DIVISION S-7—FOREST AND RANGE SOILS

Approximating Soil-Moisture Storage in Experimental Watersheds by Means of Precipitation and Streamflow Records[1]

K. G. Reinhart[2]

ABSTRACT

An estimate of total soil-moisture storage capacity of an experimental watershed is needed to understand its hydrology. A reliable estimate may be very difficult to obtain, particularly in a forested watershed, because of variations in and difficulties of measuring soil depth, root depth, texture, bulk density, stone content, and moisture content. A method is suggested for approximating storage from precipitation and streamflow records. Precipitation minus runoff—in selected periods when ample precipitation follows a dry spell—provides the estimate of soil-moisture storage capacity. The method is illustrated with data from the Fernow Experimental Forest in West Virginia and is applied to two other watersheds in the Northeast. Limitations of the method are also discussed.

In forest watershed management research, the gauged watershed continues to be a useful tool for determining the effects of treatment on streamflow. In common practice, the watershed is calibrated and treatment effects are determined from measurements of precipitation and streamflow. However, this procedure limits the possibilities of understanding the hydrology of the treatment effect. One obvious gap is the influence on treatment effect of watershed soil-moisture storage—an important factor but difficult to measure.

One way to measure soil-moisture storage is by soil sampling. But this is beset with many difficulties, which vary with the type of watershed. Direct measurement is particularly difficult in steep, stony, forested watersheds.

Stoniness makes it difficult to obtain individual samples and to determine their moisture content and moisture-holding capacity. (The neutron meter is an important advance in this field, but even with it there are still serious problems—including the cost of the meter.) Other difficulties are encountered in determining the moisture content and moisture-holding capacity of the soil profile at the sampling point. And still greater difficulties are encountered in trying to apply moisture contents from sampling points to the entire watershed, using only rough estimates of soil depth, root depth, and stone content.

So, rather than attempt direct measurements, it seemed logical to try to deduce values for soil-moisture storage by using other measurements that we have for the watershed—precipitation and streamflow records.

METHOD

The method described below is a simple application of bookkeeping procedures. For a storm period, precipitation minus runoff from the storm gives an estimate of soil-moisture recharge. If the rainfall is sufficient to wet the soil thoroughly, the soil-moisture recharge equals the field-moisture deficiency before the storm (Langbein and Iseri, 1960). The maximum value for field-moisture deficiency is equivalent to the soil-moisture storage capacity of the watershed. However, there are complicating factors in this procedure, as will be shown. To use this approach we need:

(1) *Accurate measurement of stream discharge.* Inaccurate measurement, weir leakage, or seepage to groundwater that is not measured at the streamgauging station will reduce the accuracy of the estimate.

(2) *Accurate measurement of rainfall.* Often, precipitation measurements are used as indexes; but here we are concerned with absolute amounts. Thus this approach is limited to experimental watersheds where precipitation is intensively and carefully measured.

(3) *Favorable climate-soil situation.* The ideal situation is one in which a maximum soil-moisture deficit is recharged over a short period of time by rainfall. Periods when snowfall occurs are ruled out because of the many complications introduced.

(4) *Long period of record.* The likelihood of occurrence of just the right sequence of events, and repetition for checking purposes, increases with the length of record.

(5) *Allowance for evapotranspiration.* This is one of the sticky problems in the method. With a one-day storm of sufficient precipitation to satisfy the soil-moisture deficit, we can ignore evapotranspiration. But the coincidence of heavy one-day precipitation immediately following a maximum soil-moisture deficit is unlikely; thus it is usually necessary to include in the bookkeeping account the rain occurring over a period of several days, a week, or more. Over these periods, evapotranspiration consumes part of the rainfall and must be taken into account; otherwise precipitation minus streamflow will give too large an estimate of the soil-moisture deficiency.

(6) *High infiltration capacity relative to rainfall intensity.* If part of the rainfall runs off the surface and all the soil-moisture deficiency is not satisfied, the estimated storage value will be too low.

EXAMPLES

Records of the gauged watersheds on the Fernow Experimental Forest in West Virginia are generally favorable to application of this method (Reinhart et al., 1963). We have an 11-year record on five forested watersheds, which range in size from 38 to 96 acres and total about 380 acres. The streamflow records are good. We began with 15 precipitation gauges, but reduced this number to 9 after early records indicated that the reduction would not seriously affect the accuracy of catch. Infiltration rates on these watersheds are greater than rainfall intensities. Precipitation is generally ample (annual average is about 59 inches) so that the soil reservoir is often fully recharged. Also, the soils are fairly shallow (probably averaging 3 to 4 feet deep) so that it is relatively easy to satisfy the field-moisture deficiency.

Data from two of the Fernow watersheds were used to compute watershed field-moisture deficiency. To determine streamflow resulting from rainfall of a given storm period, a discharge-depletion curve was constructed for each watershed. This was based upon the recession in flow from one day to the next in rainless periods of 5 days or more in the growing season. Using the curves, estimates could be made for any rate of discharge of the total flow expected if no further precipitation occurs. This is potential discharge.

[1]Contribution from the Northeastern Forest Experiment Station, Forest Service, USDA, Upper Darby, Pa. Presented before Div. S-7, Soil Sci. Soc. Am., Aug. 20, 1962, at Cornell University, Ithaca, N.Y. Received Sept. 25, 1963. Approved Mar. 25, 1964.

[2]Research Forester and Project Leader in watershed-management research, Northeastern Forest Experiment Station, Parsons, W. Va.

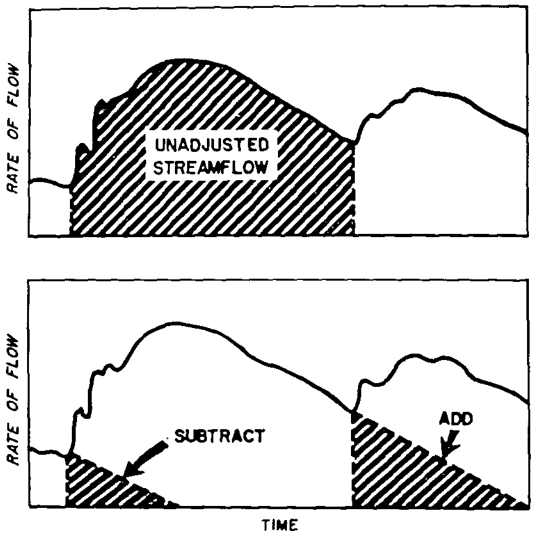

Fig. 1—Streamflow in the period of measurement was adjusted by subtracting estimated flow from prior storms and adding estimated flow that would occur after the period of measurement without further precipitation.

Daily tabulations of streamflow were used to determine the total flow for the days during any one precipitation period, plus the flow for all subsequent days until a new storm event occurred (Fig. 1). Using the discharge depletion curve, this total flow was adjusted by: (*a*) subtracting the potential discharge corresponding to the flow at the start of the precipitation period, and (*b*) adding the potential discharge corresponding to the rate of flow the day before the next storm event.

For example, the available moisture storage on Watershed 5 on September 30, 1954, was approximated as follows: Rain from October 1 through October 6 totaled 3.88 inches, and the next appreciable precipitation was on October 15. Streamflow from October 1 through October 14 amounted to 0.89 inch. The potential discharge on September 30 was only 0.01 inch; on October 14, 0.07 inch. The adjusted streamflow, or runoff, was 0.95 inch (0.89 — 0.01 + 0.07).

Evapotranspiration that occurred during the precipitation period—and thus probably made storage available for subsequent rain—was estimated at 0.67 inch. This estimate was made by using mean daily temperature and Thornthwaite procedures (Thornthwaite and Mather, 1957) on the assumption that in these periods actual evapotranspiration was very close to the potential. Precipitation minus runoff minus evapotranspiration was 2.26 inches (3.88 — 0.95 — 0.67). This was the estimated soil-moisture recharge following September 30 and might be the field-moisture deficiency on that date.

In this case, as in many cases, there is a question as to whether or not the soil-moisture deficit was fully satisfied. Runoff was not large; part of the area may have been recharged with a surplus for streamflow before recharge was completed on the rest of the area.

Therefore a second estimate of field-moisture deficiency on the same date (September 30) was made in which the next rainfall, 4.47 inches on October 15, was included in the precipitation period. Precipitation then totaled 8.35 inches; adjusted streamflow was 4.29 inches; evapotranspiration for the 15-day period was estimated at 1.38 inches.

The estimate of recharge (precipitation minus runoff minus evapotranspiration) was 2.68 inches. This was about 0.4 inch greater than the estimate for the shorter precipitation period. Streamflow was considerable—over 4 inches —and presumably the soil-moisture deficit must have been satisfied. This was probably more nearly equal to field-moisture deficiency than the previous estimate. Of course, evapotranspiration played a bigger part in this second estimate and any error in the evapotranspiration estimate would have affected the storage estimate. Estimates of soil-moisture recharge on other dates were made in the same way.

Estimates of recharge were next plotted against the rate of stream discharge, in area-inches per day, just before the storm period (Fig. 2). In this figure, points representing different estimates of soil-moisture recharge following the same day (using precipitation periods of

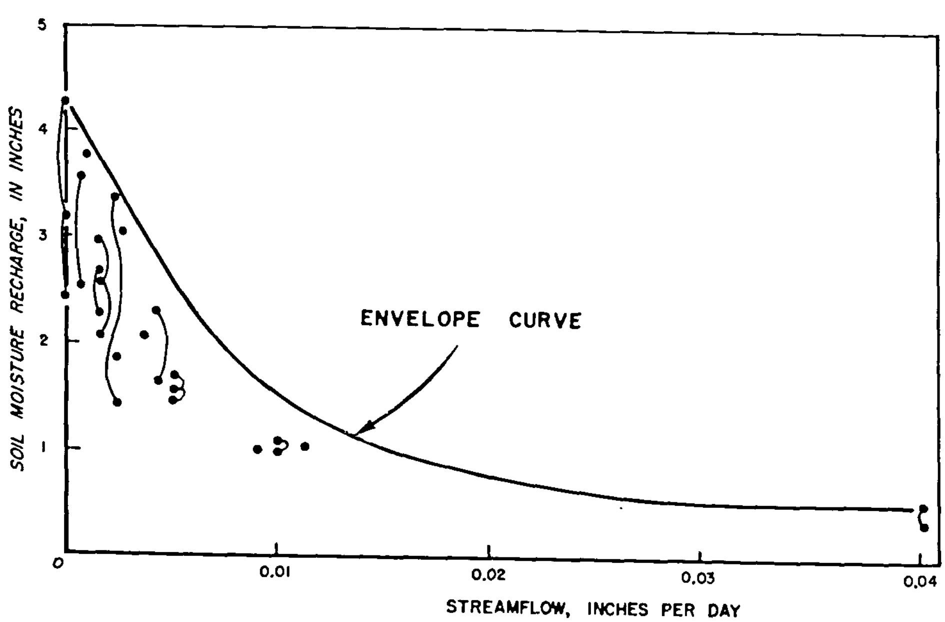

Fig. 2—Soil-moisture recharge and streamflow, Fernow Watershed 5.

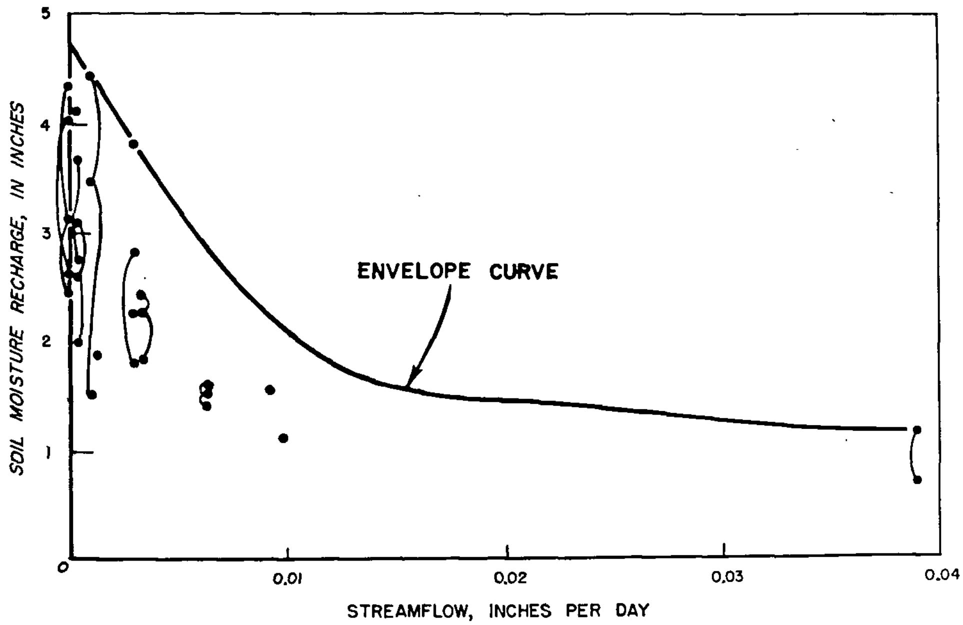

Fig. 3—Soil-moisture recharge and streamflow, Fernow Watershed 4.

different length) are connected by a line. Almost without exception, when separate estimates were made from precipitation periods of different lengths, a higher estimate was obtained for the longer period. In this figure, an envelope curve was drawn to enclose all the points. Records for Watershed 4 were analyzed in much the same way. The results are shown in Fig. 3.

DISCUSSION

A maximum value for soil-moisture recharge was determined from Fig. 2 and 3 by the intersection of the envelope curve with the ordinate, which indicated the deficiency when the stream was dry. The estimate for Watershed 5 was about 4.3 inches; for Watershed 4 it was about 4.7 inches. The difference in the estimates may be the result of deep seepage in Watershed 4, which is not measured at the weir.

These estimates of soil-moisture recharge are approximations of the maximum soil-moisture storage opportunity that occurred. They may not correspond to the theoretical soil-moisture storage capacity, which is usually determined by the difference between field capacity and wilting point, because the watershed soils probably may not have dried to the wilting point in the periods considered. Thus these estimates are for soil-moisture recharge for the driest condition experienced when followed by rainfall sufficient to satisfy the deficit.

For the reasons already given, we have not been able to make a reliable estimate of field-moisture deficiency based on soil-moisture measurement. However, on the basis of some measurements and some judgment, we have derived the following average values: field capacity, 25% by weight; wilting point, 8% by weight; bulk density, 1.2 g. per cm.^{3}; stone content, 25% by volume; soil depth, 46 inches. Using these values, we estimated about 7 inches of available water. This is about 50% higher than a corresponding approximation obtained by using precipitation and streamflow. The difference may be due largely to the fact that the soil was not at the wilting point at any of the times when field-moisture deficiency was estimated.

The streamflow measurement is probably the most reliable estimate in the method. However, there is still the question of deep seepage, which is not measured at the gauging station. In so far as such seepage occurs, it would result in a storage estimate that is too high.

The volume of streamflow is adjusted by estimates of potential streamflow at the start of the precipitation period and just before the next precipitation period. In all cases for the Fernow records, the magnitude of these adjustments was relatively small; thus the accuracy of the depletion curve used was relatively unimportant.

The precipitation measurement is a sample. Considering the relatively large number of gauges on these small watersheds, the precipitation estimate should be sufficiently accurate for the purposes of this study.

Probably the greatest room for improvement in approximating storage lies in improving the estimates of evapotranspiration. There should be no objection to equating potential and actual evapotranspiration, as has been done here, because soil moisture is not limiting in the precipitation period, even though rain does not occur on all days of the period. But there is a question as to whether the general relationship between mean temperature and potential evapotranspiration can be applied without adjustment for rainy days, especially when the duration of rainfall may vary from a few minutes to all day.

Also, in the estimates it was assumed that evapotranspiration affected the storage estimate only during the precipitation period. For the first few days after the end of precipitation, evapotranspiration probably taps both detention and retention storage. The amount of evapotranspiration from detention storage has not been estimated; it results in a reduction in streamflow, and ignoring it serves to include in the storage estimate a small quantity, but not all, of the storage above field capacity.

Estimates of interception have not been included in the bookkeeping. For short storm periods, the magnitude of interception losses would probably be small relative to the accuracy of the method being presented. Also, using Thornthwaite procedures as has been done, interception is included in the estimation of evapotranspiration.

As Fig. 2 and 3 indicate, streamflow appears to be well correlated with soil moisture, a situation recently described at the Coweeta Hydrologic Laboratory in North Carolina (Helvey and Hewlett, 1962).

After storm periods, flow of the Fernow watersheds recedes rapidly if there is no further rain. For example, the composite depletion curve for Watershed 5 showed a cumulative discharge of only 0.7 area-inch covering the period from 2 days after rainfall, when presumably the soil was fully wetted, to the cessation of flow. This discharge can be explained by soil-moisture drainage without a contribution of true groundwater, as demonstrated in the soil model described by Hewlett (1961).

If accretion to groundwater had occurred, it would in part contribute to streamflow during the measurement period and would in part be reflected in the rate of flow at the end of the measurement period. This would in turn affect the rate of discharge at the end of the measurement period (the "add" of Fig. 1). Thus, theoretically at least, water used in groundwater recharge would augment the estimate of runoff from the storm and the method should provide an estimate of soil-moisture storage, not including ground-water storage. However, with flow from both soil-moisture drainage and groundwater, estimation of potential discharge from rate of flow might be more difficult.

OTHER APPLICATIONS

The method was also applied to watersheds at two other locations. Only very general conclusions can be drawn because the author is not familiar with these areas and because analyses were not intensive.

The 1,530-acre Dilldown watershed in Pennsylvania has been gauged since 1948, and records have been published (Pennsylvania Dept. Forests and Waters, 1951-61). Approximations of field-moisture deficiency as related to streamflow were plotted.

The hydrology of the Dilldown Watershed is considerably different from that of the Fernow watersheds. For one thing, groundwater is a much more important component of flow at Dilldown. For that reason, the envelope curve could not properly be extended to the Y-axis (representing zero flow) to get an estimate of maximum field-moisture deficiency because streamflow may continue even after soils reach the wilting point.

With the Dilldown data, there was not a strong relationship between rate of streamflow and soil-moisture recharge. Apparently field-moisture deficiency was not satisfied as frequently at Dilldown as on the Fernow; and recharge does not equal the deficiency unless the deficiency is satisfied.

The highest estimate of soil-moisture recharge for the Dilldown watershed was 5.4 inches. This is very close to the estimate of 5.1 inches of live storage (field-moisture storage capacity) that has been published for this watershed (Pa. Dept. Forests and Waters, 1953, p. 24). The intensity of the rain gauge network on the Dilldown watershed—four gauges for 1,500 acres—is not as great as on the Fernow. For this reason the precipitation measurements may not be as appropriate for the analyses that have been made.

Records of a 32-acre watershed on the Hubbard Brook Experimental Forest[3] in New Hampshire were also studied. Evapotranspiration estimates were based upon mean monthly temperatures for several years of record. The approximation of soil-moisture storage capacity was a little below 4 inches. This is a glaciated watershed and no estimate of storage determined by measurement of soil parameters is available for comparison.

CONCLUSION

Soil-moisture storage capacity estimates for watersheds at the three locations in the Northeast range from about 4 to 5.5 inches. This range is not large considering that the watersheds differ greatly in soils, topography, and forest cover.

From this study, it is clear that the term "approximation" is most appropriate. However, these approximations may still be better than can be obtained by other means, especially on stony forested watersheds. This method is at best just another tool the researcher might use to understand his experimental watersheds better. And an improved understanding of the experimental watershed is essential to application of results obtained there to watersheds in general.

---

[3]Unpublished data, Northeastern Forest Exp. Sta., Laconia, N. H.

LITERATURE CITED

1. Helvey, J. D. and Hewlett, John D. The annual range of soil moisture under high rainfall in the southern Appalachians. J. Forestry. 60:485-486, illus. 1962.
2. Hewlett, John D. Soil moisture as a source of base flow from steep mountain watersheds. U.S. Forest Serv. Southeast. Forest Exp. Sta., Sta. Paper 132. 1961.
3. Langbein, W. B. and Iseri, Kathleen T. General introduction and hydrologic definitions. Manual of Hydrology: Part I. General Surface-Water Techniques. U.S. Geol. Survey Water Supply Paper 1541 A. 1960.
4. Pennsylvania Department of Forests and Waters. Forest and water research project, Delaware-Lehigh Experimental Forest. Rpt. No. 1, 1951; Rpt. No. 2, 1953; Rpt. No. 3, 1955; and Rpt. No. 4, 1961.
5. Reinhart, K. G., Eschner, A. R., and Trimble, G. R., Jr. Effect on streamflow of four forest practices in the mountains of West Virginia. Northeastern Forest Exp. Sta., Upper Darby, Pa. (U.S. Forest Serv. Res. Paper NE-1). 1963.
6. Thornthwaite, C. W. and Mather, J. R. Instructions and tables for computing potential evapotranspiration and the water balance. Drexel Inst. Tech. Climatology Pub. 10(3). 1957.