DEFORESTATION EFFECTS
on soil moisture, streamflow, and water balance in the central Appalachians

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USDA FOREST SERVICE RESEARCH PAPER NE-259
1973

NORTHEASTERN FOREST EXPERIMENT STATION, UPPER DARBY, PA.
FOREST SERVICE, U.S. DEPARTMENT OF AGRICULTURE
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MANUSCRIPT RECEIVED FOR PUBLICATION 2 JUNE 1972.

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ABSTRACT

Soil moisture, precipitation, and streamflow were measured on three watersheds in West Virginia, two deforested and one forested. Water content of barren soil always exceeded that of forest soil throughout the growing season and especially in dry weather. Streamflow increased 10 inches annually on the watersheds that were cleared, most of the increase occurring between July and October. Higher soil moisture was accompanied by large instantaneous peak flows during small storms in the growing season but this peak effect was minor in large storms and in all storms during the dormant season. With precipitation, streamflow, interception losses, and soil-moisture change estimated to comparable levels of precision, the water balance equation was solved for transpiration with sufficient sensitivity to demonstrate the effects of tree leaf growth. After tree leaves were fully grown, calculated evaporative losses from the forested watershed somewhat exceeded potential rates as long as unmeasured runoff (leakage) was disregarded. With all components of the water balance quantified, including leakage, estimated soil-moisture loss by transpiration was at rates close to potential. Estimated leakage seemed consistent with observed stream behavior.

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SMALL GAGED WATERSHEDS are frequently proposed as ideal sites for water-balance studies. Incoming precipitation and streamflow may be estimated to accuracies of $\pm$ 5 percent (*Hornbeck 1965*). The difference, precipitation minus runoff, is often equated to evaporative loss, but this reasoning is valid only if all liquid water draining from the basin is measured. Small watersheds are especially useful in forest water-balance studies because they provide real-life tree growing conditions without the confounding effects on tree growth that plague lysimetry (*Patric 1961*) or the boundary problems inherent in plot studies. However, soil-moisture variation, a key factor in computing the complete water balance on small watersheds, has proven hard to evaluate.

At least three approaches have been devised that appeared to obviate the need to quantify soil-moisture loss from forested land. Several "bookkeeping" methods were attempted (*Reinhart 1964*). Earlier forest studies showed that soil moisture in humid climates usually returns to "field capacity" during the dormant season (*Helvey and Hewlett 1962*), prompting many to disregard soil-moisture variation when computing annual water balances. But this approach is invalid in arid climates, where dormant-season precipitation is frequently insufficient to replenish soil moisture lost during the growing-season. Thornthwaite's (*1948*) concept of potential evapotranspiration allows long-term estimates of evaporative loss to be based almost entirely on climatic parameters, which are far more easily estimated than soil-moisture loss. Unfortunately, considerable uncertainty is associated with all three methods, particularly when only short-term data are available for calculating forest water balances.

The neutron-moderation technique currently offers a good approach to obtaining highly reproducible soil-moisture data from forest land. Properly installed access tubes minimally alter hydrologic performance in the surrounding soil and nearby tree roots. The measurement obtained is an integrated value for water content in about 1 cubic foot of soil. There are, of course, physical and economic limits on the numbers of tubes that can be installed, the depths to which they are inserted, and the frequency of soil-moisture measurements. Nevertheless, the technique affords easy, rapid, accurate, and relatively inexpensive collection of data on moisture variation in forest soils.

The objective of my study was to determine deforestation effects on soil moisture, streamflow, and water balance.

STUDY AREA AND METHODS

The study watersheds are located on the Fernow Experimental Forest near Parsons, West Virginia, on mountain land ranging from 2 to 3 thousand feet in elevation above sea level. The climate is perhumid (*Thornthwaite 1948*) with an average annual rainfall of 57 inches, more than twice the 22-inch average annual evaporative potential (*Patric and Goswami 1968*). Potential evapotranspiration for periods between soil-moisture readings was estimated by multiplying published pan evaporation for Parsons (*U. S. Environmental Data Service 1968 and 1969*) by the coefficient 0.78 developed by Troendle and Phillips (*1970*). The growing season is roughly May through September, with an average frost-free season of 145 days. Tree leaves emerge late in April, are fully grown by 1 June, and begin to fall in early October. About 95 percent of the tree roots are found in the upper 3 feet of soil. All the experimental watersheds were completely forested until 1964 with uneven-aged stands of hardwood species; volumes ranged from 7 to 12 MBF per acre.

Calvin soils—reddish-brown, moderately deep upland inceptisols weathered from sandstone and acid red shale—predominate on the experimental watersheds. They are well drained, moderately permeable, and strongly acid. Their natural fertility is moderate to low. Their depth to bedrock averages 32 inches, ranging from 22 to 47 inches. Subsoil texture ranges from loam to silty clay loam with silt loam most common. These soils have been described in detail by Losche and Beverage (*1967*).

Precipitation on the Fernow Experimental Forest is sampled by 14 recording and standard rain gages. Streamflow from each watershed is measured in 120° V-notch weirs. Streamflow data are processed using the program described by Hibbert and Cunningham (*1967*).

The watershed treatments have been described in detail by Patric and Reinhart (*1971*). Watersheds 6 (55 acres) and 7 (58 acres) had been completely deforested by 1967 and were maintained nearly barren thereafter by intensive use of herbicides (*Patric and Campbell 1970*). Careful logging left the forest litter relatively intact on these barren watersheds, mineral soil was seldom exposed, and overland flow was a negligible component of stormflow (*Hornbeck 1968*). Logging slash and the little revegetation not killed by herbicides provided some ground cover throughout the experiment. Herbicide application ceased in August 1969; and a dense cover of herbs, grass, weeds, tree sprouts, and seedlings has since revegetated these catchments. Watershed 4 (95 acres) remained fully forested and served as an untreated control throughout the experiment.

Twenty-five aluminum access tubes were installed at random locations on each experimental watershed. A Troxler[1] probe and scaler provided data on soil moisture content at sampling positions 6 inches below the soil surface and at 12-inch intervals thereafter to depths as much as 8 feet. The 6-inch readings were not corrected for soil-air interface effects. More than 5,200 soil-moisture readings were obtained and computer processed, using the Popham-Ursic (*1968*) program.

Soil-moisture readings were planned at 2-week intervals during the growing season, but weather and competing work often forced deviation from this schedule. A few dormant season readings were taken after prolonged rain to establish maximum soil-water content.

[1]The use of trade, firm, or corporation names in this publication is for the information and convenience of the reader. Such use does not consititute an official endorsement or approval by the U.S. Department of Agriculture of any product or service to the exclusion of others which may be suitable.

RESULTS

Deforestation Effects on Soil Moisture

After the first year of soil-moisture measurement, we concluded that water content below 2 feet remained virtually constant on the barren watersheds. Thereafter, we made soil-moisture measurements in the top 2 feet of soil on these watersheds although measurement to full access tube depth continued on the forested watershed.

The greatest fluctuation of water content always occurred in the top 2 feet. Soil-water content fluctuated more widely in forested than in barren soils (table 1), moisture in the barren soils remaining within 1 inch of the 8-inch field capacity observed by Troendle (*1970*). Although the total rainfall during the 1968 growing season was almost identical to that of the 1969 growing season (about 26 inches), its distribution was different and relatively long rainless spells caused drier soils in 1968.

Two or sometimes 3 days were needed to sample soil moisture in all 75 access tubes. Each date listed in table 1 represents, as nearly as possible, the mid-date of the sampling period. Rainfall presented a similar tabulation problem. For simplicity, the precipitation that fell on watershed 4 was listed; the rainfall on watersheds 6 and 7 varied little from these tabulated amounts.

The last two values in table 1 illustrate the interpretive problems that weather caused. A last reading for the growing season (9 October 1969) was obtained during dry weather on watersheds 4 and 6, but heavy rain fell before corresponding data were taken on watershed 7. The opposite occurred in the spring (13 May 1970), when soil moisture on watershed 7 was measured in dry weather but heavy rain fell before corresponding data were obtained on watersheds 4 and 6. The result is a spurious indication of an over-winter decrease of soil moisture on watershed 7 while the others gained small amounts of water.

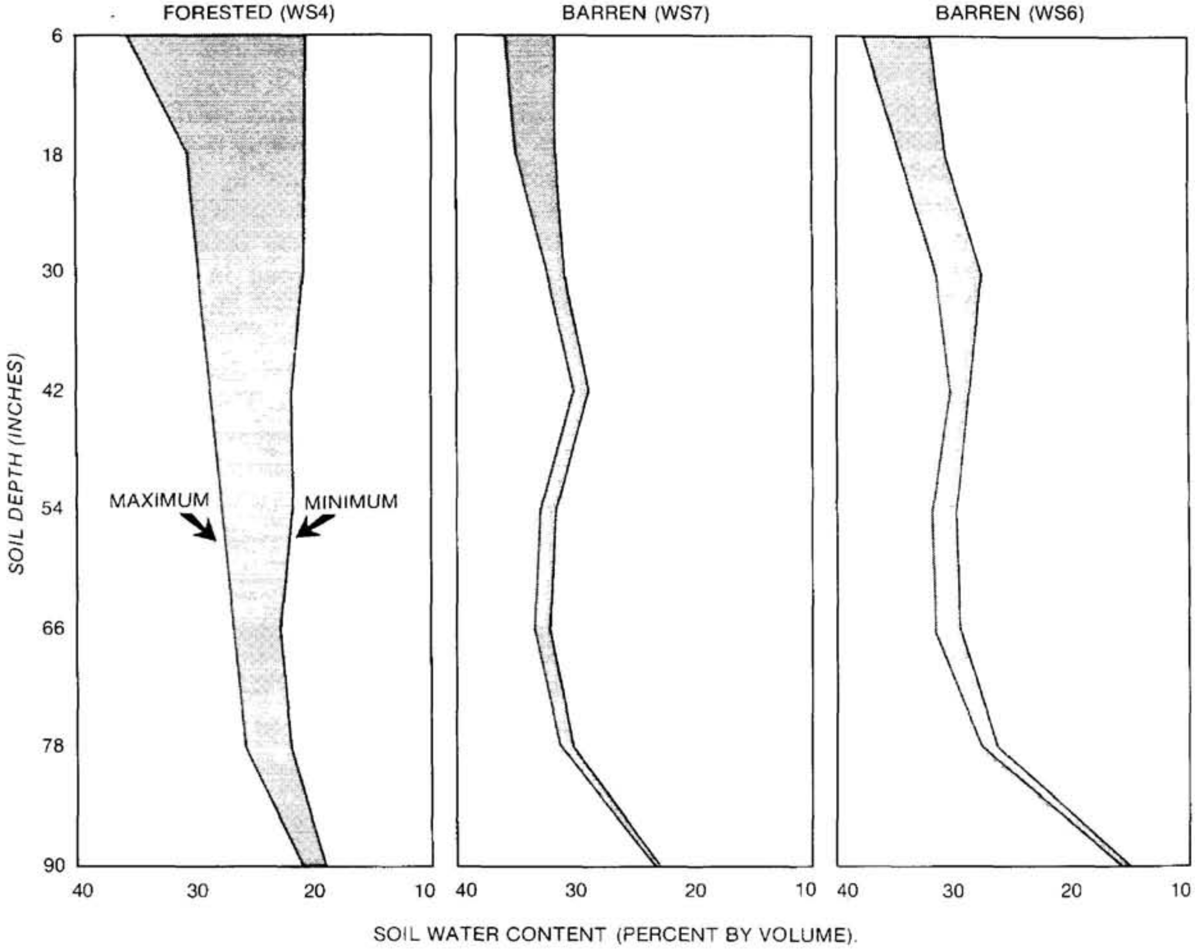
Figure 1 shows the range of soil moisture during 1968 on the three watersheds. Data were used from the six deepest access tubes on each watershed and the highest reading for each depth in any tube was plotted. The minimum readings were recorded in the same way; the width of the shaded band shows the range between observed maxima and minima. The moisture content of the soil was highest during winter and spring, lowest between June and September. As the growing season progressed,

Table 1.—Precipitation and water content in the top 2 feet of soil on the experimental watersheds, in inches

[Mean and standard deviation for 25 access tubes on each watershed]

Date	Rain since previous measurement	Soil water content		
		Forested (WS 4)	Barren (WS 6)	Barren (WS 7)
1 June 68	—	7.7±0.6	8.5±0.8	8.0±0.8
13 June 68	0.55	5.6±0.8	7.4±0.7	7.0±0.7
24 July 68	3.00	5.4±0.9	7.7±0.9	7.7±0.9
14 Aug 68	6.79	7.2±0.8	7.9±0.9	7.9±0.8
5 Sept 68	.90	5.7±0.9	—	—
17 Sept 68	3.18	6.8±0.9	7.7±0.9	8.0±0.7
8 Oct 68	.29	5.6±0.9	7.9±0.9	8.0±0.8
3 Dec 68	9.13	7.7±0.8	—	—
2 May 69	25.16	8.0±0.8	7.5±0.8	7.9±0.7
16 May 69	2.45	8.0±0.8	—	—
22 May 69	2.57	7.4±1.1	7.2±0.9	7.6±0.8
5 June 69	.42	6.7±0.8	7.1±0.9	7.1±0.8
2 July 69	3.51	6.7±1.0	7.4±0.9	7.5±0.8
16 July 69	2.88	7.5±0.9	7.4±0.8	7.6±0.7
31 July 69	2.69	7.1±1.1	7.6±0.8	7.6±0.8
9 Oct 69	15.11	7.5±0.7	7.6±0.8	8.1±0.8
13 May 70	29.33	7.8±0.9	7.9±0.8	6.8±1.0

Figure 1.—Maximum variation in soil water content observed on the experimental watersheds in 1968.



tree roots on the forested watershed took moisture from the surface soils, then from deeper soils during the periods of below-average rainfall in June (2.2 inches) and July (3.8 inches). On barren soils, even the below-average rainfall was more than enough to replace the moisture lost by evaporation. Although watershed 6 seems to have lost more moisture than watershed 7, analysis of variance did not show the differences to be significant at the .05 level. A similar comparison of the same watersheds in 1969 showed a narrower range because evenly distributed rainfall replaced soil-moisture losses sooner.

There was so much variation among the sampling sites that no consistent influence of slope position (ridge, mid-slope, or channel) on soil-moisture loss could be shown. The effects of slope on the water content of barren soil were even smaller and less consistent than on forest soil.

Soil-moisture sampling to 1 percent of the watershed mean clearly exceeded our capability (table 2). However, 25 tubes provided precision within 5 percent of the watershed mean; it was achieved at all depths in moist soil and at depths to 4 or more feet in drier soils. Under these conditions, soil-moisture data were comparable in precision with precipitation and streamflow data as measured on the Fernow Experimental Forest.

Water contents for the 6-, 18-, and 30-inch sampling depths at the driest sites were applied to moisture-release curves (*Troendle 1970*), to estimate soil-water potential. The lowest potentials in 1968 developed at the 6-

Table 2.—Number of neutron probe access tubes needed to sample watershed soil moisture to specified levels of precision at specified soil depths

Soil depth (inches)	Desired half width of 95% confidence interval					
	Moist soil[a]			Dry soil[b]		
	1%	2.5%	5%	1%	2.5%	5%
6	502	91	22	1,535	192	50
18	451	68	17	991	142	36
30	412	68	17	792	123	31
42	538	84	21	705	110	27
54	390	63	16	346	55	14
66	424	68	17	323	51	13

[a]Highest soil-moisture content, measured on May 2, 1969.
[b]Lowest soil-moisture content, measured on July 2, 1969.

inch level on forested watershed 4, ranging from −12 bars on ridges to −5 bars near streams. Water-potential values increased sharply with soil depth, never exceeding −2 bars at the 30-inch depth. This amount of drying was measured after 4 inches of water had been lost from ridgetop soils; had these soils dried to a water potential of −15 bars throughout their 4-foot depth, over 6 inches of rain would have been needed to restore the wetness observed in the springtime. In 1969, when summer rainfall was more evenly distributed, soil-water potential never fell below −10 and −0.9 bars at the driest sampling sites in ridge and channel soils, respectively.

At the 6-inch depth and below, water potentials never fell below −0.5 bars on the barren watersheds. Nevertheless, the top inch or so of soil became hot and powder-dry in summer. This dust mulch apparently combined with remnants of the organic forest soil layers to restrict soil drying almost entirely to the surface layers.

Deforestation Effects on Streamflow

During the 3 years that watersheds 6 and 7 were kept barren, average annual streamflow was 10 inches greater than it would have been had the watersheds remained fully forested. Most of the increase was in the growing season (table 3). A low rainfall trend, especially low during the mid-60's, continued through the years 1967-69. Had rainfall been more abundant and vegetation control more complete, growing-season streamflow increases might have been larger.

The increases in streamflow for each month from July through October were significant at the .05 level, although irregular rainfall caused great year-to-year variation among these increases. But when the monthly increases for watershed 6 and 7 were averaged, using data from 1967 through 1969, they seemed much more closely related to evaporation than to precipitation (fig. 2). There was, however, about a 2-month lag between the rise and fall of evaporation and the apparent response by streamflow.

Rain absorbed into already moist soils of the barren watersheds maintained the streams from these watersheds at consistently higher levels than had prevailed under the original forest vegetation. Figure 3 shows, by cumulative frequency curves, the percentage of the time that the flow equalled or exceeded specified rates. Flows greater than 5.0 $ft^3\ sec^{-1}$

Table 3.—Estimates of streamflow increase on barren watersheds,[a] in inches

Year	Watershed	Growing season (May-October)	Dormant season (November-April)	Annual
1967	6	8.9*	0.6	9.5*
	7	8.6*	1.3*	9.9*
1968	6	7.9*	2.3*	10.2*
	7	7.5*	2.8*	10.3*
1969	6	9.0*	1.5*	10.5*
	7	8.6*	.8	9.4*

[a]Derived from regression on measured flow from the forested catchment, based on pretreatment calibration period.
*Significant at the .05 level.

Figure 2.—Effects of deforestation on monthly streamflow. Each plotted point is the 3-year average monthly value for watersheds 6 and 7.

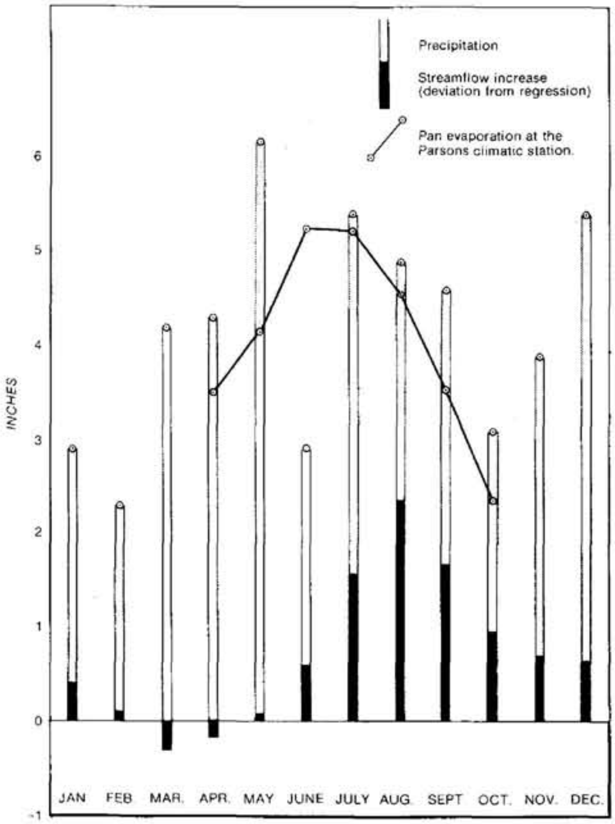
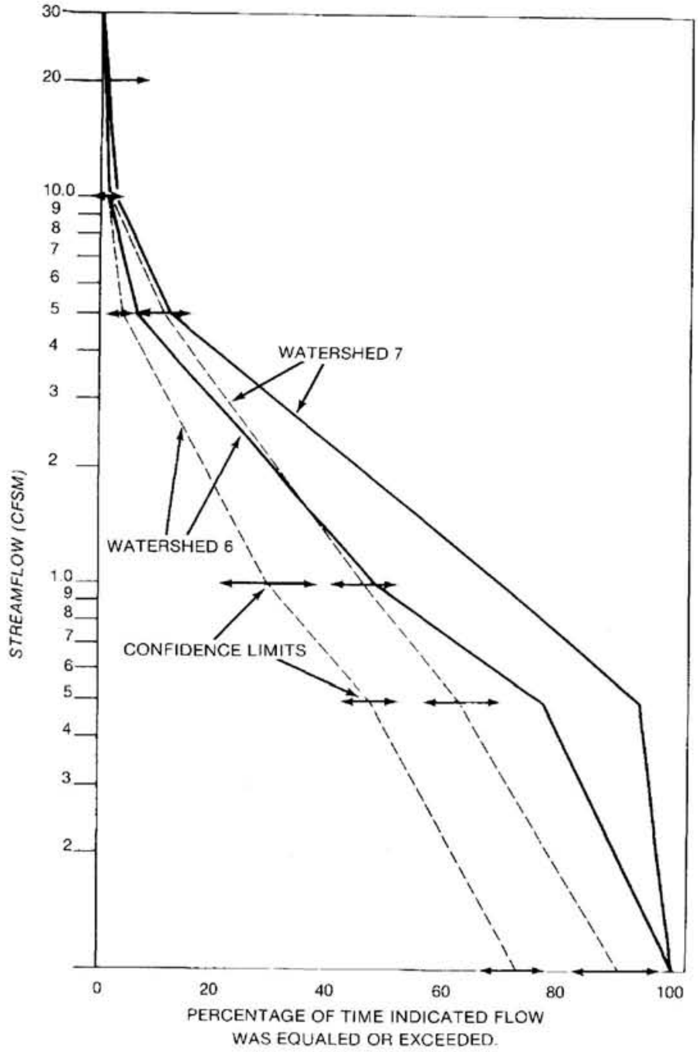


Figure 3.—Effects of deforestation on flow duration, hydrologic year 1969. Solid lines represent measured flow duration; dashed lines are estimated flow duration before deforestation, as derived from regression on measured flow duration from the forested watershed.



mile^{-2} (CFSM) were not significantly affected by deforestation. Flow from the forested watershed was much lower than that from the barren watersheds. In fact, the stream that drained the forested watershed was dry during parts of each summer from 1967 through 1969.

Quickflow is one measure of stormflow. The relationship between quickflow and delayed flow (as defined by *Hibbert and Cunningham 1967*) was probably about the same on all three watersheds when all of them were forested. After two of the watersheds were deforested, water yields from those two increased (table 3), especially during the growing season but somewhat during the dormant season, too. Thus, even though the proportion of dormant season quickflow was unchanged (table 4), absolute amounts of quickflow must have

Table 4.—Flow separation by season and vegetative cover (hydrologic year 1969)

Flow	Forested watershed (4)		Barren watershed (6)	
	In.	*Pct.*	*In.*	*Pct.*
Dormant Season:				
Quick flow	3.78	25	3.42	25
Delayed flow	11.34	75	10.29	75
Growing Season:				
Quick flow	2.29	38	2.96	28
Delayed flow	3.78	62	7.63	72

increased in proportion to water-yield increases. Wetter soils on the barren watersheds did produce a modestly increased volume of quickflow during the growing season, while delayed flow doubled in volume (table 4). These results, too, demonstrate the role of soil moisture in augmenting low flow without greatly influencing stormflow. This apportionment of quick and delayed flows also substantiates the observation that overland flow is a minor component of runoff from the barren surfaces. In fact, overland flow and soil erosion were uncommon on the barren watersheds, occurring only on and immediately below the steeper roads. Infiltration rates on the barren watersheds in 1969 far exceeded maximum rainfall intensity, varying only insignificantly from those measured on the forested catchment.[2]

Deforestation had no effect on instantaneous peak flows during the dormant season. But during the growing season, instantaneous peak flows were higher on the barren watersheds (table 5), particularly in small storms.

Table 5.—Instantaneous peak flow comparisons, 1968 growing season, in CFSM

Precipitation (inches)	Watershed 4	6	7
0.85 (small storm)	0.65	24.30	12.34
3.64 (largest storm)	86.53	105.75	112.93
26.35 (season total)[a]	9.91	20.61	19.16

[a]Average peak, all storms.

Peak, as used here, includes all measurable streamflow rises during rain or snowmelt. Regression analysis confirmed these findings, indicating a fourfold average increase in peak flow for all growing seasons after the watersheds were deforested.

Deforestation Effects on the Water Balance

The water-balance equation

$$P - (RO + I + T + L \pm \triangle S) = O \qquad (1)$$

in which

$P =$ Precipitation
$RO =$ Streamflow
$I =$ Canopy and litter interception loss
$T =$ Transpiration
$L =$ Leakage, unmeasured subsurface flow
$\triangle S =$ Gain or loss of soil moisture

is considered a suitably complete statement of rainfall disposition on the forested watershed. This equation neglects the small and relatively constant losses of soil moisture by direct evaporation, photosynthesis, and plant respiration. The water-balance equation provides a means of estimating evaporative losses from rainfall, streamflow, and soil-moisture data. For this purpose, soil-moisture changes to 4 feet were computed, providing data comparable in precision ($\pm$ 5 percent) to that on rainfall and streamflow. The 4-foot soil depth used for moisture measurement is both physically and biologically appropriate; it is about the maximum depth of Calvin soil and contains most of the forest tree roots. Interception losses were estimated from rainfall by the method of Helvey and Patric (*1965*). Transpiration was calculated by solving equation (1) for T, with L always regarded as O. If L was in fact greater than O, this method of calculating T must overestimate transpiration. Calculated evaporative losses ($I + T$) were then compared to estimated potential evapotranspiration (PET) (table 6).

PET is a useful concept because it specifies an upper limit of evaporative loss that, being controlled by solar energy, is subject to little annual variation. If we accept the 22-inch estimate of Patric and Goswami (*1968*) as a reasonable approximation of mean annual evaporative loss from forest land near Parsons, then the effect of deforestation on evaporation can be deduced. Annual streamflow increases close to 10 inches were maintained during the 3 years watersheds 6 and 7 were barren (table 3). Transpiration and interception losses greatly decreased following deforestation while evaporation from land and stream surfaces must have increased. The net effect, 10 inches more of streamflow, represents that much less evaporation from the barren watersheds. By this means, evaporative loss from

[2]From a report by J. N. Kochenderfer on file at the Parsons Timber and Watershed Laboratory.

Table 6.—Evaporative losses from the forested watershed, estimated by water balance and pan evaporation for the 1969 growing season, in inches

Soil-moisture measurement dates	Elapsed days	Water balance components						Evaporation	
		P	RO	I	ΔS	T	I+T	Pan[a]	PET
1 May	0	—	—	—	—	—	—	—	—
16 May	15	2.44	1.23	0.38	+0.01	0.82	1.20	2.62	2.04
22 May	6	.12	.15	.08	−0.56	.45	.53	.83	.65
5 June	14	.40	.10	.14	−1.20	1.36	1.50	2.59	2.02
2 July	27	3.86	.08	.79	−0.27	3.26	4.05	4.56	3.56
16 July	14	3.22	.08	.59	+1.05	1.50	2.09	2.52	1.97
31 July	15	1.72	.02	.50	−0.46	1.66	2.16	2.55	1.99
1 October	62	11.87	1.38	1.99	+0.78	7.72	9.71	8.91	6.85
Growing season	153	23.63	3.04	4.47	−0.65	16.77	21.24	24.58	19.08

[a]Published class A pan evaporation for the Parsons Climatic Station (U.S. Environmental Data Serv. 1969).

the barren watersheds was estimated at 10 inches less than 22, or 12 inches annually. This cannot, of course, be separated into interception, transpiration, and direct evaporation from soil and water surfaces.

The preceding assumptions provide annual values for all components of the water balance except leakage. Leakage was later calculated as the residue from precipitation (table 7). But here, T for the forested watershed was estimated by subtracting annual interception loss (8 inches) from PET (22 inches).

Precipitation minus streamflow (P — RO), sometimes equated to evapotranspiration, suggests that leakage from the barren watersheds has not changed since the weirs were installed. Before watersheds 6 and 7 were deforested, they averaged 37.28 and 26.89 inches, respectively, in P -- RO. After deforestation, the values averaged 24.77 and 13.78 for these watersheds. Thus, both before and after deforestation, P — RO was about 10½ inches greater on watershed 6 than on watershed 7, a finding that may be regarded as further evidence of unvarying leakage.

Table 7.—Annual water balances for the experimental watersheds for hydrologic year 1968, in inches

Watershed	Water balance components						
	P	RO	I	T	I+T	ΔS	L
Forested (4)	51	20	8	14	22	0	9
Barren (6)	49	26	—	—	12[a]	0	11
Barren (7)	48	35	—	—	12[a]	0	1

[a]Includes an unknown, probably large, amount of evaporation directly from the barren soil surface.

Table 8.—Effect of leaf development on transpiration, Spring 1969, in inches

Stage of leaf growth	Date	Mean daily evaporative loss		
		T	I+T	PET[a]
Up to half grown	5/16	0.05	0.08	0.13
Growth continuing	5/22	.08	.09	.11
Growth complete	6/5	.10	.11	.14
Mature	7/2	.12	.15	.13

[a]Potential evapotranspiration estimated from evaporation pan data.

Transpiration estimates derived from water balance (table 6) seem to reflect leaf development. Note that I + T was less than PET before tree leaves were grown, i.e. through June 5. After leaves matured (July 2), calculated evaporative losses (I + T) always exceeded PET, an effect probably caused by assuming leakage to be zero. Mean daily evaporative losses were calculated when T, I + T, and PET (as listed in table 6) were divided by elapsed days. Then the effect of leaf growth on soil-moisture loss became clearer, with daily transpiration losses increasing steadily as leaves grew (table 8). The inflationary effect of leakage on this estimate of transpiration probably persisted, although it was apparent only after the leaves had matured and daily evaporative losses (I + T) averaged 15 percent above PET.

DISCUSSION

The preceding results have important implications. Foresters know that increased soil moisture on cutover land can improve the survival and growth of new stands. Hydrologists know that added moisture in soil means higher water levels in headwater streams. Even though measurably increased flow is not expected in larger streams, conservationists should know that scattered cutting is generally beneficial to forest water resources. These benefits are not, however, uniform over time and space. Douglass (*1967*) cautions that (1) water saved from evapotranspiration is proportional to the severity of forest cutting, and (2) water savings will not persist after vigorous regrowth has started.

These results also support some older concepts of forest soil moisture and they suggest a promising direction for new study. They show that soils at Parsons always return to near-maximum water content during dormant seasons. Probably the generalization that $\triangle S = 0$ is valid for computing annual water balances in any perhumid climate. This generalization emphatically does not apply to shorter-term water balances during the growing season. The results further suggest that the concept of moisture loss rates remaining undiminished as soils dry (*Veihmeyer and Hendrickson 1955*) is valid in humid climates, as Penman (*1970*) recently concluded. Future studies of the relationship between measured soil-water loss and indexes of evaporative potential could result in more economical methods of study. Soil-moisture measurement is costly, even with neutron probes, and savings will be realized when forest water use can be estimated with confidence from the more easily measured climatic parameters.

The variable-source-area concept (*Hewlett and Hibbert 1967*) is one of the most important recent advances of thinking in forest hydrology. Source areas were not identified in this study because the access tube distribution used was not sufficiently sensitive to spatial variation in soil moisture. Soil-moisture measurement may never help to identify source areas of stormflow, because highly permeable forest soils are subject to large and rapid changes in water content during storms. Between storms, however, soil-moisture content is relatively stable, and more frequent measurements should help to identify the soils with the highest water content that are likely to supply baseflow to streams.

Long ago, deforestation was believed to cause reduced soil moisture and streamflow between storms, and disastrous flooding during storms (*Brooks 1910*). This study revealed almost opposite effects of deforestation, confirming more modern observations (*Douglass 1967*). Why has opinion changed so drastically?

Infiltration data provide clues, if not the full answer to this question. Had the deforested soils been damaged by tillage, heavy grazing, or fire, the infiltration rates would have been reduced dramatically. The soils in this study were not hydrologically damaged by deforestation, and infiltration rates remained near the high rates characteristic of forested land long after the trees were cut. Surface soils did dry quickly and severely after storms, but the dry surface protected the underlying layers from evaporation, and subsurface soil moisture remained at near-maximum levels. Hewlett's (*1961*) concept of soil as the source of headwater streams explains the resultant increase in baseflow.

Another reason that the supposed ill effects of deforestation did not materialize was that this watershed treatment was an experimental and not an economic undertaking. It did not resemble the agricultural practices on which most such ideas are based. No attempt was made to exploit the land after deforestation. Even though the barren soil surface underwent some structural change, it was far less drastic than the complete overturn and compaction that accompanies farming, and thus overland flow, soil erosion, and flooding did not occur. We are continuing to observe the hydrologic behavior of plots of deforested soil because it seems only a matter of time until infiltration will be materially reduced and overland flow and erosion will begin.

With other components of the water balance (precipitation, streamflow, interception, and soil-moisture storage) quantified within stated (.05) limits of precision, the validity of the leakage estimate is contingent on sound evaluation of the transpiration loss. The method used to evaluate transpiration requires acceptance of the assumption that 0.78 times class A pan evaporation (*Troendle and Phillips 1970*) is a reasonable approximation of potential evapotranspiration from local forests.

The class A pan has been recommended for international use as an index of evaporative demand (*Anonymous 1965*). Often considered a reliable estimator of crop water use (*Christiansen and Hargreaves 1970*), it has been used recently as a check on observed water balance in forests. For example, with plentiful (*Troendle 1970*) and average rainfall (*Waggoner and Turner 1971*), measured moisture loss from forest soil was close to evaporative potential as estimated from class A pan evaporation. These findings tend to confirm reasoning by Patric et al. (*1965*) that in humid regions adequate soil moisture to meet potential demands is available to most forest trees. But after several growing seasons of below-normal rainfall, evaporative loss from another forested watershed averaged only 68 percent of that predicted from the class A pan (*Mustonen and McGuinness 1968*). Wicht (*1967*) considered calculated evapotranspiration a valuable check on observed water balances, but Lee (*1970*) argued that empirical models provide inadequate checks. Nevertheless, numerous observations of high correlations between evaporative losses from pans and those from well-watered vegetation (*Tanner 1967*) suggest that the approach used to quantify the transpiration loss in this study is at least conceptually appropriate.

Evaporative losses from the barren watersheds could not be estimated on the basis of potential evapotranspiration because the required condition that living vegetation completely cover the soil was not met. However, wet-weather evapotranspiration at near-potential rates was expected, even without vegetative cover. Showers often wet the dark land surface, leaving water freely available for evaporation. As for dry weather evaporative loss, soil-moisture data suggest only that it occurred at rates far less than potential. A statement by Milthorpe (*1960*) probably comes as close as possible to describing weather influences on evaporative losses from the barren watersheds: "The total loss of water from soil in a season of intermittent rain and high potential evaporation is therefore independent of the potential evaporative rate and mainly a function of the rainfall, i.e. the frequency of rewetting; on the contrary, in a season of high rainfall and low potential evaporation, it is closely related to the potential evaporation and is independent of rainfall." The net effect, with rain occurring about one day out of three, was evapotranspiration at rates materially less than potential on the barren watersheds.

With all components of the water balance except leakage quantified as accurately as technique permits, how realistic is the unmeasurable residue assigned to leakage? Simple observation establishes that watersheds 4 and 6 do leak. In dry weather, flowing water is visible in stream channels above and below the concrete cutoff walls while no water passes over the weir blades. The "live" channel is about 200 yards long on watershed 6 and at least five times longer on watershed 7, suggesting that far more pervious rock underlies the former. One end of the watershed 6 cutoff wall is known to be seated in loose colluvium; dry-weather measurement shows leakage here at the rate of 2 inches per year. Perhaps the wet-weather leakage is greater.

I attribute other consistent differences in performance of watersheds 6 and 7 to leakage. For example, flow duration (fig. 3) was always greater on watershed 7. Before it was cleared, the stream draining watershed 6 dried up every summer. Watershed 7 dried up only twice during the 7-year calibration period. No-flow conditions persisted more than twice as long on watershed 6. And, if P − RO does in fact approximate evaporative losses, why should anyone expect 10 inches more of water to evaporate consistently from watershed 6? Watersheds 6 and 7 are so similar in all other respects that leakage seems the logical answer.

Although leakage has never been measured directly, its role in streamflow is well known. A case of extreme variation in runoff with small differences in climate occurred in Virginia (*Burford and Lillard 1966*). There, 52 percent of rainfall drained off one watershed and only 21 percent from another nearby. The authors attribute these differences in runoff to the soil, parent material, and basic geological characteristics of each basin. Inadequacies of the gravimetric soil-moisture-sampling method may have led Rowe and Colman (*1951*) to underestimate evaporative losses when they assigned about half the annual rainfall to leakage. Dagg and Blackie (*1965*) suspected leakage when soil-moisture sampling by electrical resistance revealed cumulative differences between measured and predicted evaporative losses. Although this is by no means a complete survey, these experiences show that leakage, endemic to small watersheds (*Hewlett et al. 1969*), greatly concerns hydrologists and must not be neglected because of the inescapable measurement difficulties.

Better progress is being made toward understanding the components of the water balance and how they relate to each other. Abbe (*1902*) once complained that evaporative loss in nature "is so mixed up with seepage, leakage, and consumption ... by plants that our metrological data are of comparatively little importance." A more recent commentary on small watershed research (*Slivitzsky and Hendler 1964*) expressed concern about "ignorance of the precise cause and effects of the different components of the hydrologic cycle." Hewlett et al. (*1969*) felt that older studies did merit such criticism, but that it is less justified in modern watershed research where "mechanisms of water loss, storage, and delivery are now studied ... with all the trappings of modern science." Ward (*1971*) went even further: "if the water balance equation ... can be solved, then it is likely that the measurements or estimations of the individual components of that balance are satisfactory." This study, at least, was not one of the many attempts to solve the water-balance equation while ignoring leakage "in the quiet hope that it is in fact zero" (*Penman 1963*). Perhaps that fact represents some progress.

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