EVAPOTRANSPIRATION IN FOREST STANDS OF THE SOUTHERN APPALACHIAN MOUNTAINS

J. L. KOVNEE
Southeastern Forest Experiment Station, Forest Service, USDA

Experiments at the Coweeta Hydrologic Laboratory in the high rainfall belt of the Southern Appalachian Mountains have demonstrated that manipulating the forest vegetation can affect streamflow from the treated watersheds. In general, it has been found that removal of the forest vegetation in part or in total produces increased streamflow. A summary of these studies was given at the 1954 meeting of the Georgia Academy of Sciences and published in the Georgia Mineral News Letter Vol. 7, No. 4, Winter 1954.

Streamflow, however, must be regarded as a residual from precipitation after the return of moisture from the ground to the atmosphere. For example, when averaged over a number of years, the storage change is negligible and the annual water balance equation may be written as

\[ P - Ev = Ro \]  

Where \( P \) = precipitation, \( Ev \) = evapotranspiration and \( Ro \) = streamflow. It is much more appropriate to consider the relationship between evapotranspiration and vegetation and to examine some of the characteristics of evapotranspiration revealed by the Coweeta studies. This is particularly important since such data are extremely difficult to obtain. It is practically impossible to directly measure evapotranspiration from forest stands. In addition, the mathematical theory of evaporation had only been developed for a free liquid surface or a permanently saturated solid surface. The extension to drying of solids as in the case of the removal of water from the soil has not been successful.

Evapotranspiration actually includes three separate processes—interception, transpiration, and evaporation. Climatologists are inclined to group all three together as being responsible for the return of moisture to the atmosphere with the accompanying latent heat energy of vaporization. In scientific investigations of natural land areas, each process will have to be studied separately in order to arrive at basic principles for manipulating vegetation in order to alter the water balance. The Coweeta experiments could not be expected to provide information of this kind with techniques of measurement presently available. In the data that follow, all evapotranspiration phenomena are combined.

The complete water balance equation for any time period, neglecting deep seepage, may be written as

\[ P = Ev + Ro + S_2 - S_1 \]  

Where \( S_1 \) and \( S_2 \) are initial and final soil reservoir storage values, respectively. Since the storage cycle normally has a 12-month period, the difference for such a period is usually small in comparison with the other quantities. This quantity can be regularly minimized by

Reprinted from the Bulletin of the Georgia Academy of Science Vol. XV; Pages 80-85
selecting an arbitrary hydrologic year. At Coweeta the period from May 1-April 30 is used since maximum soil reservoir storage regularly occurs around May 1. The greatest variability in the storage factor is due to ground-water storage. Actually this can be estimated from the derived depletion curve and used in (2). The only unknown is then the so-called retention storage, which must be very small since soils are in the range of field capacity as of May 1. The difference between annual precipitation and streamflow adjusted for groundwater storage for a hydrologic year at Coweeta is thus a good estimate of annual evapotranspiration.

\begin{figure}
\centering
\includegraphics[width=\textwidth]{fig1}
\caption{Annual precipitation and P-Ro for Watershed 2}
\end{figure}

In fig. 1 annual precipitation and P — Ro are given for watershed 2, which has an area of 31 acres and an approximate mean elevation of 2800 feet. Estimated average annual evapotranspiration is 40 inches against 29 inches of streamflow. Although precipitation during the 18-year period varied from 47.5 inches to 84.2 inches, evapotranspiration ranged from 35.6 to 43.9 inches. This conservative aspect of the evaporation process has been recognized in other humid parts of the world and has led to the observation that streamflow is a residual of the water cycle—precipitation that has escaped the attraction of the sun's energy. This fact has another implication. If evapotranspiration can be reduced, for example, by altering the vegetation to a new stabilized condition, then consistent gains in streamflow can be expected each year.

Since the evapotranspiration is dependent on climatic factors such as wind movement and humidity, it is more than likely that it will change with elevation. Fig. 2 shows the relationships between elevation and estimated average annual evapotranspiration on some Coweeta watersheds. Evidently evapotranspiration decreases with
elevation, dropping from 33 inches at 3000 feet to 20 inches at 5200 feet. This is a rather large change and may be influenced by other factors associated with elevation such as vegetation and soil depth. The relationship indicates the advantages of high elevation areas as a source of water supply. Not only are losses to the atmosphere less, but precipitation generally increases with elevation.

The breakdown of evapotranspiration into its components for a low elevation forested watershed can only be done on an approximate basis. Interception studies on forest stands similar to those on Coweeta indicate that annual interception on 70 inches of precipitation amounts to about 9 inches. On two watersheds which were clear cut the first year, increase in streamflow amounted to 17 inches. Since heavy accumulation of cut material was left on the ground, it may be assumed that interception and evaporation remained about the same and that the increase in streamflow came from cessation of transpiration. Thus transpiration may be estimated at 17 inches. The total evapotranspiration is estimated at 32 inches, leaving a balance of 6 inches for evaporation. Admittedly, this is a very rough breakdown but it does give some idea of the order of magnitude of the different components of water loss.

It is practically impossible to obtain reliable estimates of monthly values of evapotranspiration for a natural forest stand. None of the Coweeta data provides this kind of information. The annual value does provide a reliable control, however, and it is not too difficult to make a reasonable distribution of the monthly values. Thornwaite's formula for computing potential evapotranspiration was used, since soil moisture is seldom limiting at Coweeta, on account of high year-round rainfall. The formula results in a yearly total of 27.7 inches, which is somewhat low. The monthly values were adjusted accordingly.

**Figure 2. Annual P-Ro vs. elevation of watershed.**
to a 31-inch annual total, and some minor corrections made on the basis of local evaporation pan data. The distribution need only be approximate for the following discussion. The top curve in fig. 3 represents the monthly march of evapotranspiration for an undisturbed forested watershed.

The experiments at Coweeta have demonstrated that cutting the forest vegetation has decreased evapotranspiration. For example, we will consider the treatment which consisted in converting a watershed from a forested to a field condition. The change in vegetative cover has resulted in a decrease in annual evapotranspiration amounting to 10 inches. Incidentally, the previous observation regarding the conservative nature of evapotranspiration has been borne out in this experiment. Once the cover was fairly well stabilized, annual increases have regularly approximated 10 inches, ranging from 8-12 inches. Monthly values of evapotranspiration have been estimated on a basis of an annual total of 22 inches. It is assumed that the monthly evaporation values for the dormant season would remain relatively unchanged. The lower curve in fig. 3 represents the new march of evapotranspiration, and the area enclosed by the two curves represents the annual decrease of 10 inches. Fig. 3 should be a good gen-

![Figure 3](image-url)

**Figure 3.** Estimated monthly evapotranspiration before and after treatment Watershed 17.
eral description of the relative behaviour of evapotranspiration under the two kinds of vegetative cover on the watershed, since the same climatic factors controlling water transport from the ground to the atmosphere prevail.

The immediate evapotranspiration draft (except interception) is on the total soil reservoir storage which will reflect the monthly changes indicated in fig 3. Monthly increases in streamflow, how-

ever, cannot be predicted from fig 3. Fortunately, the experimental method using a control watershed makes it possible to estimate by regression analyses the average monthly increases in streamflow based on 15 years of record.

Fig. 4 is a comparison of estimated monthly increases in streamflow and decreases in evapotranspiration. It is quite obvious that the latter bear little direct relationship to the former. In fact, over half of the streamflow increase occurs during the dormant season, when the treatment actually had little direct effect on evapotranspiration. The smallest increase, 0.36 inch, occurs in April, with May about the same at 0.37 inch. The largest increase of 1.65 inches takes place in January, with December a close second at 1.50 inches. The largest
change in evapotranspiration belongs to July and amounts to 2.3 inches. Some advantage has been gained in smoothing out the evapotranspiration curve in that sizeable increases in streamflow occur during October, November, and December, when flows are normally very low.

In order to explain the monthly phase difference between evapotranspiration and streamflow, it is necessary to understand the water cycle in this region. Let us begin around the middle of April, when the whole soil profile is at "field capacity" and the water table is at its highest level. The trees begin to leaf out, intercepting sizeable quantities of precipitation, and active transpiration begins to remove large amounts of readily available water from the upper soil mass by virtue of extensive root systems, laterally and in depth. When the trees are removed, as in this experimental treatment and replaced by lesser vegetation, smaller amounts of water are removed from the soil.

Meanwhile, the water table begins to drop in the summer period as groundwater is depleted by streamflow and transpiration draft of vegetation in contact with the water table. Current net precipitation contributes very irregularly to permanent groundwater as long as it is retained and expended in the upper layers of the soil. As a result, monthly streamflow totals continue to decline without interruption at a decreasing rate right through the growing season. Each year, however, some large storms or a succession of smaller storms succeed in overcoming the moisture deficit and contributing to groundwater. Generally, the process does not take place over the whole watershed but is restricted to local areas. With a lesser type of vegetation, the direct transpiration draft is reduced and the water table does not drop so fast. This, of course, results in increased streamflow. Also, with a smaller soil moisture deficit now, a few more summer storms will contribute to groundwater and hence raise the water table above normal conditions and increase streamflow.

Beginning in the fall, around the first of October, the trend in soil moisture depletion is reversed, although streamflow may continue to drop. The forest vegetation reacts to the climatic change, notably lower temperatures, to reduce its transpiration draft. Evaporation also declines, and net precipitation, which now exceeds the combined total, is effective in reducing retention storage. Later in November frosts remove the foliage, making transpiration negligible and decreasing interception, so that the soil moisture build-up proceeds more rapidly. By the end of December, the soil mass is at "field capacity." Note that some accretion to groundwater may take place during this period and continues during the winter. The contribution accounts for increases in streamflow which show up during this period.

Streamflow in the winter months will now respond directly to the amount of precipitation, and the increases in streamflow are maintained in January and February. After the middle of March and through April the water table is at maximum rise. This accounts for the reduced increases in streamflow in March and April.