2.6 INFERRING LEAF EMERGENCE AND ESTIMATING EVAPOTRANSPIRATION FROM EDDY COVARIANCE MEASUREMENTS AND RUNOFF RECORDS

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1. INTRODUCTION

For more than a decade, eddy covariance measurements of evapotranspiration (ET) have been made at Harvard Forest, a deciduous forest in central Massachusetts. The marked increase in ET that occurs at the time of leaf emergence, previously documented to lead to measurable changes in the boundary layer temperature, humidity, and cloud based on diurnal, synoptic, and seasonal time scales. Reforestation in the eastern United States over the last century led us to believe that long−term boundary layer state variations might also occur, but reliable data are not available. We resolved to link changes in ET over the first three time scales to streamflow and water table data, in advance of extending the analysis to decadal and longer scales.

2. BACKGROUND/METHODS

a. Seasonal time scale

At the seasonal time scale, Fitzjarrald et. al. (2001) inferred leaf emergence by analyzing the trend of the tendency Bowen ratio from mean temperature and humidity data given at surface climate stations. The enhanced ET which occurs at the time of leaf emergence results in a rapid decrease in the Bowen ratio. The small amount of the altered surface flux converges into the boundary layer, which is then detected in the daily increments of mean temperature and humidity at the surface.

An important link between ET and the water budget at the seasonal time scale occurs through expression of the water budget as follows:

$$E = P - R \pm S \tag{1}$$

where E is the total evaporation, which includes ET, bare soil evaporation, and evaporative losses through interception, P the precipitation, R the runoff, and S the storage term which includes soil moisture storage and snowpack. When dealing with long−term seasonal averages, the storage term has been commonly assumed to be zero, and will be done so here. Observations of average evaporation at Harvard Forest along with the water budget estimate of evaporation from nearby long−term record precipitation and runoff stations are shown in Figure 1.

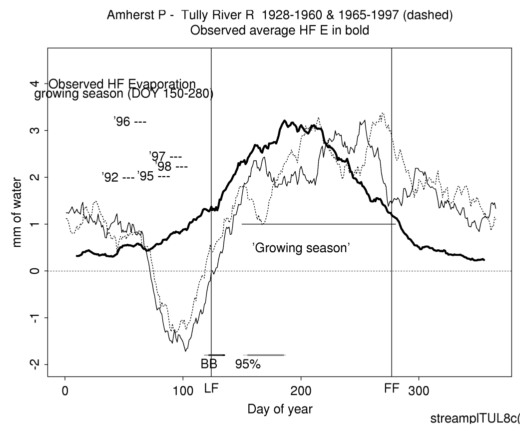

Figure 1: Observed Harvard Forest evaporation (bold) averaged from 1992 to 1998. Amherst, MA precipitation minus Tully River, MA runoff averaged from 1928 to 1960 (solid) and from 1965 to 1997 (dashed).

Note the good correspondence between the observed evaporation at Harvard Forest and the evaporation estimate given by precipitation minus runoff during the growing season (day of year 150 to 280). The climatological last freeze at Harvard Forest is given by the vertical line labeled LF at day of year 125. This occurs at the time of bud break at Harvard Forest (labeled BB) and at the time of rapid increase in evaporation. The water budget estimate of evaporation shows the effect of neglecting the storage term, as a melting winter snowpack around day 100 gives artificially negative values for evaporation. As streamflow recovers from the snowmelt pulse, runoff continues to decrease and evaporation increases concurrent with the presence of transpiring vegetation.

We identify three independent runoff characteristics that change with spring onset and leaf emergence. The first is the date of the return of the precipitation minus runoff curve to pre−snowmelt pulse values. This provides an indication of the transition from the snowmelt portion to the leaf−out portion of spring. The second characteristic lies in using daily runoff data to assess changes in streamflow recession following rainfall events. We would expect the streamflow recession time constant to decrease following leaf emergence due to enhanced ET. The third characteristic involves using 15−minute runoff data to assess seasonal changes in the amplitude of

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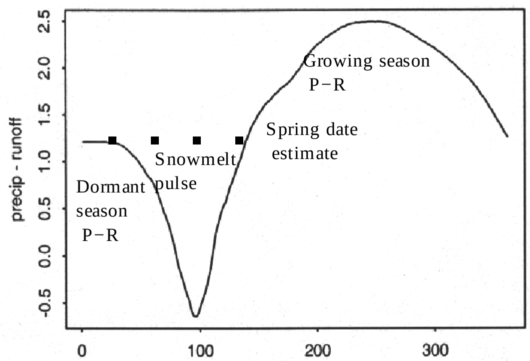
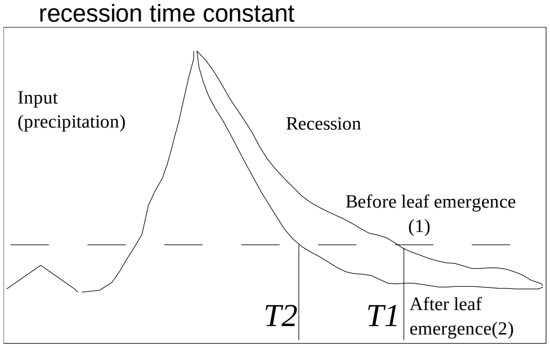
the diurnal streamflow signal observed on some watersheds. This will be discussed further in the next section. Figure 2 contains a diagram of the three methods.

a) Long−term average of annual P − R curve



b) Seasonal change in length of streamflow recession time constant



c) Seasonal change in amplitude of diurnal streamflow signal

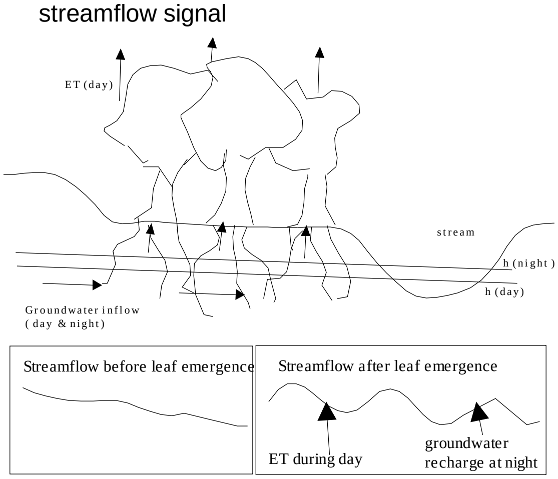


Figure 2: Three methods to detect leaf emergence from seasonal changes in runoff characteristics. The top panel shows the return of the long−term average P − R curve to pre−snowmelt pulse values. The middle panel shows the seasonal change in length of the streamflow recession time constant. The bottom panel shows the seasonal change in the amplitude of the diurnal streamflow signal.

b. Diurnal time scale

A well−defined diurnal signal in streamflow is frequently observed in small watersheds during the growing season under sufficiently dry conditions. Decreases in streamflow occur during the day due to transpiring vegetation, using some of the groundwater supply that composes the baseflow of the stream. At night, groundwater replaces some of the water that was transpired during the day. The water transpired by the vegetation is taken up by the roots through the capillary fringe by capillary action. As the water table drops, an increase in the pressure head allows for the inflow of groundwater.

For the process of the diurnal water table fluctuation, a simple water budget can be written for the watershed by continuity:

$$\frac{\partial}{\partial t}(\rho A_w h)=P+E+I-D \qquad (2)$$

where ρ is the density, h stands for the water table height averaged over the watershed, A_w the watershed area, P the precipitation, E the evaporation, I the inflow and D the drainage (outflow). Assuming constant density and dividing through by the watershed area, we get:

$$\frac{\partial h}{\partial t}=p(t)-e(t)+i-d \qquad (3)$$

Since the diurnal fluctuation is only observed during dry periods between precipitation events, we will simplify things by assuming the precipitation term is zero. We will approximate the evaporation term as the following:

$$e(t)=b\sin(\omega t)^{10} \qquad (4)$$

where b is the amplitude, and ω the diurnal cycle. Using the sine function raised to the tenth power for the evaporative forcing gives the best approximation to the observed evaporation. We can obtain an expression for the drainage term through Darcy's Law, expressed in one−dimensional form as:

$$u=-K\frac{dh}{dx} \qquad (5)$$

where K stands for the hydraulic conductivity of the soil, a measure of how well water can flow through the soil type. Substituting u back into D, and then dividing through by ρ A_w gives us:

$$d=-K\frac{A_d}{A_w}\frac{dh}{dx}\simeq-K\frac{A_d}{A_w}\frac{h}{d_w} \qquad (6)$$

where A_d is the drainage cross−sectional area, and d_w is the watershed depth. We can now introduce the time constant for the watershed, τ_w:

$$d=\frac{-h}{\tau_w} \qquad \tau_w=\frac{d_w}{K}\frac{A_w}{A_d} \qquad (7)$$

The time constant for the watershed is proportional to of the watershed volume ($A_w d_w$), and inversely proportional to the cross−sectional area of the drainage, and the hydraulic conductivity. In dealing with the inflow term, if we assume that both it and the time constant are constants, we can write $h_i = h - \tau_w i$.
Putting the terms back together, we can now write the following equation for the water table fluctuation:

$$\frac{\partial h_i}{\partial t} + \frac{h_i}{\tau_w} = -b \sin(\omega t)^{10} \qquad (8)$$

The integrating factor in this case will be exp(t/ τ_w), so we write:

$$\frac{d}{dt}\left(e^{\frac{t}{\tau_w}} h_i\right) = -b\, e^{\frac{t}{\tau_w}} \sin(\omega t)^{10} \qquad (9)$$

which yields:

$$h_i(t_f) = h_i(0) e^{\frac{-t_f}{\tau_w}} - b e^{\frac{-t}{\tau_w}} \int e^{\frac{t}{\tau_w}} \sin(\omega t)^{10}\, dt \qquad (10)$$

An analytical solution can be found to this equation, however we have not completed this task as of yet. We see that an exponential decay term will be in the solution as well as sine terms to account for the diurnal fluctuations. This model is similar in construction to the evapoclimatonomy model presented by Lettau (1969).

c. Synoptic time scale

Freedman et. al. (2001) showed through boundary layer synoptic composites done at Harvard Forest from 1995 to 1998 that the presence of a net radiation−boundary layer cumulus (BLcu) feedback ensured the appearance of BLcu on each day of a postfrontal sequence. The presence of BLcu provided favorable conditions for forest−atmosphere exchange by enhancing carbon uptake and water use efficiency. We have identified postfrontal sequences for which we plan on examining changes in the amplitude and phase of the diurnal streamflow signal as well as to develop an empirical relationship between the diurnal streamflow signal and ET. Also, these sequences are to be used for exercising our analytical model once it is solved in order to link the streamflow changes to the observed ET at Harvard Forest.

Figure 3 shows the evaporative flux measured at Harvard Forest with streamflow from Biscuit Brook in the Catskill Mountains of New York for the postfrontal sequence of July 11−17 in 1998. Note the drawdown of streamflow each day in the sequence concurrent with the increase in evaporation. The streamflow amplitude decreases later in the sequence as depletion of soil moisture and total streamflow occurs during the sequence.

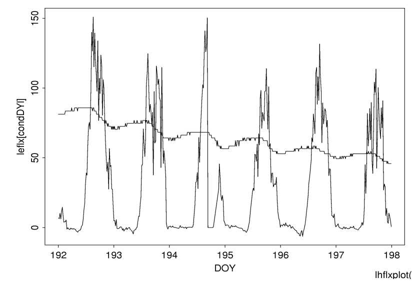


Figure 3: Evaporative flux observed at Harvard Forest (units W/m^2) and runoff from Biscuit Brook, NY (units mm on watershed, scaled by 10) for the July 11−17, 1998 postfrontal sequence.

3. DATA

Streamflow records used for the analysis of the long−term average precipitation minus runoff and streamflow recession come from a network of over 500 stream gages located from Georgia to Maine operated by the United States Geological Survey (USGS). These data are given as daily values. The records included in the analysis have a minimum of 30 years of data, although many stations have a longer period of record, up to about 100 years. The precipitation data used in the precipitation minus runoff analysis comes from another network of over 500 stations from Georgia to Maine archived at the National Climatic Data Center (NCDC). These data are also presented as daily values and have similar periods of record to the daily streamflow data.

For analysis of the diurnal streamflow signal, 15−minute streamflow discharge data is being obtained from the USGS. Currently records for over 50 stations have been obtained from the northeastern states as well as South Carolina. Records only extend from the last 2 to 10 years.

The measurements of evaporation come from the EMS tower located at Harvard Forest, obtained using the eddy covariance method. Data are available for the last decade.

4. RESULTS

Thus far the results have come from application of the methods used to infer leaf emergence from runoff records. Results of applying the first of the three methods, the return of the precipitation minus runoff curve to pre−snowmelt pulse values, are shown in Figure 4. In the northeastern states, the northward progression of the spring date is similar to the results of Fitzjarrald et. al. (2001). Note the small variation in the spring date in the southeastern states, due to the fact that the snowmelt pulse is weak in this region. The elevation dependence can be seen in the spring date. As an example, the Hudson Valley and Adirondack Mountains of New York are depicted on the map.

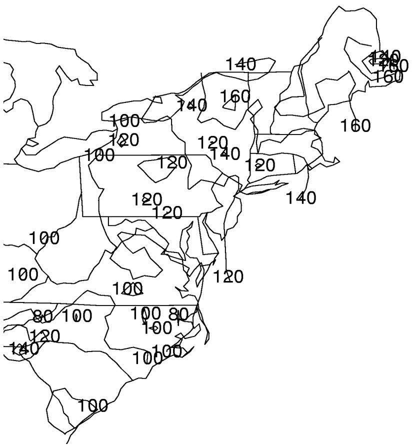


Figure 4: Date of return of the precipitation minus runoff curve to pre−snowmelt pulse values.

Figure 5 shows histograms of the streamflow recession time constant for Wappingers Creek, New York. Data from 1929 to 1998 were used to construct this figure. Note that the shortening of the time constant begins in late May with leaf emergence and progresses through early June with leaf development.

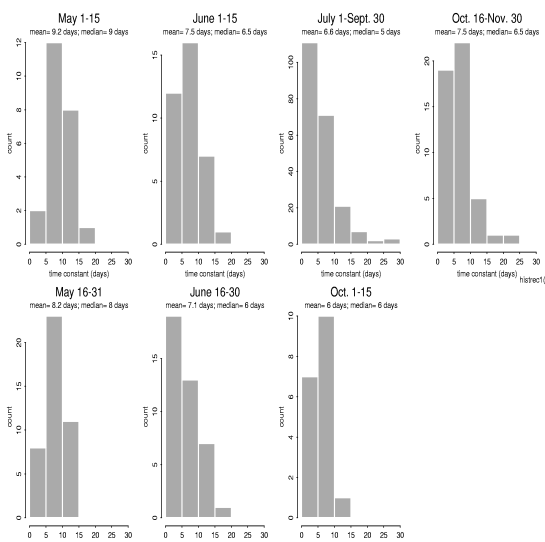


Figure 5: Histograms of the streamflow recession time constant for Wappingers Creek, NY

The top panel of Figure 6 shows the amplitude of the diurnal streamflow signal for Biscuit Brook, New York in 2001. A large increase in the amplitude occurs near day 150 (late May) around the time of leaf emergence. The amplitude remains high through June and July but decreases in August as streamflow and groundwater supply become depleted. The bottom panel shows the amplitude normalized by total streamflow. This is indicating the fraction of the total streamflow which is attributed to the diurnal amplitude signal. The normalized amplitude reaches a maximum after day 200, near the time of maximum observed ET at Harvard Forest.

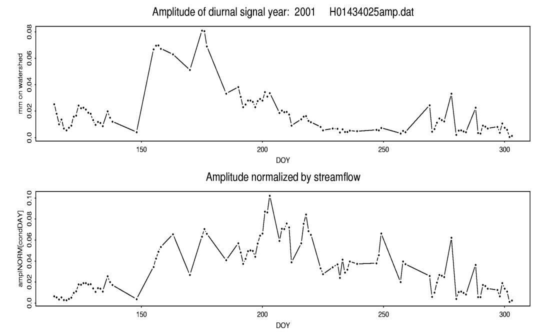


Figure 6: Top panel: Amplitude of the diurnal streamflow signal, Biscuit Brook, NY, 2001. Bottom panel: Amplitude normalized by streamflow

5. CONCLUSION

We have identified three independent runoff characteristics that change with spring onset and leaf emergence. The precipitation minus runoff method gives us a point measurement while the streamflow recession and amplitude methods give us several points throughout the season. The advantage of having both the recession and amplitude methods is that they are used at different times; the recession method is used following a rainfall event while the amplitude method is used during dry periods in small watersheds. These results add to the body of existing indicators of leaf emergence and spring onset, such as the change in tendency Bowen ratio, Normalized Difference Vegetation Index (NDVI) and Hopkins Law, but is the only one which incorporates hydrologic data.

We have also presented a simple analytical model based on a water balance. This model will assist us in relating diurnal streamflow fluctuations to observed ET at Harvard Forest as well as linking changes in streamflow and water table depth to ET at the diurnal and synoptic time scales.

6. REFERENCES

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