To the Graduate Council:

I am submitting herewith a dissertation written by Michael Wells Mayfield entitled "Variations in Streamflow Among Watersheds of the Cumberland Plateau, Tennessee." I have examined the final copy of this dissertation for form and content and recommend that it be accepted in partial fulfillment of the requirements for the degree of Doctor of Philosophy, with a major in Geography.

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The Graduate School

VARIATIONS IN STREAMFLOW AMONG WATERSHEDS OF THE CUMBERLAND PLATEAU, TENNESSEE

A Dissertation

Presented for the

Doctor of Philosophy

Degree

The University of Tennessee, Knoxville

Michael Wells Mayfield

March 1984

ABSTRACT

The purpose of this study is to evaluate the differences in streamflow characteristics among the streams of the Cumberland Plateau of Tennessee and to account for those differences. The study is limited to those gaged basins on the Plateau with watershed areas of 50 to 1000 square miles in order to avoid the extreme variations found in very small basins as well as the moderating effect of large basins.

Stream discharges at low and peak flows were obtained from published sources. In order to compare flows from watersheds of varying size, all flows were standarized to units of cubic feet per second per square mile. Graphic display of the results in the form of flow duration curves and flood flow recurrence curves reveals major differences in the hydrologic response of streams in the study area. Water budget analysis and recalculation of flood peaks using a standard time interval indicates that differences in the streamflow regimes of the various watersheds are not attributable to long-term differences in climatic factors or to chance variations of individual large storms.

Factors found to be most significant in influencing watershed runoff rates are: (1) basin storage capacity for transient surplus water; (2) the nature of groundwater storage systems; (3) slope factors; and (4) minor factors such as basin shape and land use.

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CHAPTER I

INTRODUCTION

It has previously been established that streamflow in one Cumberland Plateau basin is anomalous, that peak flows are extremely high and low flows minimal despite a large number of basin characteristics that are usually associated with stable streamflow regimes (Mayfield, 1979; 1980). In that earlier study, it was shown that the streamflow regime of the Obed-Emory was one of great extremes even though the basin has low drainage density, gentle to moderate slopes, heavy forest cover, permeable soils, and less intense rainfall than many nearby watersheds. The Obed basin was compared to two other East Tennessee watersheds of comparable size, the Collins and the Nolichucky, in terms of runoff characteristics and basin properties. It was determined that despite the aforementioned basin characteristics, the Obed watershed is able to move water rapidly through its shallow but highly permeable soils, concentrating runoff over a short period of time. The impermeable bedrock and shallow soils of the basin provide only minimal storage, so water is forced to leave the watershed rapidly as streamflow.

In this paper, the nature of streamflow is examined in other Cumberland Plateau basins in order to determine the degree to which other streams in the region conform to the pattern of the Obed-Emory and to attempt to explain the observed differences. The study area includes all of the tabular Cumberland Plateau in Tennessee (see Figs. I-1, I-2). The more heavily dissected Cumberland Mountain section

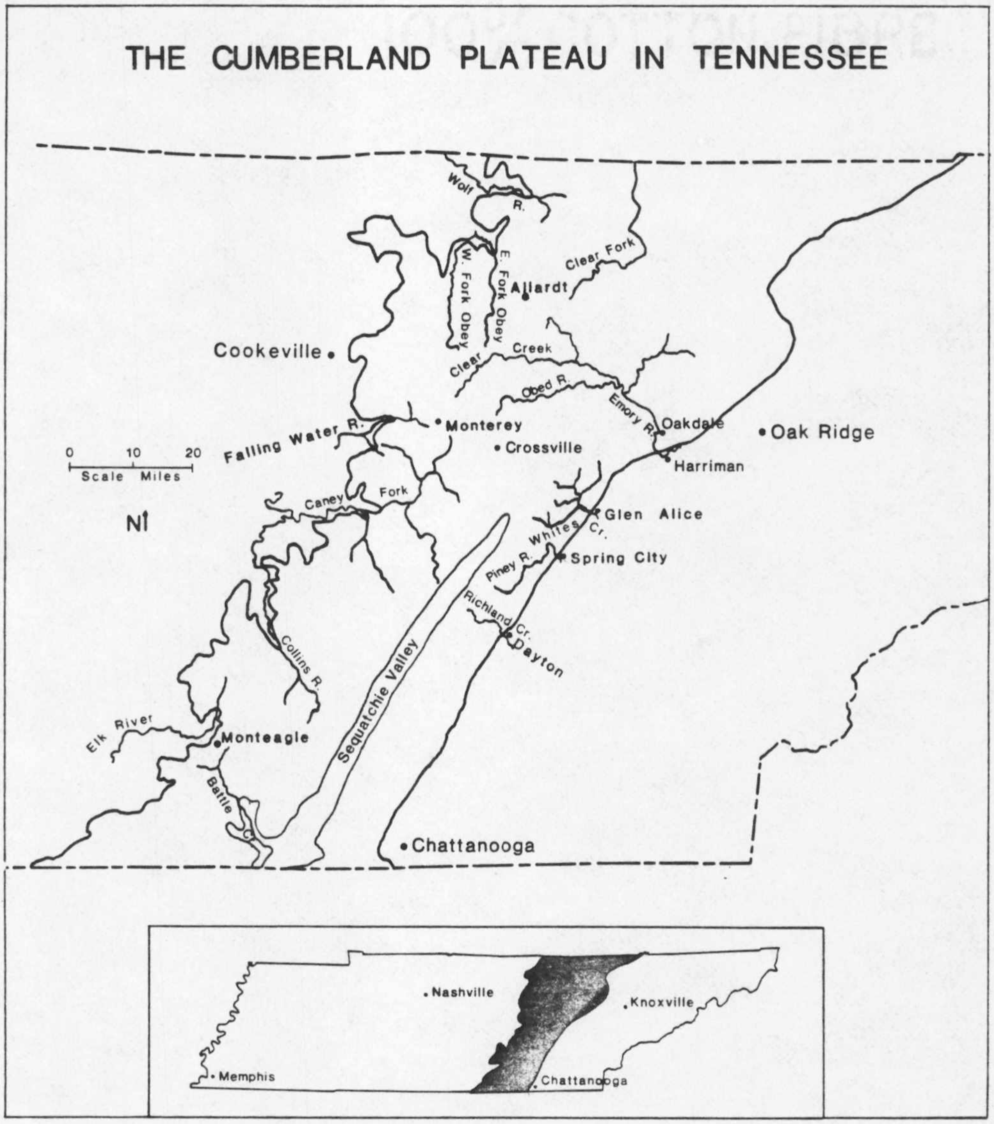

Figure I-1. The Study Area.

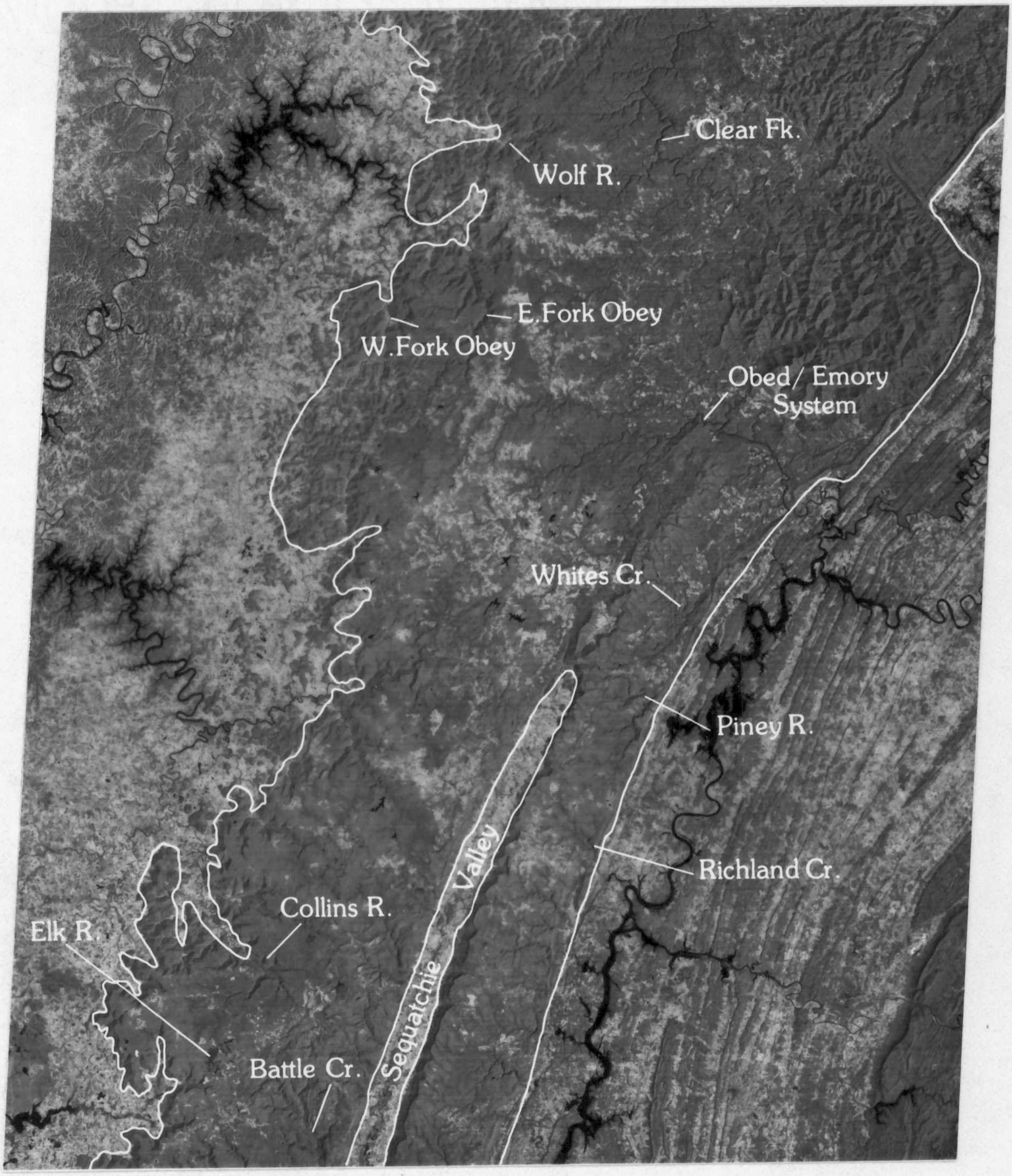

Figure I-2. LANDSAT Image of the Study Area.

is excluded from the study, because the topography is so markedly different from that of the Plateau section and because extensive mining of coal in that area has affected the hydrology (Hufschmidt, 1981). The study is limited to the state of Tennessee because of difficulties in obtaining streamflow data from the adjacent states of Alabama and Kentucky and because the topography that is characteristic of the study area does not extend greatly beyond the Tennessee border. Thirteen gaged streams with drainage areas of 50 to 764 square miles have been chosen for analysis. Additional gages exist on smaller basins, but these are not used because of the extreme variability in hydrologic response of very small watersheds.

Streams draining the tabular section of the Cumberland Plateau in Tennessee exhibit much more extreme flow duration curves than most other streams in the state; peak flows are very high and low flows are especially low (Figure I-3). As shown in Figure I-4, the mean annual flood of Plateau streams is much higher than the average for the state. This graph shows the mean annual flood for all streams in Tennessee which drain areas larger than 100 square miles and smaller than 1,000 square miles.

Regression equations developed by statewide studies for the prediction of peak flows seriously underpredict flood flows for streams of the Plateau. For example, a study by the U.S. Geological Survey projects 2-year floods of 2400 c.f.s. for Richland Creek and 18,700 c.f.s. for the Emory River, compared to observed values of 4,000 and 43,800 c.f.s., respectively (Randolph and Gamble, 1976). A similar study,

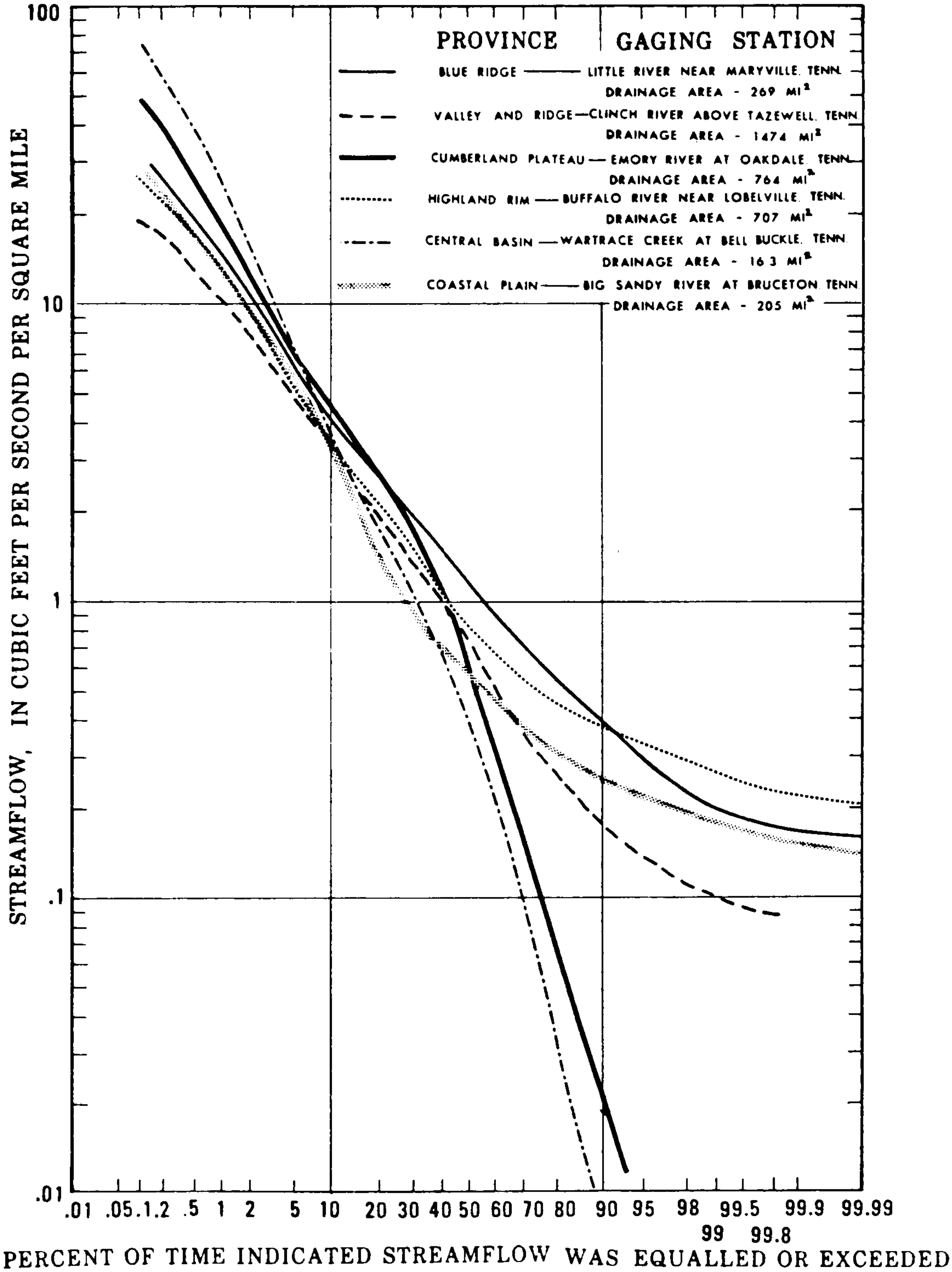

Figure I-3. Flow-Duration Curves of Tennessee Streams.

Source: A. Zurawski. 1978. Summary Appraisals of the Nation's Groundwater Resources: Tennessee Region. U.S.G.S. Professional Paper 813-L. Washington: U.S. Government Printing Office.

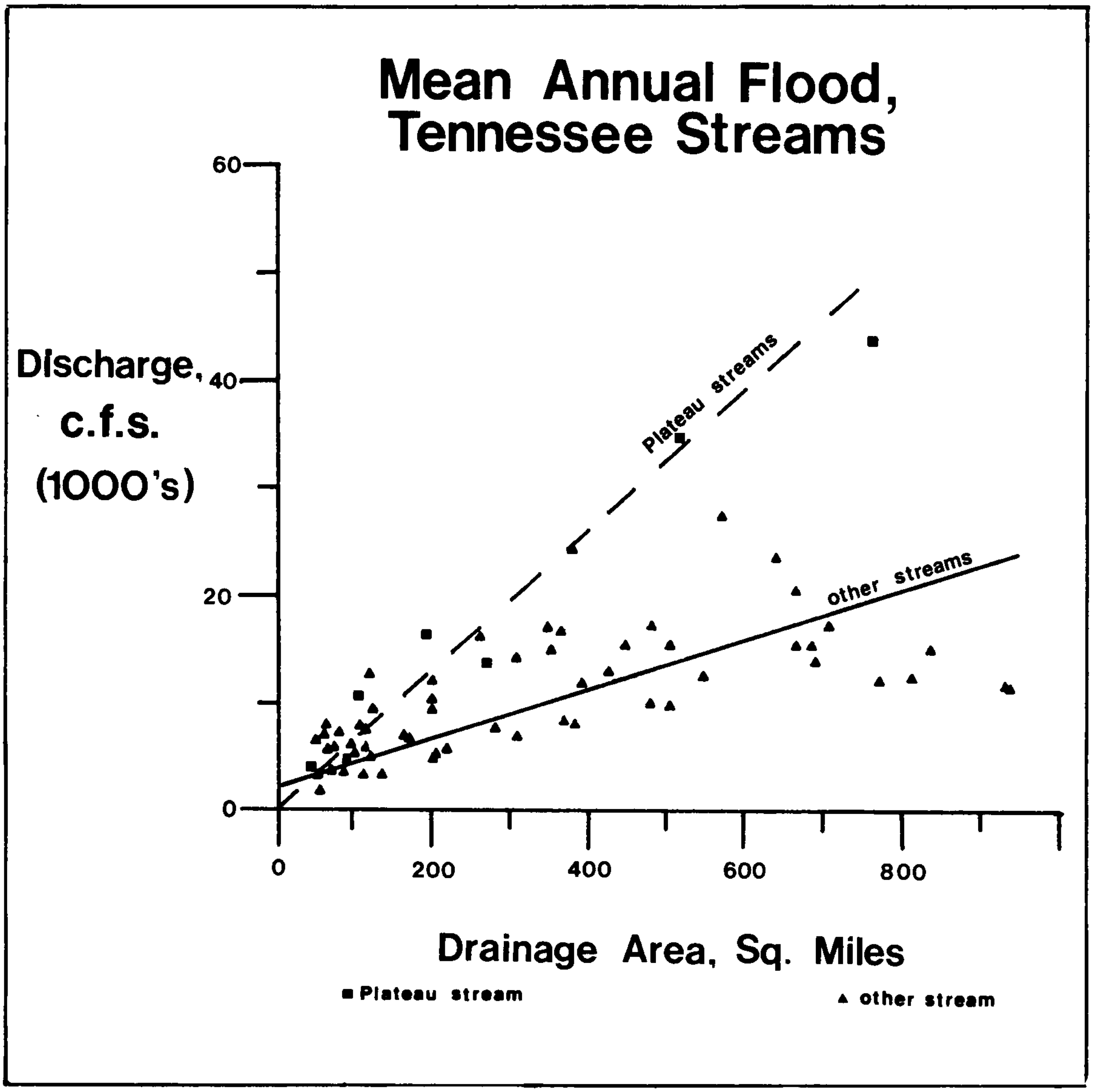

Figure I-4. Mean Annual Flood of Tennessee Streams.

Source: V.J. May, G.H. Wood, and D.R. Rima. 1970. A Proposed Streamflow-Data Network for Tennessee. U.S.G.S. Open-File Report. Nashville: U.S.G.S.

which considers more basin characteristics, indicates a 2-year flood of 17,000 c.f.s. for the Emory (May, Wood, and Rima, 1970). It is clear that the flow rates of these streams do not conform to those of other streams in the East Tennessee region.

Floods of Plateau streams have been a major problem from time to time for the cities that are built at their outlets to the adjacent Ridge and Valley Province and Highland Rim. Such cities include Spring City on the Piney River, Dayton on Richland Creek, and Oakdale and Harriman on the Emory River (T.V.A., 1961). At the same time, the extremely low summer and fall flows of these streams preclude their use as water supply streams unless they are impounded. One study showed that water-intensive industry is conspicuously absent in Plateau counties, presumably because of the difficulty of obtaining sufficient supplies of water (Garrison and Paulson, 1972).

In this paper, the differences in streamflow characteristics among streams draining the Cumberland Plateau are described, and an attempt is made to explain these differences on the basis of basin characteristics such as topography, vegetation, soils, geology, stream network configuration, and climate. Some of the geomorphic implications of the observed streamflow regimes are also investigated.

CHAPTER II

THE PHYSICAL SETTING

The Cumberland Plateau is a broad upland which extends in a northeast to southwest direction across Tennessee from the Kentucky border to the Alabama border. The Plateau does not stop at these political boundaries, but it does change substantially in character to the north and south, with the transformations in character roughly coincident with the state boundaries. In Tennessee, the Plateau is a true tableland over most of its extent, having an extensive gently-sloping upland. To the north the surface is much more dissected; to the south it is lower and discontinuous. Henceforward, references to the Cumberland Plateau are only to the Tennessee portion.

The Cumberland Plateau separates the lowlands of the Appalachian Valley and Ridge province to the east from the Highland Rim to the west. It averages 35 to 40 miles in width but is reduced to a width of only 20 to 25 miles near the Kentucky border by the uplifted Cumberland Mountain section. Upland elevations increase from 1500 feet near the Kentucky border(in the vicinity of Jamestown) to nearly 2000 feet at Monteagle in the south.

The southern half of the Plateau is divided into two parts by the linear Sequatchie Valley (see Figure II-1). Locally the eastern limb is referred to as Walden Ridge, while the wider western limb is known as the Plateau proper. The Sequatchie River occupies a breached anticline which extends for over 200 miles from central Alabama to a point near the lower Emory (Stearns, 1954). The northernmost section

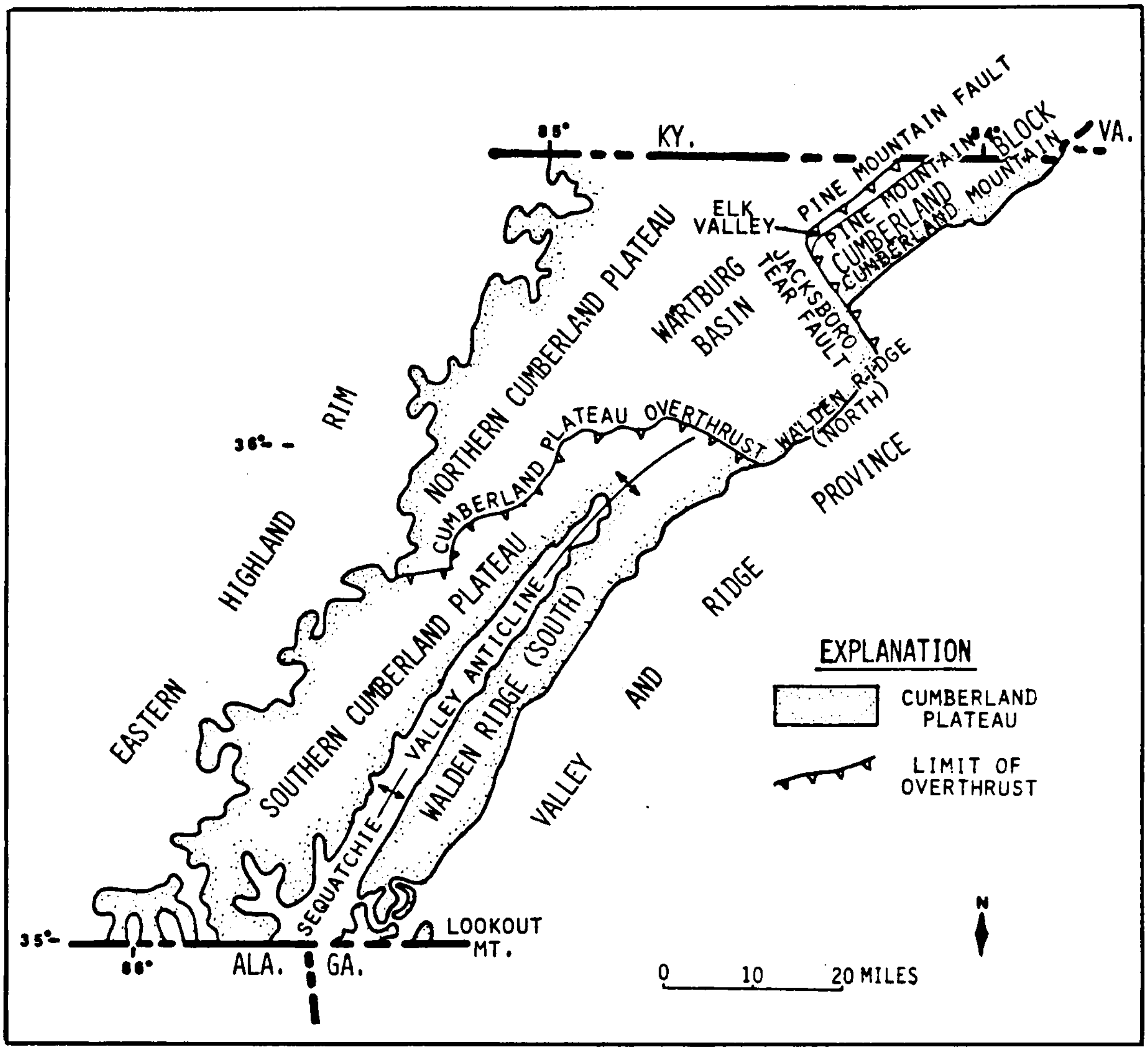

Figure II-1. Structural Features of the Cumberland Plateau.

Source: from E.T. Luther. 1959. The Coal Reserves of Tennessee. Nashville: Tennessee Division of Mines.

has not yet been breached by erosion and is expressed topographically as the Crab Orchard Mountains.

Most of the Plateau surface is gently rolling, but locally the character of the upland varies considerably (see Figure II-2). Near the Kentucky border, dissection by tributaries to the Big South Fork of the Cumberland River has produced a rather irregular surface. The central section of the Plateau, best seen along Interstate 40 from Crab Orchard to Monterey, is relatively flat. Slopes in this area are mostly in the 3 to 8 per cent range, and much of the land is cleared for agriculture. Of the two southern limbs, Walden Ridge has the more steeply sloping terrain, as short linear streams drain down the dip of the truncated Sequatchie Valley anticline. The flattest part of the Plateau is found in the southern parts of the western limb, especially in the Elk River and Collins River watersheds. Extensive areas with slopes gentler than 3% are found in Grundy and Van Buren Counties. The Plateau surface is apparently a stripped surface, as it is developed almost entirely on resistant sandstones.

The eastern escarpment of the Plateau is abrupt and nearly linear, rising 800 to 1000 feet above the adjacent valley of the Tennessee River. The valley walls of the Sequatchie are even more abrupt and undissected, as streams flow down the structure of the breached anticline, away from the valley. The Little Sequatchie and Big Fiery Gizzard are the only sizable streams that flow into the Sequatchie, which is fed mostly by springflow in its upper reaches. The western escarpment of the Plateau is also quite abrupt, but it is hardly

Figure II-2. The Plateau Surface Near Crossville.

linear. Streams such as the Elk, Collins, Caney Fork, Calfkiller, and Obey Rivers and their many tributaries have dissected this escarpment into a frayed, irregular pattern. Detached remnants of the original Plateau surface are common; examples include Alpine Mountain near Livingston, Goulden and Milksick Mountains near Sparta, and Ben Lomond Mountain near McMinnville.

Fenneman attributed the difference in character of the two escarpments to differences in structure and sapping (Fenneman, 1938). The eastern escarpment is formed along the steep western limb of an anticline, such that

> ...the beds therefore dip beneath the plateau. The fold is sufficiently simple and the dip sufficiently steep to cause the formations, when cut off by a horizontal surface, to outcrop in narrow, nearly straight bands; hence the straightness of the escarpment. The eastward dip at the western edge is slight (Fenneman, 1938).

The degree to which streams have entrenched themselves into the Plateau and dissected it varies considerably. The two large river systems of the Plateau are the Clear Fork-Big South Fork and the Obed-Emory. The former is over 90 miles in length, while the latter is over 65 miles in length. The morphology of these two systems is quite similar; both flow in shallow gorges in their upper reaches and are increasingly entrenched downstream. At their lower ends, the streams are embedded 300 to 400 feet below the flat upland surface. Vertical bluffs commonly account for more than 100 feet of this valley depth (see Figure II-3). The larger tributaries generally join the main stem at grade, but hundreds of smaller tributaries plunge over the cliffs as waterfalls, not having carved channels through the caprock. It would be difficult for the Obed or the Clear Fork to deepen their gorges up-

Figure II-3. Clear Creek Canyon.

stream significantly, because the necessary channel gradient is limiting on these long streams.

The other stream systems of the Plateau are different from these two large systems in that they have much shorter channels; each must descend the 1000 foot drop from Plateau upland to the Tennessee Valley, Sequatchie Valley, or Highland Rim in 30 miles or less. Not surprisingly, then, their channels are much steeper. These other streams lack the extensive vertical bluffs of the Obed or Big South Fork, but their gorges are much deeper, extend farther upstream, and bifurcate frequently to include even small tributaries. The gorges of Battle Creek, the Little Sequatchie, Crow Creek, Sweden Creek, and Big Fiery Gizzard Creek are all over 1000 feet deep, and those of the Elk, Collins, Caney Fork, Cane Creek, Calfkiller, Obey, and Wolf Rivers are 600 to 1000 feet deep.

The gorges of streams draining Walden Ridge are intermediate in character between the two aforementioned groups. Like the west-flowing streams, their gorges are deep and lacking in extensive vertical bluffs. Like the Obed and Big South Fork, the rivers are not deeply incised in their middle and upper reaches, and many tributaries enter discordantly. Examples of Walden Ridge streams include Whites Creek, the Piney River, Richland Creek, and North Chickamauga Creek.

Climate

The higher elevations on the Cumberland Plateau experience lower temperatures and greater amounts of precipitation than are found in the adjacent lower lands of the Highland Rim to the west or the Ridge and Valley to the east. This combination of higher precipitation and lower temperatures yields a greater water surplus for the Plateau; Crossville has an average annual surplus of 25.0 inches of water, while Knoxville's annual surplus is only 17.7 inches. Precipitation falls in greatest amounts during the winter, and fall is distinctly drier; March precipitation is nearly double that of October at each of the stations used in this study. Total annual precipitation on the Plateau varies from 52 inches at Allardt to 61 inches at Monteagle (see Table II-1).

Soils

The soils of the Cumberland Plateau are commonly "2 to 4 feet deep over rock, well drained, loamy, strongly acid, and low in natural fertility" (Springer and Elder, 1980, p. 41). On the broad uplands and relatively gentle side slopes, loamy soil grades rapidly to solid bedrock, with only a thin C horizon and rubble zone. Soils are thinner in some places but much deeper on most sideslopes of the dissected southern and western escarpment zones.

The soils of the gently-sloping Plateau surface at elevations of 1,600 to 2,000 feet are mostly of the Hartsells-Lonewood-Ramsey-Gilpin group (see Figure II-4). These soils are loamy, well-drained, 2 to 5 feet thick over bedrock, and have moderate to rapid permeability. They are generally formed from sandstone and shale residuum. These are the most favorable soils on the Plateau for agricultural use (Springer and Elder, 1980, p. 43).

TABLE II-1

AVERAGE PRECIPITATION ON THE CUMBERLAND PLATEAU

Station	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
Allardt	5.07	4.83	5.45	4.80	4.04	4.91	5.27	3.61	3.69	2.88	3.58	4.84	52.97
Crossville	5.62	5.81	5.90	4.55	3.85	4.18	4.67	4.28	3.61	2.78	4.05	5.36	54.66
Falls Creek Falls	4.98	5.36	5.33	4.27	3.78	3.86	5.60	3.91	3.35	2.38	3.85	5.01	51.68
Monteagle	5.97	6.30	6.76	5.69	4.39	4.38	5.45	4.12	3.69	3.17	4.66	5.99	60.57
Monterey	6.11	5.83	5.73	5.03	4.13	5.00	5.09	4.56	3.97	3.11	4.68	5.18	58.42

all values in inches

Source:University of Tennessee Agricultural Experiment Station. 1973. Precipitation Probabilities for East Tennessee. Knoxville: University of Tennessee.

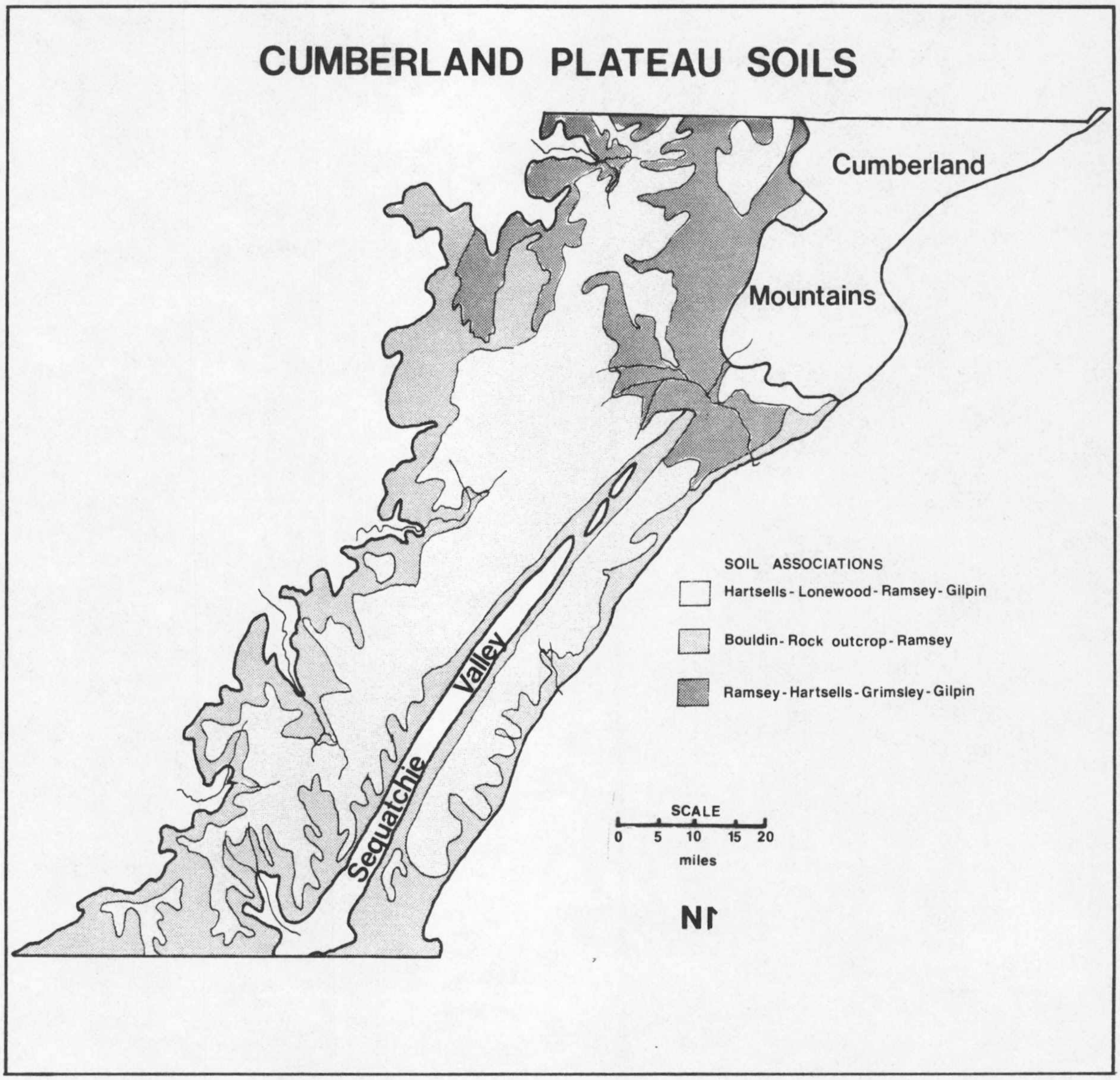

Figure II-4. Cumberland Plateau Soils.

Source: M.E. Springer and J.A. Elder. 1980. Soils of Tennessee. University of Tennessee Agricultural Experiment Station Bulletin 596. Knoxville: University of Tennessee.

On the steeper slopes of the southern half of the Plateau are found the Bouldin-Rock Outcrop-Ramsey soils. The Bouldin are deep and stony colluvial soils, while the less extensive Ramsey soils are rocky and shallow. Permeability is moderately rapid to rapid. Although visually impressive and conspicuous, the rock outcrop areas are not areally extensive (Springer and Elder, 1980, p. 43).

Ramsey-Hartsells-Grimsley-Gilpin soils occupy the steep and hilly landscape segments on the northern half of the Plateau. These are well-drained, loamy, and stony soils that range from 2 to 6 feet in depth. Approximately 50 percent of this area is covered by Hartsells and Ramsey soils. The Hartsells and Ramsey soils average only two feet in depth. Gilpin soils average 32 inches in depth and have moderate permeability, while Grimsley soils are deeper and stonier (Springer and Elder, 1980, p. 44).

Geology

The stratigraphy of the Plateau consists of a thick sequence of Pennsylvanian shales, sandstones, conglomerates, and coal underlain by massive Mississippian limestone units (see Figure II-5). Over the majority of the tabular Plateau, all but the lower Pennsylvanian beds have been removed by erosion. Shales are the most abundant rocks, in that their aggregate thickness exceeds that of the other rock types. Most of the Plateau surface, however, is developed on one of the resistant conglomerate or sandstone units such as the Rockcastle Conglomerate or the Crossville Sandstone (Luther, 1959).

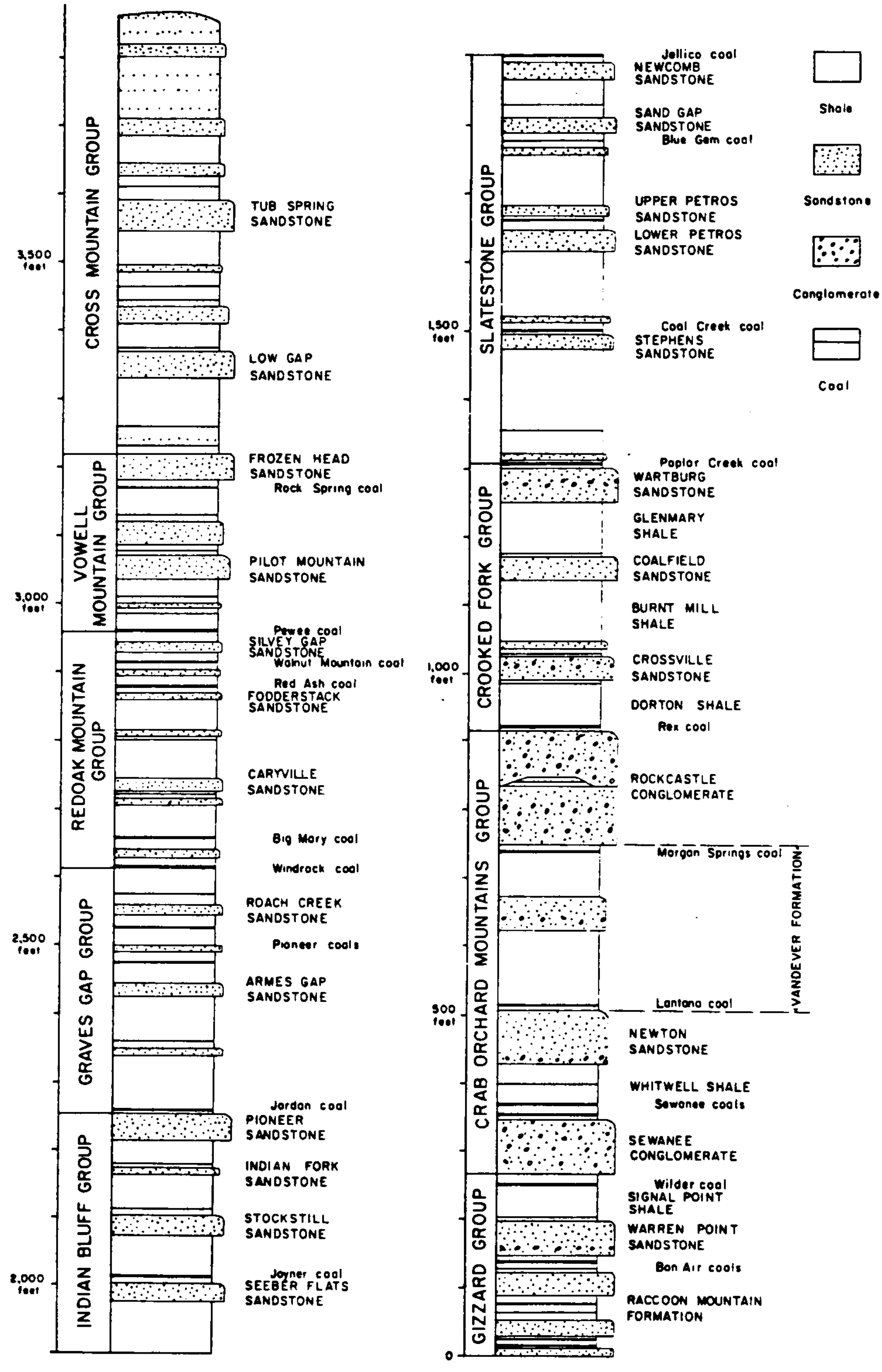

Figure II-5. Stratigraphy of the Cumberland Plateau.

Source: from R. Newcombe and O. Smith, Jr. 1958. Ground-Water Resources of the Cumberland Plateau in Tennessee. Nashville: Division of Water Resources, Tennessee Division of Conservation.

Beneath much of the Plateau rock strata dip northeastward while surface elevations remain nearly constant, so that the older Mississippian limestone units are closer to the surface on the western and southern flanks of the upland and are exposed in the valley walls of entrenched streams in those areas.

The Sequatchie Valley anticline bisects the southern two-thirds of the Plateau. It originates in northern Alabama and terminates at the Emory River fault zone near Camp Austin (see Figure II-1) (Stearns, 1954, pp. 1-3). The majority of the anticline has been breached by collapse and erosion to form the linear Sequatchie Valley. The northernmost 20 miles, which has not been breached, forms the Crab Orchard Mountains. Upper Mississippian strata are exposed extensively in the arched structure of the Sequatchie Valley and in several fensters such as Grassy Cove.

Groundwater is found in limited quantities within fractured sandstone beds that are confined between intermittent shale beds (Newcombe and Smith, 1958, p. 1). This groundwater is commonly under significant artesian pressure, so that water levels commonly rise as much as 100 feet in wells that penetrate the aquifers (Wilson, 1965, p. 37). Such artesian pressure would indicate that the groundwater system in the upper rock layers is not communicating freely with streams or spring systems.

CHAPTER III

STREAMFLOW CHARACTERISTICS OF CUMBERLAND PLATEAU STREAMS

General Characteristics

Cumberland Plateau streams differ greatly in the production of flood peaks (Figure III-1). Whites Creek is clearly the champion flood producer of the group, as its 25-year flood exceeds 300 c.f.s./square mile, which is three times the flow rate of the same flood for Clear Fork. In descending order between these two extremes are Richland Creek, Wolf River, East Fork of the Obey, Emory River, Battle Creek, Elk River, West Fork of the Obey, Caney Fork, Falling Water River, Collins River, and Calfkiller River.

These thirteen streams can be divided into two groups for general discussion, with the first group consisting of those five streams whose unit-area floods are greater than the mean for the region as a whole, and the second group consisting of the eight remaining streams. Whites Creek, Wolf River, Richland Creek, East Fork of the Obey, and Emory River comprise the first group, while the second group includes Battle Creek, Elk River, West Fork of the Obey, Clear Fork, Falling Water River, Collins River, and Calfkiller River. The second group represents a continuum of flood values from Battle Creek at the top to the Calfkiller River at the bottom, while there is considerably more variation in the upper group. (See Figure III-1, Table III-1). The low flow characteristics of these streams can also be used for categorization,

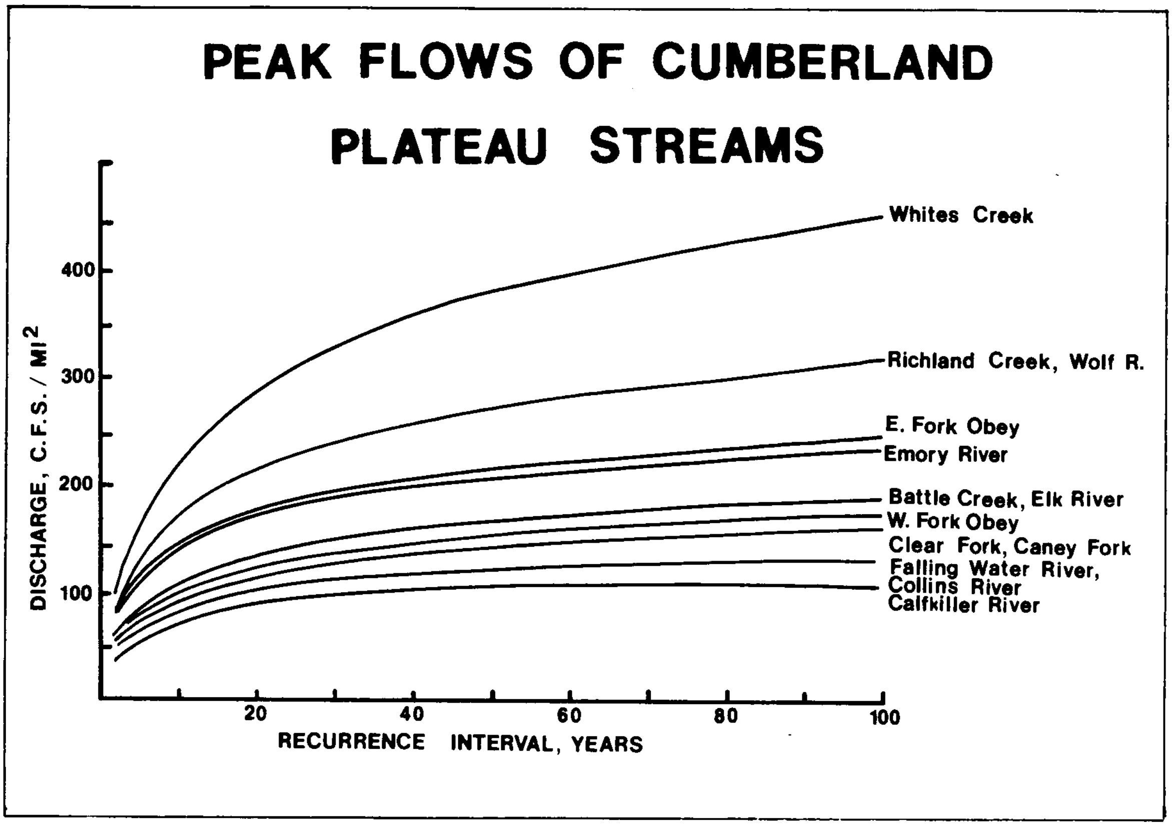

Figure III-1. Peak Flows of Cumberland Plateau Streams.

Source: V.J. May, G.H. Wood, and D.R. Rima. 1970. A Proposed Streamflow-Data Network for Tennessee. U.S.G.S. Open-File Report. Nashville: U.S.G.S.

TABLE III-1

STANDARDIZED FLOWS OF CUMBERLAND PLATEAU STREAMS

Stream	Flood Flows 2-yr.	5-yr.	10-yr.	25-yr.	Maximum Known Flood	Flow Duration, % 99	95	50	10	5	2	1	Low Flows 2-yr. 7-day	10-yr. 7-day	Ratio of March:October Mean Flow
	cfs/mi.^{2}				cfs/mi.^{2}	cfs/mi.^{2}									
Whites Creek	99	169	225	306	611	.002	.004	.51	4.9	8.1	13.8	20.1	.004	.001	
Richland Creek	81	134	173	227	279										
Wolf River	75	127	168	225	213	.053	.080	.64	4.1	6.4	11.9	17.8	.075	.048	18:1
E. Fork Obey	81	119	145	180	222	.023	.046	.74	4.7	7.9	13.6	19.7	.045	.024	18:1
Emory River	57	93	120	158	255	.002	.009	.65	4.6	7.3	12.9	19.1	.008	.001	21:1
Battle Creek	77	107	126	149	202										
Elk River	64	95	117	145	241	.027	.041	.81	5.1	8.4	14.4	20.3	.050	.021	14:1
Caney Fork	56	82	101	124	140	.001	.004	.59	4.8	7.8	13.6	20.1	.004	.001	
Clear Fork	51	77	96	121	131	.010	.021	.52	4.1	6.7	11.7	17.5	.017	.005	24:1
W. Fork Obey	62	89	109	113	132	.031	.056	.43	3.2	5.4	10.0	14.8	.040	.027	21:1
Collins River	36	57	70	88	118	.102	.139	.77	4.1	6.5	11.5	16.4	.141	.098	12:1
Calfkiller River	44	59	69	79	143	.106	.144	.85	5.0	7.8	13.3	18.1	.152	.098	12:1

Sources: V.J. May, G.H. Wood, and D.R. Rima. 1970. A Proposed Streamflow-data Program for Tennessee. Nashville: U.S. Geological Survey; U.S. Geological Survey. 1980. Water Resources Data for Tennessee. Washington: Government Printing Office; R.L. Gold. 1981. Low-Flow Frequency and Flow Duration of Tennessee Streams. Nashville: U.S. Geological Survey.

but the results are much the same; those streams with high peak flows generally have poorly sustained low flows (see Figure III-2, Table III-1). One major exception is the Wolf River, which has very high peak flows, but joins the Calfkiller, Collins, and West Fork of the Obey in having well-sustained flows.

Base flow is the flow sustained during long dry periods by seepage from groundwater and by springflow. According to May,

> Those streams on the Cumberland Plateau, however, show much lower minimum monthly flows per square mile than do streams in the other two physiographic provinces. This indicates that these streams do not sustain flow well during periods of no rainfall which is a result of either poor infiltration qualities of the land surface or the poor ability of the underground aquifer systems to store water, or both. (May, 1981, p. 30).

Presumably the differences in base flow characteristics of the Plateau streams can be explained by differences in the aquifer systems or infiltration and storage capacities of the soils and bedrock of the watersheds.

The large variation in low-flow characteristics among Plateau streams is best demonstrated by the use of a flow duration curve. Such curves are presented in Figure III-2 for six representative streams. The curve for the Collins is nearly identical to that of the Calfkiller, those of the East and West Forks of the Obey are very similar to the Elk's and that of Whites Creek is similar to the Emory's (see Table III-1). Bee Creek has by far the most poorly sustained low flows, followed by the Emory, Caney Fork, Whites Creek, Clear Fork, the East and West Forks of the Obey, the Elk River, and the Wolf River. The variation is such that the flow exceeded 99 percent of the time on the Wolf

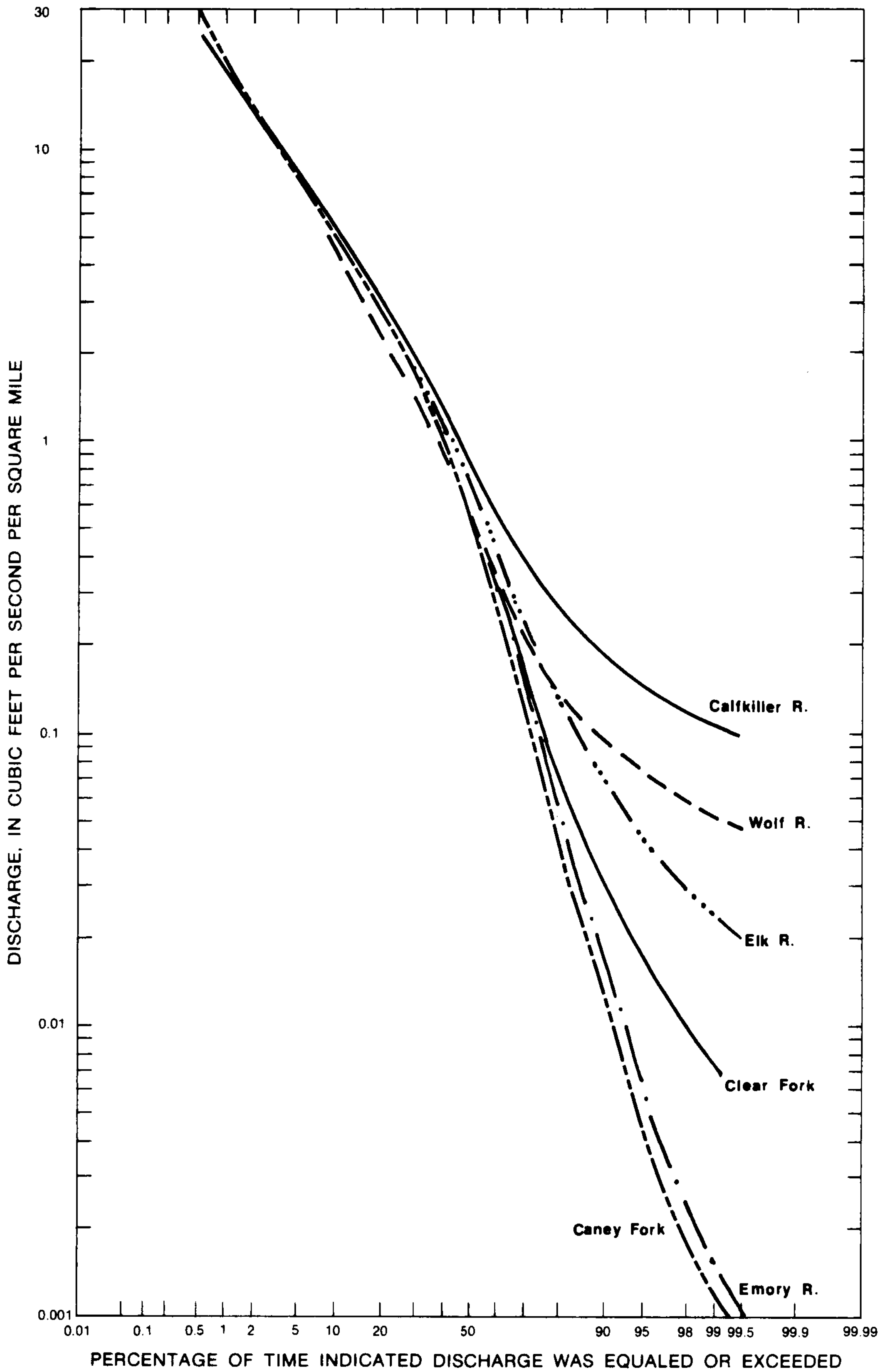

Figure III-2. Flow Duration Curves of Cumberland Plateau Streams.

River is more than 40 times as great as the same flow for Whites Creek, Caney Fork, or the Emory, but only twice as great as that of the Elk or the East and West Forks of the Obey. The low flow events, such as the 10-year, 7-day low flow, show similar patterns and magnitudes of variation among the streams being studied.

Floods of Plateau streams have occasionally caused serious damage to towns built on alluvial fans where the streams reach the gentler slopes of the adjacent Highland Rim or Valley and Ridge. By far the most damaging was the flood of March, 1929, which was caused by rainfall exceeding 15 inches in 24 hours in the Emory and Whites Creek watersheds. Other large floods occurred in 1957 and 1973. The potential for property damage and loss of life has increased greatly in some locations because of increased development in flood-prone areas, greater disturbance of watersheds, and channel encroachment. A better understanding of flood probabilities and magnitudes could help to avoid such problems.

The poorly sustained low flows of Plateau streams can cause problems for communities that rely upon them for water supply, for recreationists, and for wildlife. Better definition of low-flow characteristics is crucial to planning of water supply and wastewater treatment systems and for recreational enterprises such as whitewater rafting. Unfortunately, the Plateau streams with the best-sustained flows are ones that are generally unpopulated and are too small or too gently sloping for whitewater sport.

Flood characteristics of the streams studied vary considerably.

In figure III-1 and Table III-1, the 2-, 5-, 10-, 25-, 50-, and 100-year floods for gaged streams on the Plateau are presented. The shorter-recurrence floods are considered to be more indicative of basin flood yields, as they represent frequently-occurring events caused by more uniform storm events of moderate intensity. The infrequent floods, such as the 25- and 50-year floods, are more likely to be diagnostic of extreme storm events that have not occurred within some of the basins during the rather short periods of record. The flows for flood events are standardized to cubic feet per second per square mile of drainage area for purposes of comparison.

Low-flow characteristics of the thirteen study streams appear to be as variable as flood habits (see Table III-1, Figure III-2). Low flows are probably derived largely from groundwater seepage into stream channels, although some recent reports have suggested that delayed discharge from streamside interflow zones can be a significant component of base flow (Beaslev 1976, pp. 955-957; Whipkey, 1967, p. 257; Hewlett and Hibbert, 1967, pp. 275-289; Kirkby and Chorley, pp. 5-12; Nutter, 1973; Aubertin, 1973). The generally accepted explanation of differences in base flow characteristics of streams is summarized by Linsley, Kohler, and Paulhus:

> Basins having permeable surface soils and large, effluent groundwater bodies show sustained high flow throughout the year, with a relatively small ratio between flood flow and mean flow. Basins with surface soils of low permeability or influent groundwater bodies have higher ratios or peak to average flows and very low or zero flows between floods. (Linsley, Kohler, and Paulhus, 1975, pp. 339-340).

The streams of the Cumberland Plateau generally have a combination of these two characteristics; surface soil permeabilities are quite high, but the available volume of groundwater storage is quite small (Zurawski, 1978, p. L-15). Significant exceptions are those streams that have incised themselves into fractured and jointed limestones, in which large solution openings have often developed. It has been noted, however, that such openings do not necessarily detain groundwater and deliver it slowly. For example, one source suggests that:

> In limestone terrane, groundwater frequently moves at relatively high velocities as turbulent flow through solution channels and fractures in the limestone. Streams in limestone country often exhibit a high ratio of flood-peak flows to average flow, a condition characteristic of streams having small groundwater contributions. (Linsley, Kohler, and Paulhus, 1975, p. 225).

Such streamflow regimes are common to streams draining the limestone terranes of the Highland Rim and Central Basin of Tennessee (Zurawski, 1978). Similar conditions of large interconnected bedrock openings have, however, been linked to strongly homogeneous flow regimes in basins such as the Deschutes of Oregon (Shelton, 1981). Clearly, additional factors must be considered, such as the level of the groundwater table relative to such passages.

Low flows of Cumberland Plateau streams vary even more than peak flows of such flows are compared in the most commonly accepted manner, by plotting flow duration curves on logarithmic graph paper (see Figure III-2). The use of a logarithmic scale may, however, exaggerate differences in flow rates beyond that which is significant.

If the low flows presented in Table III-1 are plotted on arithmetic paper, most of the differences disappear, leaving only the

Collins and Calfkiller Rivers as being significantly different from the others.

Another way of comparing flows that takes drainage area into account is to determine the amount of water discharged during a given period, in inches of runoff. Using this technique it is also possible to determine the volume of water stored in each basin, to be released as baseflow. Water budgets for several stations on the Plateau indicate that no water surpluses are generated after the middle of June. Streamflow which occurs from July 1 - October 31 is, then, assumed to be water that is stored as gravity water in soils and regolith as well as groundwater that is subject to discharge into streams. For the eight study basins that have continuous gages, average discharge values for the months of July, August, September, and October have been converted into volumes of water and then into inches of water. The results are presented in Table III-2. The amount of water discharged as baseflow over this four-month period was found to range from a low of 1.24 inches for the West Fork of the Obey to a high of 2.68 inches for the Calfkiller. These figures are all probably somewhat high, because occasional unusually wet periods during the summer can raise the mean values. This is demonstrated by the fact that the mean discharge of the Emory River in September is 210 c.f.s., while the median is only 63 c.f.s. When low flows are viewed from this perspective, their variations are not nearly as great as the basin-to-basin variations in flood discharge.

TABLE III-2

MEAN RUNOFF,JULY-OCTOBER

	All Values in Inches of Runoff				
	July	August	September	October	Total
Emory River	0.68	0.40	0.31	0.23	1.62
Clear Fork	0.67	0.37	0.28	0.20	1.52
East Fork Obey	0.73	0.36	0.37	0.33	1.79
West Fork Obey	0.53	0.27	0.24	0.20	1.24
Wolf River	0.72	0.27	0.30	0.25	1.54
Calfkiller River	1.03	0.61	0.59	0.45	2.68
Collins River	0.79	0.59	0.44	0.39	2.21
Elk River	0.86	0.53	0.32	0.38	2.09

Climatic And Meteorologic Influences

In order to compare the climatically determined runoff generation characteristics of the Cumberland Plateau, water balances were calculated for Allardt, Crossville, and Monteagle (see Tables III-3, III-4, III-5). It would have been desirable to include a water budget for each of the individual basins, but detailed climatic data were available only for these three stations. These stations are representative of the northern, central, and southern sections of the Plateau, respectively.

Precipitation averages 60.6 inches per year at Monteagle, 54.0 inches at Crossville, and 53.0 inches at Allardt. Although mean monthly temperatures also decline from south to north, the resulting difference in evapotranspiration is not sufficient to counterbalance the disparity in precipitation. Thus, the calculated runoff is 30.8 inches per year at Monteagle, 25.1 inches at Crossville, and 23.0 inches at Allardt. These figures are in agreement with observed values of runoff in nearby streams; the Elk River has a mean annual discharge of 29.6 inches, the Emory's discharge is 25.0 inches per year, and that of the Wolf River is 24.2 inches (U.S. Geological Survey, 1980).

These water budgets provide a convenient check of runoff values for streams in the study area. If a stream were losing or gaining significant amounts of water to or from adjacent basins, the observed annual discharge values would differ from the calculated values significantly. Such gains and losses of water could greatly influence the low-flow characteristics of a stream and could affect peak rates to a

TABLE III-3

ALLARDT WATER BUDGET

All Figures in Inches

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Year
Precipitation	5.1	4.8	5.5	4.8	4.0	4.9	5.3	3.6	3.7	2.9	3.6	4.8	53.0
Potential Evapo-transpiration	0.0	0.3	1.2	2.3	2.6	4.8	5.6	5.2	3.7	2.0	1.0	0.3	30.0
Actual Evapotrans-piration	0.0	0.3	1.2	2.3	3.6	4.8	5.6	5.2	3.7	2.0	1.0	0.3	30.0
Change in Storage	-	-	-	-	-	-	-0.3	-1.6	-	+0.9	+1.0	-	
Storage	4.0	4.0	4.0	4.0	4.0	4.0	3.7	2.1	2.1	3.0	4.0	4.0	
Deficit	-	-	-	-	-	-	-	-	-	-	-	-	00.0
Surplus	5.1	4.5	4.3	2.5	0.4	0.1	-	-	-	-	1.6	4.5	23.0
Carryover	2.6	3.8	4.1	4.2	3.3	1.8	0.9	0.4	0.2	0.1	0.0	0.8	
Total	7.7	8.3	8.4	6.7	3.7	1.9	0.9	0.4	0.2	0.1	1.6	5.3	
Runoff	3.9	4.2	4.2	3.4	1.9	1.0	0.5	0.2	0.1	0.1	0.8	2.7	23.0

TABLE III-4

CROSSVILLE WATER BUDGET

All Figures in Inches

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Year
Precipitation	4.9	4.9	5.7	4.9	4.4	4.3	5.2	3.7	3.5	2.9	4.1	5.5	54.0
Potential Evapo- transpiration	0.3	0.5	0.9	2.0	3.6	4.8	5.6	5.2	3.7	2.0	0.8	0.3	29.7
Actual Evapotrans- piration	0.3	0.5	0.9	2.0	3.6	4.8	5.5	4.6	3.6	2.0	0.8	0.3	28.9
Change in Storage	-	-	-	-	-	-0.5	-0.3	-0.9	-0.1	+0.9	+0.9	-	
Storage	3.0	3.0	3.0	3.0	3.0	2.5	2.2	1.3	1.2	2.1	3.0	3.0	
Deficit	-	-	-	-	-	-	-	-	-	-	-	-	00.0
Surplus	4.6	4.4	4.8	2.9	0.8	0.0	0.0	0.0	0.0	0.0	2.4	5.2	25.1
Carryover	3.2	3.9	4.1	4.4	3.6	2.2	1.1	0.5	0.2	0.1	0.0	1.2	
Total	7.8	8.3	8.9	7.3	4.4	2.2	1.1	0.5	0.2	0.1	2.4	6.4	
Runoff	3.9	4.2	4.5	3.7	2.2	1.1	0.6	0.3	0.1	0.1	1.2	3.2	25.1

TABLE III-5

MONTEAGLE WATER BUDGET

All Figures in Inches

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Year
Precipitation	6.0	6.3	6.8	5.7	4.4	4.4	5.5	4.1	3.7	3.2	4.7	6.0	60.8
Potential Evapo-transpiration	0.3	0.3	0.9	2.3	4.0	4.7	5.5	5.2	3.7	2.0	0.8	0.3	30.0
Actual Evapotrans-piration	0.3	0.3	0.9	2.3	4.0	4.7	5.5	5.2	3.7	2.0	0.8	0.3	
Change in Storage	-	-	-	-	-	-0.3	-	-1.1	-	+1.2	+0.2	-	
Storage	4.0	4.0	4.0	4.0	4.0	3.7	3.7	2.6	2.6	3.8	4.0	4.0	
Deficit	-	-	-	-	-	-	-	-	-	-	-	-	00.0
Surplus	5.7	6.0	5.9	3.4	0.4	-	-	-	-	-	3.7	5.7	30.8
Carryover	3.7	4.7	5.3	5.6	4.5	2.4	1.2	0.6	0.3	0.1	0.0	1.8	
Total	9.4	10.7	11.2	9.0	4.9	2.4	1.2	0.6	0.3	0.1	3.7	7.5	
Runoff	4.7	5.4	5.6	4.5	2.5	1.2	0.6	0.3	0.2	0.1	1.9	3.8	30.8

somewhat lesser extent. None of the streams being studied shows evidence of large-scale interbasin loss or gain.

One possible interpretation of the difference in recorded flood peaks of Cumberland Plateau streams would be that there is no real difference in the hydologic response of the basins, but that flooding has varied through time. This hypothesis is worth investigating, because the periods of record on which May, Wood, and Rima (1970) based their estimates of flood flows varied from 42 years for the Emory River to 15 years for Battle Creek. In addition, the records of some stations included in the study do not overlap in time at all. The greatest observed flood peaks for most Plateau streams occurred when only a few of the stations were operative. The calculated peak flows of the Emory and Collins Rivers, whose gages were operative during the great flood of March 1929, may, therefore, be unduly biased by that single event.

In order to eliminate the possible effects of this difference in the period of record, annual flood peaks of 2-, 5-, 10-, and 25-year recurrence intervals were recalculated for the eight basins which were gaged during the period of the 1962-1979 water years. By including only the peak flows observed during this 18 year period, the non-random effects of temporal variation should be eliminated.

The method used in estimating the flood magnitudes was the Log-Pearson Type III method, which was used by May, Wood, and Rima and is recommended by the U.S. Water Resources Council (U.S.W.R.C., 1967). It is to be expected that the events of infrequent recurrence would be overestimated when using only an 18 year period of observation, because a high skew coefficient results in higher flood estimates (Linsley, Kohler, and Paulhus, 1975, pp. 339-340).

The recalculated flood flows are presented in Table III-6 and Figure III-3. Complete values for the Log-Pearson Type III calculations for each system are found in Appendix A. Based upon this analysis, the hypothesis that the differences in flood-flow magnitude among Plateau streams is due to temporal variation must be rejected. Although the overall magnitude of the values changed significantly, the relative positions and proportional differences between basins did not change significantly. These results merely reinforce the statement of spatial differences in flow characteristics discussed previously; the differences in hydrologic response must be explained by basin characteristics.

It must be emphasized that the use of this 18-year period of record significantly reduces the probable accuracy of the absolute values of flood peaks for streams such as the Emory, Clear Fork, and the Wolf, for which much longer records are available (see Figure III-4). The value of this analysis is merely that of eliminating differences in hydrologic response that may be attributable to temporal variation.

TABLE III-6

RECALCULATED FLOOD PEAKS

	Recurrence Interval				
	2-Year	5-Year	10-Year	25-Year	50-Year
Emory River	43,666(57)	75,162(98)	101,625(133)	146,218(191)	189,234(248)
Clear Fork	12,853(47)	19,187(71)	24,266(89)	31,915(117)	38,459(141)
Wolf River	9,440(89)	15,966(150)	20,324(192)	25,527(241)	29,174(275)
Whites Creek	11,668(108)	22,029(204)	31,623(293)	47,424(439)	62,373(576)
Richland Creek	5,236(104)	8,472(169)	10,641(212)	13,366(266)	15,382(306)
East Fork Obey	16,255(80)	25,527(126)	33,266(165)	45,082(223)	55,719(276)
Elk River	4,764(73)	7,998(122)	10,864(166)	15,311(233)	19,275(294)
Battle Creek	3,899(77)	5,129(102)	5,998(119)	7,145(142)	8,035(159)

Values in c.f.s. Values in parentheses standardized to c.f.s./mi.^{2}

Source: calculated from data in U.S.G.S. 1962-79(annual reports). Water Resources Data for Tennessee. Washington: Government Printing Office.

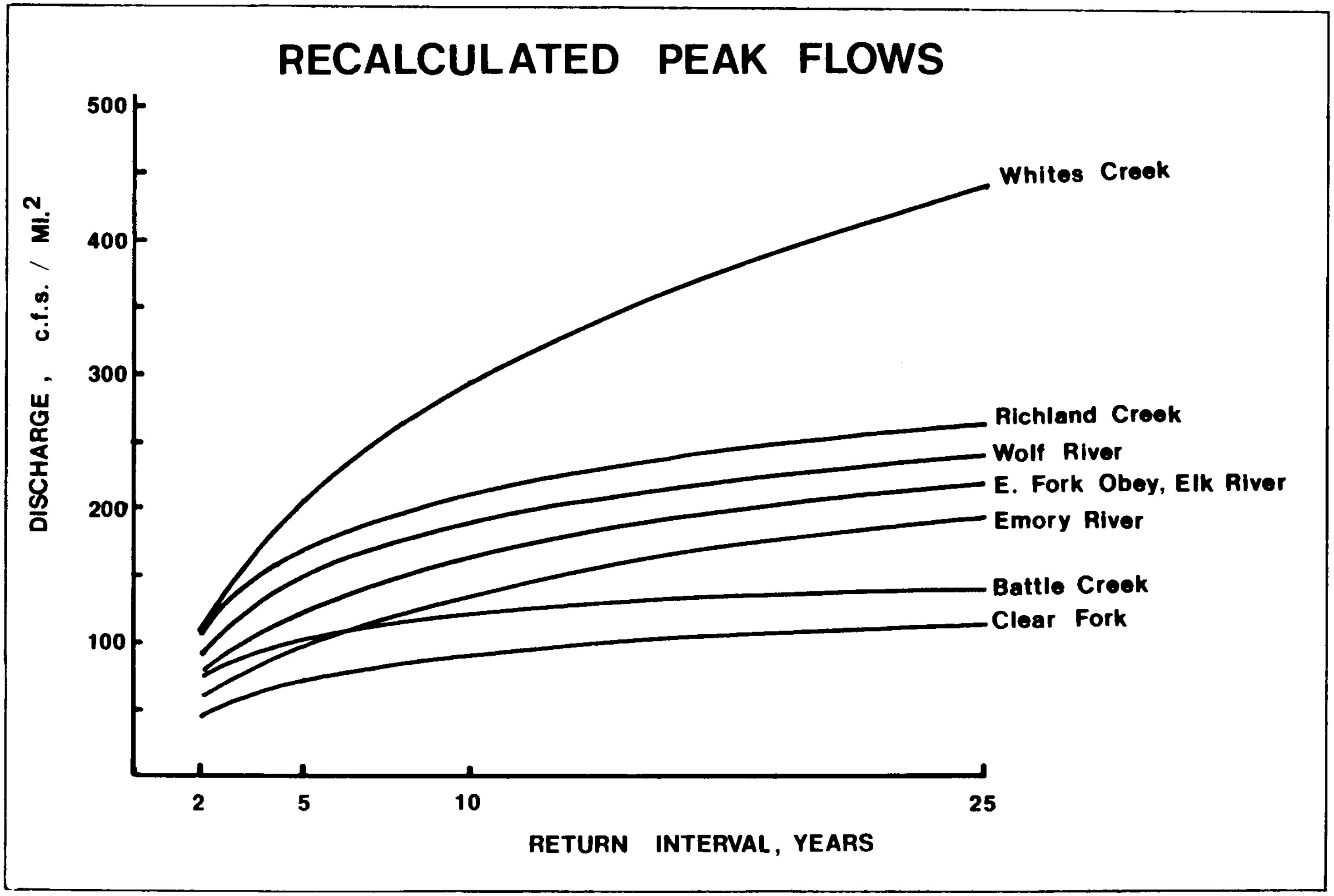

Figure III-3. Recalculated Flood Peaks.

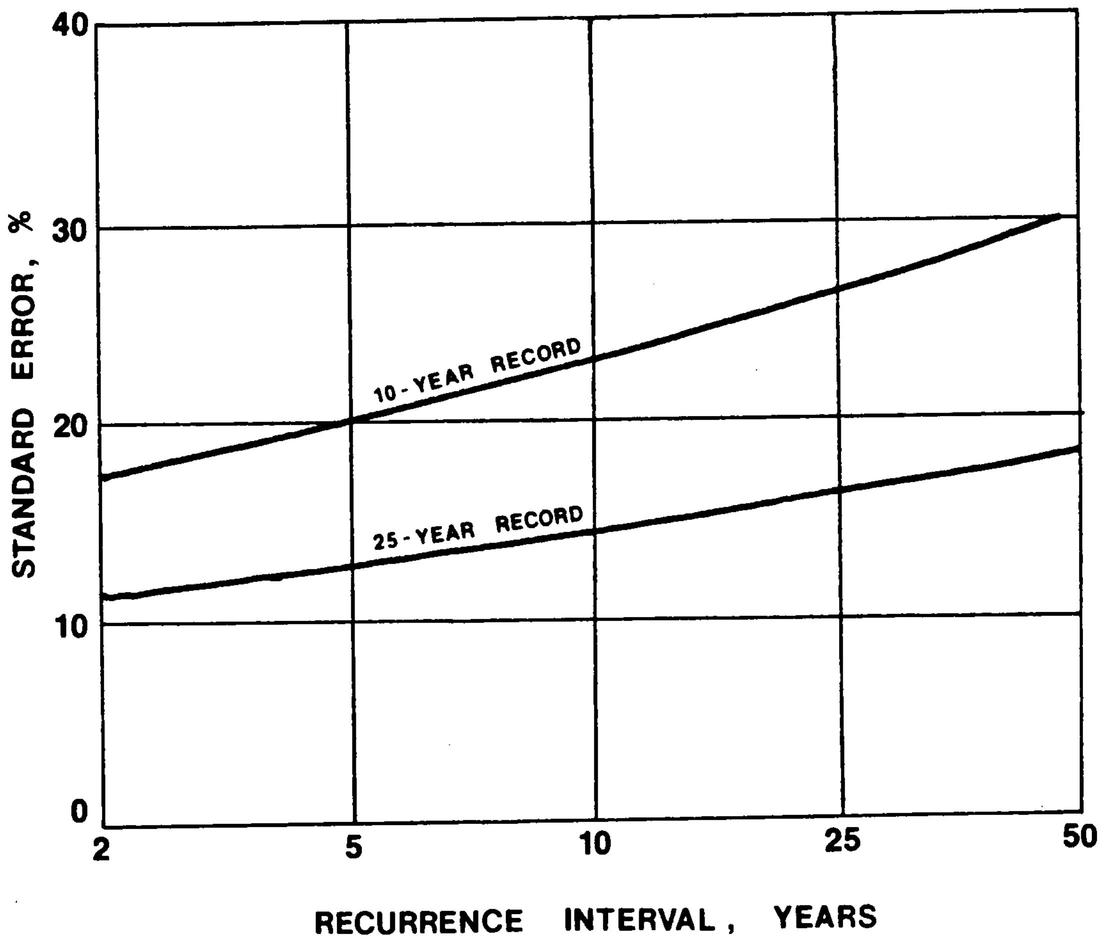

Figure III-4. Effect of Period of Record.

Source: V.J. May, G.H. Wood, and D.R. Rima. 1970. A Proposed Streamflow-Data Network for Tennessee. U.S.G.S. Open-File Report. Nashville: U.S.G.S.

CHAPTER IV

BASIN CHARACTERISTICS

The physical characteristics of the Plateau basins vary considerably. Past studies have shown basin characteristics such as basin shape, channel slope, relief, area in lakes and ponds, and land use to be important determinants of streamflow. In this chapter, the basin characteristics of each of the watersheds are presented. These data are summarized in Table IV-1.

Characteristics of Group One Watersheds

Obed/Emory System

The Obed/Emory system has the largest drainage area of any of the basins being considered in this study, with a total of 764 square miles being drained by the Obed, Emory, Clear Creek, Daddy's Creek, and various other tributary streams (see Figure IV-1). The basin is almost totally contained by Cumberland and Morgan Counties and is evenly divided between the two. The Obed/Emory basin is one of the more regular in shape, with a circularity ratio of .43 and a form factor of .77 (see Table IV-1).

Both of these parameters are indicative of the deviation of a basin from a circular shape, so that a perfectly circular basin would have a value of 1.0. Form factor is defined as $R_f = \frac{DA}{L^2}$, where DA is the drainage area of the basin and L is the length of a straight line reaching from the drainage outlet to the drainage divide adjacent to

TABLE IV-1

DRAINAGE BASIN CHARACTERISTICS

	Period of Record	Drainage Area $mi.^2$	Circularity Ratio	Form Factor	Relief, feet	Main Channel Length, miles	Main Channel Slope, ft./mi.	Area in Lakes & Ponds, %	Mean Elevation, feet	Percent Forested	Limestone at Stream Level	Area of Internal Drainage, $mi.^2$
Group One Streams												
Whites Creek	1934-82	108	.50	.88	2550			0.00	1650	87.0	No	0.0
Richland Creek		50.2	.52	.52	1450			0.00		85.0	No	0.0
Wolf River	1942-82	106	.38	.39	950	23.4	12.3	0.00	1300	88.9	Yes	0.0
E. Fork Obey	1942-82	202	.22	.26	1370	37.6	37.0	0.10	1720	92.0	Yes	6.0
Emory River	1927-82	764	.43	.77	2563	67.9	18.0	0.11	1630	87.9	No	0.0
Bee Creek	1930-37	101	.52	.89	600			0.00	1700	80.0	No	0.0
Group Two Streams												
Battle Creek		50.4	.72	.81	1560	10.9	136.0	0.00	1560	79.2	Yes	0.0
Elk River	1951-82	65.6	.52	.52	1000	15.5	78.3	0.00		87.0	Yes	0.0
Caney Fork	1930-49	111	.47	.98	515			0.20	1700	80.0	No	0.0
Clear Fork	1930-71, 1975-82	272	.48	.39	1420	36.5	12.0	0.00	1600	98.0	No	0.0
W. Fork Obey	1943-71	114	.28	.23	1370	37.6	37.0	0.00	1310	86.2	Yes	34.0
Collins River	1925-82	640	.50	.79	1556	53.3	25.9	0.01	1370	65.4	Yes	0.0
Calfkiller River	1940-71	175	.51	.45	1250	35.4	5.2	0.14	1370	60.9	Yes	64.0

Sources: columns 5-10: V.J. May, G.H. Wood, and D.R. Rima. 1970. A Proposed Streamflow-Data Network for Tennessee. U.S.G.S. Open-File Report Nashville: U.S.G.S. ; column 12: W.J. Randolph and C.R. Gamble. 1976. Technique for Estimating Magnitude and Frequency of Floods in Tennessee. Nashville: U.S.G.S., Tennessee Department of Transportation; columns 3,4: calculated from topographic maps.

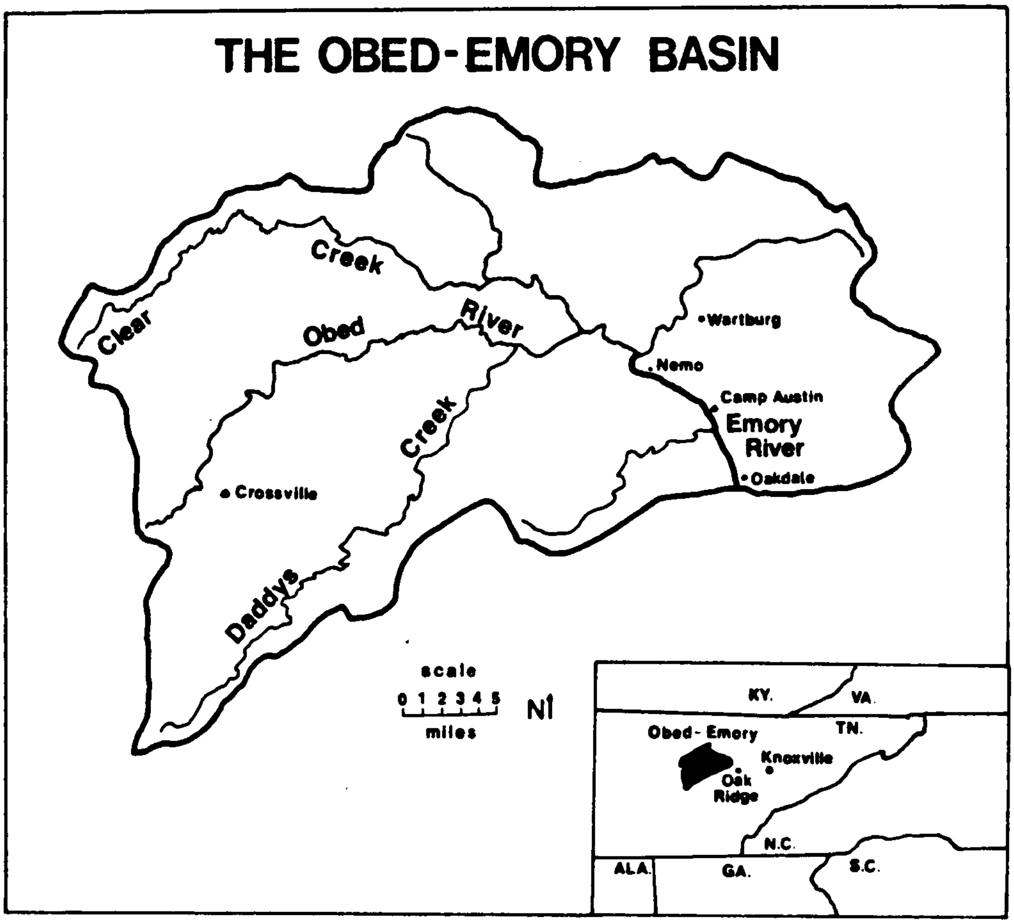

Figure IV-1. The Obed/Emory Basin.

the head of the longest stream in the basin. Circularity ratio, $Rc = \frac{DA}{Ac}$, where Ac is the area of the smallest circle that the basin would fit into. A perfectly semicircular watershed would have a form factor of 1.0 and a circularity ratio of 0.5.

The circularity ratio is somewhat misleading, in that the basin is in the form of a half-circle, so that the various tributaries are likely to deliver peak runoff at nearly the same time, thereby contributing to higher peak flow rates on the main stem of the Emory.

The slope of the river channel from points 10 and 85 percent of the distance along the channel from the gaging station to the divide is 18 feet per mile, although the slope varies greatly from 150 feet per mile to less than 10 feet (May, et al, 1970, p. A-25). The three major tributaries of the Emory have similar longitudinal profiles; none has the concave-upwards profile that is typical of graded streams (Morisawa, 1968, p.120). The depositional characteristics of these stream segments are equally variable. The upper gently-sloping segments (10-15 ft./mile gradient) of each stream have sandy-silty floodplains with steep banks cut in the depositional material. The steeper segments, however, have coarse cobble bar deposits and no clearly definable banks.

In 1970, 92 percent of the basin was forested (May, et al, 1970, p. A-15). The Catoosa Wildlife Management area covers 80,000 acres (about 16 percent of the basin) and is almost totally forested. Development within the Catoosa is minimal; a few access roads and open hunting fields are the only openings in the forest vegetation. Several strip mines operate in the basin today, and a great many more have been mined in the past. Most of the mining is in the upper Emory drainage area in the mountainous subsection of the basin. Because the areal extent of these mines is not great, their effect on quantity and time concentration of runoff is minor, although water quality has certainly suffered. Forestry activities by Hiwassee Land Company and other groups are widespread. Periodic clear-cutting may affect runoff rates in the basin to some extent, and some of the large floods of the past were probably exacerbated by extensive clear-cutting (see Figure IV-2).

Urban development in the basin is minor; Crossville and Wartburg are the only towns of note, having a combined 1970 population of 5,922. With such a small areal extent of urbanization, the amount of land in the basin that is paved or roofed is insignificant.

Resort and retirement communities have proliferated (but have not necessarily prospered) in the Crossville area of the basin in recent years. Lake Tansi, Fairfield Glade, and Holiday Hills are the largest. Approximately 24,000 acres of land in the Obed/Emory basin fall within these developments (Stroud, 1974, p.18). Because of the required scenic amenities, development and alteration of the natural terrain in these areas is not intensive. Probably the most important influences that the resorts have on the hydrology of the basin is through the numerous ponds and reservoirs that have been constructed for water supply, recreation, and scenic purposes. These ponds detain surface runoff, thereby decreasing peak flow rates slightly.

The majority of land surface in the Obed/Emory basin is gently rolling, with most slopes in the upper reaches of the basin having an

Figure IV-2. Clear-Cut Area Near Spring City.

inclination of 8 percent or less. Upland slopes of 8 percent to 15 percent are more common closer to the basin outlet. Because of gentle regional slope, elevations on the uplands decrease from nearly 2,000 feet near Crossville to 1,300 feet at Wartburg.

Several large ridges, most notably the Crab Orchard Mountains, developed on the unbreached northern end of the Sequatchie Anticline, breakup the smooth upland areas. A number of smaller, though prominent, ridges result from differential erosion on localized thrust faults. These ridges include Hatfield Mountain, Little Peavine Mountain, Lavender Knob, and Pilot Knob (Stearns, 1954, pp. 33-35).

The northeast section of the basin, drained by the upper Emory River, is a highly dissected mountainous region developed on more easily eroded shales and siltstones that overlie the sandstones of the rest of the basin. This mountainous section accounts for about 10 percent of the basin area. Coal seams within the shale units have been heavily exploited, resulting in significant local modification of the landscape. Access roads and strip mine scars along the horizontal outcrops, highwalls, tailings, and other relics of mining are common. These mining activities may have altered the runoff hydrology of the sub-basin. Peak elevations in this northeastern part of the basin are much higher than those in the plateau section; at least 20 peaks exceed 3000 feet, and Frozen Head Mountain reaches an elevation of 3324 feet, the highest point in the watershed. The lowest elevation in the basin is 761 feet, at the drainage outlet at Oakdale; thus the total basin relief is 2563 feet.

The drainage network pattern in the Obed/Emory basin is dendritic, with no evident structural control except for the linearity of certain sections of the main streams along faults. One such section is the Emory River from Nemo to Camp Austin.

The major streams flow across the sandstone caprock of the plateau in their upper reaches but become progressively entrenched downstream. The gorges are as much as 500 feet deep, and the bluffs are often vertical or even overhung. House-sized boulders have separated from the rimrock and fallen to the bottoms in many areas. The steepest gradients are not in the upper reaches of the streams, as is common, but in the lower entrenched sections. In detail, streams do not have consistent gradients but alternate between pools and rapids.

In her study of streams in the Appalachian Plateau, Morisawa found that streams have gentle gradients where they flow across resistant strata, but have steep gradients where they cut through the resistant rock (Morisawa, 1964). This relationship is generally valid in the Emory basin. Where streams cross the resistant Rockcastle and Crossville sandstones on the flat uplands, gradients are gentle. In the entrenched segments, where the streams are cutting through these same strata, gradients are much steeper.

The lithology of the Emory basin consists mostly of near-horizontal sedimentary strata of Pennsylvanian age. The complete series of rock units consists of alternating layers of sandstone and softer siltstone and shale layers.

Over most of the Emory basin, the two upper rock members have been removed, so that the Rockcastle Sandstone-Conglomerate is the most commonly exposed rock unit at the surface, especially in the broad, flat uplands. Younger shales and sandstones are exposed in the remnant hills and knobs, while older shales and sandstones are exposed in the deeply incised stream gorges. Figure IV-3 represents cross-sections taken from the Jones Knob and Spencer quadrangles and illustrates the general geology of these two sections of the Cumberland Plateau. The former quadrangle is in the Emory basin and the latter, the Caney Fork basin.

The Rockcastle Sandstone-Conglomerate has very low intergranular permeability, but some secondary permeability is provided by factures in the unit (Newcombe and Smith, 1958, p.8). As can be seen in Figure IV-3, the Rockcastle formation is underlain by the massive Vandever Shale, which is impermeable to the downward movement of water. Because of this confining underlying shale unit and thin beds of shale within the Rockcastle itself, water within the lower parts of the unit is under considerable artesian pressure. Water in wells rises as much as 200 feet when the water-bearing unit is penetrated (Newcombe and Smith, 1958, p.9). Rises of 40 to 80 feet are more common.

Water-storage capacity of the rock units underlying the Emory basin is poor, as evidenced by the poorly-sustained base flows of even the larger streams. Examination of well record reinforces this point; of approximately 700 wells in the basin, over 50 percent yield less than 10 gallons of water per minute, and only 6 percent produce more than 25 gallons per minute (Tennessee Department of Conservation, 1972,

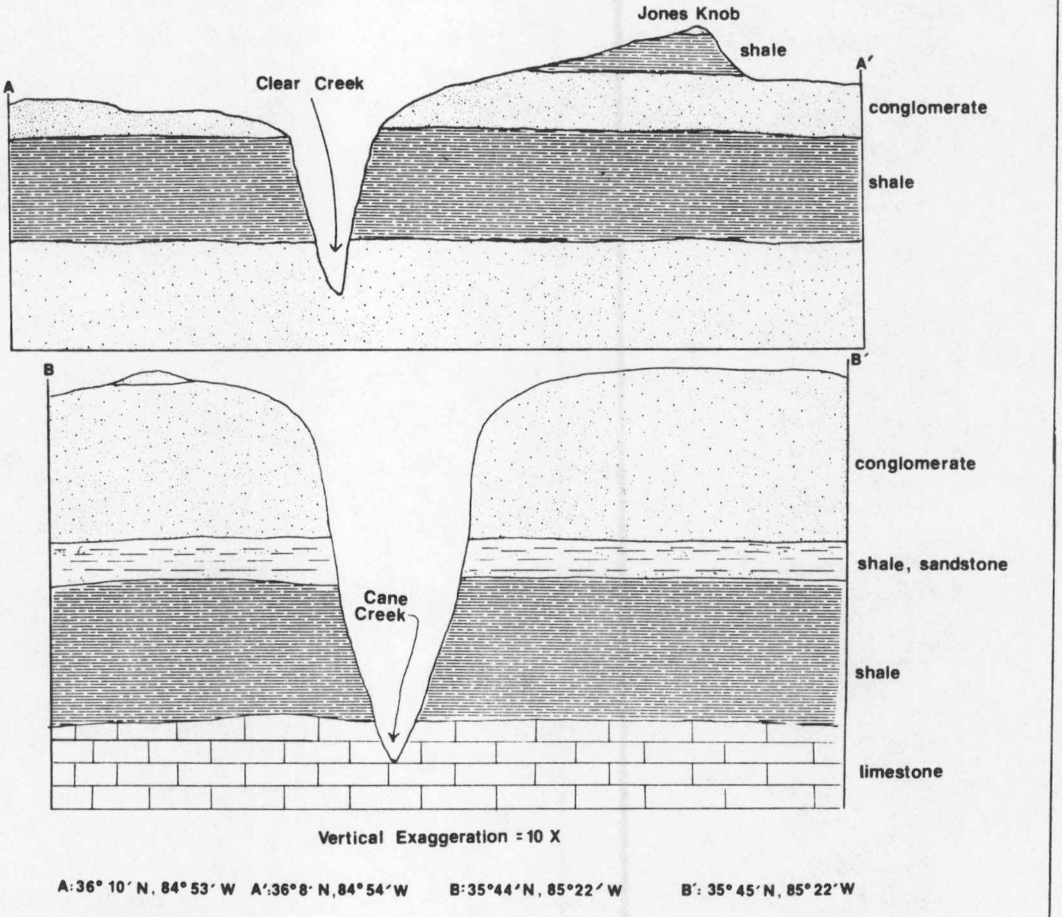

Figure IV-3. Representative Cross-Sections of Type One and Type Two Watersheds.

Sources: A.E. Coker. 1965. Geologic Map of the Jones Knob Quadrangle, Tennessee. Nashville: Tennessee Division of Geology; R.C. Milici. 1969. Geologic Map of the Sampson Quadrangle, Tennessee. Nashville: Tennessee Division of Geology.

p. 86). Many of these wells are only seasonally productive, yielding adequately during the winter and spring but going dry during the summer and early fall.

The only published detailed investigation of soils in the Obed/ Emory basin is the 1950 Soil Survey of Cumberland County (Hubbard, et al, 1950), which maps about one-half of the basin. More thorough preliminary site investigations indicate that even this survey is much too general (Lietzke, 1982). A more thorough investigation into the behavior of water in Plateau soils is now underway (Lietzke, 1982).

Two soil families, the Hartsells loams and Muskingum loams, account for most of the soils in the basin. Both series are derived principally from sandstone interbedded with thin shale layers and average only about 30 inches to bedrock (Hubbard, et al, 1950, pp. 26-28). Muskingum soils tend to be somewhat stonier, lie on steeper slopes, and have faster internal drainage. Hartsells soils are rated as second or third class, which means that they are poor to good for cropland and fair to very good for pastureland (Hubbard, et al, 1950, p. 66). The Muskingum soils are mostly in the fifth-class soil category because of their dominant stoniness and steeper slopes. Fifth class soils are defined as soils that are very poor for cropland because of steep slopes, stoniness, or susceptibility to accelerated erosion (Hubbard, et al, 1950, pp. 66-68).

The Hartsells soils are generally underlain by "a relatively thin layer of yellowish-gray sand" (Hubbard, et al, 1950, p. 42). The Muskingum series is similarly underlain by "decomposed rock mixed with many small fragments of sandstone" (Hubbard, et al, 1950, pp. 53-56). Throughflow may be concentrated at this interface between soil and bedrock.

Although the general trend of seasonal stream flow in the Obed/Emory basin is not unusual, the magnitude of the flood peaks and low flows during the seasonal fluctuations seems to be anomalous. Despite several factors which are generally considered to contribute to relatively stable stream discharges and substantial baseflow, the Emory River and it tributaries exhibit remarkably peaked flows. Winter and spring discharges are unusually high, and late summer to early fall flows are especially low. The mean discharge for March is over 20 times as great as the mean discharge for October.

Mean monthly flows are given in Table IV-2, but these averages are misleading because of the extreme variability of the summer and fall flows. For each month from June through November, the standard deviation of monthly flow actually exceeds the mean flow values, and medians are more representative of the monthly flow characteristics, as the mean is greatly influenced by unusually high flows of a few wet years. The median flow values show a more "typical" situation for summer and fall flow of the Emory River. The significantly lower median values for the months of May through November are indicative of the severity of the low flows during the warmer months.

One of the most important aspects of the runoff regime in the Obed/Emory basin is the frequent occurrence and great magnitude of flood peaks. Flood flows are significant not only because of the

TABLE IV-2

MEDIAN AND MEAN MONTHLY FLOWS OF THE

EMORY RIVER

	Mean	Median	Standard Deviation
January	2428	2720	1833
February	3337	3280	1845
March	3253	3280	1454
April	2177	2120	1000
May	1091	1160	760
June	333	539	946
July	190	448	714
August	150	267	376
September	63	210	390
October	93	153	246
November	416	844	1270
December	2340	1940	1562

all values in c.f.s.

Source: V.J. May, G.H. Wood, and D.R. Rima. 1970. A Proposed Streamflow-Data Program for Tennessee. U.S. Geological Survey Open-File Report. Nashville: U.S.G.S.

damage that is often incurred on developed flood plains, but from a geomorphic viewpoint as well. It is believed by some that flood flows are responsible for most of the geomorphic work done in a drainage basin, both in the stream channel and on the hillslopes.

Extremely high flood peaks have been recorded on the Emory River in the past one hundred years. The single highest flood event since at least 1857 occurred on March 23, 1929, when a maximum discharge of approximately 195,000 cubic feet per second (c.f.s.) was recorded, giving a gage height of 61.1 feet at Harriman. At least five other floods with peak discharges exceeding 100,000 cubic feet per second have occurred since 1900 (see Table IV-3).

Perhaps more significant than these extremely high, infrequently occurring floods are those of relatively short recurrence intervals. The two-year recurrence peak flow for the Emory River is approximately 45,900 c.f.s., and the 10-year flood exceeds twice that rate. The frequent occurrence of floods of such magnitude is unusual for a stream of this size.

The Emory River rises rapidly during the large floods. During the flood of March 23, 1929, the river rose at a rate of about 8 feet per hour, increasing from bankfull stage to the crest in only 3½ hours. Similar rapid rises can be seen in the hydrographs of other large floods on the river. These rapid increases of runoff rate indicate an extremely efficient storm water delivery system.

During the summer and fall months, potential evapotranspiration exceeds precipitation in the basin, so that the generation of runoff

TABLE IV-3

LARGE FLOODS ON THE EMORY RIVER

Date	Peak Discharge, Cubic Feet Per Second
January 2, 1929	70,600
March 23, 1929	195,000
February 3, 1937	103,000
February 13, 1948	101,000
February 1, 1951	77,500
March 22, 1955	79,700
November 19, 1959	76,700
May 28, 1973	170,100
March 13, 1975	87,200

Sources:U.S. Geological Survey. 1965. Flood Peak Runoff and Associated Precipitation in Selected Drainage Basins in the United States. Washington: Government Printing Office. U.S. Geological Survey. 1966-1978. Water Resources Data for Tennessee. Washington: Government Printing Office.

during the months of June through October is rare. The water in the streams during these months is the result of groundwater discharge, delayed vadose water discharge, and rainfall that has fallen directly into stream channels. The contribution of each of these sources is apparently small, because the flows during September and October are quite low in most years.

Streamflow is usually at its yearly minimum in October, but extremely dry conditions sometimes persist into November. On the average of every two years, the Emory River reaches a week-long average flow of only 5.0 c.f.s., which is unusually low for a drainage basin of more than 750 square miles. Since 1927, there have been two periods on no flow at all.

These frequently-occuring low summer flows preclude the use of the Obed/Emory River system's surface waters for water supply systems under natural conditions. Most water users in the basin acquire their water either from groundwater supplies or from reservoirs. The consistently poor groundwater supplies in the basin are responsible for the proliferation of small reservoirs on this part of the Cumberland Plateau in recent years.

Whites Creek

The Whites Creek watershed lies directly southeast of the Daddy's Creek watershed in Cumberland and Roane County. The basin outlet is at Glen Alice, between Harriman and Spring City. The western drainage divide follows the crest of the Crab Orchard Mountain Anticline. Basin relief is 2550 feet, which is rather high for so small a basin. Slopes are steeper than those of most other Plateau basins because of the arching associated with the Crab Orchard Anticline. The western divide of the basin is the eastern flank of the anticline and the crest of the Crab Orchard Mountains. Elevations along the divide average approximately 2500 feet.

Most of the land within the Whites Creek watershed is forested. No community in the basin has a population of more than 100-200 people, so urban development is insignificant. Numerous highway corridors pass through the basin, including those of I-40, U.S. 70, and Tennessee 68, but the total amount of paved ground is still quite small.

The basin has a circularity ratio of .50 and a form factor of .88, indicating a rather regular shape. Perhaps even more important than these measures of compactness is the actual configuration of the stream network. Piney Creek, Whites Creek, Mammy's Creek, and Fall Creek, the principal tributaries to the main stream, join within a one-and-one-half-mile stretch of the channel. This configuration should allow great concentration of runoff, resulting in high flood peaks.

The channel of Whites Creek is unusual in that bare rock is exposed in much of the streambed without a mantle of cobbles. This could help in increasing flood peaks by speeding the delivery of water that has reached the channel. No deep pools occur along the river such as are common on the Obed and the other Plateau streams; the channel is almost continuously sloping. Preliminary estimates indicate a channel slope of over 50 feet per mile.

The soils of the basin are mostly of the Hartsells-Lonewood-Ramsey-Gilpin Association. They are shallow to moderately deep loamy siliceous Hapludults derived from sandstone and shale, with some loamy siliceous lithic Dystrochrepts. Both vertical and horizontal permeability is high (Springer and Elder, 1980, p. 42).

The upland surface of the basin is developed on the Crossville Sandstone, which has been eroded from much of the Plateau. Because only thin sequences of sandstone of the Rockcastle formation are exposed in the stream gorges, fewer vertical cliffs exist in the gorges. In this section the Rockcastle formation contains thicker beds of shale than it does farther to the north. The lower unit exposed in the gorges is the Vandever formation, which consists of shale interbedded with sandstone (Milici and Swingle, 1972).

The most outstanding aspect of the streamflow regime of Whites Creek is its prodigous flooding, as indicated in Figure III-1, p. 22 . Each index flood for Whites Creek is at least 20 percent greater than that of any of the other fully gaged basins on the Plateau. The flood of March 23, 1929 had a flow exceeding 600 c.f.s. per square mile, probably the single most impressive stormflow event ever recorded on a Cumberland Plateau stream. Yet it is not only the most extreme events that have especially high peaks; the 25-year flood flow for Whites Creek is equivalent to the 100-year event for any other gaged stream in the area.

As with most streams that have especially high flood flows, Whites Creek exhibits very low base flows. The stream delivers less than 1 cubic foot per second almost 10 percent of the time. The 50 percent exceedance value is only 55 c.f.s.

Richland Creek

Richland Creek drains a 50.2 square mile area of Walden Ridge. The stream leaves the Plateau at Dayton, which is built on the alluvial fan of the river. No towns of any size exist within the basin, as most of the land is held by private timber companies. Logging is, therefore, an important activity. A fairly large amount of land is cleared for agriculture, especially on the Plateau upland. Much of the gorge of Richland Creek is maintained in undisturbed forest by Hiwassee Land Company as the Laurel-Snow Pocket Wilderness.

Relief within the basin is approximately 1500 feet. The stream channel is especially steep, averaging over 100 feet per mile. Slopes are relatively gentle over most of the basin, with the exception of the stream gorges and the Cumberland Escarpment. Slopes of less than eight percent are common on the upland surface, while slopes in the gorges are frequently greater than thirty percent. There are over 100 small (less than 1 acre) ponds on the upland. Although the areal extent of these reservoirs constitutes a very small percentage of the basin area, they may influence low flow rates in small degree by detaining excess water and allowing it to evaporate.

The basin has a form factor of .52 and a circularity ratio of .52. The stream network is such that flows are not concentrated to quite the extent that they are in the Whites Creek, Obed-Emory, or Piney River watersheds. The primary tributaries, such as Morgan Creek, Laurel Creek, Henderson Creek, and Polebridge Creek, join at well-spaced intervals.

The gorge of Richland Creek is incised into the plateau rather spectacularly; at Buzzard Point, the gorge is over 800 feet deep. Most of the streams join the canyon discordantly, dropping over waterfalls from resistant sandstone layers such as the Crossville and the Rockcastle onto the lower shale units such as the Dorton. Snow Falls and Laurel Falls are two striking examples.

The caprock for this section of Walden Ridge is the Rockcastle Conglomerate, along with the sandstone units of the Crooked Fork Group. The Vandever Formation, Newton Sandstone, and the Whitwell Formation are exposed in the valley walls (Swingle, 1963).

Soils of the basin fall into two major groups. The uplands are mantled by Hartsells-Lonewood-Ramsey-Gilpin association soils, which are shallow to moderately deep, well-drained, and highly permeable. Soils of the Bouldin-Rock Outcrop-Ramsey occur on steeper slopes in the gorges and along the Cumberland Escarpment. These soils are significantly deeper and stonier than the upland soils but are equally well-drained (Springer and Elder, 1980, p. 42).

The gage on Richland Creek is only a peak recording gage, so flow duration and low-flow statistics are not available for the stream. Richland Creek is one of the more prodigious flooding streams on the Plateau (see Table III-1, p. 23); only Whites Creek, the Wolf River, and Bee Creek show higher standardized flood flows. The peak flow of record was in February, 1903. Because flooding of the lower-lying parts of Dayton and Morgantown is a distinct possibility, flood flows of Richland Creek are critical. Floodplain occupance by low-value housing is prominent in those communities.

Becase of its extreme channel slope and its severe floods, Richland Creek is able to transport extremely coarse debris at infrequent intervals, as indicated by the photograph in Figure IV-4, taken approximately one mile above the mouth of Richland Creek Gulf. At the mouth, the channel slope changes abruptly from 133 feet per mile to 62 feet per mile. The valley bottom opens up to a width of one-half mile , and the stream bifurcates into several channels. From this point until it reaches the Chickamauga Lake embayment, the stream flows over a wide, thick alluvial fill of coarse cobbles and boulders.

East Fork of the Obey

The East Fork of the Obey River is an elongate basin (Rf = .26, Rc = .22) draining approximately 202 square miles of Fentress, Overton, and Putnam Counties. The basin shares drainage divides with the Obed, Clear Fork, Wolf, Calfkiller, and West Fork of the Obey Rivers (see Figure IV-5). There are no towns of any consequence within the basin, although there are numerous mining coummunities in various stages of occupance. Included within the basin are 6 square miles of area that drain into depressions (Randolph and Gamble, 1976, p. 31).

Forests cover 92% of the Obey drainage basin (May, et al, p. A-15). Much of the area has been contour mined for coal. The mining has clearly had serious impacts on the water quality of the basin, and it is possible that flood peaks have been pushed higher by the siltation of

Figure IV-4. Channel of Richland Creek.

EAST, WEST FORKS of the OBEY RIVER

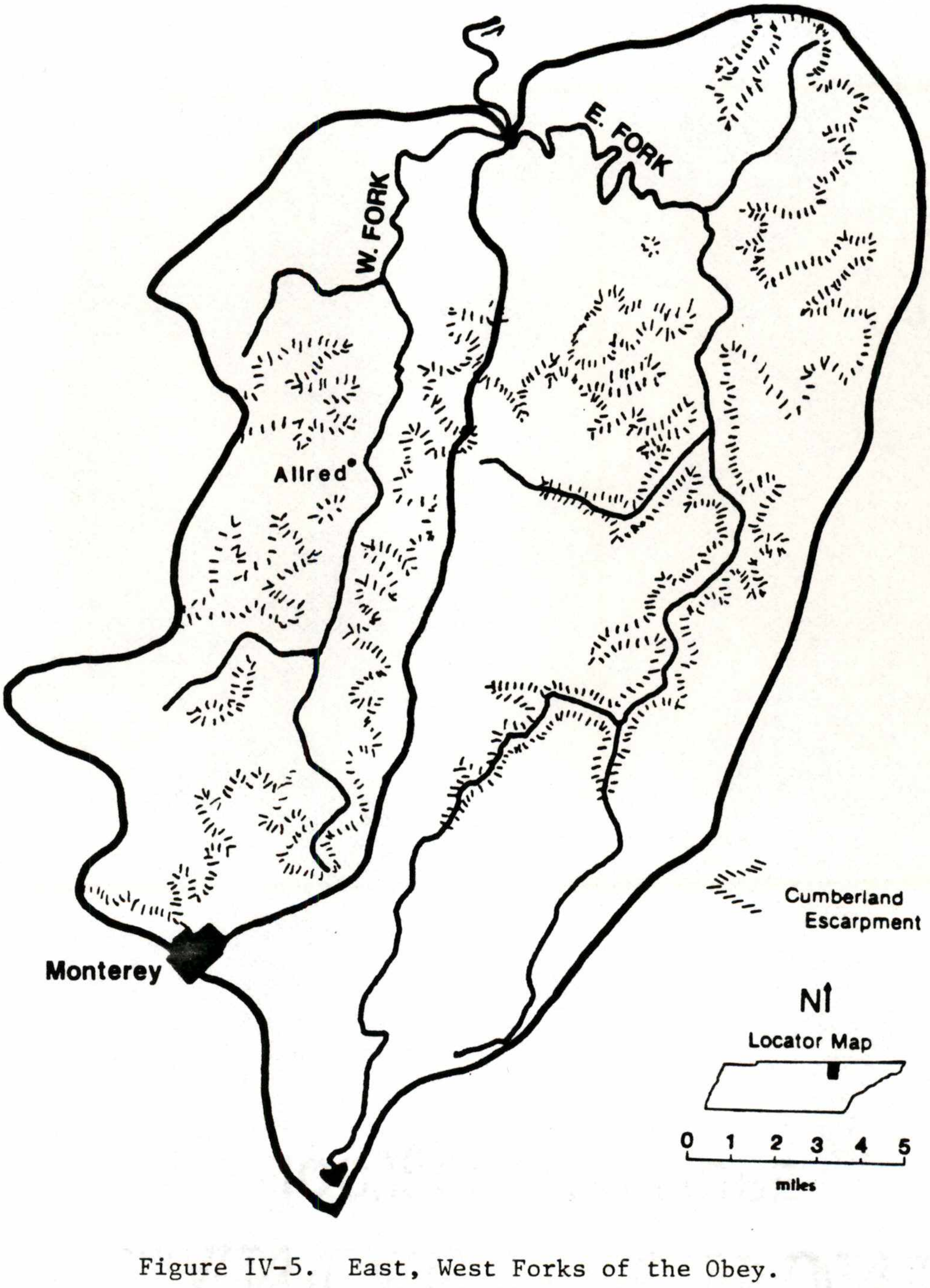

Figure IV-5. East, West Forks of the Obey.

stream beds and increased surface runoff that frequently accompany surface mining. In a study of paired watersheds on the Cumberland Plateau in Kentucky, Bryan found that mining does affect the runoff curve of a watershed, but not as dramatically or in the direction that most would expect. The following is from the conclusions section of his thesis:

> One of the first effects to be noted after stripping is an increase in variance of both stormflow and peakflow ... stormflow volumes appear unchanged ... Maximum annual stormflows were reduced, possibly due to greater storage in the benches and spoil banks. Average peakflows, on the other hand, were significantly increased by stripping ... The increase in summer peakflow will cause more flashy flows in the headwaters, and will provide the means for increased sediment export to the larger streams and rivers. The sediment exported from the stripped areas will cause the downstream channel cross sections to be reduced thereby reducing the channel's capacity to handle stormflow. The result will likely be an increase in downstream flood damage potential even though the actual stormflow volumes are unchanged. (Bryan, 1979, pp. 46-47).

The main channel of the East Fork of the Obey has an average slope of 37 feet per mile, which is rather high among Plateau streams of this size. The basin has a mean elevation of 1720 feet (second only to Daddy's Creek in mean elevation) and has 1370 feet of relief.

The slopes on the upland surface of the basin are very gentle in the southern (upstream) end of the basin but become increasingly steep towards the basin outlet, where the western margin of the Plateau is greatly dissected. The gorges of the East Fork of the Obey and its tributaries are among the more impressive in the state (see Figure IV-6). Although the walls lack the vertical cliffs prominent along the Obed or Cane Creek, the streams have cut steep V-shaped gorges that are over

Figure IV-6. Gorge of the East Fork of the Obey.

1000 feet deep in places. These deeply entrenched sections are especially prominent, because they are juxtaposed against the broad, undulating upland surface of the Plateau. It is likely that the vertical cliffs are absent because there are no thick, massive sandstone beds exposed in the valley walls; the Rockcastle Formation has been stripped off the surface in this area.

The general geology of this part of the Plateau is much like of the previously-discussed sections; the lithology consists of horizontally disposed beds of siltstones, shales, and sandstones. However, the younger Mississippian sequences are closer to the surface in this area because of the general eastward dip of the Plateau structures, so that limestone units are exposed in the lower gorges. The soils of the basin have never been mapped in detail, but the 1:500,000 map of Tennessee soils indicates that they are mostly of the Hartsells and Ramsey series (Springer and Elder, 1980).

The average annual discharge of the East Fork of the Obey is 419 c.f.s., for an annual total of 28 inches of runoff (U.S. Geological Survey, 1979). The mean monthly discharge varies from a low of 57 c.f.s. in October to a high of 923 c.f.s. in February (May, et al, p. A-25). The maximum month's flow is, therefore, 16.2 times as great as that of the lowest month. The streamflow duration curve for the East Fork of the Obey is much less extreme than that of the Emory or Clear Fork, but is more peaked than some other Plateau streams.

The 2-year 7-day low flow value for the East Fork of the Obey is 8.99 c.f.s., which is a higher low rate for this event than that of any other streams in this study except for the Collins and Calfkiller Rivers, which are clearly in a class of their own (Gold, 1981, p. 22). The other low-flow measures are similarly high for the East Fork when compared to other streams, although the values for the West Fork of the Obey and the Elk River are similar, and those for the Wolf River are even higher when the flows are standardized by drainage area.

The standardized flood flows for the East Fork of the Obey are about average for Plateau streams (see Table III-1, p.23). The mean annual flood is approximately 14,000 c.f.s. (51 c.f.s./mi.^{2}), the 5-year flood reaches 24,000 c.f.s. (119 c.f.s./mi.^{2}), the 10-year flood is over 29,000 c.f.s. (145 c.f.s./mi.^{2}), the 25-year flow is approximately 33,000 c.f.s. (121 c.f.s./mi^2), and the 50-year flood exceeds 38,000 c,f,s, (140 c.f.s./mi.^{2}) (May, et al, 1970, p. A-24). No towns within the basin are subject to flooding, and no determinations of critical values for channel modification have been made. No variations in flow charcteristics are apparent within the basin, and no basin characteristics which would cause one to expect any such variations, except in some small sub-basins where mining is or has been extensive.

Wolf River

The Wolf River drainage basin occupies an area of 106 square miles of the Cumberland Plateau and the adjacent Highland Rim near the Kentucky border (see Figure IV-7). Total relief is 950 feet, channel slope is 12.3 feet per mile, and mean basin elevation is 1300 feet (May, et al, 1970, p. A-12). There are many small communities within the

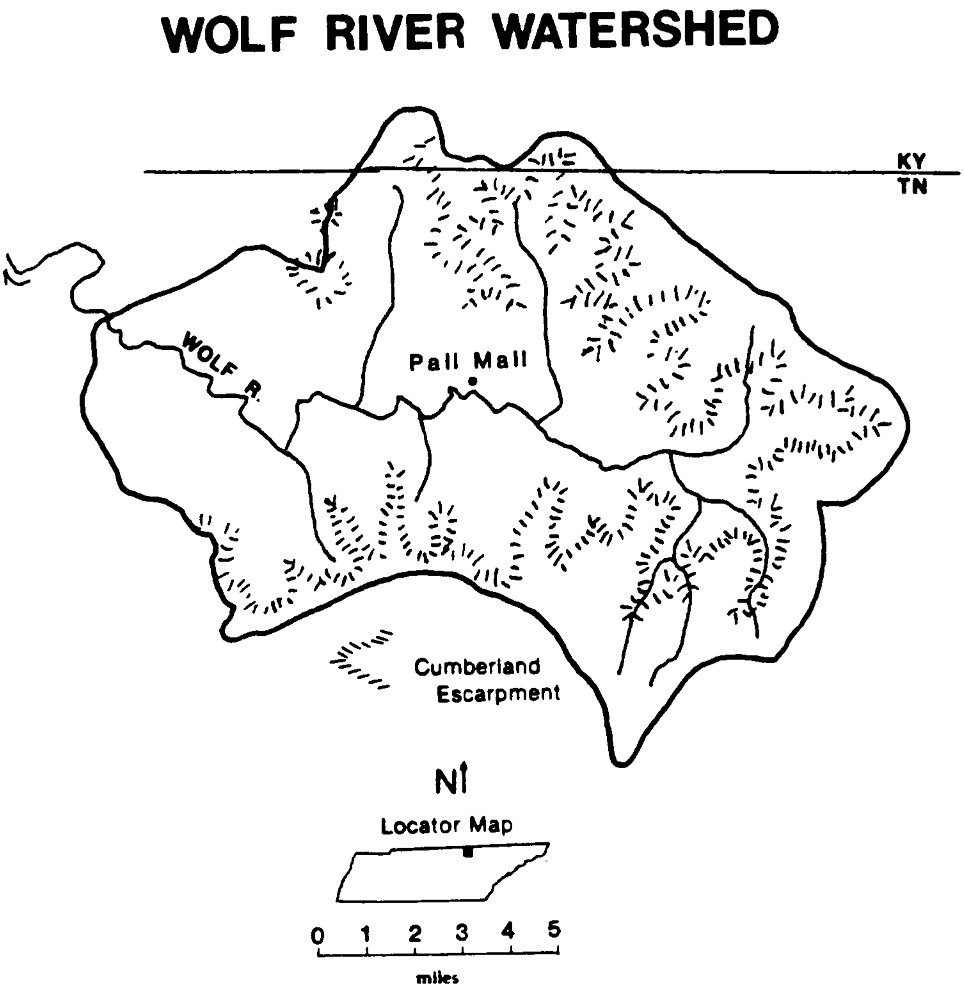

Figure IV-7. Wolf River Watershed.

basin but only Pall Mall has more than 100 inhabitants. The basin is divided between Fentress and Pickett Counties and extends about one mile into Kentucky in the vicinity of Sandcliff. Nearly 90 percent of the basin is forested, with most of the cleared land lying adjacent to the main channel of the river upstream of Ferbus, but extending across the lower slopes downstream of that point.

In the upper reaches of the basin, at elevations of 1500 to 1677 feet, slopes are rather gentle. The major tributaries of the Wolf are deeply entrenched into this upland and have dissected it into a well-developed dendritic pattern of deep valleys and steep side slopes.

The lower Wolf River has an unusually broad valley in the vicinity of Pall Mall. This valley is properly a part of the Eastern Highland Rim, as it is formed in Mississippian limstones at elevations of less than 1000 feet. Alluvial fill is extensive.

Both the geology and the soils of the Wolf River basin are quite similar to those of the East and West Forks of the Obey River; the uplands are underlain by Pennsylvanian sandstones and shales, while the older Mississippian strata are exposed in the valleys and on the Highland Rim. Soils are shallow on the Plateau surface, moderately deep on valley sides and toe slopes, and very deep over the St. Louis Limestone on the Highland Rim.

The Wolf River is unusual among Plateau streams in that it has both well-sustained base flows and extremely high flood flows. Although the base flows of the Wolf are not as great as those of the Calfkiller and Collins Rivers, they are much higher than those of streams that have comparable flood discharges (such as Whites Creek, the Obed/Emory, and Richland Creek). As noted by Carlston,

> Streamflow, therefore, would be derived from groundwater discharge plus overland flow. Both components of discharge should vary inversely in their relative contribution to stream discharge in a regular and predictable system controlled by the transmissibility of the water-table aquifer... flood runoff of streams varies in magnitude inversely with the magnitude of their base flow runoff... (Carlston, 1965).

Clearly the Wolf does not conform to this rule. The Wolf has, at the same time, rather high peak discharge rates and well- sustained low flows.

Group Two Streams

Clear Fork

Clear Fork drains an area of 272 square miles on the northern interior Cumberland Plateau of Tennessee, encompassing large portions of Fentress and Morgan Counties. The basin is somewhat irregular in shape, having a form factor of .39 and a circularity ratio of .48 (see Table IV-1, p.41) (Horton, 1932). Because a semicircular basin (such as Whites Creek) is perfectly suited to flood production, form factor is considered to be a more significant measure of basin shape. The main channel of Clear Fork has a slope of 12 feet per mile, which is relatively gently sloping for a stream draining over 200 square miles and is due to the fact that Clear Fork does not plunge off the Plateau onto the adjacent Valley and Ridge or Highland Rim but joins with the New River to form the Big South Fork of the Cumberland River, still within the Plateau Province.

Soils of the Clear Fork basin are divided nearly evenly between the Hartsells-Lonewood-Ramsey-Gilpin group and the Ramsey-Hartsells-Grimsley-Gilpin group (see Figure II-4, p. 17) (Springer and Elder, 1980, p. 44). The former group is found on the gentler sloping western portion of the basin, while the latter is more commonly found in the more steeply sloping eastern portion. Detailed information about the soils of the Clear Fork basin is not available, because soil surveys have not been produced for Morgan and Fentress Counties. The Hartsells-Lonewood-Ramsey-Gilpin soil group consists of loamy soils which are 2 to 5 feet deep over sandstone bedrock. These soils, generally formed from residuum, are highly permeable, having infiltration capacities of up to 20 inches per hour (Springer and Elder, 1980, p. 42). Soils of the Ramsey-Hartsells-Grimsley-Gilpin group are typically found on steeper slopes, are deeper than the previously mentioned group, are stonier, and have formed in colluvium (Springer and Elder, 1980, p.44).

The lithology of the Clear Fork basin is one of alternating layers of near horizontal sedimentary rocks. As in the Emory watershed, the caprock for most of this area is the Pennsylvanian Rockcastle Conglomerate, while numerous older sequences of shale, siltstone, coal, and sandstone are exposed in the valleys.

The gage on Clear Fork is located at Burnt Mill Bridge, which is 3.7 miles upstream of the junction with the New River. Flow measurements began at this site in 1930 and have continued from then until the present time, with a hiatus of measurement occurring from November 1971 until June 1975.

Average discharge for the period of record is 468 c.f.s., or 23.4 inches per year (U.S. Geological Survey, 1980). Mean monthly flow varies from a low of 46 c.f.s. in October to a high of 1110 c.f.s. in March (May, Wood, and Rima, 1970, p. A-25). Thus the mean flow for March is 24 times as great as the mean flow of October, which is an indication of the extreme seasonality of the flow regime of Clear Fork.

The flow duration curve for Clear Fork indicates that base flows are not as low as those of the Emory, Caney Fork, or Whites Creek but are significantly lower than for the other Plateau streams being examined (see Figure III-2, p. 25). Indeed, the Clear Fork plots alone on a flow duration graph, with the other basins clustered in a group above it and another group below it. This may indicate that the factors which govern flow delivery rates are unlike those operating in the other basins. Possibly the effect of the gentle slopes is partially negated by highly permeable soils and a lack of channel storage (in the form of large pools and artificial lakes).

Floods on Clear Fork are rather modest in comparison to other Plateau streams. The mean annual flood of 13,800 c.f.s. is equivalent to only 50.7 c.f.s. per square mile (see Table III-1, p. 23). The 5-, 10-, 25-, 50-, and 100-year recurrence floods have estimated values of 20,900 , 26,000, 32,800, 38,100, and 43,600 c.f.s., respectively. The highest flow of record occurred in 1939, when the peak flow rate reached was 34,000 c.f.s. A flow rate of 32,000 c.f.s. was attained in 1970, and it is quite likely that the flood of May, 1973 exceeded both of these values (during the 4-year break in gaging).

There are no communities built along Clear Fork or any of its major tributaries, so flooding is not a threat to man within the basin. It may be significant, however, in that flood flows of Clear Fork contribute to floods downstream on the South Fork of the Cumberland, which does have several settlements along it. Theshold flows for streambed modification have not been determined at this time.

Basin characteristics such as slope, geology, and land use are highly homogeneous. The southwestern corner of the basin, which drains the steep, highly dissected Cumberland Mountain section, is the only exception to the generally flat, heavily forested character of the basin. It is, therefore, assumed that streamflow characteristics of sub-basins do not vary significantly within the Clear Fork system, although the lack of additional gage sites makes it impossible to ascertain whether this is true.

West Fork of the Obey

The West Fork of the Obey shares only a common destination, a common drainage divide, and a name with the East Fork of the Obey. The West Fork drains a basin only half the size of the East Fork Basin (114 mi.^{2}) (see Figure IV-5). The basin of the West Fork is part of the heavily dissected western escarpment of the Plateau. Although the uppermost elevations of the basin are gently-sloping parts of the Plateau surface, the majority of the area is steeply sloping.

The West Fork of the Obey was gaged continuously from 1943 until 1970, when measurements ceased. This 20-year record provides a reliable data source for comparing flows with other basins. Figure III-4 indicates the importance of a long period of record to accurate determination of flow characteristics (see p. 39).

The basin has a form factor of .23 and a circularity ratio of .28 because of its elongate shape. Basin relief is approximately 1100 feet, while main channel slope is 33.6 feet per mile (May, et al, p. A-12). Such a steep channel is normally expected to produce high flood flows (Benson, 1971, pp. B-23 - B-25). Less than .1 percent of the basin is covered by lakes and reservoirs, but 30 percent of the basin drains into depressions (May, et al, 1970, p. A-12; Randolph and Gamble, 1976, p. 31). This high percentage of area demanding subsurface drainage is sure to have a strong influence on the hydrologic response of the basin.

Although most of the basin (86 percent) is forested, significant amounts of land have been cleared for agriculture, especially in the lower elevations near the outlet. The basin includes part of Monterey and many small communities. Most of the basin is in Overton County.

The uppermost land surface is underlain by the sequence of shale and sandstone members of the Pennsylvanian Crab Orchard Mountain and Gizzard Groups. In the lower slopes, however, the Mississippian sequences of shales and limestones are exposed. The gently-sloping lowlands of the Highland Rim section of the basin are underlain almost exclusively by the St. Louis Limestone, a fine-grained, medium- to thick-bedded member that commonly has karst features developed upon it (Ferguson, 1968).

Soils on the steep side slopes are generally of the Ramsey-Hartsells-Grimsley-Gilpin Association, which are 1 to 2 feet deep on upper slopes but up to 6 feet deep on toe slopes. These are loamy, permeable, strongly acid soils formed primarily in sandstone and shale colluvium. The lower limestone areas of the Highland Rim are covered primarily by the Waynesboro-Decatur-Bewleyville-Curtistown Association. These soils are clayey or loamy, well-drained, highly permeable, and very deep (Springer and Elder, 1980, pp. 34-45).

The mean monthly flows of the West Fork of the Obey are not as variable as some of the other basins being studied. The mean October flow is 19.3 c.f.s., and the February mean is 397 c.f.s., for a high month:low month ratio of 20.6 (May, et al, 1970, p. A-25). The yearly variation of flows is shown in Figure III-2, p.25). Despite the great differences in basin characteristics, the flow duration curve for the West Fork of the Obey is almost identical to that for East Fork. Low flows are also quite similar (see Table IV-2 , p. 52).

Flood flows on the West Fork of the Obey are surprisingly low, considering the steep slopes and stream channel and the lack of lakes and reservoirs. Although the low-flow statistics for the West Fork are quite similar to those of the East Fork, the standardized flood flows are much lower.

Some human occupance occurs along the West Fork of the Obey, especially in the vicinity of Allred and Shiloh, so flooding of the river could cause problems. I have not determined what river flows might be required in order to modify channel forms or to flood low-lying buildings. There are no additonal gaging sites that would be of use in determining variations in flow within the basin.

Caney Fork and Bee Creek

The upper Caney Fork was gaged from 1930 until 1949, and the adjacent Bee Creek was monitored from 1930 until 1937. Estimates of the flow rate of the 1929 flood on these streams have been made use of known high water marks. The Caney Fork basin drains an area of 111 square miles above the gaging station at Clifty, and Bee Creek drains 101 square miles above the Herbert gaging station. These streams drain a large portion of the flat uplands of the Plateau west of the Obed basin; each of them shares parts of its drainage divide with the Obed and Daddy's Creek. The relief of both basins is only 500 to 600 feet, and both are approximately 90 percent forested. Some agricultural lands have been cleared on the undulating upland surface, but little urban development exists in either basin. Bledsoe State Forest comprises a large portion of the Bee Creek basin. Strip mining of coal is common in the area, especially in the vicinity of Clifty and Eastland.

The geology of these basins is quite similar to that of the adjacent Obed/Emory basin. The flat uplands are developed on the Rockcastle Sandstone/Conglomerate, while various shale, siltstone, and sandstone,layers are exposed in the stream gorges. The lower gorges are incised into limestone, but the gages for both streams appear to be upstream of the point where limestones are first encountered.

Soils in the upland parts of the Caney Fork and Bee Creek watersheds are predominantly of the Ramsey-Hartsells Association. They are described as being "hilly and rolling soils that are shallow and moderately deep, well-drained, and have a loamy surface layer and subsoil"

(U.S.D.A., 1981, p. 12).

Along the Cumberland Escarpment, Bouldin-Ramsey soils are common. These soils have formed from colluvium and are commonly much deeper than the upland soils, with average depths of 5 to 8 feet. The texture classification of the Bouldin soils is stony loam, with the subsoil averaging 35 to 65 percent sandstone cobbles by volume. The less common Ramsey soils are much shallower (U.S.D.A., 1981, pp. 11-12).

Like the Emory, the upper Caney Fork has very poorly sustained base flows but has surprisingly low flood peaks (see Table III-2, p.30). Although the period of record is relatively short, it would seem that Bee Creek is one of the most extreme of all of the Plateau streams in its swings between low and high flows. The stream has no flow at all 2 percent of the time and produced a flood equivalent to 416 c.f.s./ $mi.^2$ during March, 1929 (Gold, 1981, p. 30). The difference in flood production between these adjacent basins is astounding. A young resident of the Caney Fork area told me that he had once swum through an underground stream near Tarkiln Ford which eventually dropped him into the Caney Fork. Small scale (1:500,000) geologic maps indicate that this area has no carbonate rock strata near the surface, but subsurface abstraction of flows could help to explain this streamflow anomaly. Unfortunately, I was unable to corroborate this report, because the area in question is posted private land.

Elk River

The Elk River drains a small (65.6 $mi.^2$) basin on the southwestern flank of the Cumberland Plateau in Grundy County. The only towns within the basin are Pelham, Elkhead, and Payne Cove. The basin shares drainage divides with the Collins River, the Little Sequatchie, and Big Fiery Gizzard Creek (see Figure IV-8).

The upland Plateau surface is unusually flat in the Elk watershed, and it is heavily forested. Almost no land is cleared on the Plateau surface, but much land is cleared along the river on the Highland Rim (see Figure IV-8). In the heavily dissected western escarpment coves are steep-sided and extend almost to the drainage divide.

Slopes on both the Plateau upland portion and the valley bottom are rather flat, but the dissected escarpment is quite steep (see Fig. IV-9). Over 75% of the basin is in slopes of less than 8%. The watershed has a circularity ratio of .52 a form factor of .58, and local relief of approximately 1000 feet. The upper Elk watershed is L-shaped, having only two major tributaries that join two miles above the basin outlet. Hence the basin effectively has the concentration timing characteristics of a basin with a much higher circularity ratio.

Strip mining of coal has affected approximately five percent of the surface in the southeastern corner of the watershed, in the vicinity of Coalmont (see Figure IV-8). Although this activity has certainly had adverse effects on water quality, it is too limited in extent to affect runoff rates significantly.

On the generally flat Plateau upland, soils of the Hartsells-Ramsey-Lonewood association are found. Hartsells soils make up 50% of the association, Ramsey, 25%, and Lonewood 25%. Hartsells and Lonewood soils occur on the very gentle slopes and are formed in residuum, while

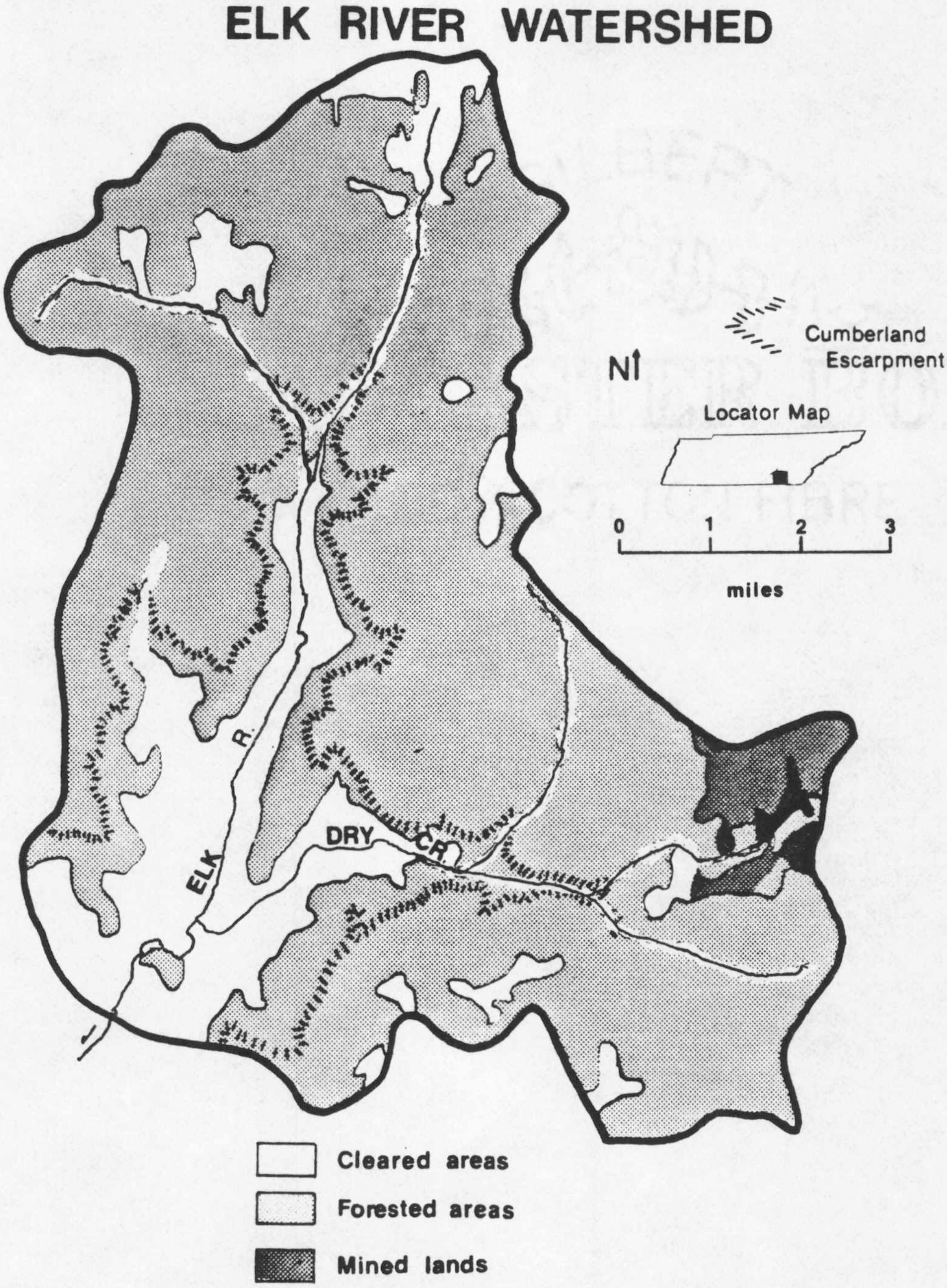

Figure IV-8. Elk River Watershed.

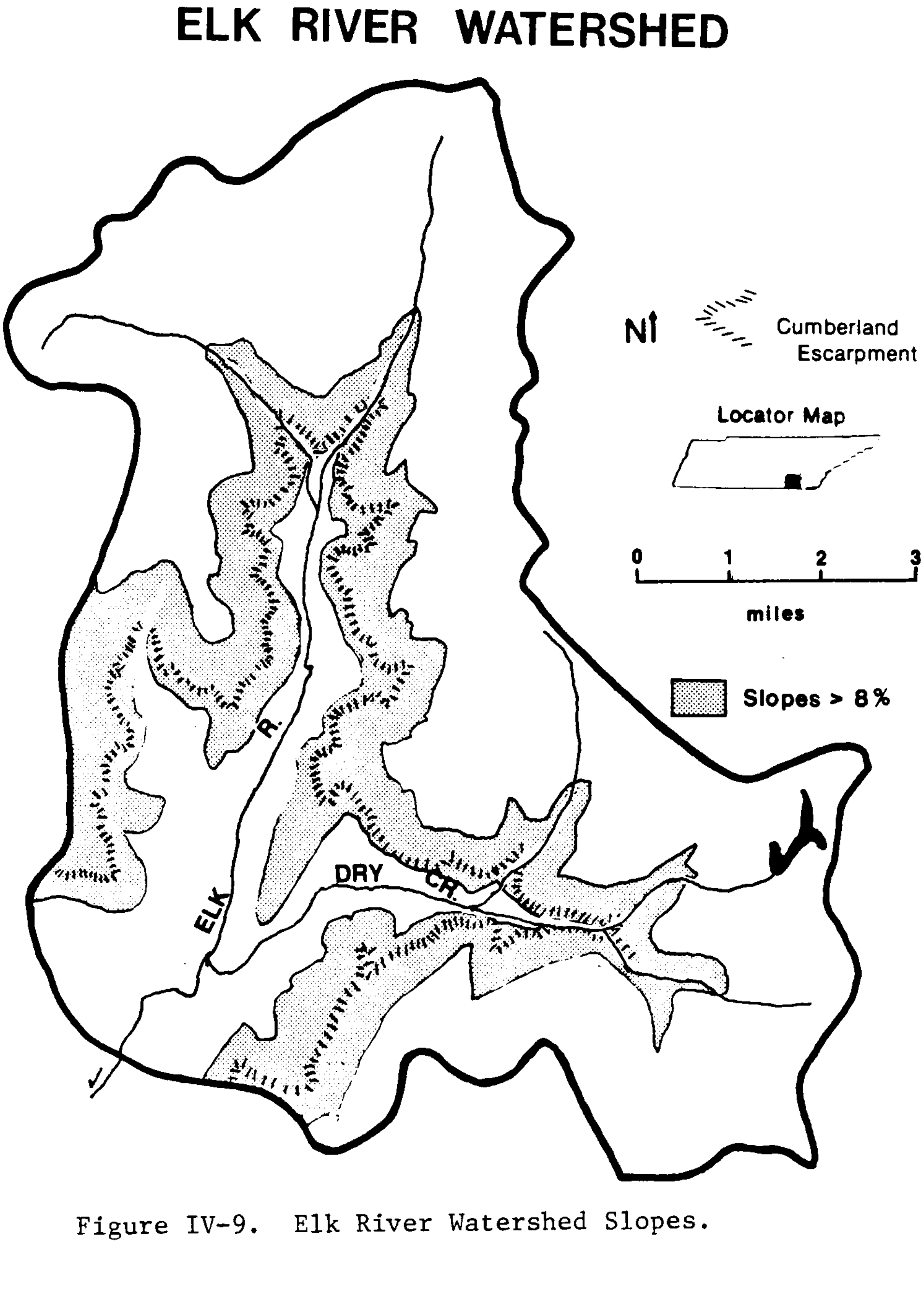

Figure IV-9. Elk River Watershed Slopes.

Ramsey soils are found on steeper slopes and are formed in part of colluvium, with much included stony material. Hartsells soils average only 12 to 20 inches in depth, Lonewood over 60 inches, and Ramsey soils are highly variable in depth (U.S.D.A., 1979).

In the steep stream gorges, soils are generally of the Bouldin-Allen/Rock Outcrop association or the Colbert/Rock Outcrop association, all formed in colluvium. The most common series in the gorges is Bouldin, with soils 5 to 15 feet deep over bedrock and containing a large proportion of sandstone fragments (greater than 35% by volume). The Allen series is quite similar but includes a much smaller percentage of sandstone fragments. The Colbert/Rock Outcrop association is found on the toe slopes of the Plateau escarpment. These soils have silt loam textures and are 10 to 60 inches deep over bedrock, except where bedrock is exposed at the surface (U.S.D.A., 1979, pp. 4-7).

The main channel of the Elk slopes 15.5 feet per mile, which is a rather low figure; only Battle Creek has a gentler sloping channel (May, et al, 1970, p. A-18). The uppermost channel of the Elk and its tributaries, such as Dry Creek, are much steeper, the Elk showing a gradient of over 300 feet per mile for a two mile stretch. This section of extreme gradient does not affect the overall gradient figure, because the commonly-used measure of gradient is based on the elevation at points 15 and 85% of the distance from basin outlet to drainage divide.

The Elk has a rather subdued flow duration curve in comparison to many other Plateau streams (see Figure III-2, p.25). Its low flows are better sustained than those of the other streams in this study except for the Collins and Calfkiller , as indicated by the extreme low-flow statistics. Peak flows of the Elk are moderate; one- half of the watersheds in the study have higher peaks and one- half have lower unit floods than the Elk (see Tables III-1, p.23, III-2, p. 30). The only community within the basin that may be subject to flooding is Elkhead.

Numerous springs feed the Elk. These include Big Spring, Sartain Spring, Gravel Spring, and Blue Spring. Flow from these springs contributes to the well-sustained low flows of the Elk.

Battle Creek

Battle Creek drains a 50.4 square mile area of the western limb of the Plateau in Marion and Franklin Counties (see Figure IV-10). The northwestern divide of the watershed passes through the town of Monteagle. Interstate 24 follows Battle Creek from Dixie Lee to Monteagle. Downstream from the river gage, Battle Creek joins with Big Fiery Gizzard Creek and subsequently joins the Tennessee River at South Pittsburg. The average elevation of the watershed is 1560 feet, and the main channel has a slope of 136 feet per mile (May, et al, 1970, p. A-18). The basin has 1350 feet of relief, a form factor of .81, and a circularity ratio of .72 (see Table IV-1).

Approximately 80 percent of the Battle Creek watershed is forested, and only two lakes or ponds of significance occur in the basin. The paved areas in the town of Monteagle and Interstate 24 probably constitute less than 1 percent of the surface area of the watershed, so

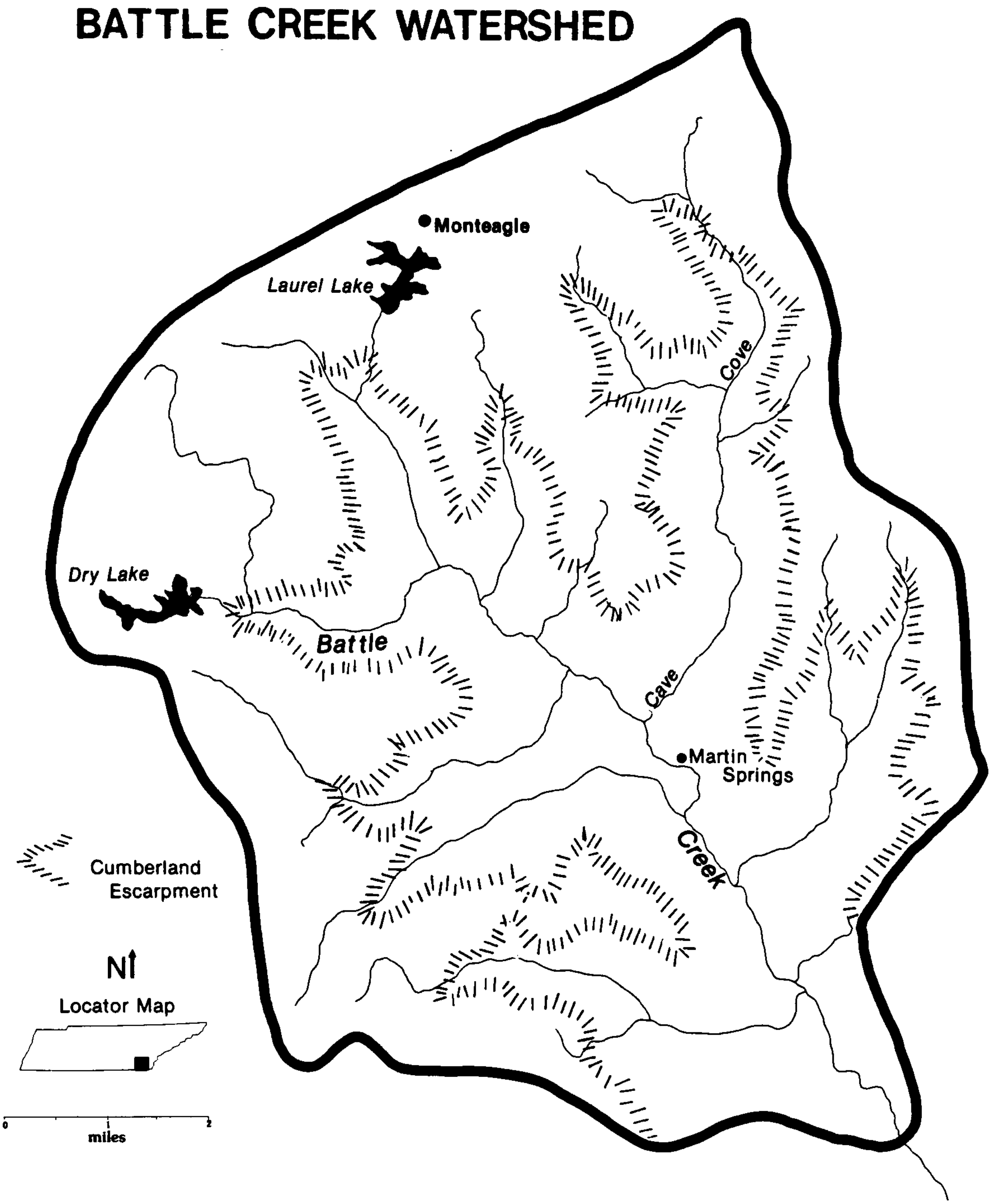

Figure IV-10. Battle Creek Watershed.

urban runoff effects are not likey to be significant. Crude calculations indicate that storm runoff from I-24 could increase streamflow by approximately 250 c.f.s. during a 2-inch rainfall of brief duration. Almost all of the cleared land is found along the stream channels.

This part of the plateau is capped by the Sewanee Conglomerate, which is 40 to 140 feet thick; nearly one-half of the Battle Creek watershed is directly underlain by this unit. The Sewanee is underlain by the Warren Point sandstone, which is expressed extensively as vertical cliffs at the top of the deeply incised coves. The uppermost unit exposed in the gorge walls is the Raccoon Mountain Formation, consisting mostly of sandstone that is commonly 100 feet thick. A thick shale formation, the Pennington, is expressed as the broad, less steeply sloping valley sides from elevations of 1200 feet upwards. Beneath the Pennington formation are extensive limestone units, including the Bangor, Monteagle, and St. Louis limestones, with a composite thickness of 600 feet or more. The valley floors have a wide, thick covering of Quaternary alluvium (Moore and Briggs, 1979). Numerous springs emerge from the limestone.

The Battle Creek watershed has a higher percentage of its area in steep side slopes than most of the Plateau basins, because of the advanced degree of stream dissection. Less than half of the basin is made up of flat upland surfaces. Also, the valley floor of Battle Creek is much narrower than that of the Elk River.

Soils on the flat Plateau surface are predominately of the Hartsells-Muskingum-Cotaco-Atkins association (see Figure IV-11). Depth to bedrock generally ranges from 18 inches to 5 feet. Much deeper

BATTLE CREEK WATERSHED

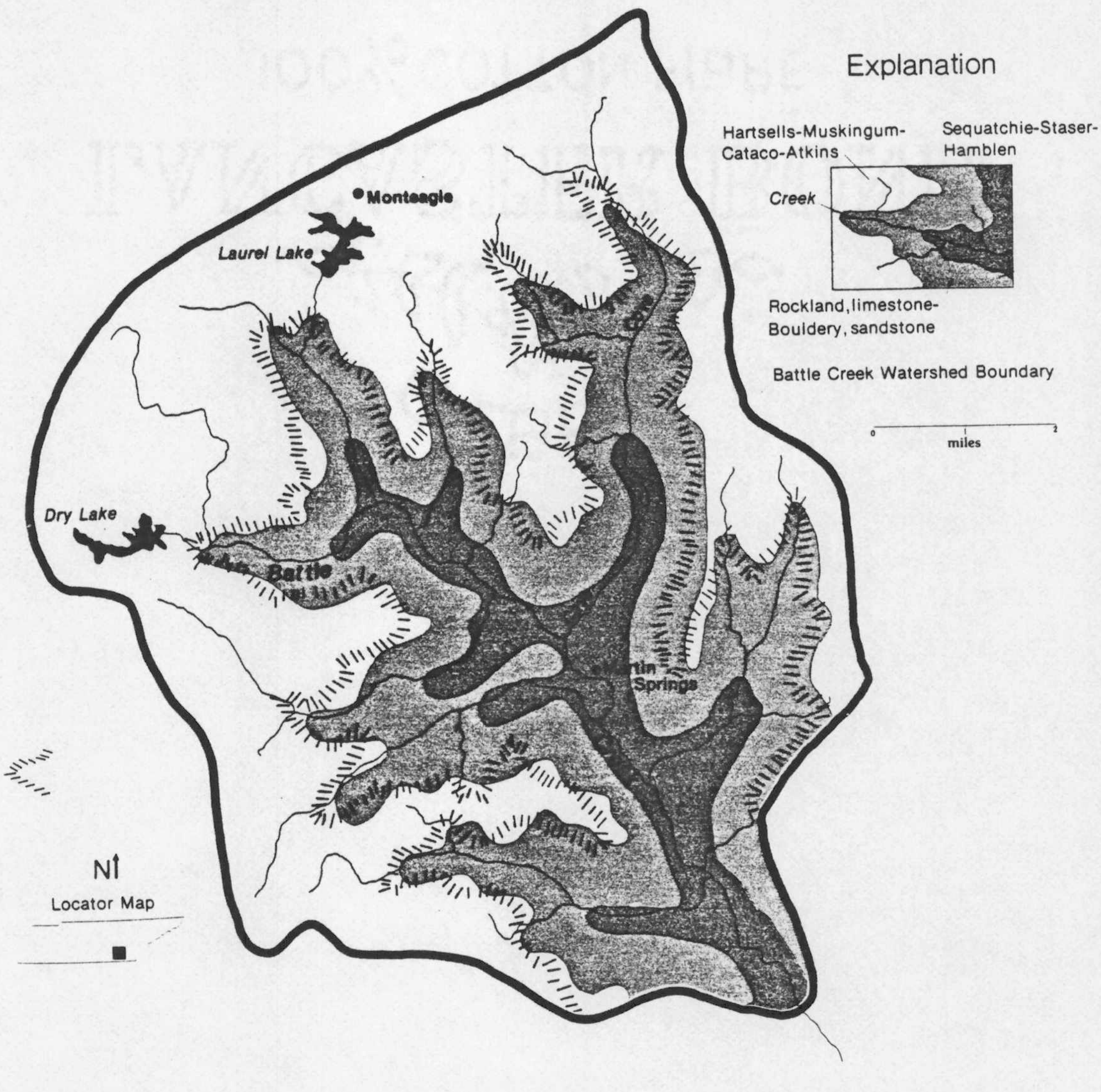

Figure IV-11. Battle Creek Soils.

soils are found on the steep valley sides and alluvial floors.

Floods on Battle Creek are much lower than on many other Plateau streams (see Table III-1, p. 23 , Figure III-1, p. 22). The 25-year recurrence flood of Battle Creek has a unit flow only one-half that of Whites Creek, for example (May, et al, 1970, p. 34). Low-flow values are not known, because the gage on Battle Creek is only a peak-event recorder. Battle Creek is, however, well-sustained by flow from several large springs, most notably Martin Springs. Given the relatively low peak flows of Battle Creek, it is reasonable to expect that its base flows will be relatively high, as there is generally an inverse relationship between low flows and peak flows of streams. During field investigations in September 1982, it was found that flows were being sustained almost entirely by springflow from Martin Springs and Tate Cave Spring (see Figure IV-12). Local residents familiar with the springs indicated that springflow at these sites is highly stable. Stream channel morphology below Martin Springs supports this contention; there is no evidence of strong high water discharges from the springs.

Collins River

The upper Collins River and its tributaries, Savage Creek, Big Creek, and Dry Creek, drain an area of approximately 400 square miles in Grundy and Warren Counties (see Figure IV-13). Most of this drainage area is within the Cumberland Plateau. Unfortunately, the long-term gaging station at McMinnville is 2 miles downstream from the confluence of the Collins with the Barren Fork, which drains an area of approximately 250 square miles on the eastern Highland Rim. Therefore, the streamflow data for the Collins are somewhat "contaminated" by flows

Figure IV-12. Martin Springs.

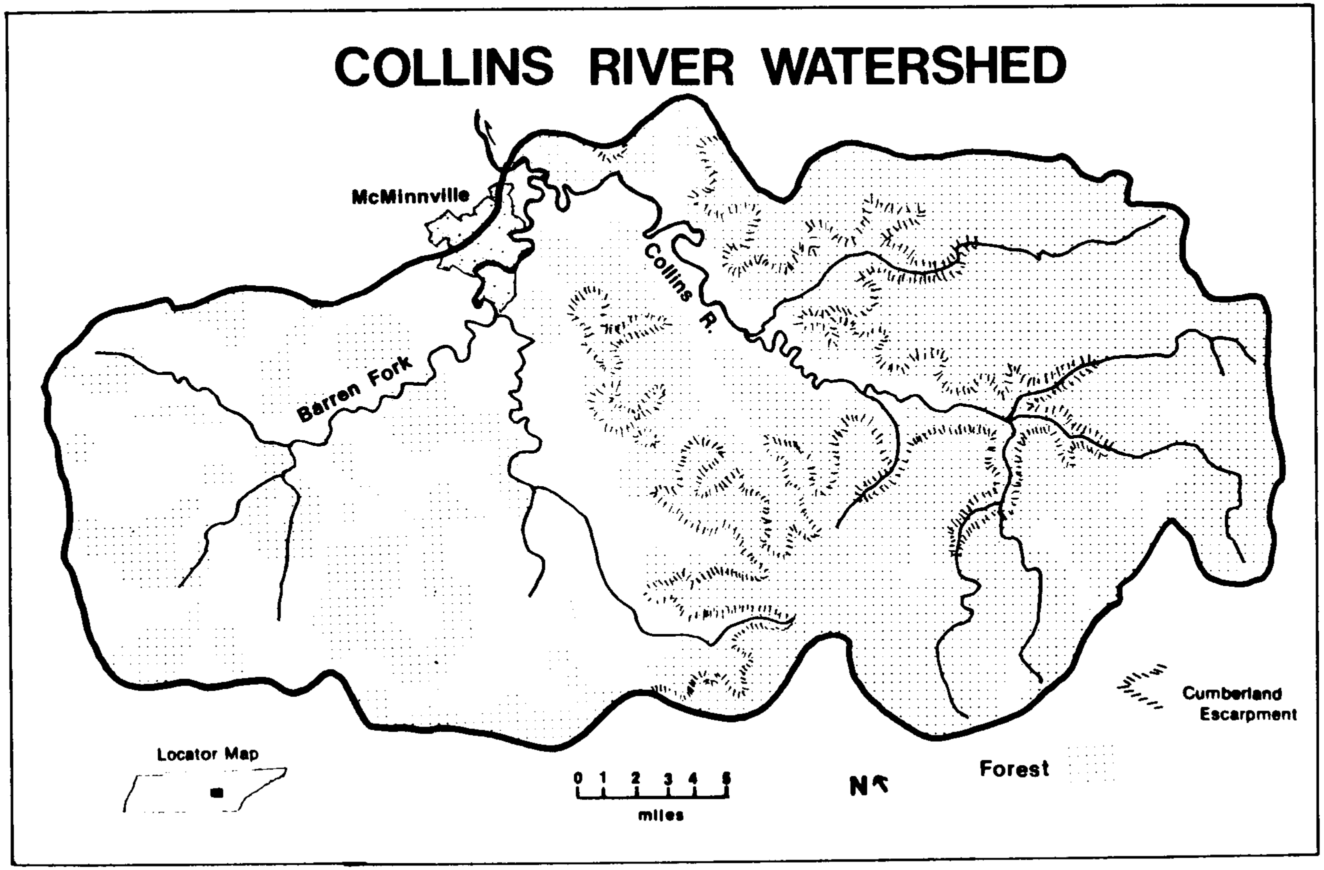

Figure IV-13. Collins River Watershed.

from a relatively large tributary from outside the region. The Collins was included in the study, even though runoff characteristics from the Plateau portion of the basin may be distorted somewhat by runoff from the Highland Rim, because the Collins is one of the largest basins draining the Western Plateau.

The Collins basin has a form factor of .79 and a circularity ratio of .50, indicating that the stream network should be capable of delivering channel flows in such a way as to increase flood peaks by concentrating flood runoff from the various tributaries. Channel slope for the Collins averages 26 feet per mile, which is relatively high for such a large stream. Less than 1% of the basin is covered with lakes, indicating that runoff is not likely to be delayed by artificial storage. The average elevation is 1370 feet, and basin relief is 1556 feet, both of which are close to the average for all of the Plateau basins. Approximately 65% of the basin is forested, but nearly all of the cleared land is on the Highland Rim, within the drainage of Barren Fork (May, et al, 1970).

Land cover on this part of the Plateau is dominated by forest; over 90% of the area is forested (see Figure IV-13). On the Plateau surface, most cleared land is adjacent to the drainage divide on gently sloping terrain where the hydrologic impacts of land clearing are minimal. This cleared land is mostly along Highways 108, 56 and the Cagle Highway. The valley of the Collins downstream of Tarlton is largely cleared and used for nurseries, pasture, and corn. The deep, rugged upper gorges of the Collins, Big Creek, and Savage Creek are managed by the state of Tennessee as the Savage Gulf Natural Area and the Great Stone Door Environmental Educational Area. Coal is strip-mined along the periphery of the watershed, especially in the vicinity of Coalmont and the upper reaches of Dry Creek. The effects of coal mining are chiefly upon water quality, but runoff rates can also be affected. May has reported that the water table is sometimes lowered enough that streamflow augmentation by groundwater discharge is eliminated in parts of the Collins and Caney Fork watersheds (May, 1981, pp. 6-7). It is possible, then, that low flows have been reduced by mining activities, but because of the limited area affected by mining, it is not likely that even low flows are greatly diminished.

Urban development within the basin is minimal. Population concentrations are especially sparse on the Plateau portion of the basin; the two larges towns are Palmer and Beersheba Springs, with populations of 1069 and 577 respectively. It is highly unlikely that this minor amount of urbanization has measurably affected runoff rates or volumes. The Highland Rim portion of the watershed is much more densely populated but the overall degree of urbanization is still quite low.

Soils of the gently-sloping uplands are mostly of the Hartsells-Ramsey-Lonewood association, which are well to excessively drained and shallow (except for the Lonewood soils, which are much deeper). Seventy-five percent of the area covered by this association has soils that are only 12 to 20 inches deep to bedrock. Clearly these soils do not provide much storage capacity for surplus water. The second most common soil type on the Plateau upland is the Ramsey/Rock Outcrop-

Hartsells-Gilpin Association, which is found in a narrow strip 1 to 2 miles wide along the courses of streams such as Big Creek, Savage Creek, and Firescald Creek. These soils are also relatively shallow, but not as shallow as the previous group. Some of the more steeply sloping upland areas have soils of the Gilpin-Hartsells-Ramsey Association. These soils average 30 to 40 inches in depth and are well drained. Along the lower reaches of some stream segments and some of the steeper colluvial slopes are found the deep(5 to 15 feet), stony Bouldin-Allen/ Rock Outcrop soils. Along the flat-floored alluvial bottoms are several groups of deep (6 feet and greater) soils formed in terrace deposits. These two latter groups of soils clearly have the ability to store much more water than the thin soils of the Plateau surface. The soils of the Highland Rim portion of the Collins basin are predominantly of the Waynesboro-Decatur-Bewleyville-Curtistown group. These soils are very deep over limestone bedrock (Elder and Springer, 1980, pp. 34-35).

The lithology of the Collins River watershed is essentially the same as that of the Elk River; the Plateau surface is capped by Pennsylvanian sandstones, while thick sequences of Mississippian limestones and shales are exposed in the stream gorges and on the Highland Rim.

The Collins River exhibits a rather subdued hydrograph (see Figure III-2,Table III-2, pp. 23, 25). The low flows of the Collins are better sustained than those of any other stream on the Plateau except for the Calfkiller, to which it is nearly identical in most flow characteristics. The 99.5% exceedance flow of .092 c.f.s. per square mile is two orders of magnitude greater than the flows of the same recurrence in the Emory, Caney Fork, and Whites Creek basins. The 3-day 20-year low flow of the Collins of 49 c.f.s. is equal to .076 c.f.s. per square mile, while many of the other Plateau streams have essentially no flow at similar recurrences.

Floods in the Collins watershed are also less extreme than those of most other Plateau streams (see Figure III-1, p. 23). The discharge of the mean annual flood is equivalent to a flow of only 36.5 c.f.s. per square mile. The Collins has the lowest unit-area floods of any of the streams in this study.

Water moves extensively through cavernous limestones in the Collins watershed. The flows of the upper Collins, Savage Creek, and Big Creek are totally subducted to underground networks prior to their junction three miles southeast of Beersheba Springs, so that the channel of the Collins is totally dry at the Highway 56 bridge between Beersheba Springs and Tarlton during all but the highest flow conditions. All of this runoff is returned to the stream channel above the gage at McMinnville. This is evidenced by the fact that the annual runoff at that gage is equal to 24.9 inches per year, which is in accord with the amount of runoff predicted by a water budget (see Table III-5, p. 34).

Calfkiller River

The Calfkiller River drains an area of 175 square miles on the western edge of the Cumberland Plateau and the adjacent Highland Rim. This watershed shares drainage divides with the East and West Forks of the Obey, the Caney Fork, and the Falling Water River and lies entirely within White and Putnam Counties. The gaging station for the Calfkiller is at the southern edge of the city of Sparta.

Approximately 61% of the Calfkiller watershed was forested in 1970 (May, et al, 1970). Most of the forested land is on the Plateau uplands and steep escarpment, while the gently-sloping alluvial lowlands are used for grazing and row crops. Only a few small lakes, such as Monterey Lake, exist in the basin, covering only 0.14% of the land area in the basin (May, et al, 1970). Some surface mining of coal and building stone occurs, but to a much lesser extent than in the Caney Fork or Obey basins. Urban development is very limited. Sparta is the only town with a population of over 500, although part of Monterey is within the basin. Except for city streets in Sparta and Monterey, only 30 miles of roads are paved within the basin. It is unlikely that artificially impermeable surfaces greatly affect runoff rates.

The Calfkiller watershed is fairly regular in shape, having a circularity ratio of .51 and a form factor of .48. These values are about average, so basin shape is probably not an important factor in determining runoff rates. The main channel has an average slope of only 5.2 feet per mile between the points 10 and 85 percent of the distance from basin outlet to divide, which is by far the lowest value of all the streams in the study. This may be somewhat misleading, however, because some of the tributaries to the Calfkiller, for example, Blue Spring Creek, Doe Creek, Buck Creek, have gradients of over 300 feet per mile. A more complex method of calculating basin channel slopes by considering the slopes of tributary streams would give a better estimate of average channel slope in the case of the Calfkiller and Wolf Rivers. Langbein suggested one such method (Langbein, 1947). The Calfkiller basin has an average elevation of 1370 feet and a total relief of 1250 feet (May, et al, 1970). As with the other Plateau basins, the Calfkiller has few surface impoundments to regulate surface flow; only 0.14 percent of the land area is covered by lakes and ponds. Sixty-four square miles of land on the Plateau surface, escarpment, and the Highland Rim segment of the basin drain into depressions. This represents 37 percent of the land area in the watershed. The water forced into underground flow systems is not lost, but eventually joins the Calfkiller through seeps and springs. This is evidenced by the fact that runoff at the Sparta gage averages 28.7 inches per year, which is close to the amount of runoff predicted by a water budget for Crossville.

Streamflow characteristics of the Calfkiller are nearly identical to those of the Collins; floods are relatively low and low flows are well-sustained (see Table III-1, p. 23 , Figure III-1, p. 22 , and III-2, p. 25). The mean annual flood of the Calfkiller has a discharge rate of only 43 c.f.s. per square mile and the 10-year flood produces a flow of only 73 c.f.s. per square mile. The 50-year flood of the Calfkiller is equivalent to the mean annual flood of Whites Creek. Low flows of the Calfkiller are among the best-sustained of all the Plateau streams. The 99.5 percent exceedance flow of .094 c.f.s. per square mile is two to three orders of magnitude greater than those of the Emory, Caney Fork, and Bee Creek (see Table III-2, p. 30). The same is true of the 2-year 3-day low flow and most other measures of minimal flows. The month-to-month variability of streamflow is relatively low; the average

March discharge is only 12.2 times that of the October discharge.

Other Streams

Many other streams that drain the Cumberland Plateau have only peak discharge gages or no gages at all. The largest area without gaged streams is the southern two-thirds of Walden Ridge where no continuous record stations exist. Whites Creek, the Piney River, Richland Creek, and Soddy Creek have peak stage recorders but no continuous record of low-flow conditions. The largest watershed on Walden Ridge without any type of gaging station is the North Chickamauga Creek basin.

On the western flank of the Plateau, streams without continuous record gaging include the Little Sequatchie, Rocky River, Cane Creek, and the middle section of the Caney Fork. All of these streams except the Caney Fork have partial-record stations. The Plateau is well represented in accurate streamflow information everywhere except for Walden Ridge.

Gold has produced estimates of low flows for partial-record stations by using occasional low flow measurements on those streams and correlating those flows with observed low flows at nearby continuous record gaging stations (see Table IV-4) (Gold, 1981). Such estimates are certainly less accurate than those made for continuous record stations but are probably sufficiently accurate for purposes of comparison.

These provisional data generally reinforce the earlier conclusion that the streams draining eastward off the Plateau have poorly sustained low flows and high peak flows, while the westward-draining streams

TABLE IV-4

ESTIMATED LOW-FLOW VALUES FOR PARTIAL-RECORD STATIONS

Stream	Drainage Area(mi.^{2})	Low Flows, c.f.s. 3-Day 10-Year	Low Flows, c.f.s. 3-Day 20-Year
Spring Creek	50.3	3.60 (0.72)	3.20 (0.640)
Cane Creek	134.0	0.00 (0.00)	0.00 (0.000)
Rocky River	78.8	3.40 (0.43)	3.00 (0.038)
Collins River	157.0	0.00 (0.00)	0.00 (0.000)
Collins River	174.0	0.70 (.004)	0.50 (0.003)
Whites Creek	108.0	0.73 (.007)	0.44 (0.004)
Piney River	95.9	0.00 (.000)	0.00 (0.000)
Little Sequatchie River	116.0	4.00 (.034)	3.45 (0.030)
Battle Creek	117.0	3.30 (.028)	2.85 (0.240)

Values in parentheses are in c.f.s./mi.^{2}

Source:R.L. Gold. 1981. Low-flow Frequency and Flow Duration of Tennessee Streams. U.S. Geological Survey Open-File Report 78-807.

have much better sustained low flows but lower flood peaks. Seeming exceptions to this statement are Cane Creek and the two gages on the Collins River. Each of these stations is located along a segment of the stream where streamflow has been diverted underground temporarily; flows return to the channel a few miles downstream and are much better sustained at these downstream points.

The Little Sequatchie drains an area of 116 square miles in Marion and Grundy Counties, flowing due south to join the Sequatchie four miles northeast of Jasper. As indicated in Table IV-4, the Sequatchie has very well-sustained low flows. Field investigations during a period of extended dryness revealed that Whites Creek and the Piney River were nearly dry, but the Little Sequatchie was discharging approximately 200 c.f.s. It was found that almost all of this flow was being generated by a group of 5 springs at a point 5 miles upstream from the gage at Highway 27. Several smaller springs contribute lesser amounts of discharge. The watershed of the Little Sequatchie is nearly identical to that of Battle Creek in all important aspects except that its valley floor is not nearly as wide. Because of this similarity of basin characteristics and the great similarity of baseflow characteristics, it is assumed that peak flows of the Little Sequatchie are similar to those of Battle Creek.

The second partial-record gage site on Battle Creek is downstream from the junction of Battle Creek and Big Fiery Gizzard Creek. Like the Little Sequatchie, Battle Creek has very well sustained low flows. Field studies revealed that essentially all baseflow is supplied by springflow. The springs of upper Battle Creek have already been discussed. During periods of low flow, the bed of Big Fiery Gizzard Creek is essentially dry from a point one-half mile upstream of Denny Cove. At this point, water flows at a rate of 50 c.f.s. or more from a cavernous opening in the eastern valley wall. This flow is dependable, as it is used to support a fish farm. Within 2½ miles, the channel is once again dry, however. It was found that at a point 2 miles downstream from the spring at the fish hatchery, Big Fiery Gizzard Creek makes a bend to the east and abuts against the valley wall. A cavern in the valley wall at that point consumes all of the stream's discharge at low flow. A dry streambed form this point to the junction with Battle Creek accomodates higher flows. It is not known where this diverted streamflow is discharged, but it may return to the surface at Gilliam Spring, which feeds Battle Creek one-half mile below the confluence of Battle Creek and Big Fiery Gizzard Creek.

The Piney River drains an area of approximately 96 square miles on the northern part of Walden Ridge. The watershed's western boundary is the upturned edge of the Sequatchie Valley and the stream enters the Ridge and Valley at Spring City. The basin has nearly 2300 feet of relief, and the main channel of the Piney has a slope over 100 feet per mile. The Plateau upland has a regional slope of 3% from the rim of the Sequatchie Valley to the top of the Cumberland Escarpment, and general landslopes are only slightly steeper. The major streams do not become entrenched into the Plateau until they reach a point approximately 10 miles upstream from Spring City but quickly incise themselves from that point onward. The gorge of the Piney reaches a depth of 800 feet two miles upstream of Spring City. Soils on the upland are generally thin Hartsells loams (Springer and Elder, 1980). The watershed is not remarkably circular or irregular in shape, having a form factor of .68 and a circularity ratio of .41. Much of the watershed is owned by forest products companies, and the forests are managed accordingly. Nonetheless, over 90% of the watershed is forested at any given moment in time. A few square miles are kept cleared for row crops on the Plateau surface, mostly along the Wash-Pelfrey road.

Streamflow in the Piney is extremely irregular; the stream frequently ceases to flow altogether, and floods can be extremely high. The basin has only a peak stage recording gage, and stage-discharge relationships have not been established, but a rough estimate of the discharge value of the mean annual flood is 10,600 c.f.s. (111 c.f.s. per square mile). This estimate is based on an extrapolation of known stage-discharge relationships at lower flows, and is probably a low estimate, because with rising water levels the discharge increases at an increasing rate. Even if the estimate of 111 c.f.s. per square mile is correct, the mean annual flood of the Piney is greater than that of any other Plateau stream. Both the flood of March 1929 and that of November 1957 produced discharges exceeding 400 c.f.s. per square mile. The storm of 1957 deposited roughly equal amounts of precipitation on the Piney watershed and the Whites Creek watershed (T.V.A., 1961), and the peak discharges of those two basins were essentially equal. The flood of 1929 had a much higher peak on Whites Creek, but precipitation was much higher in that watershed.

Floods of the Piney are a continuing problem for the city of Spring City, which occupies an alluvial fan built by the river. It is likely that this fan, which extends from the mouth of the Piney Gorge to the Watts Bar Lake impoundment (a distance of 2 miles), is a relic of Pleistocene flooding. Neither the 1929 flood nor the 1957 flood significantly altered the feature, despite the fact that each had an estimated recurrence interval of over 50 years.

Spring Creek and the Rocky River are tributaries of the Caney Fork, each draining westward off the Cumberland Plateau. The Rocky River drains a large area of flat Plateau upland between the Caney Fork and Collins watersheds. Spring Creek drains an area of highly dissected escarpment terrain east of Cookeville. Both streams have well-sustained low flows (see Table IV-4). In this respect, these watersheds are quite similar to the Collins, Wolf, and Elk Rivers.

CHAPTER V

RELATIONSHIPS OF DISCHARGE TO CHANNEL MORPHOLOGY

The stream channels draining the Cumberland Plateau contain large quantities of coarse debris. Much of it is too coarse to be transported by even the most extreme events (see Figure V-1) and must await further weathering and size reduction. A somewhat typical pool and riffle sequence characterizes most of the Plateau streams, but the sequence is frequently interrupted and modified by bedrock steps in the channel (small falls and rapids) and by large blocks of rimrock that have fallen into the stream. The riffles and point bars are composed of well-sorted sandstone cobble and boulders with dimensions of up to 1'x1'x2½'. Most of the point bars have a sparse covering of saplings.

In order to gain a better understanding of the fluvial processes a section of the Obed was chosen for examination as a case study. The site chosen was Canoe Hole, a large pool in the Obed below Clear Creek. This site was chosen because of its relative accessibility and my familiarity with it.

Channel depths were measured in cross-section and along the presumed thalweg by sounding with a depth finder from a canoe. I originally intended to make sequential cross-section profiles at 30-foot intervals for the length of the pool. After making several such measurements, however, I found that the cross-sections were highly similar except where interrupted by large boulders. These subaqueous boulders are apparently of the same origin as those found along the stream banks. Their tre-

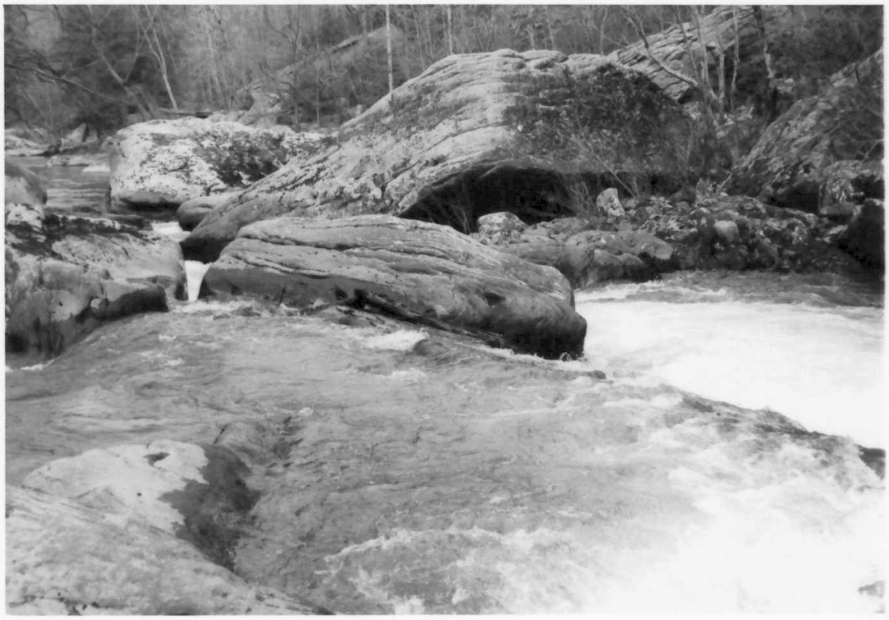

Figure V-1. Large Boulders on the Obed.

mendous size precludes entrainment by the stream. The average depth of the deepest channel section was found to be 22-23 feet. Such a great depth is indicative of the high scouring power of the peak flows on the Obed.

After closely inspecting Canoe Hole and other pools downstream, I concluded that the pool and riffle sequence of the Obed is not that of the typical river. The pools are not evenly spaced products of standard fluvial processes. Rather, it would seem that the pool and riffle sequences are controlled by randomly occurring rock falls from the sandstone bluffs, which effectively dam the river and are incorporated as the upstream and downstream limits of the pools. At Canoe Hole, both ends of the pool are marked by jumbled groups of sandstone boulders. Although there was evidence of some movement and imbrication of the smaller boulders, it is apparent that the river is not competent to move these blocks as normal bedload. The river has adjusted to these obstructions, however, as if they were normal gravel bar riffles; the pool between the obstructions is a typical scour pool with large boulders thrown in. The major difference between this pool and a normal scour pool is that its dimensions are not determined by the stream. It cannot move in an upstream-downstream direction because of the boulder falls, and it cannot move to either side because of the confining canyon walls.

A tenet of fluvial geomorphologists is that most "geomorphic work" is accomplished by frequent floods (recurrence interval of ½ to 1 year), rather than by infrequently-occurring floods of great magnitude. This school of thought is based upon research by Wolman and Miller who examined the amount of work done by streams in relation to prevailing flow rates (1960). Wolman and Miller suggest that the relative work done by extreme events and commonly-occurring events can be compared "in terms of (1) the relative amount of 'work' done on the landscape, and (2) the formation of specific features of the landscape". The authors suggest that the suspended load of a stream is an accurate measure of the former, while channel geometry is a measure of the latter. Later research by Dury (1973), Costa (1974), Gupta and Fox (1972), McPherson and Rannie (1969), and others seems to agree almost totally with Leopold and Wolman. In each of these studies, events of extremely high magnitude were shown not to have modified fluvial landforms to a degree comparable to their extreme magnitude.

Baker (1977) has recently shown that these studies may not be universally applicable, pointing out that the Wolman-Miller principle "requires modification for application to stream channels in certain climatic and physiographic settings." In regions with climatic conditions which favor infrequent floods of great magnitude, separated by long periods of quiescence, the standard explanation may not be valid. Baker points out that most of the aforementioned studies focused on streams with large drainage basins in humid climates. The segments studied were almost all in low-gradient alluvial environments. The Nene River, used in Dury's study, has a gradient of only 2.6 ft./mi. The great flood (100+ years recurrence interval) examined by Dury was equivalent to a flow rate of only 22 c.f.s. per square mile. Comparable figures for Cumberland Plateau streams are generally in excess of 300 c.f.s. per square mile. Wolman and Miller (1960) studied such large alluvial streams as the Colorado River (drainage area 137,8000 miles^2), the Rio Puerco (5,160 miles^2), and the Cheyenne River (8,710 miles^2). The authors did take note that smaller drainage basins are less sensitive to change by frequently-occurring events, however.

Wolman and Miller (1960) defined a stream's "work" as the amount of suspended and solution load it carries. They recognized the fact that bedload movement is a part of the work accomplished by a stream but argued that it could be disregarded because it is of relatively minor importance.

Because of the well documented phenomenon of downstream reduction in particular size, it stands to reason that bedload transport of coarse debris should become an increasingly important part of the work done by streams as one approaches the source of the stream. Thus lowland alluvial rivers should transport their fine particle load in suspension, while upland streams should be expected to move a large proportion of their total load by traction and saltation. While it is quite difficult, if not impossible, to measure directly the amount of bedload being transported by a stream, it is quite possible to examine in detail the depositional features of an upland stream.

Alluvial deposits along the Obed, East Fork of the Obey, Clear Fork, Piney, Whites Creek, Richland Creek, and Caney Fork Rivers consist almost entirely of coarse cobbles and boulders. The typical point bar deposit occurs in places where the overall channel of the stream curves somewhat, with the point bars on the inside of the bend. The high-water channel generally has much less curvature than the low-water channel (see

Figure V-2). These point bar deposits are commonly composed of imbricated sandstone boulders of approximately 2 feet by 2 feet by 1 foot (see Figure V-3). These are the lowest-lying deposits along these streams and therefore the features most frequently modified by fluvial transport. Very few of these deposits show signs of frequent modifications, however; even-aged beech, sycamore, and birch saplings are well established on most of them despite the absence of fine grained material at the surface.

By counting the annual rings of these trees, it is possible to date the most recent flood event which has strongly modified such a deposit , as such a flood would uproot and destroy pre-existing trees. It might be argued that the trees themselves influence the susceptibility of the deposits to modification in that mature stands would protect the deposits from entrainment by anchoring the alluvium and by reducing flow velocity. If this were the case, recently modified deposits would be more susceptible to rearrangement by ensuing floods of lesser magnitude. While this argument is probably valid for alluvial streams of lower gradient with finer-grained bar deposits, it is probably not a significant factor for the Cumberland Plateau streams. The alluvial deposits are extremely coarse, the trees are small and well spaced, and large floods are two to three times as high as the tree tops. Therefore, the threshold for entrainment is probably not greatly affected by tree maturity.

Trees were sampled from only two point bar deposits, but extensive fieldwork in the study area revealed that these two deposits were typi-

Figure V-2. Point Bar on the Emory at Nemo.

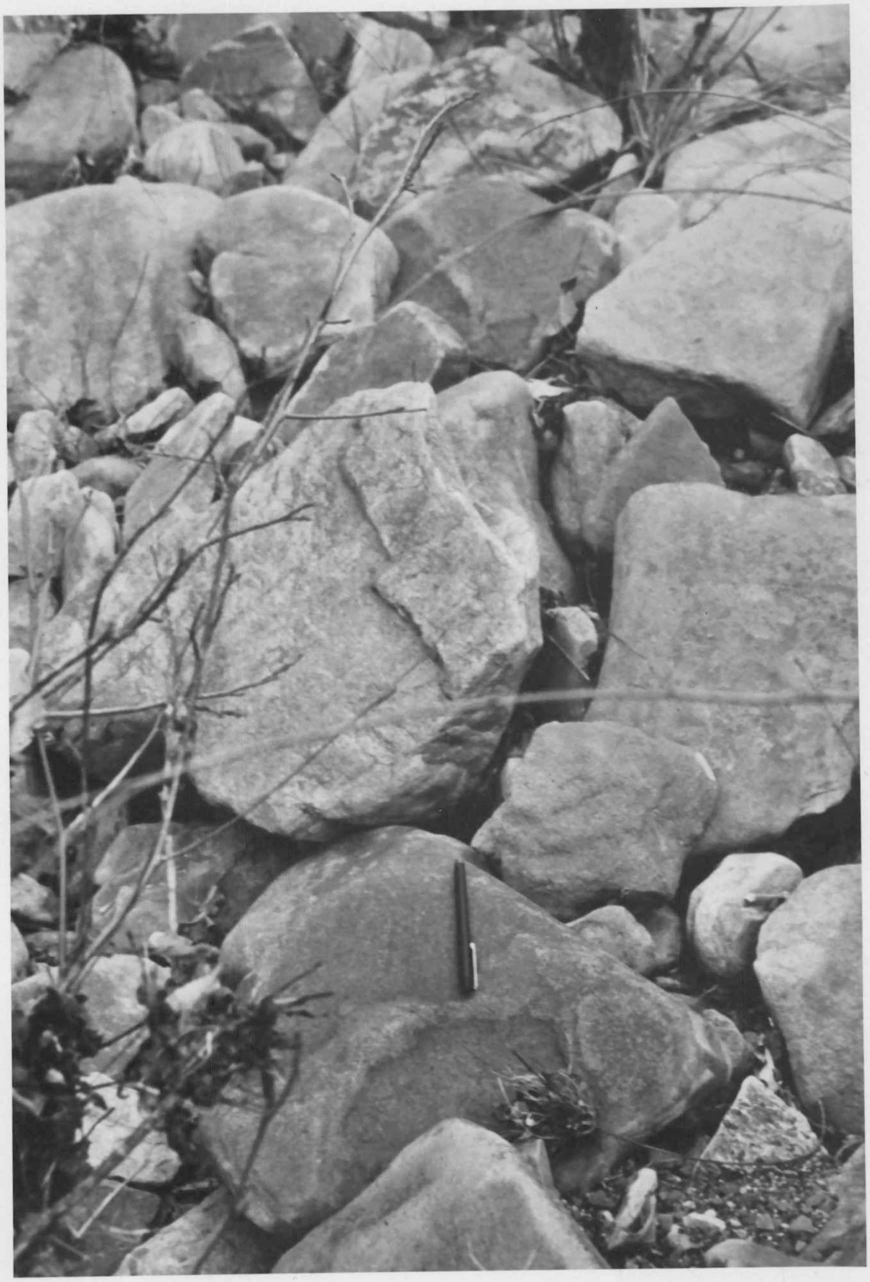

Figure V-3. Boulders in the Point Bar at Nemo.

cal of all such deposits. Almost every point bar is somewhat sparsely covered with even-aged stands of saplings that are 2 to 4 inches in diameter and approximately 15 feet tall.

The first site studied was the point bar immediately (200 yards) downstream of Nemo Bridge on the Emory River. The drainage area above this point is approximately 650 square miles, and the stream gradient in this river section is 10 feet per mile. The river drops approximately 8 feet in traversing this bar, thereby creating Nemo Rapid. Most of the trees on the bar are of uniform size and height, although a few larger trees grow on the inside of the deposit. During six years of intermittent observation of this deposit (1976-1982), no discernible changes have occurred, although water has overtopped the deposit numerous times during that period.

A 1952 topographic map (1:24,000 Lancing Quadrangle) depicts the feature as a treeless island, with clear channels on both sides of it. A 1967 photorevision indicates that vegetation had become established on the island. The deposit is no longer an island; the southern channel has been closed off such that the bar is contiguous with the south bank of the river at most flow levels.

Three saplings were sampled on this deposit on January 9, 1981. Two were birches and the third was a sycamore. All but one of the trees on the bar were of the approximate size of these sampled or smaller. Each of the trees had nine annual rings, indicating that growth began in 1973. Clearly, then, there was a major flood in late 1972 or 1973 that cleared off this point bar. Streamflow records reveal that the flood of

May 18, 1973 reached a peak discharge of 170,000 c.f.s. (223 c.f.s. per square mile). Nemo Bridge was one of the few bridges in the watershed that survived this flood. The return interval of this flood is approximately 100 years, so it is not surprising that such a flood modified its channel deposits. What is surprising is that subsequent floods have not been competent to do the same. The Emory attained a discharge rate of 107,000 c.f.s. (140 c.f.s. per square mile) on April 4, 1977. This flood has an approximate return interval of 15 years. This would indicate that the Emory only adjusts its point bar deposits every 15 to 100 years. Such stability over long periods of time is all the more remarkable in light of the prodigious flooding of the Emory.

A second site for point bar investigations was chosen along Whites Creek, one-quarter mile downstream from the junction of Whites Creek and Piney Creek. The site chosen is part of Beech Bottom, a terraced boulder bar deposit which extends to both sides of the river. The highest terrace levels are approximately 30 feet above the low-water pool upstream from the major riffle at the bar. This uppermost level is heavily forested by mature trees of various ages and by a dense cover of low bushes. This bar deposit is made of poorly sorted debris of various sizes. The largest have dimensions of up to 15 feet by 5 feet by 2 feet, yet movement of these irregular boulders is evidenced by well-developed imbrication. A sycamore that appeared to be representative of the oldest class of vegetation on the bar was cut down and its rings examined during a field trip in August of 1982. Ten annual rings were found in this tree, indicating that this deposit was last worked over in late

1972 or early 1973.

The peak flood of May 1973 crested at a discharge of 62,500 c.f.s. on Whites Creek. This remarkable flood was equivalent to a discharge of 579 c.f.s. per square mile, with a recurrence interval of more than 200 years. Floods on Whites Creek since 1973 have been rather high relative to the longterm average, yet they have not modified the low-lying deposits. During the 6 years of record since 1973, peak annual floods have exceeded the 5-year flood four times and the 10-year flood twice.

Whites Creek currently flows on the north side of Beech Bottom; the bar deposit examined in this study is on the south of the river. A topographic map of the site (Roddy Quadrangle), however, shows the main stream channel on the south side of Beech Bottom, with a small side channel on the north side, leaving a small tree-covered island in the middle. This map was produced from aerial photographs taken during 1973. Apparently the photographs were taken prior to the flood of May 28, which realigned the stream channel and stripped the vegetation from the mid-channel bar.

It is now evident that even the lowest-lying floodplain deposits of at least some of the major Cumberland Plateau streams are relatively immobile, long-lived features which are modified by only the most extreme flood events. This evidence agrees with Baker's (1977) contention that some upland streams accomplish more enviromental work during rare events of great magnitude. While Baker stipulated that such events be separated by long periods of quiescence, this is clearly not the case with Plateau streams, which produce relatively high floods every year. These lesser floods, however, are apparently geomorphically insignificant in that they do not attain the threshold value of shear stress necessary to entrain the coarse debris which makes up these bar deposits.

If Cumberland Plateau floods were of only the broad regional average in magnitude, their ability to do geomorphic work would be greatly diminished. Several sources indicate that the floods of the Plateau have discharge rates approximately twice as great as the regional average (May, et al, 1970; Randolph and Gamble, 1976). Therefore, a flood of a given magnitude would occur much less frequently in streams outside the Plateau. Using broad regional flood averages, the flood necessary to entrain debris on the Nemo point bar would have a recurrence interval of over 100 years rather than the current estimate of 25 years. This could mean that geomorphic development of the Cumberland Plateau would be proceeding at a much slower rate were it not for the great floods produced on the Plateau. Alternatively, it is conceivable that lower floods would result in an increase in channel slope because of overloading by side streams, therby allowing the system to move the same amount of debris with lower flood rates.

CHAPTER VI

POTENTIAL EXPLANANTIONS FOR THE OBSERVED PATTERNS OF RUNOFF

Regolith and Lithology

A recent paper on the hydrology of the Cumberland Plateau and the substantial variability in flood production among its streams states:

> These differences are undoubtedly caused by a combination of factors ... The dominant factors are probably soil types and geology which are not readily susceptible to mathematical analysis because they are difficult to quantify ... (May, 1981, p. 36).

While it may not be possible to quantify the effects of varying geology and soil characteristics, I believe that a systematic examination of these and other factors will lead to a viable explanation of the variation in flow characteristics among streams draining the Plateau.

My preliminary investigation indicated that bedrock lithology was the most important variable in determining basin response on the Cumberland Plateau (Mayfield, 1981). I still believe this to be true, although some important aspects of the regional hypothesis do not fit this explanation. The most important differences among the basins appear to be their bedrock lithology and the thickness of soil and regolith. The streams with the highest flood peaks are all on the northern and eastern flanks of the Plateau and drain areas of Pennsylvanian conglomeratic bedrock formations exclusively. Of the five streams that produce the lowest flood peaks (as expressed by discharge per unit area), four drain areas on the southern and western flanks of the Plateau, where the regional dip of the bedrock formations has brought Mississippian limestones nearer the surface. In upland areas where the caprock is sufficiently fractured to allow percolation of water into the limestone members, groundwater is detained and slowly paid out over long periods of time through springs along escarpments and stream gorges. More commonly, water is abstracted from stream channels where the streams are incised into limestone. Water lost from stream channels in such a manner is commonly returned to the river channel via large springs. These springs, such as Virgin Falls along Caney Fork, Big Spring on the Collins, and Fishcamp Spring on the Little Sequatchie, serve to dampen the extremes of stream flow. The streams draining the western limb of the Plateau also drain significant areas of these limestone terranes, where excess water is more likely to reach the groundwater table and be significantly delayed. In some of these streams, such as Caney Creek, large amounts of water are lost from flowing streams when the Bangor Limestone is encountered. The "lost" streamflow is presumably diverted to and stored in underground solution openings and eventually returned to the stream at some later time.

Perhaps more important for the southwestern section is the unique form of interaction of groundwater and streamflow common to karst regions. Gross permeability of the limestone terrane varies considerably according to the degree of joint enlargement by solution. Zones of high permeability tend to be in those places above the water table where flow is concentrated. Therefore, permeable zones are found adjacent to valley networks of surface drainage. According to LeGrand,

> The major trunk stream tends to have a high base flow because groundwater continues to drain towards it after the

> water table falls below the level of some tributary streams, at which time the tributary streams go dry ... The high permeability in the unsaturated zone tends to cause flood stages to be smoothed out to the extent that the floods are lower in height and longer in time than those in non-carbonate regions. The smoothing out of flood stages occurs because: (1) precipitation is quickly drained underground where it has a somewhat slower movement through solution channels to a surface stream than would overland flow; and (2) water in a rising stream tends to spread laterally during a flood into adjacent air-filled cavities and thereby to subdue the potential flood. (LeGrand, 1983)

LeGrand's first point is not especially applicable to the streams in the study area, because only limited areas of the basins have limestone exposures at the surface. His second point is more applicable to the Plateau streams and helps greatly in understanding the more subdued flows of the second group of streams.

The degree to which this flow dampening mechanism is effective in each basin must depend upon the volume of solution openings available for storage of effluent channel flow and the height of the fluctuating water table relative to that storage system. Only by intensive long-term studies of each drainage system, including dye tracings and detailed sequential gaging of discharge, could these properties accurately be determined, but the data in Table III-3(p. 32) provide good first estimates. From those data it is evident that the Calfkiller, Collins, and Elk basins have considerably more long-term storage than the others. Perhaps even more significant to the reduction of flood peaks, however, is the short-term storage, as a delay (storage) of only a few hours or days will substantially reduce peak flow rates. Such short-term storage is evidenced by the much gentler recession limbs of stormflow hydrographs of the second group of streams.

In conjunction with this, the highly dissected basins have larger areas of deeper soils and regolith, which detain and store larger amounts of excess water than the thin soils and clastic rocks have thick colluvial soils, and the flat-floored valleys of the Highland Rim have thick accumulations of residuum and alluvium. This explanation works well in general, but there are exceptions. Whites Creek and Richland Creek, the champion flood-flow streams of the Plateau, drain relatively undissected uplands with Pennsylvanian bedrock sequences and thin soils. Clear Fork, however, has similar basin attributes yet the most subdued flow characteristics of any of the study streams. Similarly, although the Wolf River basin is highly dissected and is incised into limestone, the Wolf is one of the more prodigious flood producers on the Plateau. Clearly, additional factors are involved in the observed flow regimes.

Within the Cumberland Plateau region, the total volume of available storage for free water in soil and other unconsolidated material is probably the single most important factor in determining the rate at which surplus water leaves the drainage basin via stream channels. Basins with abundant storage potential should be able to detain larger volumes of water for longer periods of time than basins with little unconsolidated rock and soil material. Within such basins, flood peaks are likely to be lower and the low flows better-sustained than in those watersheds which have only a thin mantle of soil. Because detailed soil surveys are available for only three of the study basins, it is not possible to analyze this storage element in detail. Estimates of the volume of soil and loose regolith were made for two study basins to compare Group One streams with Group Two streams.

Most of the watershed of Whites Creek is within the bounds of Cumberland County, for which a soil survey was completed in 1950 (Hubbard, et al, 1950). Soil series within the county were mapped at a scale of 1:48,000. From that map, areas having soils of six different families were calculated, using a dot planimeter. Results of that measurement are found in Table VI-1. The average depth of each of these soil families above bedrock was determined from the soil survey. The volume of soil material of each soil family was then calculated by multiplying the area of that soil type by the average depth of the soil type. Using this method, a soil volume of approximately 250 million cubic yards was estimated for the Whites Creek watershed. This value is equivalent to approximately 2.3 million cubic yards of unconsolidated material per square mile of drainage area.

Battle Creek drains an area of 50.4 square miles, essentially all of which is within the boundaries of Marion County. Using the 1958 soil survey for Marion County, soil volume was estimated for the watershed, using association- level mapping at a scale of approximately 1:160,000 (U.S.D.A., 1958). A total of 390 million cubic yards of unconsolidated material were estimated to exist within the Battle Creek watershed, which is the equivalent of 8 million cubic yards per square mile of drainage area (see Table VI-2).

The roughly measured difference in soil volume between these

TABLE VI-1

CALCULATED SOIL VOLUME IN THE WHITES CREEK WATERSHED

Soil Family	Area, $mi.^2$	Average Depth, ft.	Volume, $yd.^3$
Crossville Stony loam	4.5	2.00	9,000,000
Crossville Loams	2.9	2.66	8,000,000
Hartsells	17.9	3.00	54,000,000
Muskingum	66.4	2.33	160,000,000
Rough stony land	15.7	1.00	16,000,000
		total	247,000,000

Source: calculated from U.S.D.A. Soil Conservation Service. 1950. Soil Survey: Cumberland County, Tennessee. Washington: Government Printing Office.

TABLE VI-2

CALCULATED SOIL VOLUME IN THE BATTLE CREEK WATERSHED

Soil Association	Area, $mi.^2$	Average Depth, ft.	Volume, $yd.^3$
Sequatchie-Staser-Hamblen	5.54	8.00	45,750,000
Hartsells-Muskingam-Cotaco-Atkins	28.22	3.22	94,700,000
Rockland, limestone-Bouldery colluvium-Rockland, sandstone	16.13	15.00	250,000,000
		total	390,450,000

Source: calculated from U.S.D.A. Soil Conservation Service. 1958. Soil Survey: Marion County, Tennessee. Washington: Government Printing Office.

two basins is indicative of the observed differences in streamflow characteristics between the two. Battle Creek has considerably more storage available than Whites Creek and has lower flood peaks and better sustained low flows. The calculated volumes of soil material are unlikely to be highly accurate because of the smallscale mapping used for the area determinations and the considerable variation in depth to bedrock within some of the mapping units. The threefold difference in calculated soil volume between the two basins, however, is certainly too great to be due to measuring error.

Perhaps more important than the total volume of potential storage is the location of that storage. In the Whites Creek watershed, soils are relatively thin throughout the basin. There are no large areas of deep soil, colluvium, or alluvium, so the storage potential of the soils is distributed fairly evenly throughout the basin. In the Battle Creek watershed, upland soils of the Plateau surface are only slightly deeper than those within the Whites Creek watershed, but there are considerable differences on the valley slopes and alluvial valley floors. The relatively deep alluvial soils of the Sequatchie, Staser, and Hamblen series surround the stream channels and are therefore properly located so as to provide, in conjunction with the sometimes cavernous limestones found in the valley, the regulating hydrologic function described by LeGrand (1983). A similar combination of storage locations occurs in the valleys of all the Group Two streams except for Clear Fork.

Channel and Basin Topography

Several important studies of flood rates have found that channel slope is an important factor in the determination of flow rates(Benson, 1962; May, et al, 1970). Many such studies have concluded that channel slope is second only to drainage area in explaining flood peaks. The steeper the channel, the greater the flood peak. This relationship is presumably due to the ability of a steeper channel to deliver surplus water from all parts of the basin to the outlet more rapidly. The slope factor would seem to help in explaining the observed pattern of variation. Whites Creek and Richland Creek have rather steep channels, as do the Elk River and Battle Creek. In the first two, basins already expected to have high peak flow rates are given one more reason to have them, while the steep channels of the latter pair may serve to offset partially the effects of deep soils and limestone terranes. The unusually gentle slope of Clear Fork may play a role in moderating its flood rates.

Channel storage capacity varies considerably among watersheds, as some have pool and rapid sequences while others slope more uniformly. The deep, wide pools of streams such as the Obed may serve to delay runoff, thereby reducing flood peaks. This, together with gentler channel slope, may explain in part why the Obed's flood flows are so much lower than those on adjacent Whites Creek. Although other characteristics are similar, Whites Creek lacks deep pools and has a much steeper channel.

Basin shape factors are sometimes important in determining runoff concentration. Long, narrow,basins generally have lower peak flood rates than circular basins because of the timing of tributary input (Ward, 1978). This appears to be true of the Plateau streams. The elongate basins, such as the East and West Forks of the Obey, have relatively low flood peaks, while some of the more compact basins, such as Whites

Creek, Bee Creek, and Richland Creek, are heavy flood producers. Exceptions are Battle Creek and Caney Fork, which have high circularity ratios and form factors but rather low unit floods.

Drainage area is the most important of all factors affecting flood peaks; large basins produce high flow rates and small basins produce low peaks. The relationship, however, is not linear; small basins produce higher flows per unit area than large basins (see Figure VI-1). In order to avoid the effects of drainage area, all flows were standardized to unit drainage area, and only those basins with drainage areas of 50 to 800 square miles were included in the study. If the large watersheds were consistently underproducing and the small ones were overproducing, it might be necessary to consider further adjustments to flows based on drainage area (such as a logarithmic standardization of flows). This was not true for Plateau streams, however (see Figure III-1, p. 22).

The degree of inclination of the land surface exerts an important influence on the rates of delivery of excess moisture to stream channels, especially in forested basins where subsurface flow is dominant. The Clear Fork watershed has the highest percentage of area in gentle slopes. The Obed/Emory, Whites Creek, Richland Creek, Caney Fork, and Bee Creek watersheds are also relatively undissected. The East Fork of the Obey, Wolf River, Elk River, Collins River, and Battle Creek basins have broad flat uplands and also a large area of dissected escarpment. The basin of the West Fork of the Obey River consists almost entirely of highly dissected Plateau remnants, with only small areas of gently sloping up-

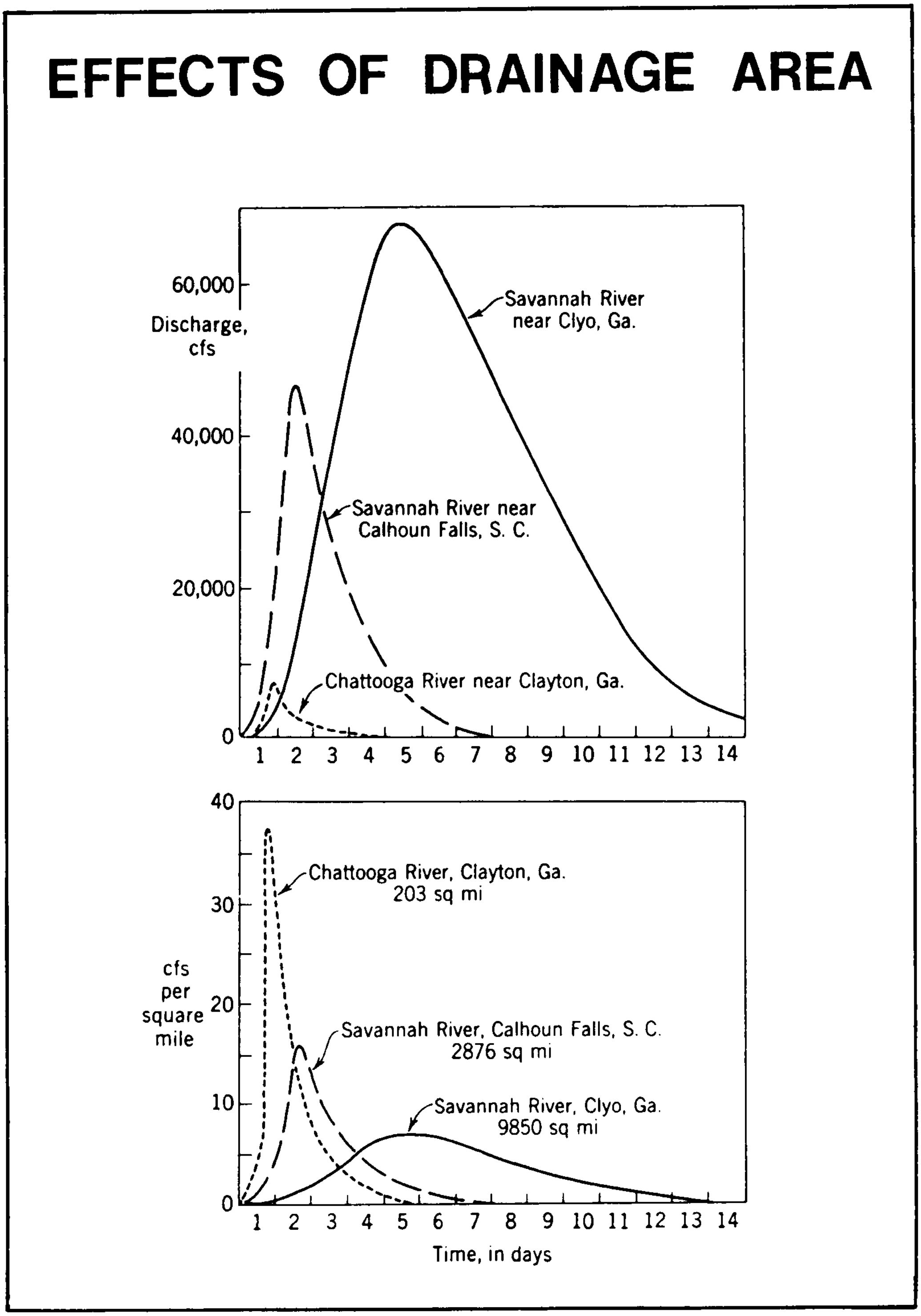

Figure VI-1. Effects of Drainage Area.

Source: from W.G. Hoyt and W.B. Langbein. 1955. Floods. Princeton, N.J.: Princeton University Press.

lands and floodplains. Slope does not seem to correspond well with general flood peaks in the study area. This is probably because the steep, dissected western escarpment is covered by thick colluvium that detains and stores significant volumes of water (Zurawski, 1978). Within Group One, however, the basins with the steepest slopes have the highest flood peaks. This suggests that once bedrock and soil characteristics have been taken into account, slope becomes an important factor in the determination of flood peaks.

In many studies, it has been found that high flood peaks are produced by drainage basins with high drainage densities. For example, Carlston (1963) stated that the mean annual flood is directly proportional to the square of drainage density and that base flow is directly proportional to the inverse square of drainage density; that is, a basin with high flood peaks and minimal base flows should have high drainage density. Cumberland Plateau streams generally have low drainage density values. Carlston's reasoning was that "Drainage density is adjusted to the most efficient removal of flood runoff" (Carlston, 1963). The reasoning behind this argument is that Horton's "length of overland flow" is inversely proportional to drainage density (Horton, 1945).

> Thus, the higher the drainage density, the shorter the time required for runoff to reach a channel. This reasoning has suggested that basins with higher drainage densities should have higher flood peaks (per unit area), hydrographs with shorter time bases and faster recessions, a larger ratio of direct runoff to base flow, and high average runoff. (Dingman, 1978).

This argument is reflective of the continuing dependence of geomorphologists on Horton's concepts of runoff generation, which have been essentially refuted as they relate to humid, well-vegetated terrain. Horton's concept was that rainfall intensity often exceeds the infiltration capacity of the soil, resulting in puddling at the surface and eventual direct surface runoff (Horton, 1932). This direct surface runoff was believed to be the major component of runoff and the sole contributor to storm flows.

The traditional Horton model of inadequate infiltration rates yielding overland flow has been increasingly questioned by hydrologists in recent years (Horton, 1945). Beaseley (1976), Whipkey (1967), Hewlett and Hibbert (1967), Kirkby and Chorley (1967), Nutter (1973), and Aubertin (1971) have argued that subsurface flows contribute greatly to stormflow. In his article on "The Role of Soil Water in the Hydrologic Behavior of Upland Soils," Nutter noted the persistence of the overland flow concept in spite of a lack of corroborative evidence and made these statements:

> Except during the most extreme storms, all the precipitation falling on well-vegetated slopes infiltrates ... On permeable upland slopes in humid regions, overland flow is rarely observed and subsurface flow in the vicinity of the stream channel accounts for much of the storm runoff (stormflow). (Nutter, 1973).

In their study of runoff in small watersheds, Hewlett and Hibbert (1967) challenged the traditional Hortonian approach to runoff production in humid forest lands:

> For example, the statements or assumption that all floods are due to surface runoff has persisted in hydrology papers and even in some hydrology textbooks despite much evidence to the contrary in forestry and agriculture research. Virtually all techniques now in general use for estimating flood peaks and volumes are based on the assumption that a quick rise in streamflow proves that rainfall

> is failing to infiltrate and is running over the surface of the ground to stream channels.
>
> Usually a discussion of the runoff from a watershed begins with the assumption that direct runoff is the product of overland flow and that other types of flow are mere exceptions to the general rule. Perhaps the opposite approach is more logical in the case of the forest; that is, to begin with the assumption that all flow is subsurface until there is evidence otherwise.

These new concepts of runoff production must, then, throw doubt upong the relationship between drainage density and streamflow, because it has been shown repeatedly that Horton overland flow is not essential to the production of flood peaks. Dingman has recently noted:

> In spite of the widespread use of drainage density, there are aspects of its definition and its relation to streamflow characteristics that do not appear to have been critically examined in the context of a modern understanding of runoff generation processes. (Dingman, 1978).

Conditions exist on the Cumberland Plateau which would permit surplus water to arrive quickly in stream channels despite the relatively gentle slopes, low drainage densities, and highly permeable soils. It is, in fact, the high permeability of those soils that would permit water to move rapidly laterally through the matrix of the soil and through existing soil macro-channels. In laboratory tests,samples of the Hartsells soil series, which is the dominant series on the Plateau, have saturated permeabilities of 6.8 to 10.4 inches per hour in the A horizon and 3.3 to 8.8 inches per hour in the B horizon (Longwell, Parks, and Springer, 1963). Such rates of movement would still only result in a movement of 13 to 18 feet in a 24-hour period if applied to horizontal movement of free water in the soil. These tests were accomplished on prepared soil samples, however, and indicate only rates of movement through the soil fabric. Beasley (1976) has demonstrated that movement of stormflow is largely accomplished within macropores generated by burrowing animals and decaying plant roots. Such macropores are not present in prepared laboratory samples of soils. Lull and Reinhart (1975) cite numerous field studies which reported infiltration rates of over 50 inches per hour in undisturbed forest cover. Such rates could clearly afford sufficient time for most runoff to reach a stream channel within one to two days. The thin rubble zone at the interface between the soil and the sandstone bedrock has even higher permeability rates (Lietzke, 1981). While the possibility of saturation overland flow need not be eliminated, it is probably not common on the rolling terrain and sandy loam soils of the Plateau. Dense drainage networks are not required in order to move excess water quickly and efficiently to the main stream channels.

The permeable soils of the Plateau generally overlie a well indurated sandstone formation interbedded with thin shale layers. Low permeability of these shales is suggested by the artesian pressure encountered by wells in the Rockcastle formation. Low intergranular permeability of these formations is indicated by the low rates of delivery from wells. Of the wells tested on the Cumberland Plateau, one-half yield less than 10 gallons per minute, and only fifteen percent yield as much as 25 gallons per minute (Tennessee Division of Water Resources, 1972). Field investigation during the present study encountered no evidence of significant emergence of water from the bedrock along the walls of the larger valleys which drain to the east. With shallow, permeable soils and bedrock of modest permeability interbedded with impervious shales, the Plateau would seem unusually well fitted to deliver infiltrated water quickly to nearby stream channels via subsurface flow.

Because the bedrock tends to exclude water, while the soils are highly conducive to lateral movement, surplus water naturally follows the lateral route. There is little opportunity for storage and detention of water in the shallow soils and relatively impermeable bedrock of the Plateau, so surplus water moves quickly into channels and is rapidly removed from the basin. Because these flows are concentrated over a brief period of time, flood peaks are high and base flows are low.

None of the streams has more than 0.2 percent of its drainage area in lakes and ponds, so surface detention is not likely to be an important determinant of flood peaks and low flow characteristics. Urbanization is not great enough in any of the basins to affect streamflow significantly. Other types of land modification vary more extensively among basins and may have had some influence on past flow conditions. Strip mining is much more extensive in some basins (such as the East Fork of the Obey) than the others, but Bryan's (1981) investigations indicate that the impact of mining on flow rates is not great. Logging is an important activity on the Plateau and has been practiced throughout the period of record. It has been shown repeatedly that deforestation causes an increase in runoff, particularly during stormflow (Ward, 1978). While it is likely that some of the extreme flows have been exacerbated by intense deforestation, it is unlikely that any significant differences in flow patterns are attributable to such activities, as all basins in the study area have comparable areas of their uplands in forest cover. Increased flood heights (rather than volumes) have been attributed to the damming effect of sawn logs picked up by flood waters which lodge against bridges (T.V.A., 1930).

Field Investigations

In order to determine the sources of streamflow during low-flow conditions, field invsetigations were undertaken. Streams representing type one and type two runoff patterns were examined.

Streamflow was measured at three sites on Whites Creek on July 21, 1982. Heavy rains on the night of July 20 had caused the stream to rise by 6 inches or more at some sites, as indicated by low water marks on boulders in the channel. Discharge was measured at the point where the stream emerges from the Plateau, where Black Creek Road ends. At this point, streamflow was measured at 47.6 cubic feet per second (c.f.s.). Flow was then measured at a point 1.5 miles upstream, just below the confluence of Piney Creek and Whites Creek. Flow at this site was measured at 53.6 c.f.s. The difference in measured flow getween the two sites is negligible, as errors of measurement of this magnitude are easily possible. Both measurements were made at the head of riffle segments, where the stream flows over coarse sandstone cobbles. The presence of these cobbles makes depth measurements difficult and causes considerable turbulence, which can make flow rate determination difficult. The reach between the two sites has only two tributary streams, Ford Branch and Laurel Creek. The discharge of Whites Creek from these two streams was immeasurably small at the time that these measurements were made.

A third flow measurement was made on Piney Creek 100 yeards upstream of its confluence with Whites Creek. The flow at this point was found to be 39.8 c.f.s., which is a reasonable proportion of the total flow, based upon the drainage areas of the two sub-basins.

The discharge measurements made on Whites Creek indicate that it is neither significantly influent nor effluent along its lower reaches. This is not surprising, because the stream is incised into shale and sandstone-conglomerate, with very little alluvial fill in the valley bottom. The consecutive reach flow survey indicates that there is not large contribution of groundwater to low flows from concentrated sources such as springs or massive seepage zones. These findings are consistent with the suggestion that Group One streams have little storage of surplus runoff that would supplement low flows.

Field investigations of the flow rates and sources of water for the Wolf River were made during August of 1982. All streams on the Plateau were flowing at rather low rates at that time. At this time, the main tributaries to the Wolf, such as Delk Creek, Dry Creek, and Little Dry Creek were all completely dry, and only a trickle of water was flowing down Poque Creek. Wolf River has no other tributaries, but was flowing at a rate of approximately 40 to 50 c.f.s. at the Highway 28 crossing near Pall Mall.

The source of this discharge was found to be two large spring systems. The first is actually a large, deep cavern system know as Blowing Cave, through which a small stream flows (see Figure VI-2). Only about 5 c.f.s. was flowing from Blowing Cave. The large volume of air blowing through the cave is evidence of large openings to the surface elsewhere

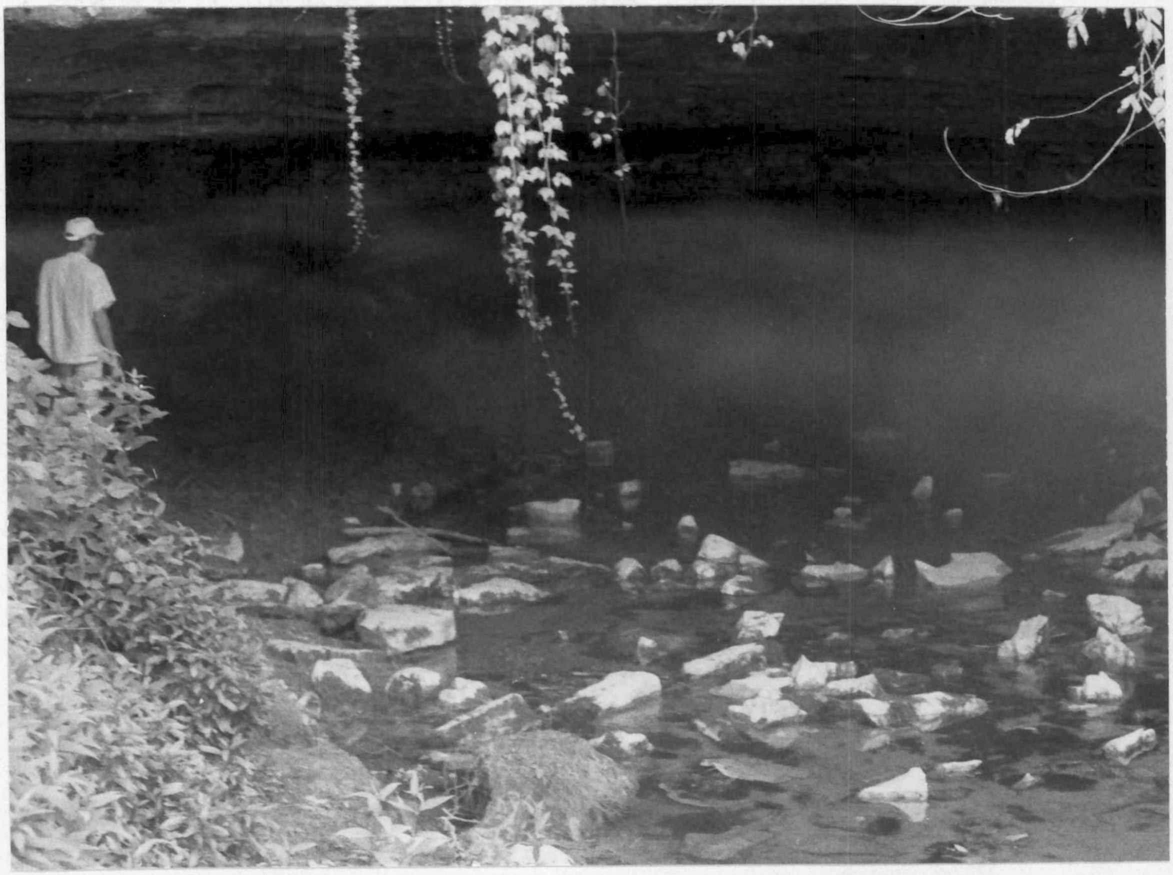

Figure VI-2. Blowing Cave in the Wolf Watershed.

in the cave system. Rather than acting as a sink for large volumes of water, this cave is probably not much different from a surface stream in hydrologic behavior.

The primary source of low-flow discharge is another mile upstream at Tater Cave (see Figure VI-3). This spring was discharging water at a rate of 15 to 20 c.f.s. Even though the discharge from Tater Spring is much more reliable than that of Blowing Cave, evidence in the form of mud stains on rocks indicates that the stream occasionally flows at least 4 feet deep.

The pattern of high baseflows and hig peak flows in the Wolf River basin is more easily understood in light of these spring systems which sustain it during times of low runoff. Unlike the spring systems of the Collins, Elk, and Calfkiller Rivers, the springs of the Wolf appear to be rather like open tubes that divert surface water for a brief time but do not store large volumes of water. The total amount of water discharged by the Wolf over a period of 1 month at a rate of 10 c.f.s. (the average 1-month low flow) is equal to only 0.11 inches of water. Despite the presence of an active streamflow-groundwater exchange system in the Wolf watershed, floods are very high. It would appear, then, that either the volume of storage is not very large or that the water table is deep enough that the storage is not usable.

During a field trip to the Collins watershed on August 4, 1982, streamflow rates were either measured or estimated at several points along the Collins River and its major tributaries in order to determine the places where water was being lost to subsurface systems or added to the

Figure VI-3. Tater Cave in the Wolf Watershed.

stream by spring discharge. All three of the upper tributaries of the Collins (Savage Creek, Big Creek, and the upper Collins) were flowing at an estimated rate of 5 to 10 c.f.s. It was not possible to actually measure discharge rates from these streams because of the shallow, rocky nature of the channels at low flow.

The channel of the Collins at the Highway 56 bridge, 2 miles northeast of Beersheba Springs, was completely dry, and trees with ages of 3 to 5 years are growing in the channel (see Figure VI-4). Mud stains on the tree trunks indicated that moderate channel flows had occurred within the last year. One-half mile farther downstream, the channel of the Collins was totally clear of vegetation and had some standing pools of water. The channel in this section of the river is covered by coarse cobbles and has a dramatic pool and riffle sequence, with the dry pools lying 10 feet and more below the level of the adjacent riffles (see Figure VI-5). This channel configuration indicates that relatively high discharge rates are experienced from time to time, while the channel only one-half mile upstream is rarely subjected to flows of sufficient depth and velocity to keep itself cleared of vegetation. Between these two points, a large boulder-choked blowhole was found along the west side of the channel. No water was being discharged from the blowhole on August 4, but mud stains on the boulders all around it indicate that flows from this subsurface system are rather substantial at times.

Big Spring is located along the west side of the Collins River, several hundred yards up a small tributary and 2 miles downstream from the Highway 56 bridge. Discharge from the springs was determined to be

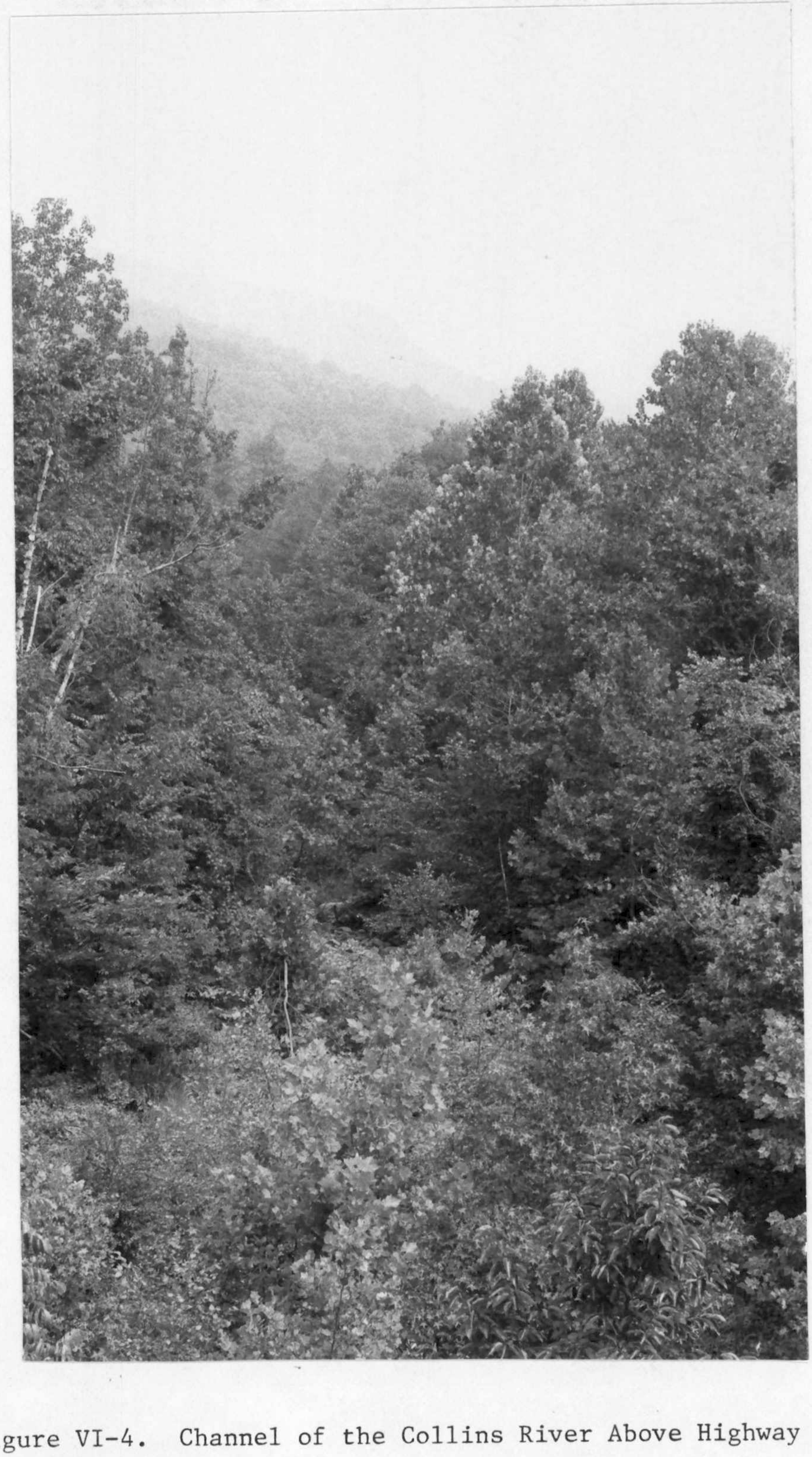

Figure VI-4. Channel of the Collins River Above Highway 56.

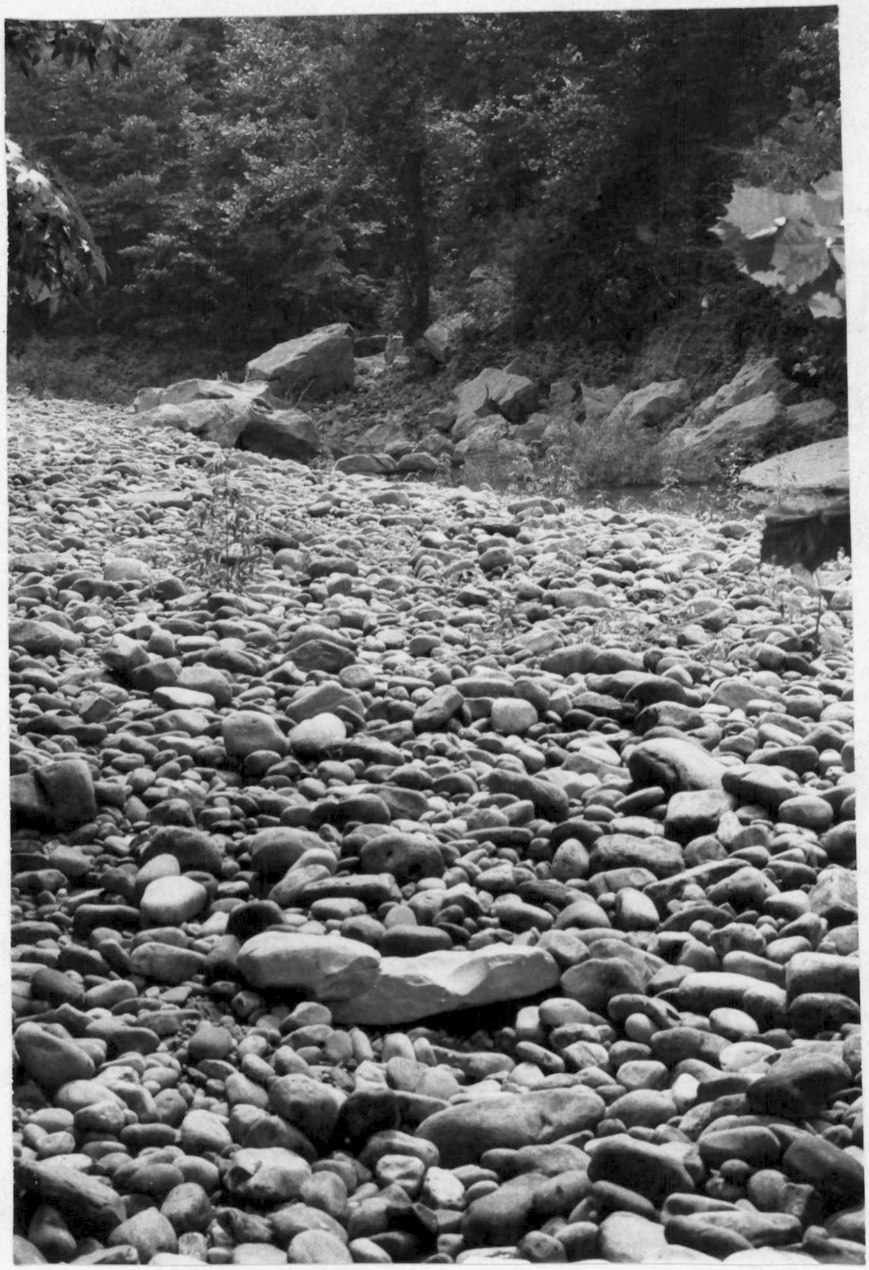

Figure VI-5. Dry Channel of the Collins River.

at least 40 c.f.s. on August 4, 1982 (see Figure VI-6). During low flow conditions, the majority of the water in the Collins reaching McMinnville is derived from this spring.

The discharge of Big Spring was greater than the sum of the discharges of the tributaries above the point where the cavernous limestones are encountered. This is direct evidence that the high baseflows of the Collins are supported by delayed groundwater discharge from spring systems. The blowhole system apparently serves to divert streamflow during times of high discharge, bypassing the middle section of the channel. This system does not contribute to low flows at all and probably causes little delay in stormwater discharge. The Collins and Calfkiller Rivers provide the best examples on the Plateau of the diverted streamflow mechanism for runoff dampning suggested by LeGrand (1983).

In summary, it has been shown that there is a critical shortage of water storage capacity across the entire upland of the Cumberland Plateau. Within parts of the region, this is at least partially offeet by the availability of storage in the soil-regolith zone adjacent to streams and in locally cavernous limestones. This storage system serves to regulate runoff rates by detaining surplus water for varying periods of time. The Group One streams, however, lack such storage, while the Group Two watersheds have considerably more regulating storage available. Within Group One and Two, channel slope is of some importance, as are land surface slope and basin shape. Each of these factors, however, is of secondary importance compared to the aforementioned storage factors.

Figure VI-6. Measurement of Discharge from Big Spring.

Some basin characteristics that have been shown to be important determinants of streamflow in other areas were not deemed significant in this study, because their values vary but little from basin to basin within the study area. These include forest cover, area in lakes and ponds, and various climatic factors. Drainage density was shown to be an insignificant, or even misleading, factor in consideration of actual runoff processes.

CHAPTER VI

SUMMARY, IMPLICATIONS, AND DIRECTIONS FOR FURTHER RESEARCH

It has been shown that the flow regimes of Cumberland Plateau streams are unlike those of most other streams in Tennessee (see Figure I-3, p. 5). Only the streams of the Nashville Basin have similar regimes. Streams of the Plateau have high peak flow rates, as indicated by floods of 2- to 50-year recurrence values (see Figure III-1, p. 22, Table III-1, p. 23). Low flows of Plateau streams are poorly sustained, as indicated by flow duration curves for streams that are representative of the various regions within Tennessee (see Figure I-3, p. 5). Only in the Nashville Basin do streams have comparable low flows.

Within the Cumberland Plateau, stream response varies. Peak flow rates for the 50-year flood range from a low of 100 c.f.s./mi.^{2} for the Calfkiller to a high of 375 c.f.s./mi.^{2} for Whites Creek, with the entire range between the extremes being represented. Low flows are equally variable in magnitude in Plateau streams, although the absolute differences in flows are not great (see Figures III-2, p. 25 , III-3, p. 38 , Table III-1, p. 23). At the two extremes, some very large streams such as the Obed and Caney Fork cease to flow entirely at times, while the Collins and Calfkiller Rivers have relatively well sustained low flows.

In an earlier study of the flow regime of the Obed-Emory system, it was found that although many of the basin characteristics of that watershed are those that are generally believed to produce stable flow regimes, the Obed has a very erratic hydrograph, with unusually low base flows and extremely high floods (Mayfield, 1980). One need not deny, however, the importance of channel slope, basin shape, area in lakes and ponds, drainage density, and other factors suggested by numerous researchers as being important determinants of flood discharge. It must now be realized that these factors are important in contributing to differences within the region, but that an overriding factor sets the region apart from surrounding areas (Dalrymple, 1960, p. 29; Benson, 1962, pp. 21-31). This overriding factor of watershed hydrology is the volume water storage available within the basin. As stated by B. J. Knapp (1979),

> The nature of the response of any stream depends on: (a) how quickly storage can be filled up thus allowing an increase in flow; and (b) how much usuable storage there is to sustain flow once rainfall has ceased. Before a stream can respond to rainfall in a catchment with permeable bedrock, a much larger storage capacity has to be filled than is the case with impermeable bedrock.

While many researchers have recognized the importance of storage volume and discharge systems, they have generally chosed to ignore it, because it means that differences in flood rates are controlled by factors not easily measured. As stated in a recent U.S.G.S. report, "The dominant factors are probably soil types and geology which are not readily susceptibile to mathematical analysis because they are difficult to quantify" (May, 1981). Similar statements have been in numerous other reports (Dalrymple, 1960, p. 30; Benson, 1962, p. 30). If a dificiency of storage results in poorly sustained low flows, it must conversely result in more rapid runoff as well.

The Cumberland Plateau has been shown to be an area which has only limited amounts of storage, either in the bedrock system or in regolith and soils. This element of the physical geography of the Plateau distinguishes its hydrologic response from those of all other physiographic provinces in Tennessee.

Within the Plateau region, too, differences in amounts of usable storage are in part responsible for major differences in runoff regimes. The streams which drain the western and southeastern flanks of the Plateau have larger volumes of storage in their deeper soils and limestone bedrock, while the streams draining the eastern flank have shallow soils, little ground water storage, and much flashier streamflow regimes. Detailed analysis of the Whites Creek and Battle Creek watersheds emphasize such differences. Within the Cumberland Plateau itself, the traditionally emphasized factors of basin shape, channel slope, and land slope are also significant. Of the watersheds with small volumes of storage, the basins with steep slopes (Whites Creek, Richland Creek) have much higher peak flows than those with significantly gentler land and channel slopes (Obed-Emory, Clear Fork).

Implications of this Study

An understanding of flooding characteristics of a region is directly useful for land use planning. Because earlier state-wide reports have tended to underestimate the peak discharges of Cumberland Plateau streams, critical mistakes could be made in determining the likelihood and height of flooding of floodplains of ungaged streams draining the Plateau.

Most textbooks of physical geography continue to emphasize Horton overland flow as the primary component of stormflow, despite the fact that many recent field investigations in Europe, Australia, and the United States have failed to substantiate its existence under natural conditons. This study shows that watersheds with characteristics favoring shallow subsurface movement of water are capable of producing high flood peaks, contrary to prevailing notions of stormflow in commonly read textbooks. These characteristics include highly permeable soils, heavy forest cover, and generally gentle slopes.

Geomorphic implications of the frequent high flood peaks of Plateau streams are briefly examined in Chapter V. The frequent occurence of large floods on such streams as the Obed and Whites Creek would suggest a high potential for geomorphic work in the form of fluvial erosion and transport. The coarse materials on the floodplains of these streams can only be moved by the highest floods, however. Using the DeBoys criterion for boundary shear stress generated by flowing water, the ability of a stream to transport coarse debris may be calculated. The 2-year flood of the Emory generates a shear stress 40% greater than would be generated by a 2-year flood on an "average" East Tennessee stream (May, et al, 1970; Randolph and Gamble, 1976). At the Oakdale gage, this would be equivalent to an increase in maximum diameter of mobile particle from 12 inches to 20 inches (Baker, 1975). Despite this high level of competence, field investigations have revealed that streams of the Plateau modify their bank deposits infrequently because of the coarseness of materials carried.

This study also has implications in the field of physical hydrology. Most significantly, it confirms the possibility of significant storm flows from watersheds that favor rapid, shallow throughflow rather than overland flow (Beasley, 1976). The highly permeable soils, gentle slopes, impermeable bedrock, and nearly complete forest cover all work to insure that water moves beneath the ground surface.

Directions for Future Research

Many additional questions concerning the regional hydrology of the Cumberland Plateau are yet to be answered. Some could be answered only by further detailed field study of individual basins, while others can best be pursued by more refined analysis of the data already available.

One way in which this study could be expanded would be to enlarge the study region. Because there are significant topographic, geologic, and land use differences to the south in Alabama and to the north in Kentucky, it would be interesting to extend the study into those states and even farther north to determine the degree of similarity between streamflow regimes in the various subsections of the Cumberland-Allegheny Plateau. With an increased number of streams to study, statistical analysis of the importance of individual watershed characteristics would be appropriate. It would then be possible to define more precisely the importance of individual characteristics such as channel slope, soil water storage volume, land slope, basin shape, and forest cover.

Within the present study area, an analysis of land use change and runoff rates would be informative. The Plateau has been through several cycles of dramatic change inland use over the last century. These changes include massive clear-cutting in land use over the last century. These changes include massive clear-cutting, periods of significant deep mining and strip-mining, subsequent regrowth of forests, and most recently, extensive clearing for row crops and amenity developments. Thousands of small ponds and reservoirs have been built on the Plateau in recent years to supply water for communities, farms, and resorts. Peak flows are probably not affected significantly by such changes, but low flows may be altered appreciably. On the upper Obed River, low flows have been consistently higher in recent years because of regular discharge of water from the Crossville sewage treatment plant. The changes in land use could be determined from historic maps, logging and mining records, and aerial photography. It might then be possible to relate these changes to observed streamflow patterns and determine how land modification affected streamflow. There is not evidence to this point that land use patterns within the basins of the Plateau are responsible for much, if any, of the observed difference in streamflow characteristics, however.

Only a limited amount of useful groundwater data is available for the Plateau, because the number of monitored observation wells is small. Nonetheless, the records of seasonal water table fluctuations in watersheds which have such information could be tied with streamflow data and water budget calculations to determine more closely the timing and volume of groundwater recharge and subsequent discharge to streams. This would require calculation of such aquifer characteristics as porosity, permeability, transmissivity, and specific yield. It would be difficult to obtain accurate values for these parameters, because aquifer tests have not been made for the Rockcastle or the lower limestone aquifers, but even general ranges of values determined by comparing well yields to the changing water table could be useful in developing a better understanding of the relationships between streamflow and the groundwater systems of the Plateau.

Within individual basins, a number of research projects could be pursued that would better define routes of water movement, as well as the relationships between infiltration, overland flow, throughflow, stormflow, and sustained low flows. Most such studies would require significant funding for instrumented tracking of water movement. Dye tracing or radioactive isotope tracing of water is a commonly-used method for the determination of the routes which water follows in limestone terrane. Nick Crawford of Western Kentucky University has done some tracing work on the western limb of the Plateau, but with the purpose of understanding routes followed by subducted streams rather than understanding the storage and discharge of groundwater (Crawford, 1979). If such studies were coupled with measurements of discharge rates at the springs where the water is returned to the surface, more could be learned about the hydrologic systems of those watersheds.

Another approach would be to do detailed consecutive reach studies, whereby the zones of streamflow gain and loss could be examined in detail. This would show where streamflow was being gained or lost from the groundwater system over large areas rather than from well defined springs. Such studies were done on a small scale for Whites Creek and Collins Rivers, but more accurate surveys with better instruments should be completed, with measurements made at various rates of discharge.

The timing of discharge from individual watersheds could be examined in much greater detail if precipitation data could be obtained at intervals of 1 hour or less. By studying the time of concentration and lag factors for various watershed, a better understanding of the importance of the aforementioned factors of slope and basin shape could be realized. Going one step farther, standard flow routing techniques could be employed to understand the importance of discharge rates from various areas of the watershed.

Conclusions

Detailed study of the runoff regimes of the streams of the Cumberland Plateau has revealed significant variety in the pattern of streamflow response. Factors responsible for these differences appear to be primarily those that govern the volume of storage available in the basins, including groundwater storage systems, transitory storage systems, and soil-regolith systems. Factors used more commonly by hydrologists to explain runoff regimes, such as basin shape, channel slope, and terrain slope, were also found to be important determinants of peak flow rates, but only after the aforementioned storage factors were considered.

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APPENDIX

TABLE A-1

LOG-PEARSON TYPE III CALCULATIONS: EMORY RIVER

Year	Peak Discharge	Log Q	Log Q-$\overline{\text{Log Q}}$	(Log Q-$\overline{\text{Log Q}}$)^2	(Log Q-$\overline{\text{Log Q}}$)^3
1962	53,200	4.726	.020	.0004	.0000
1963	50,500	4.703	-.003	.0000	.0000
1964	42,400	4.627	-.079	.0062	-.0005
1965	45,100	4.654	-.052	.0027	-.0001
1966	25,500	4.407	-.029	.0894	-.0267
1967	52,300	4.719	.013	.0002	.0000
1968	34,900	4.543	-.163	.0266	-.0043
1969	33,300	4.522	-.184	.0339	-.0062
1970	110,000	5.041	.335	.1122	.0376
1971	27,000	4.444	-.262	.0686	-.0180
1972	32,400	4.511	-.195	.0380	-.0074
1973	170,000	5.230	.524	.2746	.1439
1974	63,600	4.803	.097	.0094	.0009
1975	87,200	4.941	.235	.0552	.0130
1976	41,000	4.613	-.093	.0086	-.0008
1977	107,000	5.029	.323	.1043	.0337
1978	35,300	4.548	-.158	.0250	-.0039
1979	44,000	4.643	-.063	.0040	-.0003
		$\overline{\text{Log Q}}$=4.706		.8593	.1609

Standard Deviation= .2248
Skew Coefficient= .9340

TABLE A-2

LOG-PEARSON TYPE III CALCULATIONS: WHITES CREEK

Year	Peak Discharge	Log Q	Log Q-$\overline{\text{Log Q}}$	(Log Q-$\overline{\text{Log Q}}$)^2	(Log Q-$\overline{\text{Log Q}}$)^3
1962	8,800	3.945	-.143	.0204	-.0029
1963	10,200	4.009	-.079	.0062	-.0005
1964	8,400	3.924	-.164	.0269	-.0044
1965	10,100	4.004	-.084	.0071	-.0006
1966	3,350	3.525	-.563	.3170	-.1785
1967	10,400	4.017	-.071	.0050	-.0004
1968	13,800	4.140	.052	.0027	.0001
1969	5,730	3.758	-.330	.1089	-.0359
1970	28,000	4.447	.359	.1289	.0463
1971	9,700	3.987	-.101	.0102	-.0010
1972	6,900	3.839	-.249	.0620	-.0154
1973	62,500	4.796	.708	.5013	.3549
1974	29,000	4.462	.374	.1399	.0523
1975	17,100	4.233	.145	.0210	.0030
1976	17,800	4.250	.162	.0262	.0043
1977	24,000	4.380	.292	.0853	.0249
1978	6,000	3.778	-.310	.0961	-.0298
1979	missing				
	$\overline{\text{Log Q}}$=	4.088		1.5651	.2164

Standard Deviation= .3128
Skew Coefficient= .4680

TABLE A-3

LOG-PEARSON TYPE III CALCULATIONS: RICHLAND CREEK

Year	Peak Discharge	Log Q	Log Q-$\overline{\text{Log Q}}$	(Log Q-$\overline{\text{Log Q}}$)^2	(Log Q-$\overline{\text{Log Q}}$)^3
1962	4,720	3.674	-.028	.0008	.0000
1963	4,180	3.621	-.081	.0066	-.0005
1964	1,860	3.270	-.432	.1866	-.0806
1965	6,240	3.795	.093	.0086	.0008
1966	2,030	3.307	-.395	.1560	-.0616
1967	2,260	3.354	-.348	.1211	-.0421
1968	2,420	3.380	-.318	.1011	-.0322
1969	3,680	3.566	-.136	.0185	-.0025
1970	10,500	4.021	.319	.1018	.0325
1971	2,860	3.456	-.246	.0605	-.0149
1972	4,350	3.638	-.064	.0041	-.0003
1973	10,500	4.021	.319	.1018	.0325
1974	10,800	4.033	.331	.1096	.0363
1975	8,330	3.921	.219	.0480	.0105
1976	8,680	3.939	.237	.0562	.0133
1977	8,280	3.918	.216	.0467	.0101
1978	8,450	3.927	.225	.0506	.0114
1979	6,200	3.792	.090	.0081	.0007
	$\overline{\text{Log Q}}$=	3.702		1.1867	-.0866

Standard Deviation= .2642
Skew Coefficient= -.3115

TABLE A-4

LOG-PEARSON TYPE III CALCULATIONS: EAST FORK OF THE OBEY

Year	Peak Discharge	Log Q	$\text{Log } Q-\overline{\text{Log } Q}$	$(\text{Log } Q-\overline{\text{Log } Q})^2$	$(\text{Log } Q-\overline{\text{Log } Q})^3$
1962	20,300	4.307	.074	.0055	.0004
1963	17,700	4.248	.015	.0002	.0000
1964	7,900	3.898	-.335	.1122	-.0376
1965	16,400	4.215	-.018	.0003	.0000
1966	9,120	3.950	-.272	.0745	-.0203
1967	21,200	4.326	.093	.0086	.0008
1968	8,910	3.950	-.283	.0801	-.0227
1969	24,000	4.380	.147	.0216	.0032
1970	30,800	4.489	.256	.0655	.0168
1971	17,400	4.241	.008	.0001	.0000
1972	11,400	4.057	-.176	.0310	-.0055
1973	44,800	4.651	.418	.1747	.0730
1974	23,100	4.364	.131	.0172	.0022
1975	31,700	4.501	.268	.0718	.0192
1976	9,280	3.968	-.265	.0702	-.0186
1977	24,700	4.393	.160	.0256	.0041
1978	10,600	4.025	-.208	.0433	-.0090
1979	16,300	4.212	-.021	.0004	.0000
		$\overline{\text{Log } Q}$=4.233		.8028	.0812

Standard Deviation= .2173
Skew Coefficient= .5217

TABLE A-5

LOG-PEARSON TYPE III CALCULATIONS: WOLF RIVER

Year	Peak Discharge	Log Q	Log Q-$\overline{\text{Log Q}}$	$(\text{Log Q}-\overline{\text{Log Q}})^2$	$(\text{Log Q}-\overline{\text{Log Q}})^3$
1962	15,000	4.176	.231	.0530	.0120
1963	5,880	3.769	-.176	.0310	-.0055
1964	1,850	3.267	-.678	.4597	-.3117
1965	6,950	3.842	-.103	.0106	-.0011
1966	3,260	3.513	-.432	.1866	-.0886
1967	10,600	4.021	.076	.0058	.0004
1968	5,280	3.723	-.222	.0493	-.0109
1969	5,320	3.726	-.219	.0480	-.0105
1970	11,600	4.064	.119	.0414	.0017
1971	16,800	4.225	.280	.0784	.0220
1972	4,580	3.657	-.288	.0829	-.0239
1973	19,100	4.281	.336	.1129	.0379
1974	15,900	4.201	.256	.0655	.0168
1975	21,000	4.322	.377	.1421	.0536
1976	8,630	3.936	-.009	.0001	.0000
1977	18,100	4.258	.313	.0980	.0307
1978	6,480	3.812	-.133	.0177	-.0024
1979	16,200	4.210	.265	.0702	.0186
		$\overline{\text{Log Q}}$= 3.945		1.5532	-.2529

Standard Deviation= 0.3023
Skew Coefficient= -0.6058

TABLE A-6

LOG-PEARSON TYPE III CALCULATIONS: CLEAR FORK

Year	Peak Discharge	Log Q	Log Q-$\overline{\text{Log Q}}$	(Log Q-$\overline{\text{Log Q}}$)^2	(Log Q-$\overline{\text{Log Q}}$)^3
1962	15,400	4.188	.060	.0036	.0002
1963	13,600	4.134	.006	.0000	.0000
1964	7,940	3.900	-.228	.0520	-.0118
1965	14,000	4.146	.018	.0003	.0000
1966	8,140	3.911	-.271	.0471	-.0102
1967	18,400	4.265	.137	.0188	.0026
1968	9,690	3.986	-.142	.0202	-.0029
1969	8,090	3.908	-.220	.0484	-.0106
1970	32,000	4.505	.377	.1421	.0536
1971	9,050	3.957	-.171	.0292	-.0050
1972	missing				
1973	missing				
1974	missing				
1975	missing				
1976	19,700	4.294	.166	.0276	.0046
1977	27,800	4.444	.316	.0999	.0316
1978	12,500	4.097	-.031	.0010	.0000
1979	11,400	4.057	-.071	.0050	-.0004
	$\overline{\text{Log Q}}$= 4.128			.4882	.0536

Standard Deviation= 0.1938
Skew Coefficient= 0.6609

TABLE A-7

LOG-PEARSON TYPE III CALCULATIONS: ELK RIVER

Year	Peak Discharge	Log Q	$\text{Log Q} - \overline{\text{Log Q}}$	$(\text{Log Q} - \overline{\text{Log Q}})^2$	$(\text{Log Q} - \overline{\text{Log Q}})^3$
1962	3,580	3.554	-.133	.0177	-.0024
1963	9,050	3.957	.270	.0729	.0197
1964	4,570	3.660	-.027	.0007	.0000
1965	3,500	3.544	-.143	.0204	-.0029
1966	3,940	3.595	-.092	.0085	-.0008
1967	2,180	3.338	-.349	.1218	-.0425
1968	5,230	3.719	.032	.0010	.0000
1969	4,230	3.626	-.061	.0037	-.0002
1970	6,340	3.802	.115	.0132	.0015
1971	3,600	3.556	-.131	.0172	-.0022
1972	1,430	3.155	-.532	.2830	-.1506
1973	15,800	4.199	.512	.2621	.1342
1974	6,060	3.782	.095	.0090	.0009
1975	5,770	3.761	.074	.0055	.0004
1976	7,720	3.888	.201	.0404	.0081
1977	15,600	4.193	.506	.2560	.1296
1978	3,450	3.538	-.149	.0222	-.0033
1979	3,130	3.496	-.191	.0365	-.0070
	$\overline{\text{Log Q}}$=	3.687		1.1910	.0825

Standard Deviation= 0.2647
Skew Coefficient= 0.3151

TABLE A-8

LOG-PEARSON TYPE III CALCULATIONS: BATTLE CREEK

Year	Peak Discharge	Log Q	$\text{Log Q} - \overline{\text{Log Q}}$	$(\text{Log Q} - \overline{\text{Log Q}})^2$	$(\text{Log Q} - \overline{\text{Log Q}})^3$
1962	3,200	3.505	-.095	.0090	-.0009
1963	3,700	3.568	-.032	.0010	.0000
1964	3,130	3.496	-.104	.0108	-.0011
1965	3,450	3.538	-.062	.0038	-.0002
1966	3,440	3.537	-.063	.0040	-.0003
1967	2,700	3.431	-.169	.0286	-.0048
1968	5,030	3.702	.102	.0104	.0011
1969	2,880	3.459	-.141	.0199	-.0028
1970	5,550	3.744	.144	.0207	.0030
1971	2,700	3.431	-.169	.0286	-.0048
1972	4,100	3.613	.013	.0002	.0000
1973	7,000	3.845	.245	.0600	.0147
1974	6,250	3.796	.196	.0384	.0075
1975	5,030	3.706	.106	.0112	.0012
1976	missing				
1977	5,200	3.716	.116	.0135	.0016
1978	2,770	3.442	-.158	.0250	-.0039
1979	4,750	3.677	.077	.0059	.0005
	$\overline{\text{Log Q}}$=	3.600		.2910	.0108

Standard Deviation= 0.1349
Skew Coefficient= 0.3116

VITA

Michael Wells Mayfield was born in Charlotte, North Carolina on August 17, 1954. He attended elementary and secondary schools in Charlotte and graduated from Myers Park High School in June, 1972. He attended Western Carolina University from 1972 until 1976 and received a Bachelor of Science degree with a major in Geography from that institution. He accepted a teaching assistantship at the University of Tennessee in September, 1976 and received the Master of Science degree with a major in Geography in March, 1980. Mr. Mayfield worked for for an environmental consulting firm part-time while continuing coursework at The University of Tennessee-Knoxville. He received the Doctor of Philosophy degree with a major in Geography in March, 1984.

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