec basin, one with considerable pastureland and the other dominated by forested land, which are compared to the main stem of the upper Meramec River that in all key aspects (see Table 2) has intermediate character (Fig. 6).
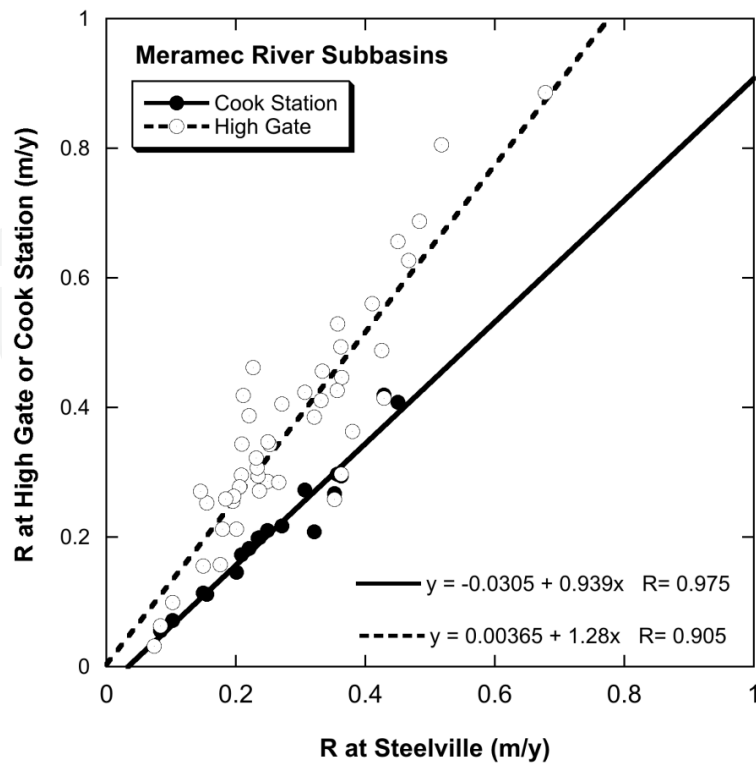
Figure 6. Comparison of runoff from three subbasins in the upper Meramec River basin. Compared to the reference basin near Steelville, the upper Bourbeuse River above High Gate (open circles) has much more pastureland, while the upper Meramec River above Cook Station (closed circles) is dominantly forested. Note that the slope for High Gate is > 1 but that for Cook Station is < 1, indicating that forests have high physical ET. The y-intercepts are small, but the intercept for Cook Station is clearly negative.

3.2. Interpretation of y-intercept in Runoff vs. Runoff plots

It is both expected and observed that y-intercepts are normally small on runoff-runoff plots (e.g., Fig. 6). In fact, the errors in the regression equations may normally overwhelm any small actual differences in the ET_0 values of the watersheds that would dominate the magnitude of this quantity (eqs. 6 and 9a). However, in cases where significant storage effects occur or where there are large transfers of water into or out of a watershed, the y-intercepts can be large and significant. In such a case, ΔW in eqs. 1, 2, and 7a cannot be neglected.

Transfers of water between proximal subbasins are affected by elevation. That is, high areas tend to lose water to the groundwater system that flows to regions of lower head and normally, but not necessarily, is discharged at lower elevation along the same stream. Data for numerous gauged sites in Missouri show the tendency for high altitude subbasins to have below average runoff, illustrating this effect. For example, compared to lower basin runoff near Eureka, the upper subbasins of the Meramec River (Irondale, High Gate, Cook Station, and Steelville) all have negative y-intercepts relative to the downstream site near Eureka (Table 1).

An extreme example of an interbasin transfer is provided by the Chicago Sanitary and Ship Canal in Illinois. Most surface waters and all wastewaters in the Chicago region flow away

from, or are diverted away from, Lake Michigan into the Illinois River system. Ostensibly, the area of the contributing watershed at the canal gauging station is 1,914 km^2; however, this canal also receives wastewater discharge from the Stickney Treatment Plant, whose average output of ~ 6 million m^3/d ranks it as the world's largest. That output, representing an average of ~ 70 m^3/s, primarily represents water originally drawn from Lake Michigan that is subsequently treated to provide the municipal water supply of Chicago. Following use and then cleanup at Stickney, all this water is diverted from the Great Lakes watershed into the Mississippi River watershed, via the canal.

Fig. 7 compares runoff for the Chicago Sanitary and Ship Canal (USGS #05536995) to that of the Vermilion River in east-central Illinois. The regression is poor because approximately 75% of the flow in the canal is treated wastewater, derived from outside the basin. Nevertheless, that is the relevant point. The y-intercept in this case is huge, greatly exceeding the total rainfall normally delivered to this "watershed," and its value independently quantifies the total, average, man-made contribution to the canal's flow as ~ 80 m^3/s.
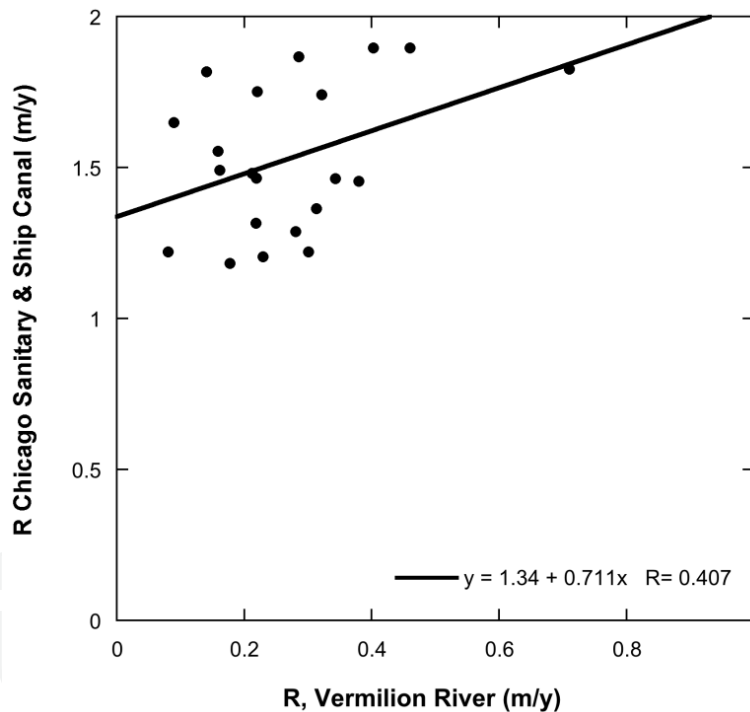

Figure 7. Comparison of runoff in the Chicago Sanitary and Ship Canal (USGS #05536890) to that in the Vermilion River at Pontiac, Illinois (USGS #05554500). The huge y-intercept represents the artificial transfer of ~ 80 m^3/s of water into the canal.

3.3. Land use and ET

Different land use practices have different effects on the ET and runoff rates. For example, it is well known that conversion of forested areas to urban or agricultural areas causes increased runoff (Bosch & Hewlett 1982). This effect is due to the reduction in ET in urban

environments because of the decreased vegetative coverage. As noted above, in forested areas physical evaporation is also enhanced by leaf area, due to "interception" of precipitation. Crops and grasses can also intercept and transpire rainwater, but at lower rates than in forested ecosystems.

Land use data were compiled for all the subwatersheds in the Meramec basin (Table 2). The data confirm that the more heavily forested the subbasin, the lower the runoff (Fig. 8). Subbasins with higher percentages of pasture/hay and cultivated crops had increased runoff relative to forested areas; a result of the smaller surface area from which water is transpired by these grasses and smaller plants. However, urban land use had the largest impact on ET rates and strongly increased runoff when compared to forest.

Station	Open Water	Developed	Barren Land‡	Forest	Wetland	Grass-land	Scrub/ Shrub	Cultivated Crops	Pasture/ Hay
Irondale	0.50	3.10	0.02	70.99	0.21	1.81	0.43	0.02	22.92
Richwoods	0.56	6.03	0.34	72.19	0.40	2.71	0.30	0.14	17.33
Byrnesville	0.63	6.39	0.29	72.38	0.54	2.57	0.24	1.00	15.96
High Gate	0.43	7.72	0.08	50.74	0.30	2.44	0.00	0.58	37.71
Union	0.49	6.30	0.07	56.51	0.61	2.04	0.09	2.08	31.81
Cook Station	0.21	3.47	0.06	77.39	0.34	1.81	0.01	0.09	16.62
Steelville	0.18	4.95	0.04	65.98	0.34	2.25	0.00	0.12	26.14
Sullivan	0.22	4.45	0.11	73.16	0.33	2.16	0.88	0.08	18.61
Eureka	0.40	4.94	0.16	73.56	0.46	2.06	0.70	0.66	17.06
All Basins	0.56	7.35	0.21	67.01	0.56	2.18	0.35	1.33	20.45

‡Rock, sand, and clay.

Table 2. Percentage of various types of land use in the different subbasins of the Meramec River watershed.

Another important factor that dramatically affects ET is bedrock geology. Distinct trends in the slope (Table 1) and the land use (Table 2) were observed in the western, shale-rich subbasins and the eastern, carbonate-hosted subbasins (Fig. 8).

3.4. Runoff dynamics of small and urban watersheds

Many have argued that impervious surfaces such as buildings, roads, parking lots, and other structures enhance runoff because these structures prevent water from infiltrating. Consequently, surface runoff is directly conveyed into stream and river channels. St. Louis is ideal for such study given the large number of small watersheds that are gauged. Examination of runoff relationships in all 39 small, gauged basins in City of St. Louis and St. Louis County revealed many large and sometimes inexplicable differences. Surprisingly, the area-weighted average runoff from all these basins was only slightly higher (~ 35%) than that for the Meramec basin (~ 32%). This result reveals a major complication. In urban areas, ΔW in the water balance equations can be highly important, as storm sewers can cause large intrabasin and interbasin
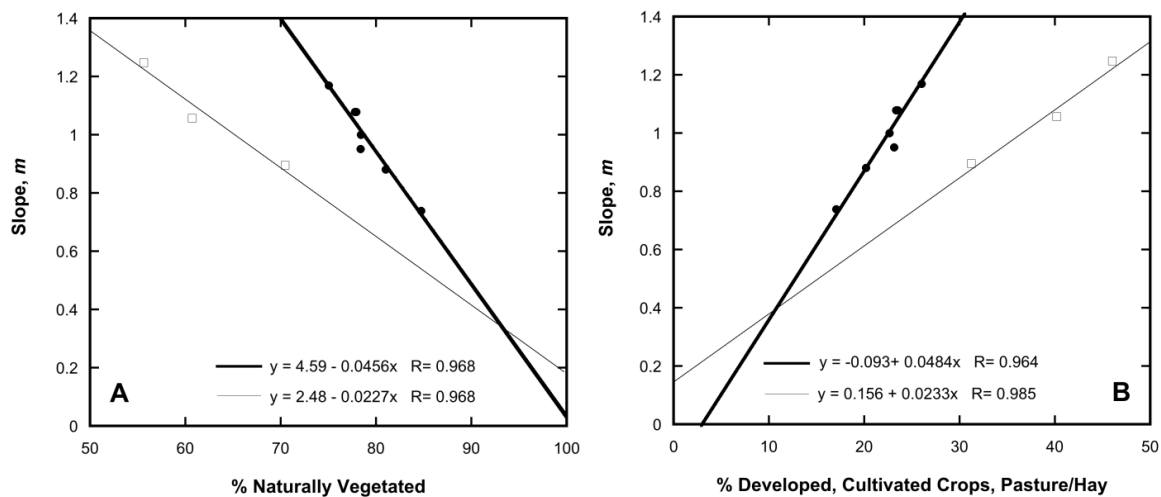
Figure 8. Plots of slopes (Table 1) vs. land use (Table 2) for the Meramec River basin (Fig. 1). (A) Demonstrates that the more heavily forested subwatersheds have the lowest runoff, while (B) shows that the urban development, cultivated crops, and pasture/hay increase the percentage of runoff. A distinct difference was observed between the behavior in the eastern (closed circles) and the western (open squares) portion of the watershed, attributed to lithologic differences.

transfers. This situation is particularly magnified by combined sewer systems, where stormwaters and sanitary discharges enter the same pipes, ultimately to be treated and released into major rivers. Moreover, in many residential areas, lawn irrigation can supply the equivalent of several inches of rainfall during summer months (Spronken-Smith & Oke, 1998).

We found the runoff-runoff plot to be particularly useful in interpreting discharge data in these small, moderately to intensely developed watersheds. The predominantly residential Creve Coeur watershed was selected as the reference basin as it was relatively large and behaved similarly to several other small watersheds in the area. Each graph in Fig. 9 contrasts runoff from Creve Coeur Creek (hereafter, CCC) to runoff from selected watershed pairs that display contrasting characteristics.

Fig. 9A compares runoff from the Kiefer Creek watershed and the Fishpot Creek watershed to CCC. The y-intercepts are large but one is positive and the other is negative. This feature exemplifies the large and opposite values for ΔW; in this case caused not by storm sewers, but by karst groundwater flow. That is, these contrasting sites respectively represent gaining and losing stream reaches, whereas no large springs occur in the CCC basin. This comparison is compelling because land cover is predominantly residential in all three watersheds. Fig. 9B compares runoff from the Black Creek watershed, which has extensive areas of impervious surface, and the mostly forested Williams Creek watershed to CCC. The profound difference in land cover is clearly reflected in the slopes of the regression lines. The y-intercepts are probably also significant; Williams Creek is gauged below several significant springs, while Black Creek contains several combined sewer lines and combined sewer overflows, and is probably losing stormwater runoff to the Mississippi River, where treated stormwaters are discharged. Fig. 9C compares runoff from the upper River des Peres watershed near University City and the Sugar Creek watershed to CCC. The upper River des Peres basin is residential and commercial, while the Sugar Creek basin has low-density residential development, so the

former would be expected to have the highest slope, yet the opposite is seen. Both basins are underlain by shale-rich Pennsylvanian strata, so compared to the CCC watershed that has much more limestone, the slopes would be expected to be > 1. The likely cause of these disparate slopes is that the topographic relief in the upper River des Peres watershed is very low compared to CCC and Sugar Creek. Thus, the slopes of the trend lines suggest that the effect of topographic relief on runoff generation is more important than that of land cover in this case.
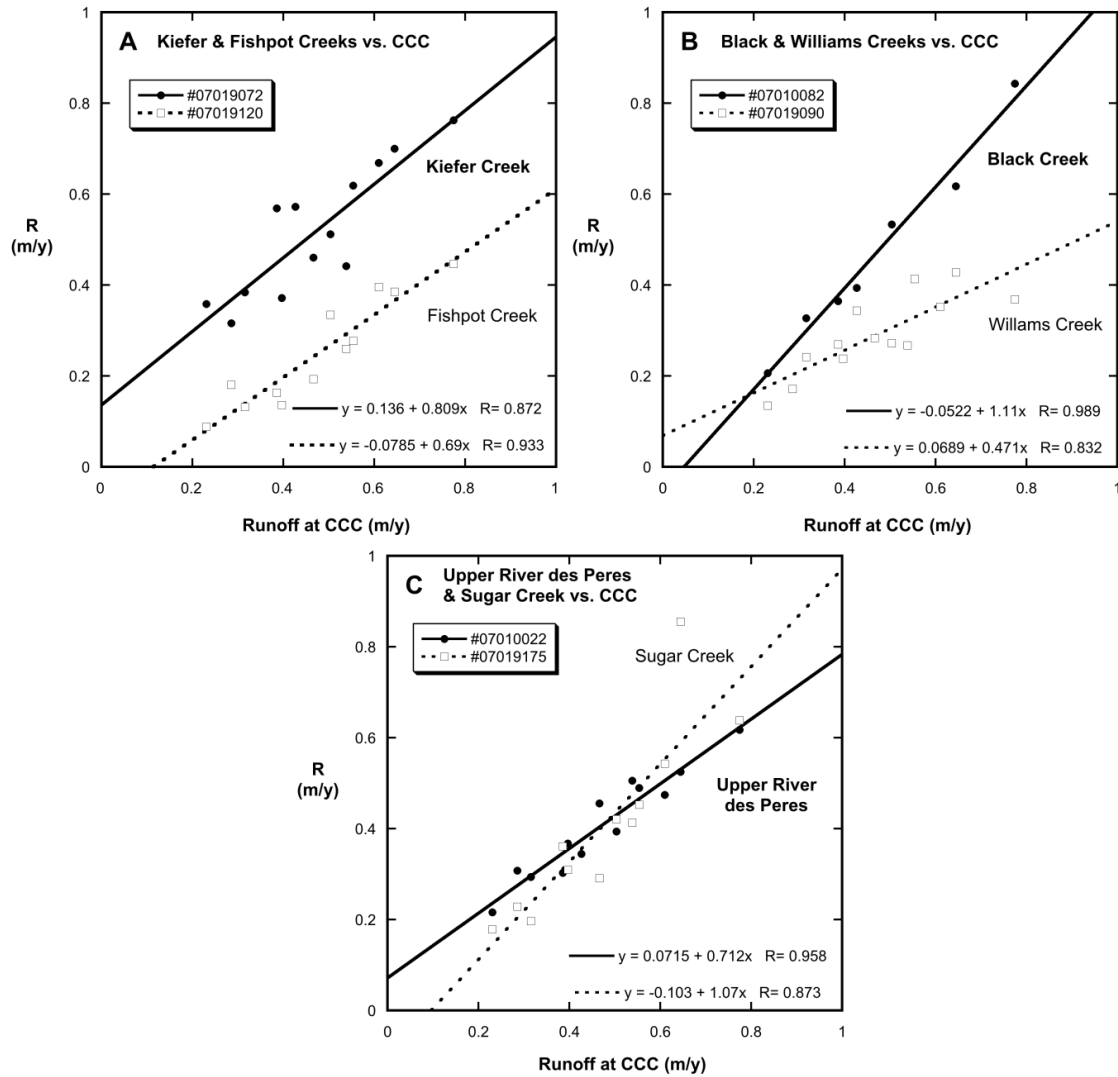

Figure 9. Comparison of runoff in several small watersheds to that of the 57 km^2 CCC watershed (USGS #06935890). (A) Runoff in the 10 km^2 Kiefer Creek watershed (closed circles) gauged just below a significant spring and that of the proximal 25 km^2 Fishpot Creek watershed (open squares) gauged just above a spring, both vs. CCC. Note the positive and negative effects of groundwater transfer on the y-intercept for these proximal watersheds. (B) Runoff in the 15 km^2 Black Creek watershed (closed circles) that has large areas of impervious surface and the mostly forested, 20 km^2 Williams Creek watershed (open squares) gauged below a spring, both vs. CCC. Note the effects of land cover on the slopes for these watersheds. (C) Runoff in the 23 km^2 watershed of the upper River des Peres watershed (closed circles) and that of the 13 km^2 Sugar Creek watershed (open squares), both vs. CCC, contrasting the effects of flat and steep topographic slopes.

4. Conclusions

The water balance equation provides an effective means to calculate the ET rate if long-term data precipitation and stream discharge are available for a given watershed. Simple estimates of ET can be made by subtracting the long-term mean values of runoff (R) from precipitation (P). Much richer information is provided by the fundamental graph, R vs. P, which depicts the relationship between these measured, independent variables over time intervals of interest. Annual data for R and P display good linear trends on this graph, but they show several surprising characteristics. First, a large x-intercept is seen on this graph, indicating that significant precipitation is "lost" before any runoff is generated; second, even after this demand is satisfied, the "excess" rainfall is subject to additional losses, so a 1:1 slope between R and P is not realized. We show how the y-intercept and slope of this plot can be used to deconvolve ET into different components that approximate physical evaporation and biological transpiration. We also show how the magnitudes of these quantities vary with basin character and throughout the year. We confirm standard expectations for runoff generation from different landscapes, such as high runoff fractions shed from impervious surfaces and low runoff generated by forested watersheds. However, over an annual cycle, we find that physical evaporation effects from natural basins are strongly skewed compared to the symmetrical, bell-shaped curves defined by pan data, an effect we attribute to changes in leaf area. Moreover, we show that short-term changes in soil moisture confuse monthly ET analyses.

Another graph, a direct graphical comparison of annual runoff from different, proximal basins, is very useful for estimating relative ET differences. More importantly, the y-intercepts on such plots both identify and quantify out-of-basin gains or losses of water. Such interbasin transfers can be very significant in karst areas due to groundwater flows, as well as in developed areas due to storm sewers, especially combined sewer systems.

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