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3. Climatology and Hydrology

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In a problem statement that preceded establishment of Coweeta Experimental Fo Hursh (1932a), recognized two problems for forestry research in the sout Appalachian Mountains: (1) how is streamflow regulated by vegetative cover, an(how can erosion be controlled by vegetative cover? In early reports on the new rese program, Hursh champions studies to describe and develop understanding of clin precipitation, soils, and topography of these mountains and how they interact forest vegetation to produce streamflow. The state of our knowledge of climate streamflow in the Coweeta Basin and our understanding of interactions with topc phy and vegetation are subjects of this chapter.

History of Climatic and Hydrologic Measurements

Climatic Stations

During the last 50 years, 30 climatic stations were operated at Coweeta. Most active for only a few years to measure the influence of various watershed treatm(Six stations were sited around the valley from 1936 to 1939 to establish the climato of the basin. Only the main Climatic Station 01 (CS01) has been continuously a(since the initiation of precipitation measurements there in August 1934. In April 1 air and soil temperature, relative humidity, wind travel, evaporation, and cloud c observations were added. Some early temperature and precipitation measurem

were taken near the CCC Camp but the major site has always been in the Labora
administrative area at elevation 686 m. Initially, CS01 was on the south side of Sl
Fork at the approximate location of a parking lot below the present office. In 1963
site was moved 85 m NE across the stream. Solar radiation measurements were a
in 1965 and conversion from manual and strip chart observations to automatic data
gers began in 1974.

Although many Coweeta streamflow and precipitation stations have long and
tinuous records, climatic stations outside the administrative area do not. A go
establish continuing climatic measurements above the valley bottom on north-, so
and east-facing slopes is being realized. In 1969 and 1974, records began at two
elevation sites (Table 3.1) and in 1985, we initiated climatic measurements high o
east-facing slope. An additional station is registering the climatic changes assoc
with clearcutting and deciduous forest regrowth on WS 7 and another began in
adjacent *Pinus strobus* plantation in 1986. All of these sites now use battery-pow
electronic data loggers.

Precipitation Stations

Over 135 precipitation sites have been gaged at various times in the last 50 years.
sity of gaging in the main Coweeta Basin was greater than is reported for any
mountain area of similar size. On several experimental watersheds, density was as
as 1 gage for every 4 to 5 ha while the base network of 50 gages covered the entire
with a density of 1 per 32.5 ha. This network was maintained for 20 years and, thr
much of this period, the 8-inch standard gages were measured after each storm. I
reduced in 1958 to a long-term network of 9 sites, each with a pair of gages. The am
in the 8-inch standard gage is the most accurate measure of the week's total preci
tion and observations from the recording gage are adjusted to this amount. The rec
ing gage trace for each storm is broken into unequal-duration intensity periods
maximum 5-, 15-, 30-, and 60-min intensities are interpolated for all storms over 6
total precipitation. Daily (midnight to midnight) totals are also calculated fron
intensity period data.

Table 3.1. Active Climatic Stations at Coweeta Hydrologic Laboratory

Station Name	Elevation (m)	Slope	Type of Site	Record B
		Permanent Stations		
CS01	685	Valley floor	Grassed field	August 1
CS17	887	North-facing	Forest opening	October
CS21	817	South-facing	Forest opening	July 1974
CS28	1189	East-facing	Forest opening	May 198
		Study-Related Stations		
CS20	866	South-facing	Regrowing forest after logging	August 1
CS24	740	South-facing	White pine canopy	April 198

Streamflow Stations

Permanent weirs were constructed on 32 Coweeta streams with records initiate 9 streams by 1934. By 1941, 28 streams were gaged on watersheds ranging from to 760 ha. Presently, 16 streams are being gaged. Head or flow depth was reco on strip charts until 1964. Now, head is recorded on 16-channel punched tape e 5 min. Recorders are checked by hookgage weekly. Chart records were manually grated until 1958, when computer processing of flow records began. Methods of extraction, flow calculation, and storm analysis developed at Coweeta have l utilized by hydrologists throughout the world. Details of climatic and hydrologic m urements and analyses are given by Hibbert and Cunningham (1966) and Swift Cunningham (1986).

Climate

The climate of the Appalachian Mountains is distinguished from that of surroun lowlands by characteristics of precipitation and temperature or evaporation. Ui Köppen's system, Coweeta's climate is classed as Marine, Humid Temperature ((because of high moisture and mild temperatures (Critchfield 1966; Trewartha 19 The lower elevations of Coweeta are borderline, because the basis for separating Marine climate from the adjacent Humid Subtropical is the mean monthly air temp ture and whether it is below or above 22°C in the warmest month (see Table 3.2). Thornthwaite's original scheme, Coweeta is in the wet, mesothermal, adequate rai (AB'r) climate. Thornthwaite's modified classification (1948), based on water sur or deficiency calculated from precipitation and potential evaporation, gives essenti the same result—perhumid, mesothermal with water surplus in all seasons.

Precipitation

Precipitation, as modified by topography, is a dominant factor of the Coweeta clim At the basin scale, precipitation consists of frequent, small, low-intensity rains ii seasons with little snow. Occasional large storms occur at shorter return intervals t experienced by adjacent, less mountainous areas.

In general, precipitation increases with elevation (about 5% per 100 m) along east-west axis of the Coweeta valley but changes little with elevation over north- south-facing side slopes (Figure 3.1). The pattern is fairly regular, except near ri lines where gage catch is consistently lower than at sites immediately downsl Hibbert (1966) found reductions of 30% in a 30 m change of elevation extending ab 100 m either side of a ridge. A compact array of 18 gages installed near the top of W after it was cleared showed 50% less rainfall in a cove facing the prevailing wind wh a strong updraft was observed. Ninety-three percent of Coweeta's precipitation f when winds are calm or blowing less than 2.2 m/s from southwest, south or south (Swift 1964). High precipitation points occur on the lee side of major gaps in the mo tains that rim the south and west boundary of Ball Creek. We hypothesize that moving across the ridges is accelerated, particularly through the gaps, and point

Table 3.2. Averages and Ranges of Monthly Means for Climatic Variables at Coweeta CS01

Climatic Variable	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
Precipitation (mm/month)												
Average	174	181	203	156	140	130	144	139	125	112	139	178
Minimum	50	28	72	13	32	32	9	41	6	1	24	23
Maximum	371	423	433	287	477	318	300	340	333	291	478	388
Evaporation (mm/month)												
Average	34	43	77	97	104	107	108	98	81	67	46	30
Minimum	16	11	50	57	83	79	88	77	60	48	34	14
Maximum	60	76	132	137	133	157	135	120	100	83	57	57
Air temperature (°C/day)												
Average	3.3	4.4	8.0	12.6	16.4	20.0	21.6	21.3	18.4	13.0	7.8	4.1
Minimum	−4.0	−0.4	1.7	10.0	13.4	17.4	19.6	19.5	16.5	10.8	4.8	−0.6
Maximum	10.9	8.8	12.8	15.0	18.8	22.9	23.1	22.8	20.4	16.0	10.9	8.5
Solar radiation (MJ/m^2/day)												
Average	8.0	11.1	14.1	17.8	18.2	19.2	17.9	16.2	14.0	12.5	9.0	7.0
Low	6.0	9.0	10.3	12.9	15.4	14.5	13.9	12.5	12.3	9.2	7.2	5.2
High	10.8	13.6	17.6	21.3	21.4	22.5	24.3	19.8	16.9	17.7	11.1	8.7

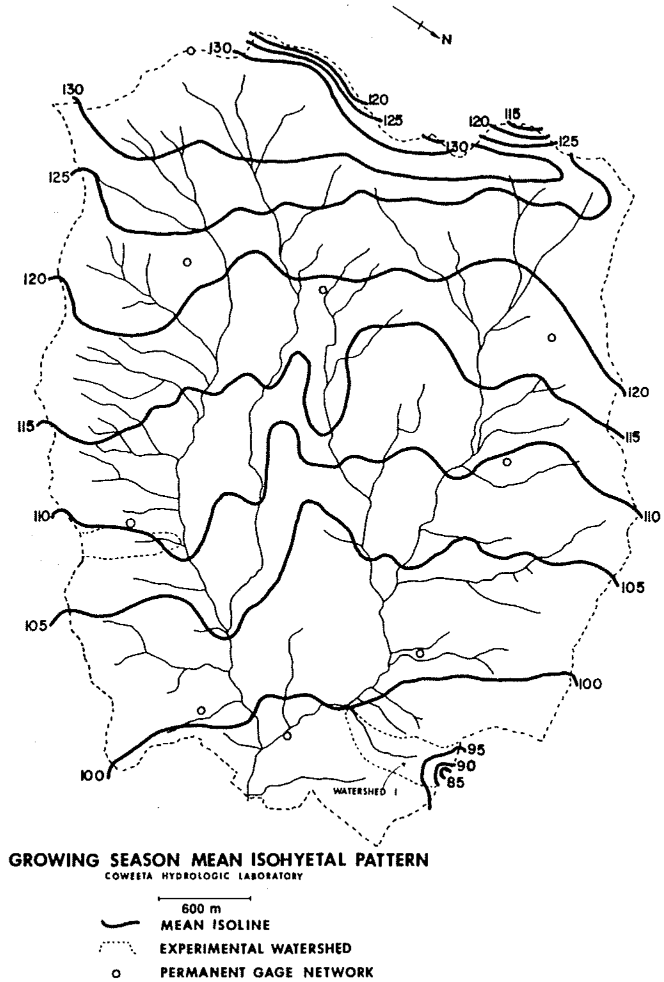

Figure 3.1. Growing season mean isohyetal pattern for Coweeta Hydrologic Laboratory. Va on mean isolines represent 100 times the ratio between season total precipitation at each m ured point in the basin and the catch at Standard Gage 15.

high precipitation represent dump zones for rain transported away from the rid; Based on long-term data and these studies of precipitation distribution, Swift (19 developed mean isohyetal maps and from these derived weighting factors for e mating watershed precipitation. Unlike the previously used mean polygon weigh system, isohyetal weighting uses ridgeline and dump zone patterns. Isohyetal lines w based on ratios of precipitation at each site to precipitation at a base or normali: gage. March et al (1979) extended the normalizing procedure while relating precip tion pattern for six storm types to topography.

Monthly mean precipitation amounts for the upper and lower elevations (Figure : show a consistent difference in all seasons. Precipitation is greater in late winter

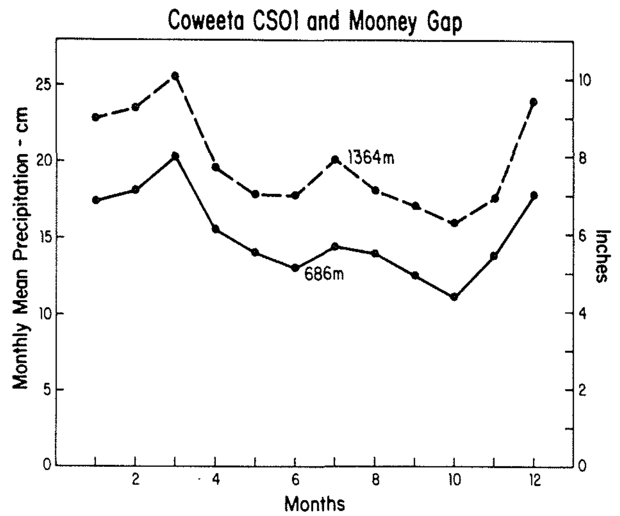

Figure 3.2. Long-term monthly mean precipitation at two gages near the elevational extr
at Coweeta. Mooney Gap (Gage 31) is near the Appalachian Trail at 1364 m above mean sea
(47-year record). Climatic Station 01 (Gage 6) is at 686 m (50-year record).

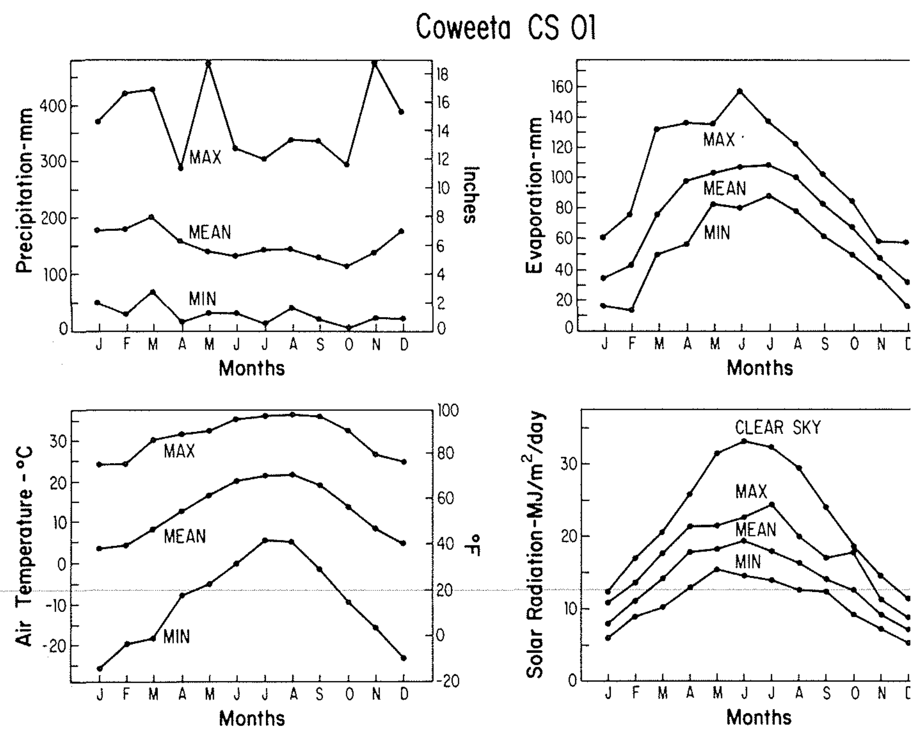

Figure 3.3. Mean monthly precipitation, evaporation, air temperature, and solar radiati
Climatic Station 01 compared with maximum and minimum monthly values.

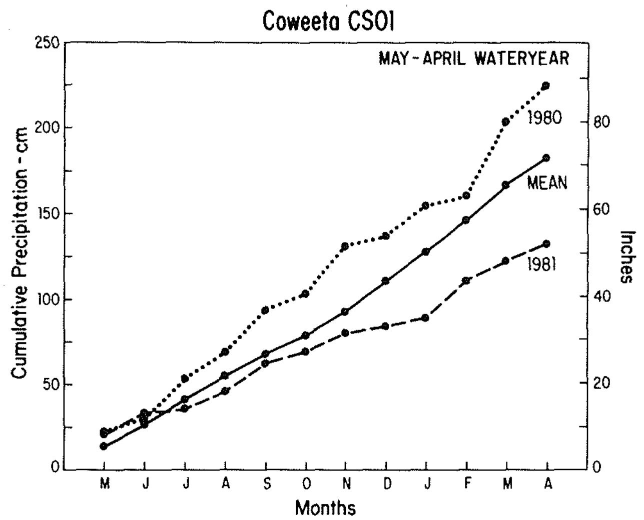

Figure 3.4. Cumulative precipitation at Climatic Station 01 for 2 extreme years compared cumulative monthly means. The May 1979 to April 1980 water year had the largest total prec tation out of 50 years (224.7 cm). The May 1980 to April 1981 year had the third lowest t (132.5 cm). The minimum year at Coweeta was 1986 with 122.9 cm.

spring with March the wettest month. Although fall months are the driest, half of record high rain days have been in the fall, typically caused by tropical storms. I amounts of 50 to 100 mm/day have been measured in every season. Rain falls in months. Snow comprises only 2 to 10% of annual precipitation at lower elevat CS01. Precipitation averages 152 mm/month at CS01 and has ranged from 0.8 mn October 1963 to over 475 mm in November 1948 (Table 3.2). Figure 3.3 shows means and extremes for monthly precipitation totals at this site, whereas Figure compares the cumulative precipitation for our wettest (1980) and third driest (19 years with accumulated mean precipitation. Except for the dry year, accumulation most rapid in the dormant season (November through April). High precipitation in in this low evapotranspiration period provides fairly consistent soil moisture recha by early spring.

The probability of measurable precipitation (>0.254 mm) on any date ranges fr 5% on November 30 to 73% on February 6 but a 5-day moving average puts the pro bility at around 40% for over 6 months of the year (Figure 3.5). On the average, storms are recorded annually. A storm is defined as a period of precipitation separa by rainless periods of at least 4 hr. Many storms are short and small (Figure 3. Consequently, interception loss accounts for over 20% of rainfall in half the storms e year. For discussion on interception losses see Helvey and Patric (Chapter 9) and interception processes see Murphy (1970). Daily precipitation amounts are skev toward 0 and fit the S_B, a three-parameter log-normal distribution (Swift and Schreu 1981), whereas groups of consecutive rain days best fit the gamma distribution.

Precipitation intensity is often used in simulation models to estimate streamflow a soil erosion, although Hewlett et al (1984) strongly question the value of intensity

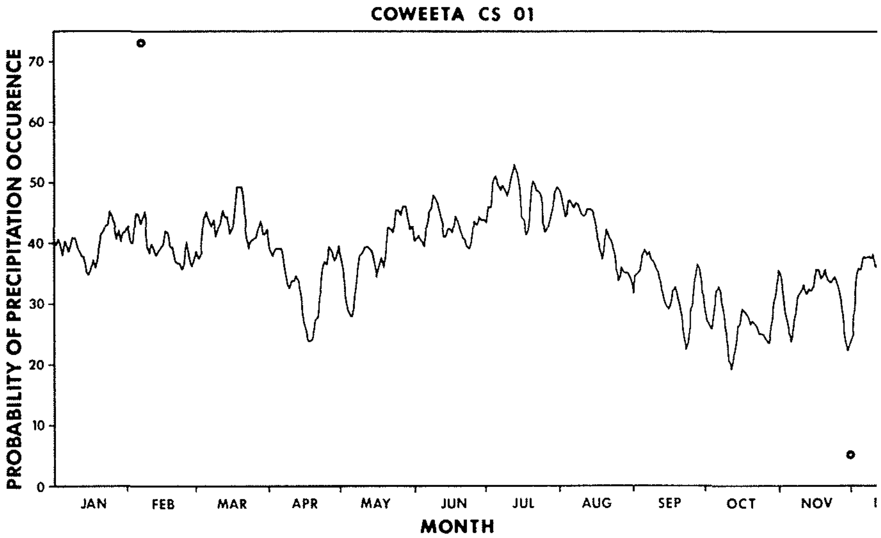

Figure 3.5. Five-day moving average of probability of precipitation occurrence for each d the year. Extremes are 73% on February 6, and 5% on November 30.

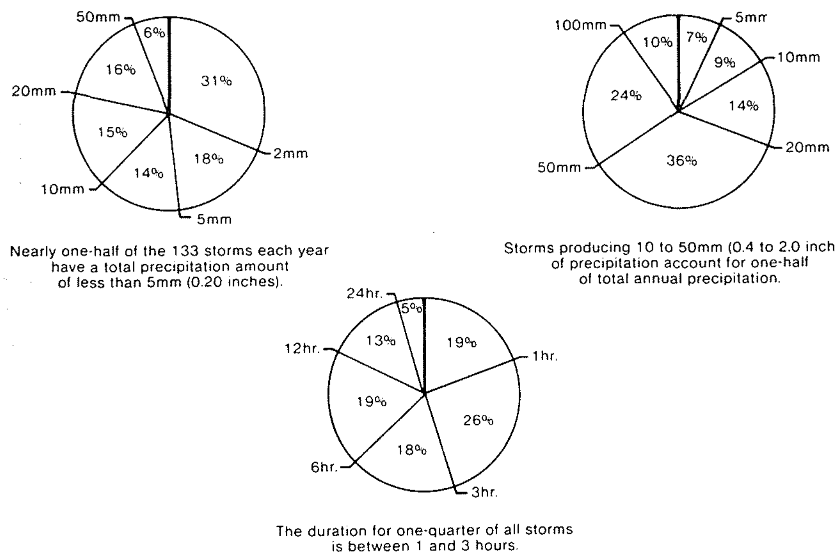

Figure 3.6. Storm size and duration relationships at Coweeta Climatic Station 01. Diagram based on 40 years' data from Recording Gage 6.

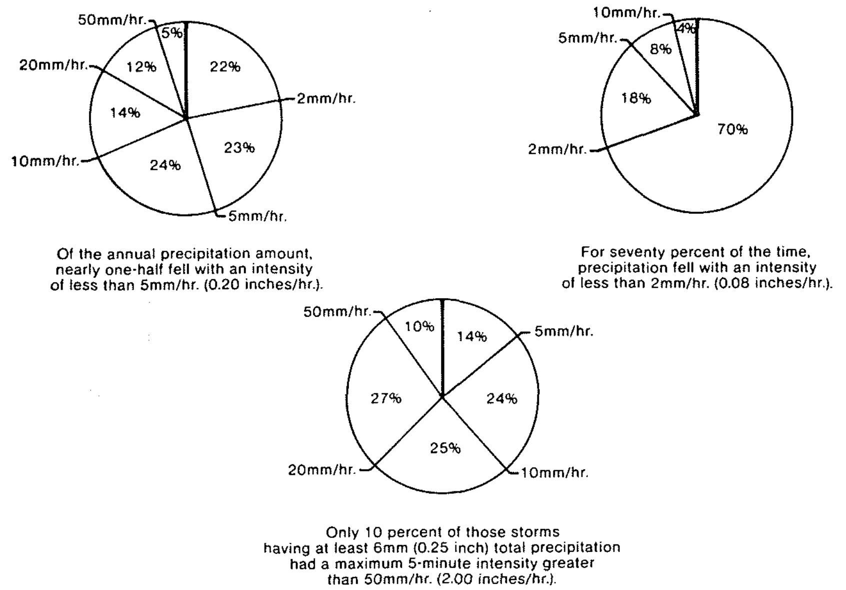

Figure 3.7. Precipitation intensity relationships at Coweeta Climatic Station 01. Diagrams based on 40 years' data from Recording Gage 6.

a predictor of stormflow. Figure 3.7 suggests a contributing cause for their conclus —nearly three-quarters of Coweeta's precipitation falls with an intensity of less tl 10 mm/hr, and only 10% of storms have maximum intensities over 50 mm/hr. Howe\ intensity and raindrop size are useful for estimating erosion rates. Drop size dis butions were measured at Mooney Gap (Gage 31) for 102 storm periods during months. Drop sizes ranged from 0.4 to 5.4 mm with the peak frequency at 1 to 2 r in diameter (Cataneo and Stout 1968). Drops of 1.5 mm have a terminal velocity about 650 cm/sec.

Because of point-to-point variability in mountain topography, standard tables : maps seriously underestimate rainfall intensities and derived parameters for Cowe and possibly other upper elevation sites in the southern Appalachian Mountains. erosivity index for the Universal Soil Loss Equation is based on both the intensity : amount of rain. The index for Coweeta is given as 300 in a frequently referenced n in Wischmeier and Smith (1978). However, annual erosivity indexes for Coweeta r Gage 6 give typical values of 400 to 500. Extreme rainfall frequency maps (Weat Bureau 1961) predict intensities of annual maximum storms for various return peri that are uniform over all the Coweeta Basin and surrounding terrain. These clos match the storm history for the low-elevation gage at CS01. In contrast, Bradfo (1977) depth–frequency–duration curves for high-elevation Gage 31 at Mooney ((Figure 3.8) are more similar to the Weather Bureau prediction for a high-rainfall z southeast of Coweeta. Earlier publications contain less detail and are not suitable use in the mountains because they were based on limited data from first-order stati such as Asheville, where precipitation is 53% of that at Coweeta. Greater precipitat

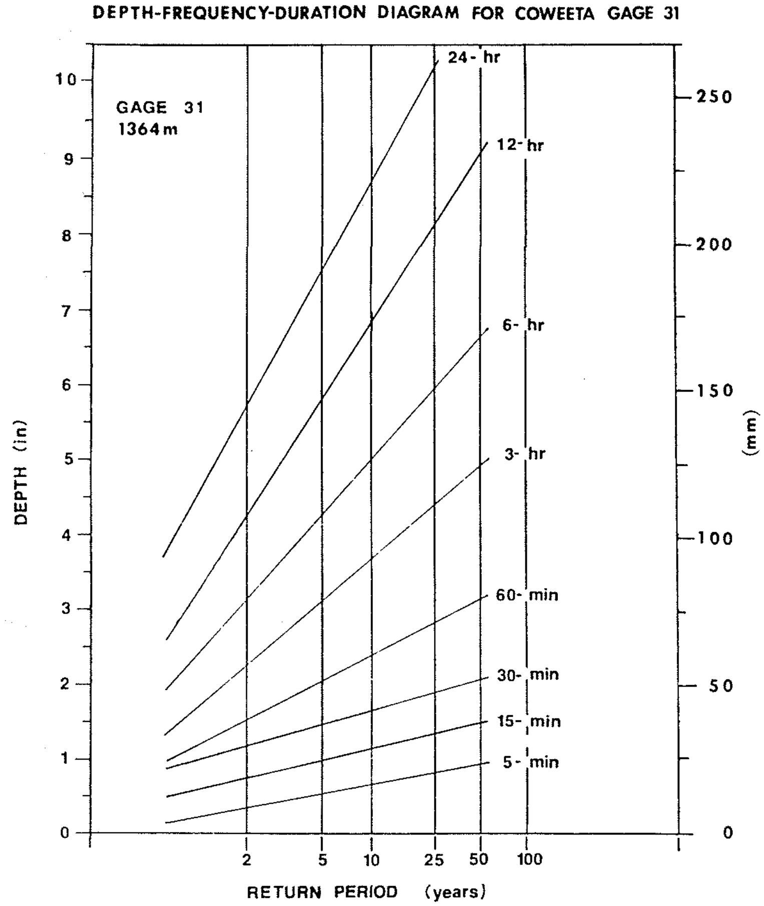

Figure 3.8. Precipitation depth-frequency-duration diagram for Rain Gage 31 at Mooney (Bradford 1977.)

at higher elevation sites along the southeastern edge of the Appalachian Mounta attributed to uplift and frontal contact of moist air masses moving inland. For 11 years (1963 through 1979) Coweeta Gage 31 (elevation 1364 m) caught more rain any other gage in the Tennessee Valley Authority system and was in the top fiv most of the other years. Thus, rainfall amount and intensity at the upper elevatio Coweeta are probably similar to those at Rosman, Highlands, and several mounta gages along the southeastern front of the Appalachian Mountains which regu report high annual precipitation totals.

Solar Radiation, Temperature, Wind, and Evaporation

Coweeta's temperate climate is due to the combined influences of elevation (6 1592 m) and latitude (35°N). The low latitude increases the potential solar radi available for sensible heat, plant growth, and evaporation, but the actual energy a point is influenced by the inclination and aspect of mountain slopes, shading by ac cent hills, and cloud cover. The seasonal difference between the mean monthly glo solar radiation and the curve for clear skies (Figure 3.3) is due to frequent afternc clouds in the summer months. Only 10 to 15% of the summer solar radiation is tra mitted through the fully leafed canopy, but in the dormant season about half penetra to the forest floor (Swift 1972). The record at Climatic Station 01 is slightly fau because the site is shaded by adjacent mountains from direct beam radiation apprc mately 60 and 90 min at the beginning and end of a midwinter day when radiatioı least. Nevertheless, radiation on mountain slopes can be estimated from the CS record (Swift and Knoerr 1973) using an algorithm for calculating potential radiat (Swift 1976). Mean monthly recorded solar radiation is approximately 45% of poten radiation. The algorithm has been useful in modeling evapotranspiration (Huff a Swank 1985; Swift et al. 1975), moisture content of leaf litter (Moore and Swa 1974), and streamflow increases due to forest cutting (Douglass and Swank 1975)

The majority of solar radiation retained on site is converted to heat or evapotrans ration. Figure 3.3 and Table 3.2 show the annual cycles of air temperature and eva ration at CS01. Warmest months are June through August, while December throu February are the coldest. Pan evaporation follows the same trend except that it pe a month earlier in the summer. The 50-year record temperature extremes w −27.8°C in January 1985 and 36.7°C in July 1952. Growing season degree d accumulate, on the average, from early May through early October (Figure 3.9). Ho ever, frost has been recorded at CS01 as late as June 2 and as early as September

Relative humidity in the forest drops to an average 50% during the day and rises 75% on winter or spring nights and over 90% most summer and fall nights.

Little is known about wind speed and direction around the Coweeta Basin. Lo term records show that wind speeds are low at CS01, except during some sto periods, probably because of the sheltered position of this valley bottom site. M

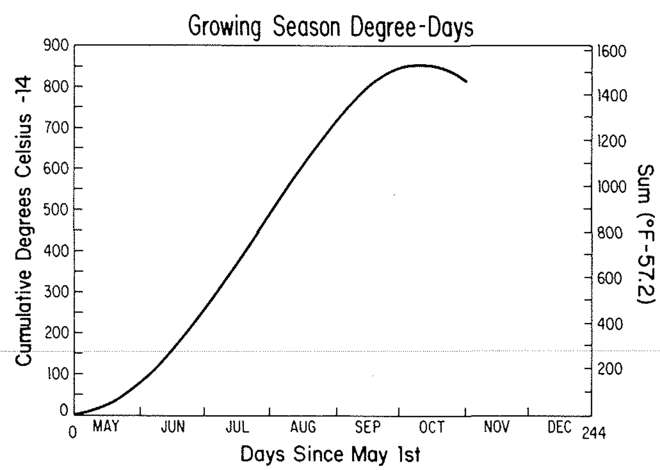

Figure 3.9. Growing season degree-days at Climatic Station 01, based on temperature mean each date over the period of record.

complete data are forthcoming from upper elevation stations CS 17, 21, and 28, equipped with moderately sensitive anemometers and vanes. In a detailed s Mowry (1980) consistently found linear wind profiles over the pine forest on proving that the theoretical logarithmic profile does not universally apply.

Annual evapotranspiration can be estimated several ways: from the evaporatior at CS01, empirical and energy balance equations, and the water balance of co watersheds. None of these methods are error free (Lee 1970), but several approxi 900 mm at the lower elevations. The mean annual total for the evaporation p 892 mm. With an admittedly low estimate for winter, Thornthwaite's PE e 696 mm while the PROSPER simulation (Swift et al. 1975) estimated ET, usi Penman-Montieth equation, to be 890 and 910 mm for 2 specific years. Av precipitation minus runoff values (P − RO) approximate 90 and 50 cm per year re tively for low-elevation WS 2 and high-elevation WS 36, 2 undisturbed watershe Coweeta (Figure 3.10). Huff and Swank (1985) present an application of long- Coweeta climatic data to simulate evapotranspiration and streamflow.

Hydrology

Hydrologic behavior of a watershed is determined by precipitation input, as mod by gradients of soil moisture storage and by evapotranspiration use. These combi define the annual hydrograph (cycle of monthly flow totals), the storm hydrograph characteristics for flow frequency and peak discharge for each stream.

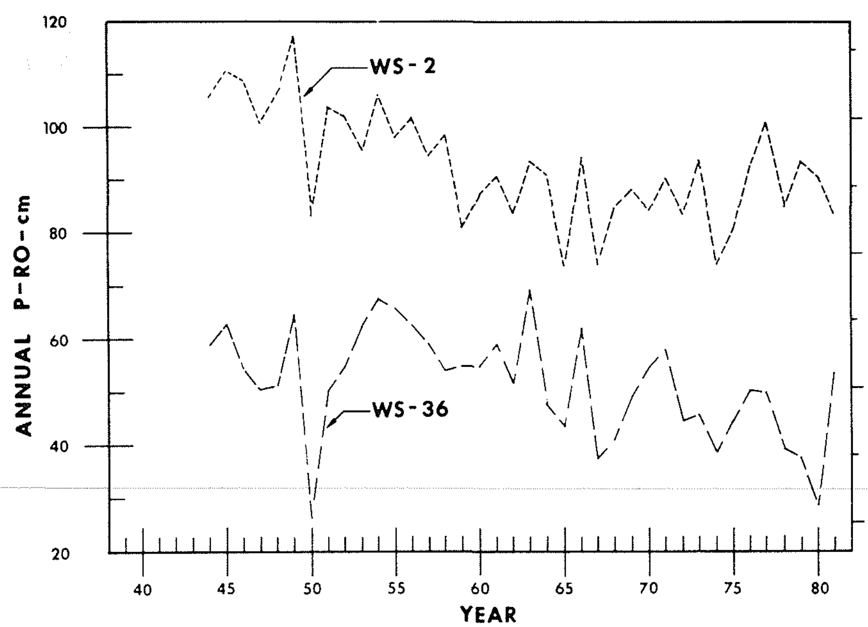

Figure 3.10. Range of annual precipitation minus runoff values for two control watersh Coweeta. Midelevations are 857 m for WS 2 and 1281 m for WS 36.

Hydrograph Analysis

Hursh (1932b) concluded that Horton's concept of rainfall excess and overland flow \ not applicable to most forest land because virtually all rain infiltrates into the s Thus, his early work was in the area of infiltration, storage, and movement of wate forest soils. Hursh (1942a) felt that runoff coefficients (such as total runoff/precip tion, storm runoff/precipitation, and storm runoff/total runoff) could be used characterize watersheds. Hursh and Brater (1941) discussed the need for and meth of separating hydrographs into direct and delayed (base) flow components and the r tive importance of the different sources of water that contribute to runoff fr Appalachian forests. With computers, consistent, systematic separation of the hyd graph into components became a practical reality (Hibbert and Cunningham 1966) runoff responses to rainfall could be examined in detail. Hewlett and Hibbert (19 found that many traditional runoff coefficients gave highly variable results betw watersheds, even within the comparatively small Coweeta Basin. Hewlett fin selected the ratio of annual mean quickflow (storm runoff) to mean precipitation as most useful response characteristic and used it to develop a response map, firs Georgia (Hewlett 1967) and then of the entire eastern United States (Woodruff Hewlett 1971). Such maps are useful for land management planning and for estima the quickflow response to precipitation on ungaged watersheds.

Hoover and Hursh (1943) noted that large differences in runoff occur betw watersheds in the Coweeta Basin. They attribute runoff differences primarily to dif ences in rainfall amount, soil depth (i.e., soil water storage and release characteristi and topography. Although Hursh and Brater (1941) describe the components of water flow that contribute to the storm hydrograph and Hursh and Hoover (1942) cussed the importance of water holding and transmitting characteristics of soil profi proof of the importance of unsaturated flow through soils on steep slopes as a sourc baseflow was obtained only after a series of soil model studies (Hewlett 1961; Hew and Hibbert 1963; Scholl and Hibbert 1973). Hydrologic models had been using Hortonian concept of effective rainfall, a fixed percentage of precipitation, as input, Hewlett (1961) and Hewlett and Hibbert (1966) postulated the now generally accep concept of a dynamic, variable sized area adjacent to stream channels and drain ways as being the source of quickflow during rainfall events (see Chapter 8).

The historical records of control watersheds at Coweeta allow examination of a c siderable range of hydrologic responses from relatively undisturbed forest syst within the 1625-ha basin. Some physiographic and hydrologic characteristics of control watersheds are presented in Table 3.3. The hydrograph data for each waters are averages of 1440 to 3555 separate storm events. No more detailed data base for ested watersheds exists anywhere in the world, either in length or in quality of reco

Hydrologic Response

Figure 3.11 shows the average hydrograph for four of the six watersheds presente Table 3.3 (the other two watersheds fall within the extremes set by these four). figure illustrates the large range in response to rainfall that occurs at Coweeta. W and 18 typify low elevation, deeper soiled and gentler sloped watersheds. WS 27 36 are the contrast—high elevation, steeply sloping watersheds with shallow soils high rainfall that yield a greater percentage of their precipitation as streamflow. L

Table 3.3. Some Physiographic and Mean Hydrologic Characteristics of Coweeta Conrol Watersheds

Characteristic		Watershed 2	14	18	34	27	36
(1) Area	ha	12.26	61.03	12.46	32.70	39.05	48.60
(2) Maximum elevation	m	1004	992	993	1184	1455	1542
(3) Minimum elevation	m	709	707	726	852	1061	1021
(4) Land slope	%	60	49	52	52	55	65
(5) Record length	yr	37	44	45	18	35	39
(6) Mean annual precipitation	cm	177.17	187.55	193.90	200.94	245.08	222.25
(7) Mean annual runoff	cm	85.39	98.81	103.42	117.47	173.74	167.51
(8) Precipitation-runoff	cm	91.78	88.74	90.48	83.47	71.34	54.74
(9) Hursh's runoff coefficient $= \frac{(7)}{(6)}$		0.482	0.527	0.536	0.585	0.709	0.754
(10) Initial flow rate	l/s/km²	24.52	29.46	29.54	36.57	36.30	40.81
(11) Quickflow before peak	cm	2.22	3.30	2.73	1.83	17.65	11.33
(12) Quickflow after peak	cm	5.93	6.62	7.00	3.74	34.15	25.86
(13) Delayed flow	cm	77.20	88.90	93.69	111.91	121.94	130.33
(14) Storm duration	hr	13.78	13.45	13.50	11.10	29.93	26.75
(15) Time to peak	hr	4.20	4.65	4.11	3.95	9.31	8.02
(16) Absolute peak	l/s/km²	68.78	81.12	79.97	75.65	242.60	154.05
(17) Recession time	hr	9.57	8.81	9.39	7.15	20.62	18.73
(18) Mean runoff events/year		66	80	79	80	68	68
(19) Response factor $= \frac{(11)+(12)}{(6)}$		0.046	0.053	0.050	0.028	0.211	0.167
(20) Hursh's storm runoff ratio $= \frac{(11)+(12)}{(7)}$		0.095	0.100	0.094	0.047	0.298	0.222

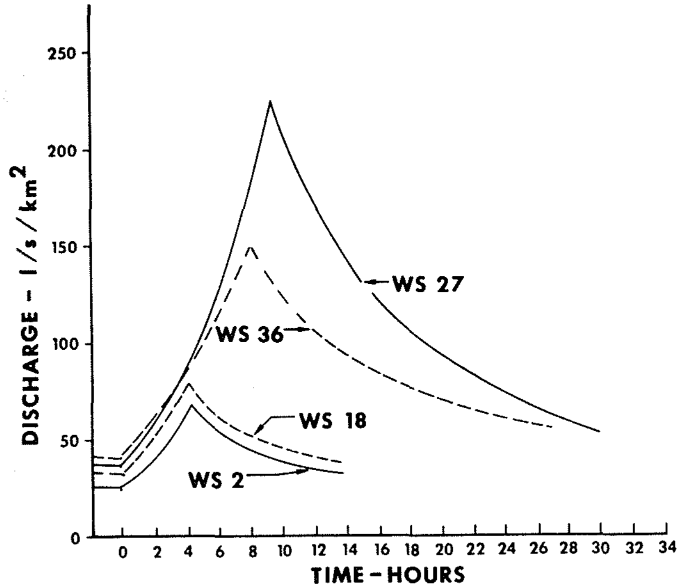

Figure 3.11. Average hydrograph for each of four control watersheds at Coweeta.

elevation watersheds release about half of their rainfall as runoff (Table 3.3, line 9), a 5% as quickflow (or storm runoff) (line 19). High-elevation watersheds with st slopes and shallow soils yield 75% of their rainfall as runoff with 20% of rainfall leav as quickflow. Thus, over a horizontal distance of 4.5 km and a vertical distance 915 m, average annual runoff response to precipitation (Hursh's runoff coefficie increases by a factor of 1.5. Over this same distance, the time span from the beginn of storm runoff to peak runoff rate doubles from about 4 to nearly 9 hr; storm durat doubles from about 14 to 28 hr and peak discharge increases by a factor of 2.6. Volu of quickflow, which is the primary factor influencing downstream flooding, increa about fivefold.

The large differences in response to rainfall shown in Table 3.3 and Figure 3 reflect physical differences between watersheds. Figure 3.1 shows that rain increases with elevation. Drilling to determine depth of regolith indicated a gen decrease in soil depth with elevation. Although weakly correlated, slope steepr increases with elevation; thus, the hydrologic head driving water through increases. Shallow soil at upper elevations provides a smaller cross-sectional a through which subsurface water moves. The combined result is that rate of discha and volume of storm runoff both increase with elevation. Hence, elevation represent the integrated effects of physiographic and hydrologic variables which de mine rates and volumes of discharge.

Hewlett (1967) showed that runoff response was a hydrologic characteristic wh could be mapped. His quickflow map (Figure 3.12) indicates a 10-fold increas potential flood waters between low and high elevations in the basin, an astounding 50 cm range.

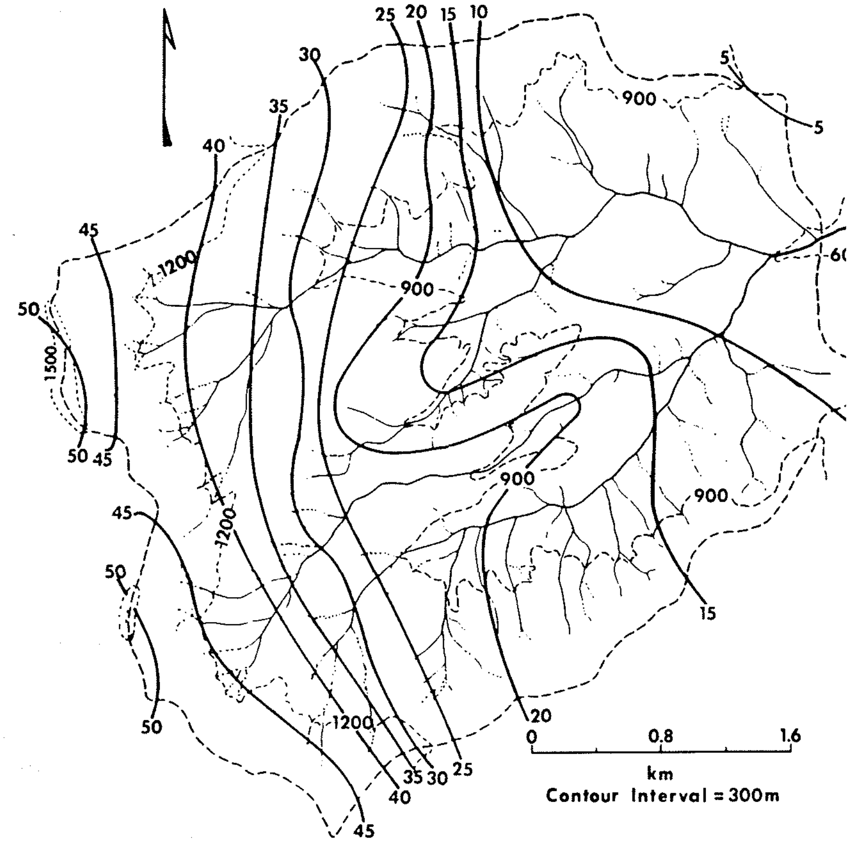

Figure 3.12. Runoff map of Coweeta Basin (Hewlett 1967). Units are cm of mean annual q
flow.

Peak Discharge

Although average response factor is a useful watershed characteristic, the ran; streamflow response is also needed to define the expected minimum and maxi storm events. The minimum stream response to a small storm may appear to be : but some minor response due to channel interception does occur. The maxi response is of more interest because it is related to maximum peaks and storm rı volumes. The maximum storm response for WS 2 occurred in a frontal storm on 27–29, 1976. This storm of 30 hr duration came 7 days after a 15-cm storm; thus moisture was recharged and the second rain produced approximately a 100-year rı event (Douglass 1974). Table 3.4 shows the quickflow reponse of 4 control waters to this storm. Twenty-six percent of the precipitation left each low-elevation water as quickflow compared with 50 to 54% of rainfall from the high-elevation waters

Table 3.4. Maximum Quickflow Responses of Four Control Watersheds to the May 2
1976, Frontal Storm at Coweeta

Watershed	Midslope Elevation (m)	Precipitation (cm)	Quickflow (cm)	Quick / Precipi
2	857	20.02	5.21	0.2
18	860	20.81	5.41	.2
27	1257	25.86	13.92	.5
36	1281	23.80	11.99	.5

Although precipitation was 3.4 to 5.4 cm greater on the upper watersheds, quicl was 6.6 to 8.6 cm greater. Again, upper elevation watersheds yield a greater perce of rainfall, but the range in response between watersheds for this large event is less the 10:1 ratio shown for average annual response.

Douglass (1974) used elevation as an integrator of climatic and physiographic fa to develop equations for predicting peak discharge from forested watersheds c Appalachian Mountains of North Carolina. These equations were derived from Coweeta's experimental watersheds and from larger, predominately forested w sheds (up to 13,400 ha in size) elsewhere in the mountains of North Carolina correlation indices (r^2) of equations for peak discharge versus watershed are: maximum watershed elevation all exceeded 0.98 for the 2.33- to 50-year recur intervals. Figure 3.13 illustrates a set of curves for the 20-year equation. While ε

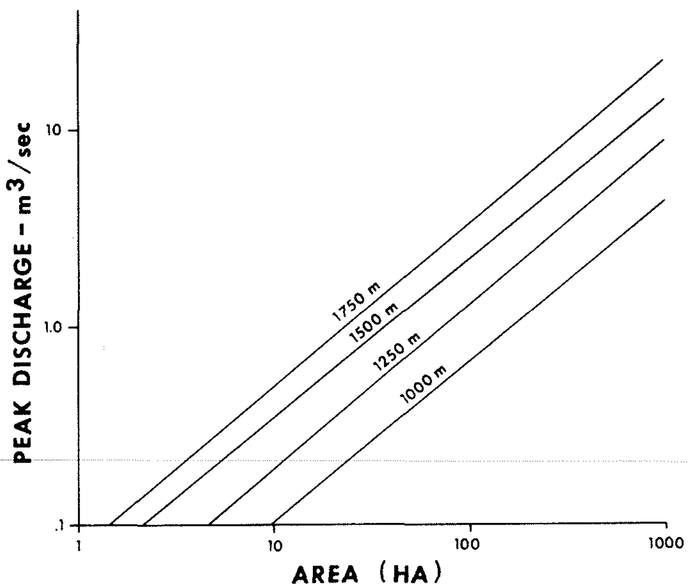

Figure 3.13. Peak discharge from forested watersheds in the Southern Appalachians f 20-year return interval as a function of area and maximum elevation of watershed.

tion is a convenient integrator of several effects, the individual contribution of vari such as slope, precipitation, and soil moisture storage have yet to be quantified.

Another flow characteristic of watersheds useful in engineering applications i distribution of flow. Flow frequency or the percent of time a given discharge ra equalled or exceeded is one measure of flow distribution. Figure 3.14 is the ave flow frequency distribution for the same four control watersheds at Coweeta. The tern for flow frequency follows the same general order shown by Figure 3.11 and T 3.3 and 3.4. The smallest rates of discharge for all exceedence percentages is obse on low elevation WS 2 and 18, which have less rainfall, gentle slopes, and deeper s

Annual Hydrograph

The annual hydrograph tracks monthly total streamflows. Figure 3.15 shows n mum, mean, and minimum monthly total flows for WS 2. The minimum trace s

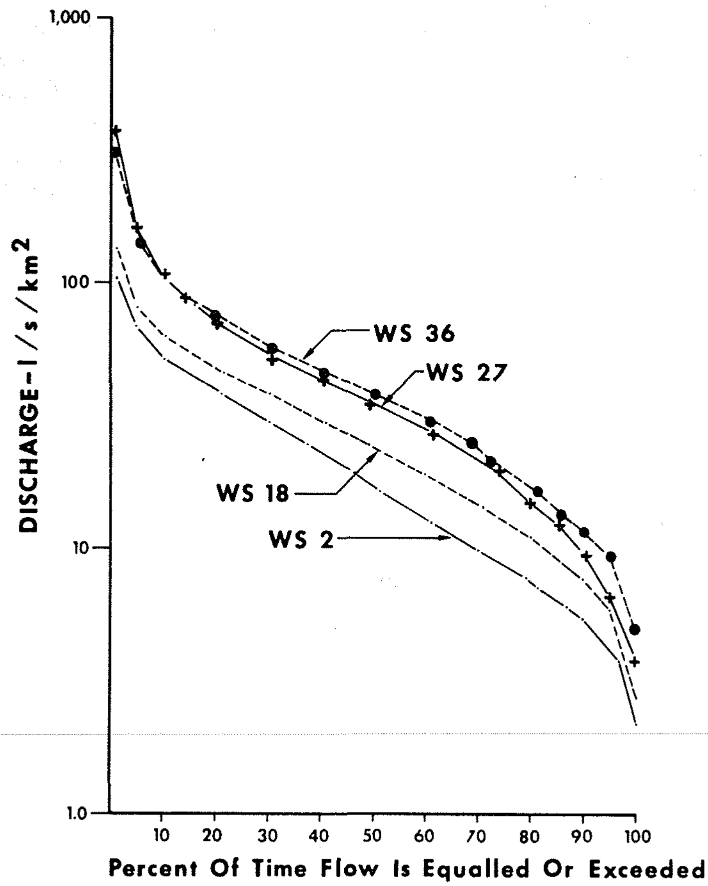

Figure 3.14. Flow frequency distribution for 4 control watersheds at Coweeta showing inc ing flow rates with increasing elevation.

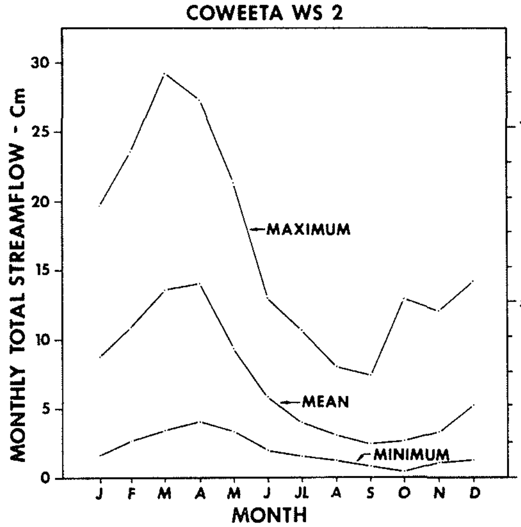

Figure 3.15. Mean monthly total streamflow from control WS 2 compared with the maximum and minimum monthly observed.

that in any month of the year, the total flow can be less than 4 cm, but flow does cease. Comparing the minimum and mean, low flows in late summer and fall have b half the mean flows for those months. The maximum monthly runoff for WS 2 rar from 7.5 cm to over 25 cm/month during the March to April recharge period, wh ends the dormant season. Rainfall for WS 2 during the peak flow month of April a ages 16 cm; thus, 87% of the rain falling on this watershed during April appear streamflow during that month. As a watershed becomes recharged, rainfall added to soil surface quickly displaces into the stream an equivalent amount of water fı storage in the soil profile.

The long-term P − RO value for each control watershed is given in Table 3.: varies from about 55 cm for more responsive high elevation watersheds, which ha shorter growing season and generally cooler and wetter climate, to 90 cm for low ele tion watersheds with their longer growing season. P − RO values for 39 years for W and 36 (Figure 3.10) illustrate the very large year-to-year variation, typical also other watersheds at Coweeta. One might expect that the year-to-year variation in RO for a given watershed could be related easily to amount and timing of precipitat However, fluctuations in successive years sometimes may be positively and other ti negatively correlated with precipitation while adjacent watersheds may not resp alike in the same year. Nevertheless, the P − RO value is a summary of the waters response to climate and hydrologic processes and with the length of record: Coweeta, detailed investigation of the causes for P − RO variation over time between watersheds could be a fruitful area of study.

Summary

Streamflow from an undisturbed forested watershed is the net result of the physio phy of the catchment and its climate. Various hydrologic response characteris correlate well with watershed elevation and it has been used as a predictor of stre

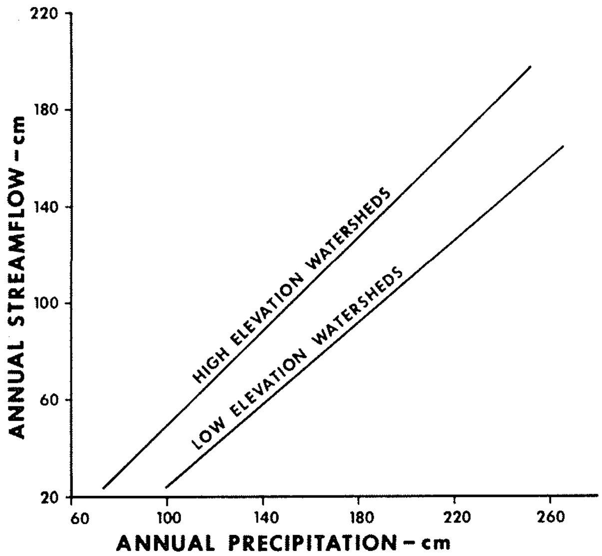

Figure 3.16. A streamflow versus precipitation relationship for forested watersheds ov range of elevations at Coweeta.

flow response. Elevation, however, is a surrogate variable representing several variables whose exact physical relationships to streamflow are incompletely defir this time.

Precipitation amount and timing have the greatest influence upon stream Precipitation increases with elevation along the east-west axis of the Basin, but i closely correlated with elevation along the slope for most of the north- and south-f watersheds.

At all elevations, precipitation is distributed fairly evenly throughout the year large individual storms occurring in nearly every month. Generally, monthly precipitation is less in April, late summer, and fall. Lagged streamflow response snowmelt is a minor factor, because heavy snows and long lasting snowpacks are The majority of rains have short durations and low intensities, but when large, intensity storms do occur, they occur with shorter recurrence intervals than pred from standard reference works.

On an annual basis, precipitation exceeds evapotranspiration demand and str flow yearround. Solar radiation is the primary source of energy for evapotranspira This energy input is influenced by land slope, aspect, topographic shading, and gr fog. Thus, radiation on various watersheds is only partly correlated with eleva However, solar energy is converted to heat and in this way influences tempera humidity, and wind, variables which do exhibit elevation effects. Because evapc spiration rates are sensitive to temperature, humidity, and wind, evapotranspir also can be correlated with elevation.

Soil depth decreases and slope steepness increases with elevation. Both fa reduce the ability of the watershed to retain precipitation and thus increase the pe tage that appears as streamflow. Upper elevation watersheds have lower precipitat runoff (P − RO) factors because they have less soil moisture storage capacity, ret a higher percentage of precipitation as quickflow, and have less evapotranspira demand to create soil moisture storage opportunity before a rain. Figure 3.16 propo a relationship between annual streamflow and precipitation for forested wateshed the high and low elevation extremes in the Coweeta Basin. Here, elevation a represents the net effect of several watershed physical factors, precipitation input, evapotranspiration demand. Other, fully forested deciduous forests in the eas United States, for which we have data, generally fall within the bounds of the two l in Figure 3.16.