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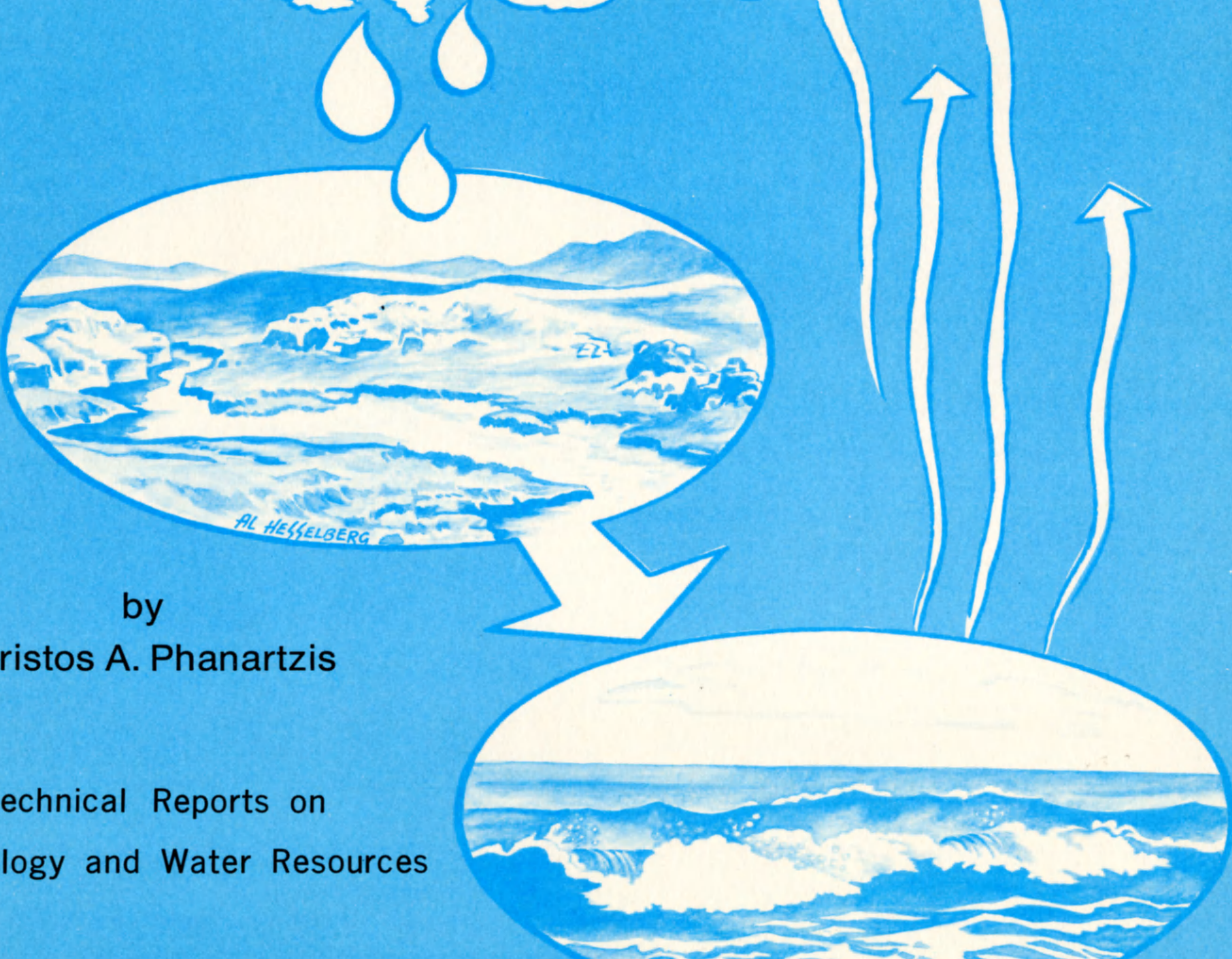
Spatial Variability of Precipitation in the San Dimas Experimental Forest and Its Effect on Simulated Streamflow



by
Christos A. Phanartzis

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Hydrology and Water Resources

The University of Arizona
Tucson, Arizona 85721

SPATIAL VARIABILITY OF PRECIPITATION IN THE SAN DIMAS EXPERIMENTAL FOREST AND ITS EFFECT ON SIMULATED STREAMFLOW

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Christos Apostolou Phanartzis

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PREFACE

This report constitutes the master's thesis of the same title completed by the author in June, 1972 and accepted by the Department of Hydrology and Water Resources.

This report series constitutes an effort to communicate to practitioners and researchers the complete research results, including economic foundations and detailed theoretical development that cannot be reproduced in professional journals. These reports are not intended to serve as a substitute for the review and referee process exerted by the scientific and professional community in these journals. The author, of course, is solely responsible for the validity of the statements contained herein. A complete list of currently-available reports may be found in the back of this report.

ACKNOWLEDGEMENTS

For the completion of this study I am deeply indebted to both Dr. Chester C. Kisiel and Dr. Hasan K. Qashu to whom I extend my hearty appreciation: to Dr. Kisiel for his encouraging guidance, interest and critical advice and to Dr. Qashu for his useful and practical suggestions and arrangements for securing the required data.

I wish to express my thanks and appreciation to the Pacific Southwest Forest and Range Experiment Stations, U.S. Department of Agriculture Forest Service, for allowing the use of their data, maps and publications. Special acknowledgement is extended to Dr. Raymond M. Rice and in general the personnel of the San Dimas Experimental Forest for their valuable assistance in making the above information available to me.

I further wish to express my gratitude and appreciation to the Food and Agriculture Organization (FAO) of the United Nations for providing me with a fully-financed scholarship and to the Government of Cyprus for granting me the required study leave. Last, but not least, I wish to extend this appreciation to their personnel and especially to Mr. McLaughlan, who handled my scholarship matters with great patience.

All computer programs were executed on the CDC 6400 computer installed at the Computer Center of The University of Arizona.

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Hydrology and Water Resources," sponsored by the Office of Water Resources Research, U.S. Department of the Interior.

TABLE OF CONTENTS

TABLE OF CONTENTS--Continued

LIST OF TABLES

LIST OF TABLES--Continued

LIST OF ILLUSTRATIONS

LIST OF ILLUSTRATIONS--Continued

ABSTRACT

The effect of altitude on individual storm precipitation in some of the San Dimas experimental watersheds is investigated. It is found that there is a well-defined increase of storm precipitation with altitude for storms greater than one inch. This increase is a linear function of storm depth.

Using 41 storms of different magnitudes, a precipitation-altitude relationship is derived for a small area in the San Dimas Experimental Forest. The regionalization of this relationship and its transferability are tested by analyzing differences (errors) between computed and observed storm precipitation values in each case. In testing the regionalization of the precipitation-altitude relationship by computing <u>mean areal</u> storm precipitation over a larger area the standard error of estimate is around 11 percent. In transfering the same relationship the results are not as good and give a standard error of 16 percent. For individual points, however, the error is much higher. A rainfall-runoff model is used as a tool for evaluating the effect of precipitation errors, on simulated streamflow, in a watershed of 4.5 square miles. For annual flows, errors range between 3.4 and 12.8 percent while errors in simulated monthly flows are as high as 22 percent. It is also evident that there is a strong dependence of the error magnitude on the state (wet, dry, etc.) of the preceding year or months, whichever is applicable. An error propagation is observed as a result of consistently over-estimating the

precipitation input to the model. This evaluation is more of a qualitative nature and the values of error given should be viewed in this sense.

CHAPTER 1

INTRODUCTION

The accelerated development of hydrologic models, largely as a result of the introduction of high-speed electronic computers, has been emphasized by several authors engaged in hydrologic research. Hydrologic models, and specifically rainfall-runoff models are progressively becoming more sophisticated. In general, the more sophisticated models require more data, of higher accuracy, and the results from such models, whether the models are stochastic or deterministic, depend on the amount and accuracy of the available data.

The applicability of sophisticated rainfall-runoff models to real-world hydrologic problems has not yet been firmly established. Dawdy, Lichty and Bergmann (1970), commenting on rainfall-runoff simulation models have proposed the following three applications, one or more of which must be achieved if modelling is to be considered practically useful (p. 23):

1. A rainfall record can be used to add to the information content of a streamflow record which has a shorter period of record than the rainfall record.

2. Model parameters for ungaged sites can be estimated on the basis of those derived for gaged sites, and information can be gained at the ungaged sites through the use of recorded or simulated rainfall and estimated parameters at the ungaged sites.

3. The effect of man-made changes on a basin can be related to changes in model parameters so that measured "before" conditions can be compared with simulated "after" conditions of sufficient accuracy for planning purposes.

The same authors proceed to emphasize that in the process of choosing or building a model the following criteria must be met. The model must a) require only input data which are generally available, b) be simple enough to be easily understood and operated by the user and c) provide the output desired at an acceptable level of accuracy for the application for which the model is used. Additional constraints of equal or lesser importance are the availability of funds and the length of the time allowed for attaining the goal for which the model will be used. Closely associated with these and the biased tendency to use existing models familiar to the user, is criterion (b) which is very important in the case of time and fund limitations. On the other hand, criterion (a) immediately eliminates the use of sophisticated models, since in most cases the available data is not sufficient to adequately describe the hydrological processes and their interactions; lack of sufficient data leads to empiricism and lumping, both spatial and temporal, which in turn has an impact on the level of attainable accuracy. However the purpose of this paper is not to go into the model choice question. The paper by Kisiel and Duckstein, "Economics of hydrologic modelling" (1972), is a good presentation of a rational approach to this problem.

The shortage of hydrologic data in areas under development is a problem commonly encountered in the assessment of water resources and the determination of their distribution and variability in time and space; knowledge of the latter is a very important factor in the planning and economic appraisal of water development projects. In such cases the transferring of information to ungaged or partially gaged watersheds becomes desirable. For some watersheds only a short record of streamflow may be available, for others only precipitation records, and in the worst case, watersheds may lack both streamflow and precipitation records.

In sampling precipitation, its variability in space and time as measured by the density of raingage networks is of varying importance depending on the hydrologic variable of interest. This is also pointed out by Eagleson (1967) where three such variables are listed: precipitation, which is the input signal itself, precipitation averages and streamflow. Not only do these three variables necessitate different sampling frequencies, but for each variable tolerable sampling intervals may differ depending on the nature of the system and the accuracy required.

The present study deals with the investigation of the spatial variability of precipitation in the San Dimas Experimental Forest in California, with special emphasis on the effect of altitude on precipitation amounts. A precipitation-altitude relationship is derived using storm data in a small watershed, and the transferability of this relationship is tested by applying it to a neighboring watershed. The effect of precipitation errors, arising by the transfer of the precipitation-altitude relation, on simulated streamflow is further investigated by using a simple rainfall-runoff model of daily events, a full description of which is given in Appendix 1. Monthly volumes of streamflow are considered to be the variable of interest in evaluating the effectiveness of the precipitation relationship when transferred to another watershed.

Apart from the introduction, the study, as presented in this thesis, is composed of four parts. In the first part, Chapter 2, a description of the San Dimas Experimental Forest is given with special reference to Watershed 5 which is the watershed used in the section of the thesis on streamflow simulation. The second part of the thesis deals with precipitation in the same area and gives an account of the previous work done by other investigators followed by a full description of the work and findings of this investigator.

The third part, Chapter 3, describes the application of a rainfall-runoff model to study the effect of errors in precipitation input on simulated streamflow volumes. Two sets of precipitation inputs are used, covering the same period of time. The first set which is considered to represent the actual precipitation in Watershed 5 consists of 24-hour precipitation values at raingage I35 in the same watershed (see Appendix 2 for justification of this). The second set consists of 24-hour precipitation values, over the same period, computed by using the corresponding values at raingage I35 and the precipitation-elevation relationship derived in Watersheds 2 and 4 (Equation [4]). Following calibration of the rainfall-runoff model, the simulated streamflow from the first set of precipitation input is considered the reference for evaluating errors in simulated streamflow when the second set of precipitation input is used. Figure 1 is a schematic illustration of the two sets of Input-System-Output with some explanatory notes. In other words, a deterministic precipitation model is used to compute values of precipitation, which in turn, form the input into a deterministic watershed model.

Input Set 1

24-hour precipitation at station I35 (considered as actual on Watershed 5)

Output Set 1

Simulated streamflow (considered the reference streamflow for comparison purposes)

SYSTEM

Watershed 5 described by rainfall-runoff model. Model is calibrated by using rainfall at raingage I35.

Input Set 2

Computed 24-hour precipitation from data at I35 and Equation [4] (errors due to transferring)

Output Set 2

Simulated streamflow to be compared with Output Set 1 (contains errors due to errors in Input Set 2)

Figure 1. Schematic representation of watershed system and the two sets of input with their corresponding output.

The fourth part, Chapter 5, is a summary of the conclusions arrived at with respect to errors in precipitation and their effect on the simulated streamflows. In some cases a short discussion is also presented.

CHAPTER 2

THE SAN DIMAS EXPERIMENTAL FOREST

General Information

The San Dimas Experimental Forest was established in 1933 to serve as a center for watershed research, at which time a research program for the accurate determination of precipitation on mountain watersheds was initiated. The latter was part of a program on hydrologic research with the objective of developing principles and practices for the management of chaparral-covered watersheds. This program was started by the California Forest and Range Experiment Station of the U. S. Forest Service.

The area in question is in the San Gabriel mountain mass in Southern California and has an area of approximately 26.5 square miles. It was originally divided into ten "intermediate"-size watersheds of varying sizes (Figure 2), of which Watersheds 2, 4 and 5 are referred to in this investigation, the first two associated mainly with the rainfall part while number 5 was used in the rainfall-runoff part of the study.

Geomorphology

The San Dimas Experimental Forest is, for the most part, very rugged and is characterized by very steep slopes, at places gradients exceed 100 percent. Elevations (Figure 2) range between 1500 feet and 5400 feet, the higher elevations being found in the eastern part of the area.

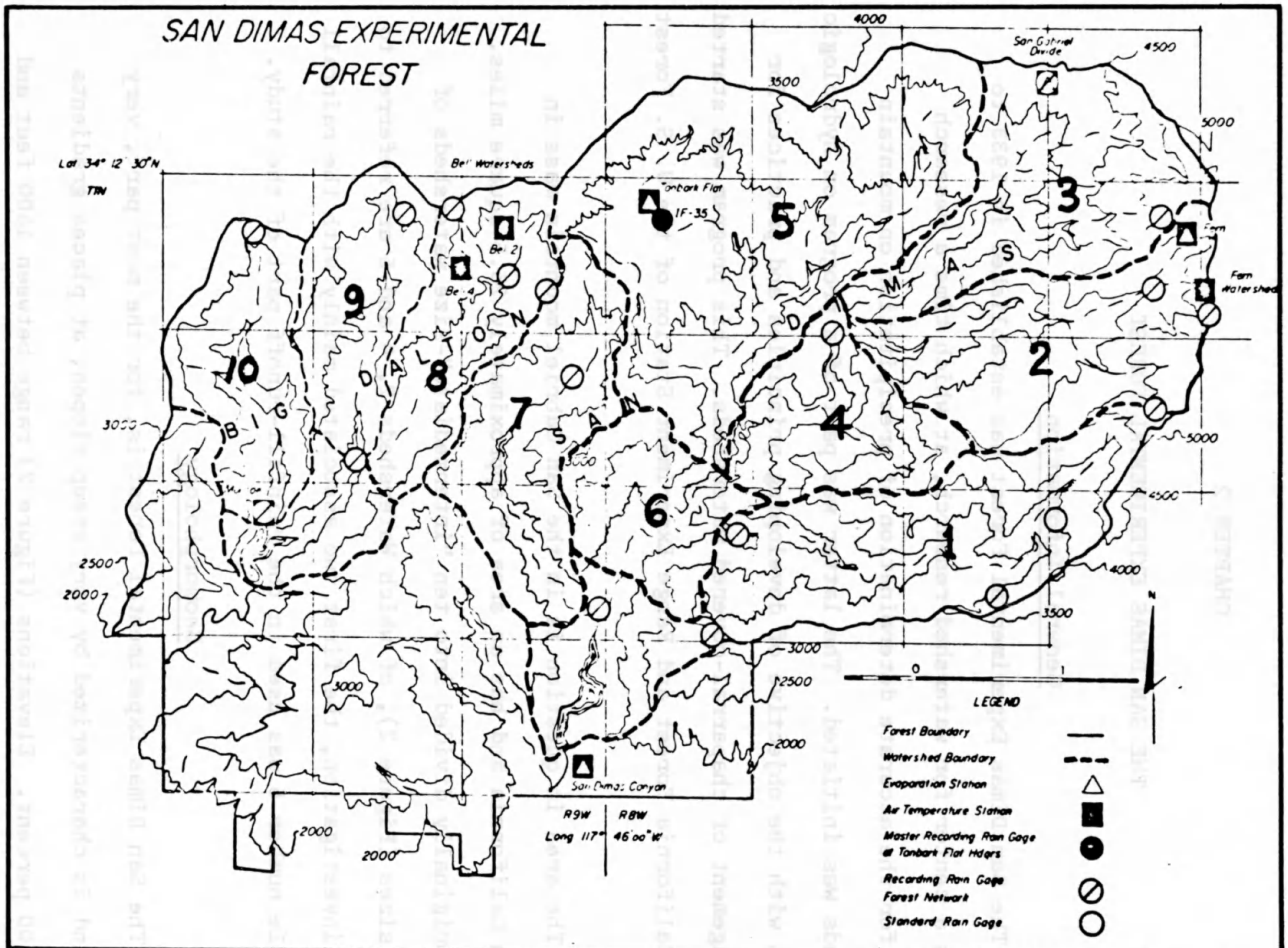


Figure 2. San Dimas Experimental Forest in the San Gabriel mountains of southern California showing watersheds and elevation contours.

Watersheds 2, 3, 4 and 5 which are used in the present study are dominated by steep slopes (Figure 3), and gradients range from 55 to 70 percent. Areas with moderate (0 to 55 percent), and extremely steep (greater than 70 percent) gradients are also present especially in Watershed 5, but these are of lesser extent.

Geologically the area consists of faulted igneous and metamorphic rocks which are also extensively, and at places, deeply fractured. On the eastern part the dominant rock is Gneiss, while Watershed 5 in the central part of the area consists of diorite and metadiorite with some Gneiss and Schist and several dacite dikes.

Soils

The soil cover is made up of generally medium-to-coarse textured, well-drained, sandy loam which overlies the parent rock. Soil depths range from very shallow (less than 1 foot deep) to deep (over 2 feet deep), but the "very shallow" and "shallow" are much more dominant (Figure 4) and cover 74 percent of the area; shallow soils are, in general, associated with very steep slopes. In Watershed 5 the soil depth is mostly less than 2 feet, except in the Tanbark flats area and the northeastern corner where medium and deep soils may be found.

Vegetation

The vegetation prevailing prior to the 1960 fire was a dense growth of chaparral and woodland chaparral. The importance of this vegetation in the disposition of rainfall, especially with respect to amounts of moisture in the soil horizon, was recognized very early and appropriate research was set up for its study. Relevant studies have

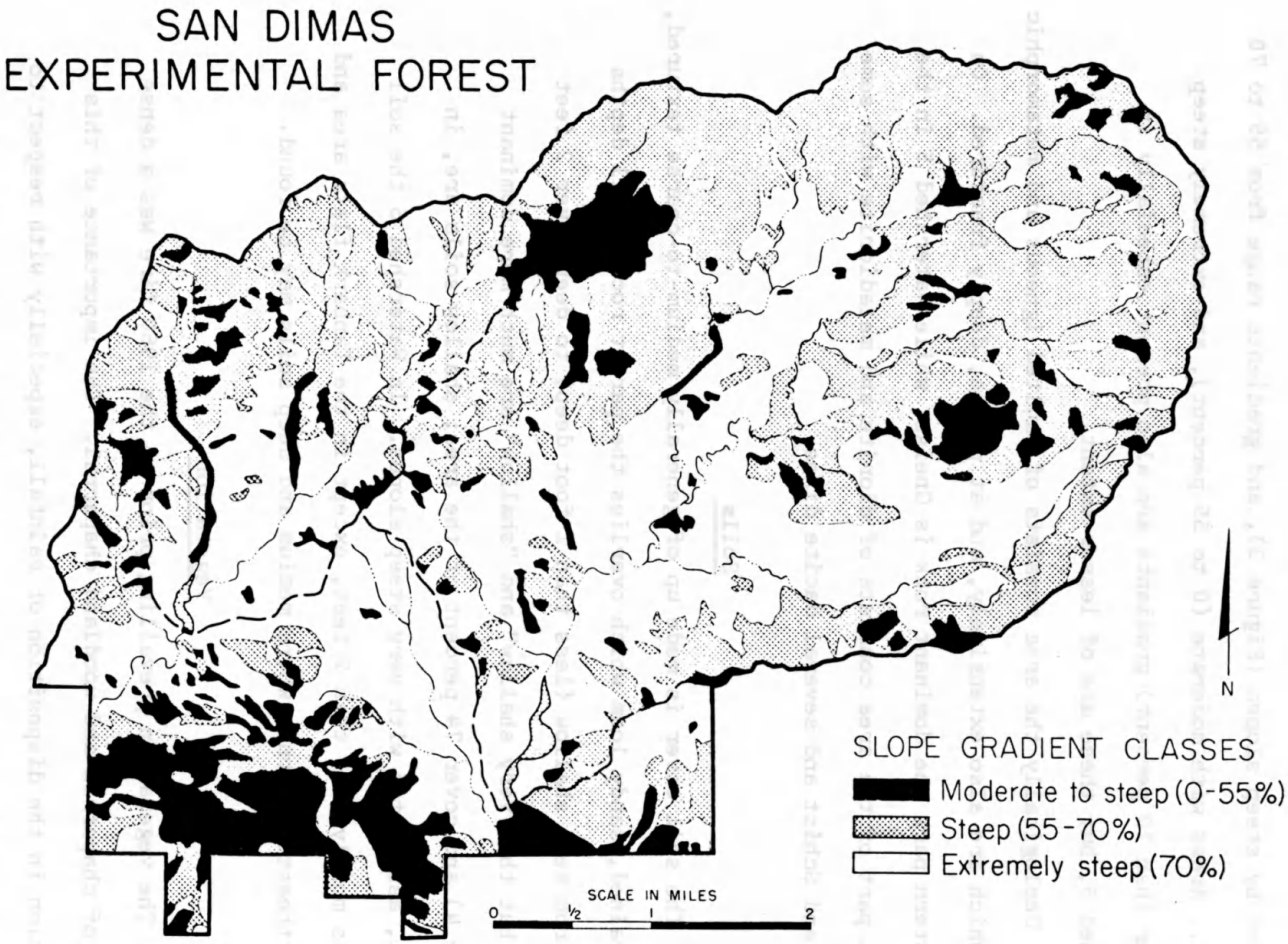


Figure 3. Slope gradient classes on the San Dimas Experimental Forest.

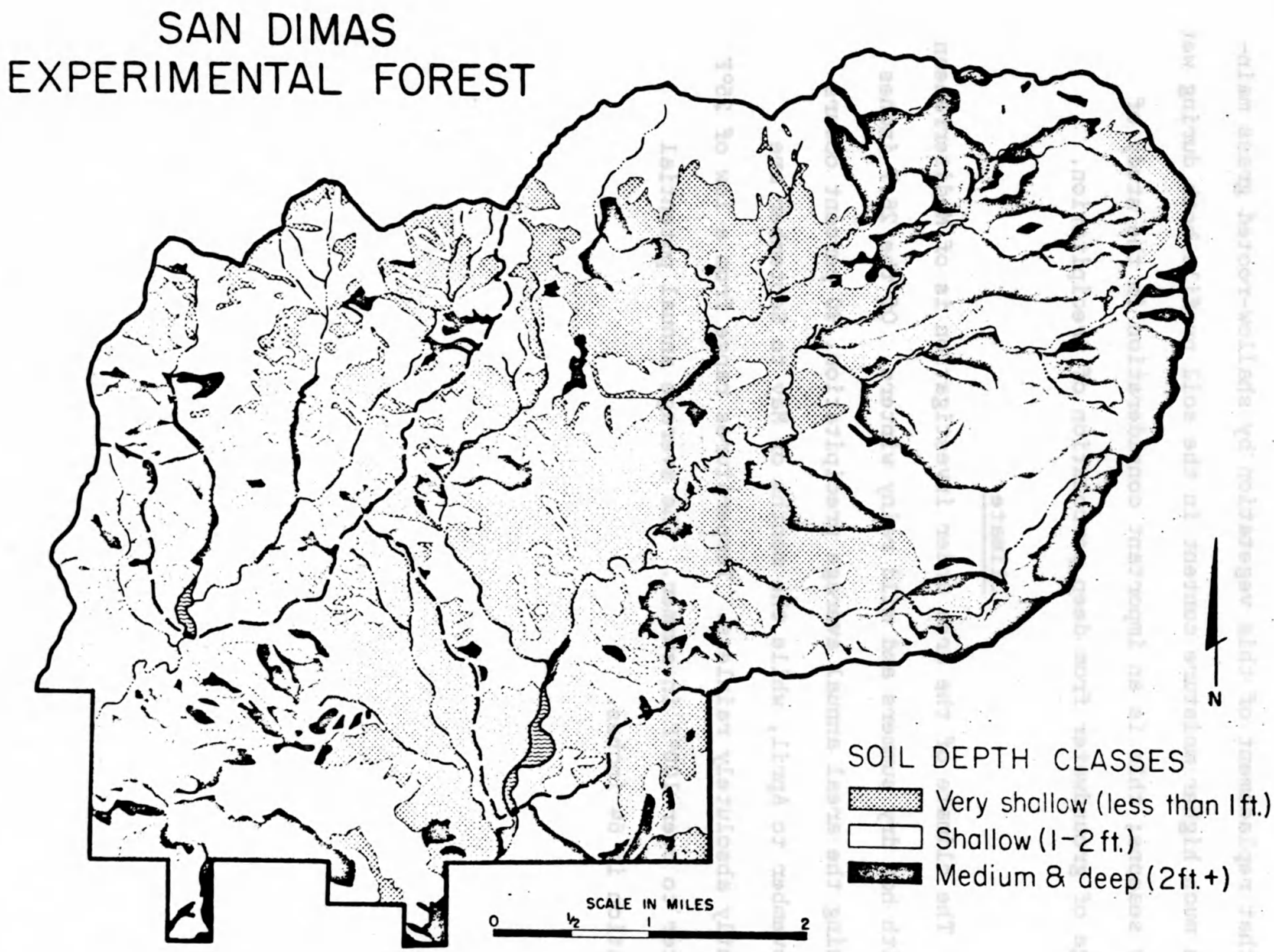


Figure 4. Soil depth classes on the San Dimas Experimental Forest.

shown that replacement of this vegetation by shallow-rooted grass maintains a much higher moisture content in the soil profile both during wet and dry seasons; this is an important consideration in the study of recharge of groundwater from deep percolation of precipitation.

Climate

The climate of the area under investigation is of Mediterranean type with hot dry summers and mild rainy winters. Of the 26.7 inches comprising the areal annual average precipitation, 90 percent occurs from November to April, while the months of May to September are frequently absolutely rainless. Temperatures range from a low of 25°F in winter to over 100°F in summer. The average annual potential evaporation is 64 inches.

CHAPTER 3

PRECIPITATION

Precipitation Characteristics

Precipitation in California is dominantly a result of cyclonic storms originating in the North Pacific (Storey, 1939). Storms in the San Dimas Experimental Forest were classified as northerly or southerly, depending on the direction from which the rain comes. Northerly storms are accompanied by low velocity winds and produce low amounts of precipitation at small inclinations from the vertical. On the other hand, southerly storms are characterized by higher intensities, higher wind velocities, and the angle at which precipitation falls as compared with the vertical, is considerably higher. Southerly storms account for more than 60 percent of the total long-term precipitation. Precipitation in the form of snow is an infrequent occurrence and when it happens it affects mostly the higher elevations of the San Dimas watersheds. Prior to the present study several other studies were made of precipitation measurement and behavior in this area (Wilm, Nelson and Storey, 1937; Storey, 1939; Storey and Hamilton, 1943; Storey and Wilm, 1944; Hamilton, 1954; and Hamilton and Reimann, 1958).

Definition of "Storm"

In order to maintain consistency in compiling and publishing of precipitation data, the Forest Service has adopted the following definition of a storm, as it appears in publications (Reimann and Hamilton, 1959, p. 17) of the precipitation data from the San Dimas Experimental Forest:

> A meteorological disturbance producing at least 0.10 inch of precipitation in the master raingage number I35 at Tanbark Flat field headquarters. A period of 24 hours without rainfall and with clearing conditions marks the end of a storm. A new storm, however, may be designated within 24 hours of the cesation of rainfall if attendant meteorological conditions warrant it.

This definition of a storm implies that the number of storms during a specified period is the same for all points and all elevations in the area, and prohibits an approach to modeling mountainous precipitation similar to that mentioned in the paper by Davis, Duckstein and Kisiel (1972), where the number of storms increases with altitude.

History of Raingage Network and Previous Studies

The raingage network has gone through several dramatic changes since 1933, reaching a maximum number of 318 gages which were eventually reduced to 21 intensity or recording raingages.

The original network of 318 gages provided for the plain distribution of these gages at 1/2-mile intervals on contour trails at 2100, 3100, 4100 and 5100 feet, and on all other available access trails and roads. These gages were placed vertically according to the U.S. Weather Bureau specifications. It was then hoped that by placing the gages in clearings within the chaparral cover the sheltering requirements would be satisfied.

First Study

The first study (Wilm et al., 1937) was an analysis of two years of precipitation data collected from the original raingage network, and the purpose was to determine whether the spatial distribution of samples was statistically adequate to give a "precise" measure of rainfall depth by watersheds.

For the two small watersheds of "Bell" and "Fern" the gage distribution was much denser (Figure 5) and it was concluded that the raingage network was adequate for sampling of rainfall. This conclusion was reached after showing that the average catches, for each of ten storms, as computed from isohyets, agreed closely to the arithmetic means for the same storms.

The same criterion was also used in deciding that sampling of precipitation in the larger watersheds, where gage density was not as high, was adequate. In addition, the standard error (SE) was used as a criterion of the variability of the average rainfall depths. Concerning the number (N) of gages required to provide a pre-determined degree of variability (SE), the authors suggested the following relation:

$$N = \left(\frac{SD}{SE}\right)^2 \qquad [1]$$

where N is the number of raingages in a watershed required to attain a specified reliability, SE is the standard error assumed to describe this reliability and is taken as a fraction, say 0.05, of the mean catch for a storm, and SD is the standard deviation of the same storm as sampled by the number of existing gages in the same watershed. Equation [3] implies that each set of raingage data is independent of sets of data at all other points in space. This is not so in this case due to the large areal extent of the storms in this area.

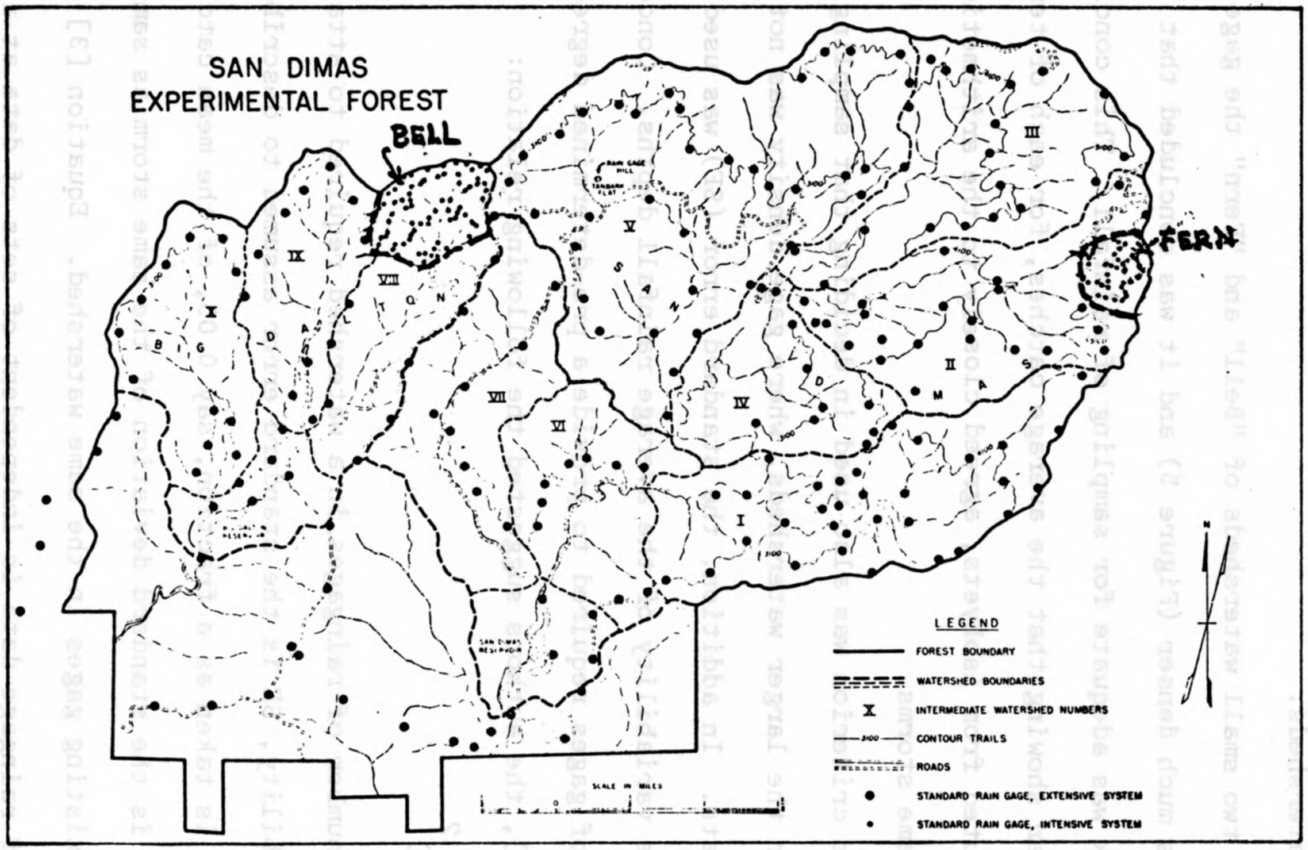


Figure 5. Original network of 318 raingages.

Since nothing was said about the spatial distribution of the N raingages, it is assumed that the latter are distributed as in the original network.

Upon examining rainfall records for 54 years at the nearby station of Glendora, the authors observed that small storms, which exhibit the greatest spatial variation, accounted only for a small part of the total precipitation. This led to their decision to weight the effect of the storms and the adoption of the following arbitrary requirement to describe the accuracy of storm sampling:

Storm Magnitude	SE/Storm Mean
Less than 0.5 inch	10.0 percent
Between 0.5 and 1.0 inch	5.0 percent
Greater than 1.0 inch	2.5 percent

Toward the end of 1935 the raingage distribution in the first eight out of ten intermediate watersheds was extended to comply with the set requirements as listed above. An analysis of twelve subsequent storms ranging in magnitude from about one-fourth inch to over ten inches was made and the standard errors (SE) were found to be well within the specified limits for most storms. Storms of small magnitudes gave higher values of SE.

The study in question led to the following conclusions, as listed by the authors:

1. The gage system used in the study gave accurate results for most storms measured, within practical limits.

2. In order to avoid excessively large numbers of gages, the accuracy requirements for averages should be modified in inverse relation to the size and importance of storms.
3. With a system of gages designed to sample rainfall variation as thoroughly as possible, a simple average of their readings will agree within close limits with rainfall catch as computed by isohyets.

The conclusions reached by the authors are not convincing enough such that the network studied is also the "minimum" gage network for the required accuracy of storm measurements. For example, conclusion (3) does not exclude the possibility that a less dense network, properly designed, will give equally good estimates of watershed precipitation.

Second Study

Storey (1939), made a study on the influence of topography on precipitation, and he claimed that this influence is more than simply an increase in elevation. He presented isohyetal maps of the "Bell" watersheds, of about 250 acres in area and equipped with 57 raingages, for two storms. In both cases, precipitation was shown to be greater in the valleys and lower on the ridges. He then concluded that the spatial variation of precipitation on rugged terrain is very high, even in small areas as the one studied. He attributed this to the wind and stated that this kind of distribution is caused by the wind carrying rain past the ridges and depositing it on the lee side and in sheltered canyons.

However, investigations during more recent studies rendered these findings meaningless.

Additional Studies

It was eventually recognized that although the analysis presented in the first study indicated that the rainfall variation was sampled reasonably well, it gave no evidence that the samples themselves were accurate. In other words, it was not certain that the spatial variation which was studied and on the basis of which the raingage network was improved really existed and did not arise from sampling and measurement errors. Divergences of rain catch had until then been attributed to topographic effects. Extreme variations however initiated further investigations of the behavior of rainfall on rugged terrain.

Following a review of literature on the effect of wind speed and direction on rainfall depths as measured by gages on steeply sloping terrain, three types of experiments were designed and carried out for the purpose of:

1. Determining how physical factors like wind speed and direction might influence rain catches.
2. Comparing rainfall samples from a number of gages with different exposures and placement, with actual rainfall depths as measured on the ground.
3. Comparing rainfall depth between paired raingages, one tilted and one vertical, for 22 locations in a 100-acre watershed.

Two papers (Storey and Hamilton, 1943; and Wilm, 1944) were written on the results of the above experiments. Hamilton (1954), in a fifth paper, summarized these results as follows:

1. A vertical raingage did not give an accurate sample of the rainfall actually reaching the ground on an adjacent control catchment surface, whereas a gage tilted and oriented normal to the slope and aspect of the control surface gave a very good sample.
2. The distribution of 19 vertical raingages in a 100-acre watershed caught 15 percent less rain than did a parallel network of 19 gages tilted normal to the slope. The inadequacy of the vertical gage was related to the speed of wind which causes rain to be greatly inclined as it falls.

It was also determined from the experiments that the incidence of rain on the experimental watersheds was most of the time southerly and generally at a considerable inclination from the vertical.

Hamilton (1954), examined the data collected during the experiments and noticed that storms of less than one inch in depth were generally coming from the north, east or west and that storms greater than one inch were almost exclusively coming from the south. By analyzing 30 southerly storms he found that the tilted gages showed an average catch 16.5 percent higher than the vertical gages. For four years of observations, it was found that the annual excess of tilted versus vertical gage catches ranged from 12 to 21 percent. Individual storms, however, were as much as 25% higher when measured with tilted as compared to vertical gages.

The study by Hamilton (1954) shows that although, according to the first study, the original 318 raingage network was found statistically adequate as to number and distribution of raingages (Wilm et al., 1937), it failed to provide a correct measure of rainfall depth in some places. For the correction of old rainfall data from the vertical gages, Foucarde's equation (see Equation [2]) was proposed for storms greater than one inch. The use of this equation necessitates knowledge of the prevailing wind speed and direction.

After applying such a correction to the vertical gage catches which described the storm of April 7, 1935, a profile of which was presented by Storey (1939), an entirely different picture evolved, and the extreme variability of rainfall with space, as concluded upon by same, disappeared. The original cross-section showing rainfall depth as recorded by the vertical gages and the same after correction, are shown in Figure 6. It became evident from this plot, that rainfall variation, in space, in the San Dimas Experimental Forest watershed is less than it was originally thought.

Following the above findings, all storms greater than one inch and sampled by vertical gages were corrected by using Foucarde's equation,

$$r = R + R\tan\alpha \tan i \cos(\beta - \omega) \qquad [2]$$

where r = true rainfall

R = rainfall sample from vertical gage

α = gradient of slope being sampled

β = aspect of the slope being sampled

ω = average storm direction

i = angle of inclination of rain

} average values

For storms less than one inch, the equation provided no improvement.

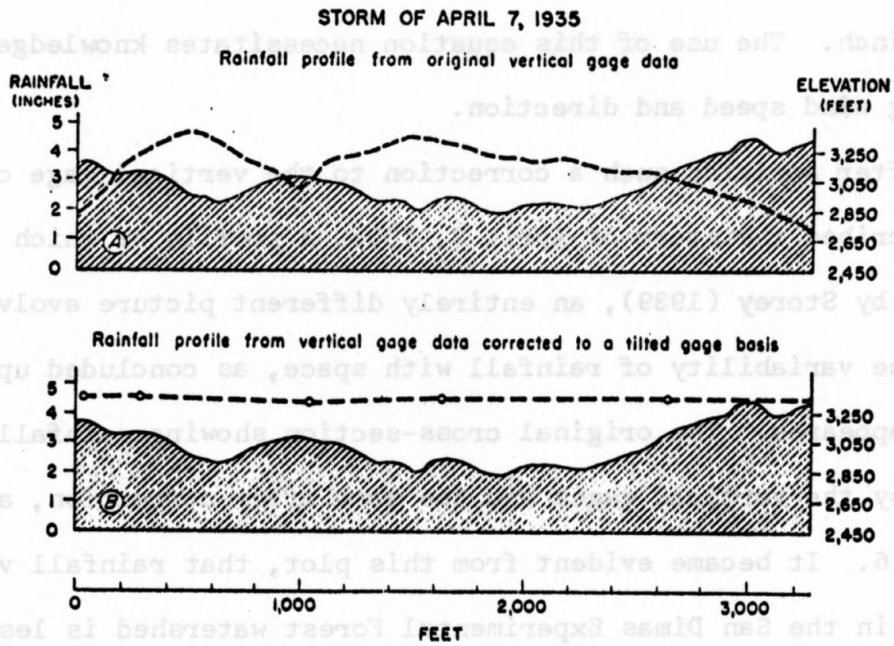


Figure 6. Original and corrected profiles for the storm of April 7, 1935.

With respect to ω and i, not only do their values vary during the same storm, but they are also expected to vary spatially, especially in rugged areas like the San Dimas Experimental Forest.

Improving the Rainfall Sampling Network

Based on the findings of the studies previously mentioned, a new raingage network was designed (Hamilton, 1954), employing only 72 raingages, all of them tilted according to local slopes and located in facets representative of areal topography. The new distribution of gages was by elevation and followed closely the proportion of total area in each altitudinal zone as indicated in Table 1.

The new network was opened together with the old one for three years and the results indicated that average watershed precipitation was higher when measured by the new network of tilted gages. Furthermore, isohyetal maps based on the new network were found to be less complicated and made more sense with respect to altitude and topography. Figure 7 shows the facets and new raingage network.

Further Simplification of the Raingage Network

In 1958 the San Dimas raingage network was further studied by Hamilton and Reimann in an attempt to show that average watershed rainfall could be sampled by a smaller number of raingages and the findings were published (Hamilton and Reimann, 1958). More specifically, this work was concerned with the study of a new "minimum" raingage network consisting of 21 raingages. For more than two-thirds of the storms greater than one inch, the new network gave values of average watershed rainfall within 5 percent or less of the corresponding values determined

Table 1. Improved distribution of raingages by altitudinal zones.

Altitudinal zone (feet)	Area (percent)	Raingages (percent of total)
1500-2500	27	28
2500-3500	44	45
3500-4500	19	18
4500-5500	10	9

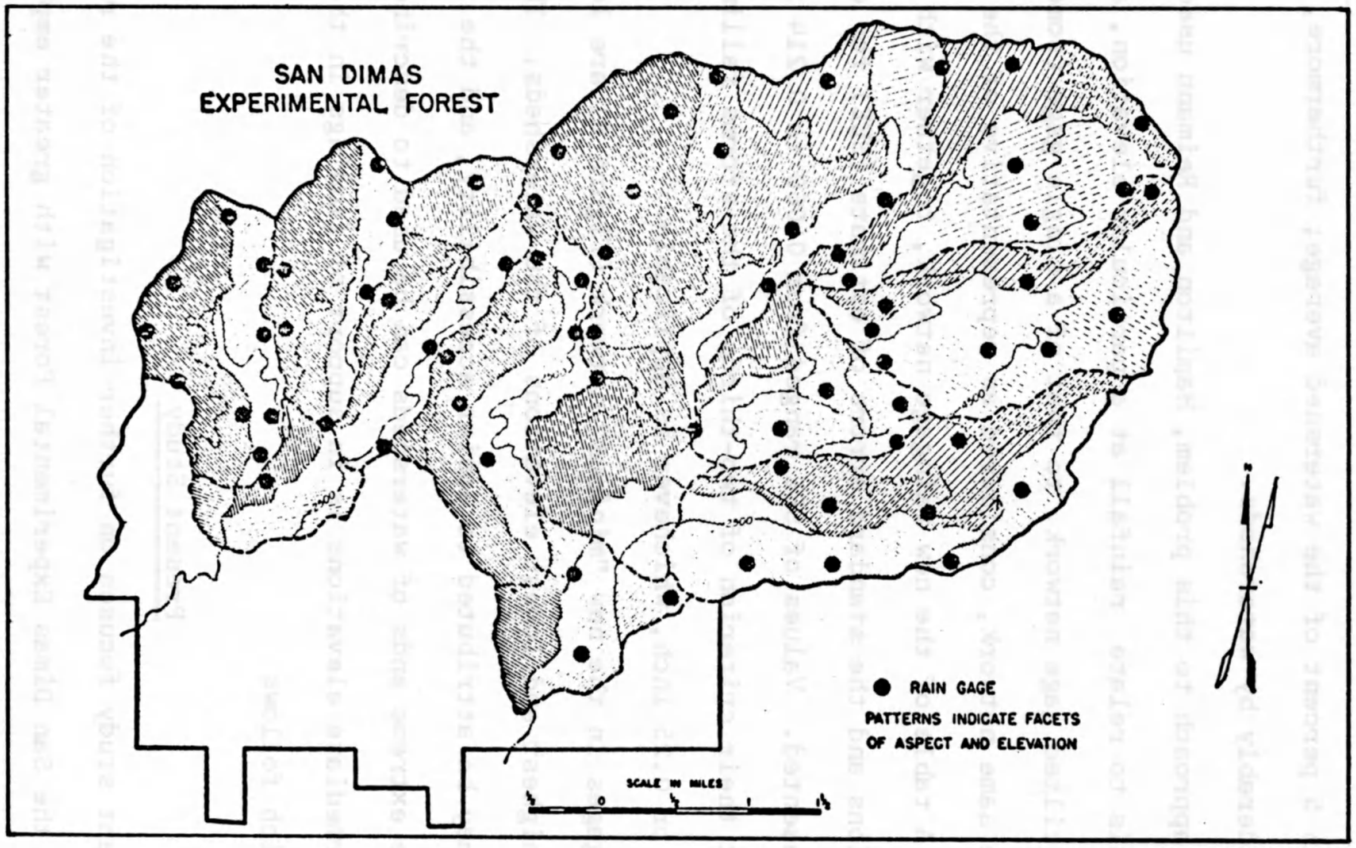


Figure 7. Tilted raingage network established according to facets of uniform slope and aspect.--Raingages are distributed by altitude within intermediate watersheds.

from the tilted raingage network. For small storms (below one inch), however, observational errors became important, and only half of the storms were within 5 percent of the watershed average; furthermore, they also varied considerably by watersheds.

In their approach to this problem, Hamilton and Reimann used regression analysis to relate rainfall at a particular elevation, as estimated by the tilted gage network, to that of a single gage (sometimes two), part of the same network, considered as representative of the elevation zone. A table of the new raingage network, together with the regression equations and the standard errors of estimate (SEE) for each elevation was presented. Values of SEE ranged from 0.060 to 0.214 inches and more than met their criterion of two-thirds of the storms falling within 5 percent or 0.25 inch, whichever is applicable.

The raingages in the new "minimum" network (Figure 8) are located mostly near the highest and lowest elevations of the watersheds. The reason for this may be attributed to easier accessibility, and the fact that gages at the extreme ends of watersheds can be used to describe rainfall at intermediate elevations is in support of findings in the present study which follows.

Present Study

The present study focuses on further investigation of the rainfall behavior in the San Dimas Experimental Forest with greater emphasis on the relation of altitude to precipitation amounts and the use of this relation to further reduce the raingage network. It has been shown by several researchers that the reduction of the number of observation

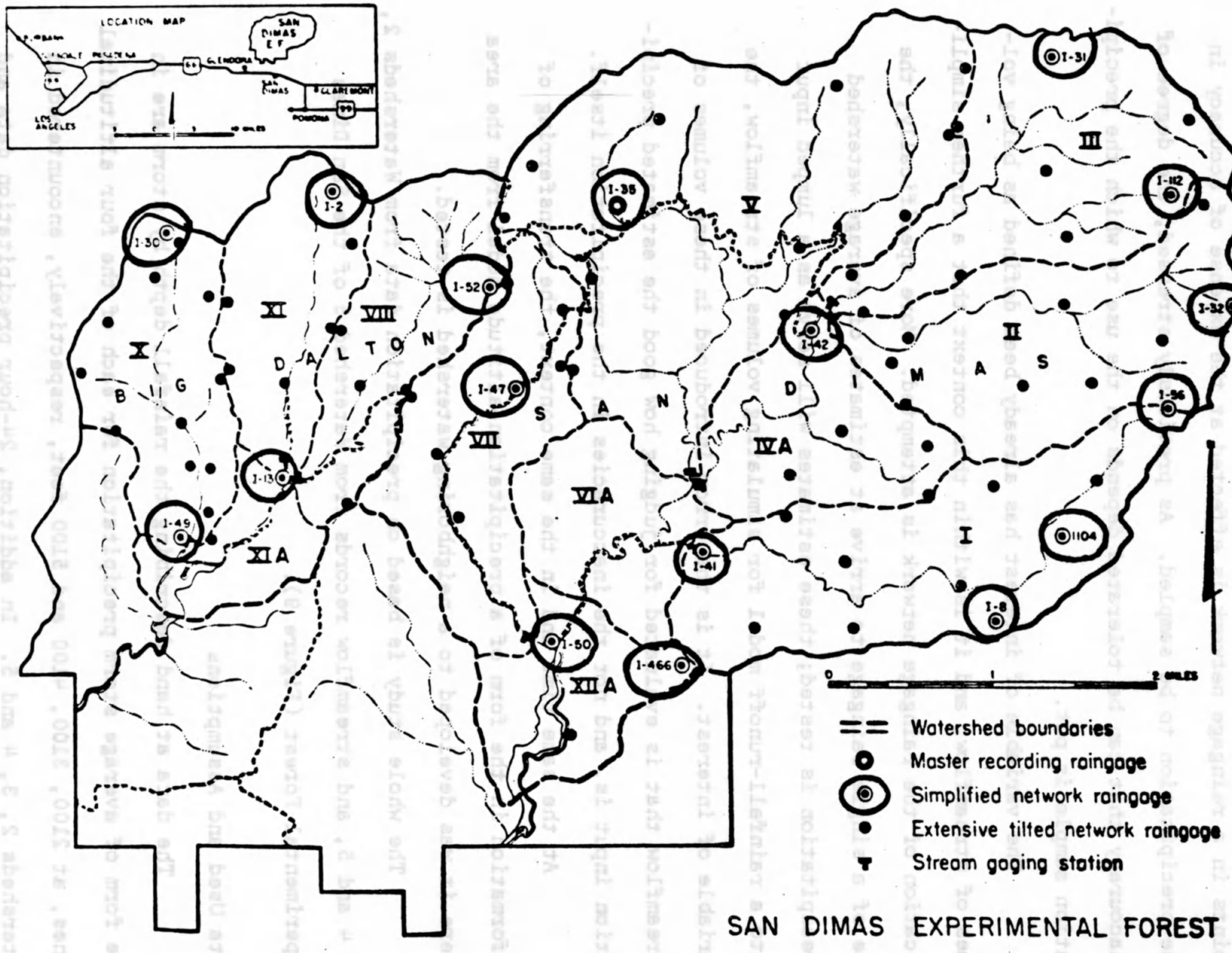


Figure 8. Distribution of 17 gages superimposed on the 77 tilted gage network.

points in a raingage network is effected at the expense of accuracy in the precipitation to be sampled. As previously stressed, the degree of inaccuracy that can be tolerated depends on the use to which the precipitation sample is put.

The variable of interest has already been defined as being volumes of streamflow, and it is within this context that a further simplification of the raingage network is attempted. More specifically, the use of a single raingage to arrive at estimates of average watershed precipitation is tested; these estimates will serve as a lumped input into a rainfall-runoff model for simulating volumes of streamflow, the variable of interest. It is the error introduced in these volumes of streamflow that is evaluated for judging how good the estimated precipitation input is, and not the inaccuracies in the precipitation itself.

At the same time and in the same context, the transferring of information in the form of a precipitation-altitude model from the area where it was developed to a neighboring watershed is tested.

The whole study is based on precipitation data from Watersheds 2, 3, 4 and 5, and streamflow records from Watershed 5 of the San Dimas Experimental Forest (Figure 9).

Data Used and Assumptions

The data at hand describing the rainfall depth by storm are in the form of average storm precipitation for each of the four altitudinal zones, at 2100, 3100, 4100 and 5100 feet, respectively, encountered in Watersheds 2, 3, 4 and 5. In addition, 24-hour precipitation data and

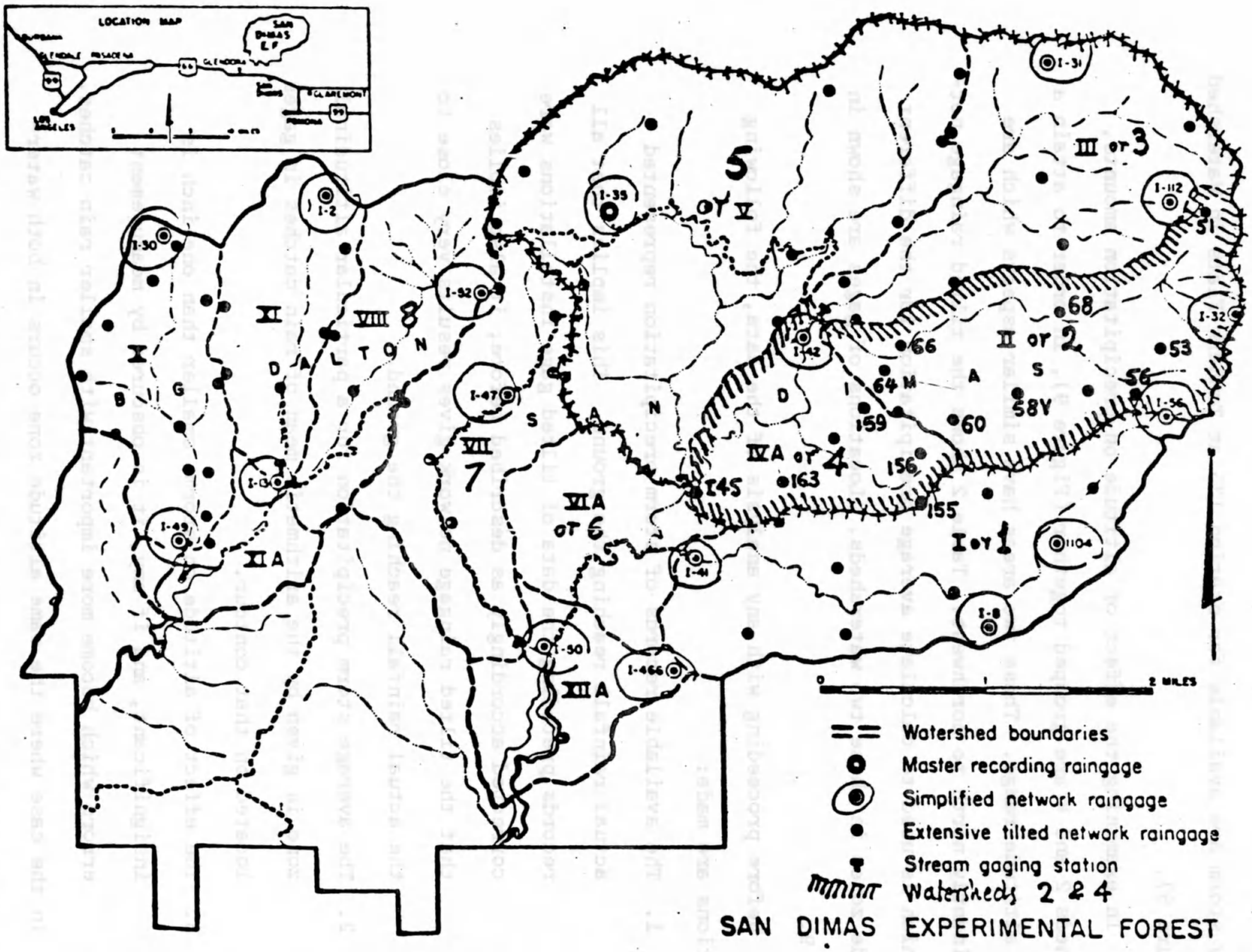


Figure 9. Location of Watersheds 2, 3, 4 and 5 of the San Dimas Experimental Forest.--Raingages in Watersheds 2 and 4 numbered according to Table 2 are used in the precipitation-altitude relationship.

also by storm are available for station I35 at Tanbark Flats in Watershed 5 (Figure 9).

In examining the effect of altitude on precipitation amounts, Watersheds 2 and 4 are grouped together (Figure 9), in order to attain a larger altitude range. These two areas have similar aspects which are predominantly north to northwest. Table 2 shows the tilted raingage network which is used to calculate average precipitation for the different altitude zones in these two watersheds. Locations of gages are shown in Figure 9.

Before proceeding with any analysis of the data, the following assumptions are made:

1. The available records of storm precipitation represented actual rainfall reaching the ground. This implies that all records preceding the data of tilted gage installations were corrected accordingly as described before; it also implies that the tilted raingage network gives results very close to the actual rainfall reaching the ground.
2. The average storm precipitation for a particular altitudinal zone is given by the arithmetic mean of rain catches in gages located on that contour.
3. The effect of altitude on storms smaller than one inch is insignificant, and if any, it is obscured by measurement errors which become more important with smaller rain catches.

In the case where the same altitude zone occurs in both watersheds, a weighted average of the given altitude storm depth is taken; the

Table 2. Altitudinal distribution of raingages in Watersheds 2 and 4.

	Raingage Nos.	
Altitude (feet)	Watershed 2	Watershed 4
2100		I45
3100	66, 64	159, 163
4100	68, 58V, 60	156, 155
5100	51, I32, 53, 56	

weighting is based on the number of raingages in each watershed for the same altitude zone and is based on assumption 2.

Methods of Analysis

In order to demonstrate the influence of altitude on precipitation depth, 41 storms of different magnitudes, all greater than one inch, were selected from the record between 1938 and 1951 and most of them are plotted individually with altitude (Figure 10). The selection is not random since the scarcer storms of large magnitudes are purposely included in order to cover the largest possible range in storm size and storms lower than one inch are ignored. By no means should these 41 storms be taken as a representative sample from the storm population in the San Dimas area.

Storms below one inch in depth are not included in the present analysis due to the absence of any consistent altitude effect on such storms as found by the previous studies (see assumption 3), and a preliminary examination of the data by the writer; furthermore, these storms are of lesser importance in the study of runoff-producing storms, and small errors in determining area precipitation for such storms are even less important.

It is obvious from the plot of the 41 storms (Figure 10) that for any particular storm, the rainfall amount increases with altitude, and by visual examination it may be further concluded that this increase is linear.

Following the plot mentioned above, a mathematical relation is sought between precipitation at a base station, in this case I45 at 2100

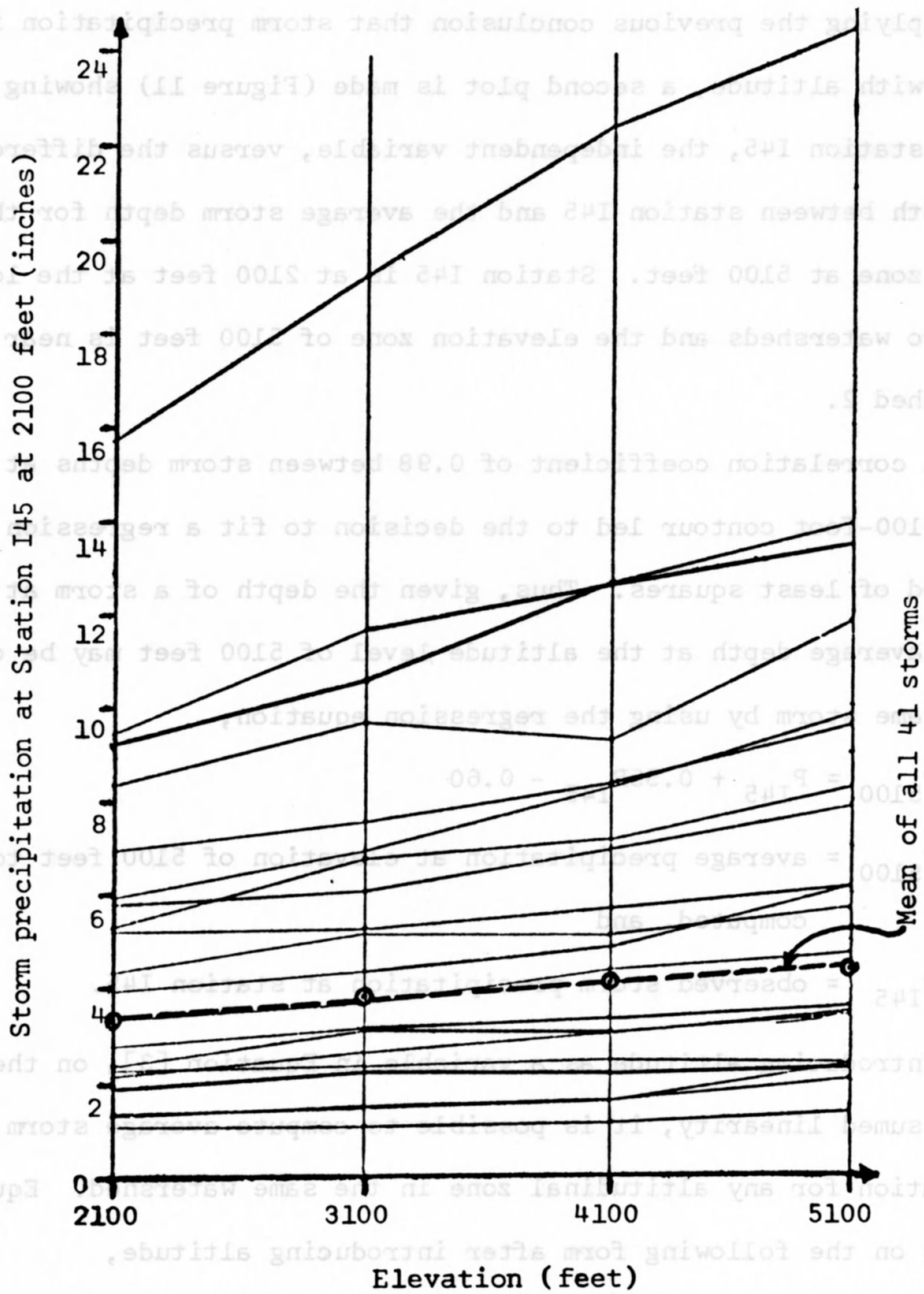


Figure 10. Storm depth versus altitude for 41 storms larger than one inch in Watersheds 2 and 4. (The plot does not show all 41 storms).

feet in Watershed 8, and the average storm precipitation at any altitude zone. Applying the previous conclusion that storm precipitation increases linearly with altitude, a second plot is made (Figure 11) showing storm depth at station I45, the independent variable, versus the difference in storm depth between station I45 and the average storm depth for the altitude zone at 5100 feet. Station I45 is at 2100 feet at the lower end of the two watersheds and the elevation zone of 5100 feet is near the top of Watershed 2.

A correlation coefficient of 0.98 between storm depths at station I45 and 5100-foot contour led to the decision to fit a regression line by the method of least squares. Thus, given the depth of a storm at station I45, the average depth at the altitude level of 5100 feet may be computed for the same storm by using the regression equation,

$$P_{5100} = P_{I45} + 0.55P_{I45} - 0.60 \qquad [3]$$

where P_{5100} = average precipitation at elevation of 5100 feet to be computed, and

P_{I45} = observed storm precipitation at station I45.

Introducing altitude as a variable in Equation [3], on the basis of the assumed linearity, it is possible to compute average storm precipitation for any altitudinal zone in the same watershed. Equation [3] takes on the following form after introducing altitude,

$$P_H = P_{I45} + 0.183(H - 2.1)P_{I45} - 0.2(H - 2.1) \qquad [4]$$

where H = any altitude in 1000 feet, and

P_H = storm precipitation at altitude H, to be computed.

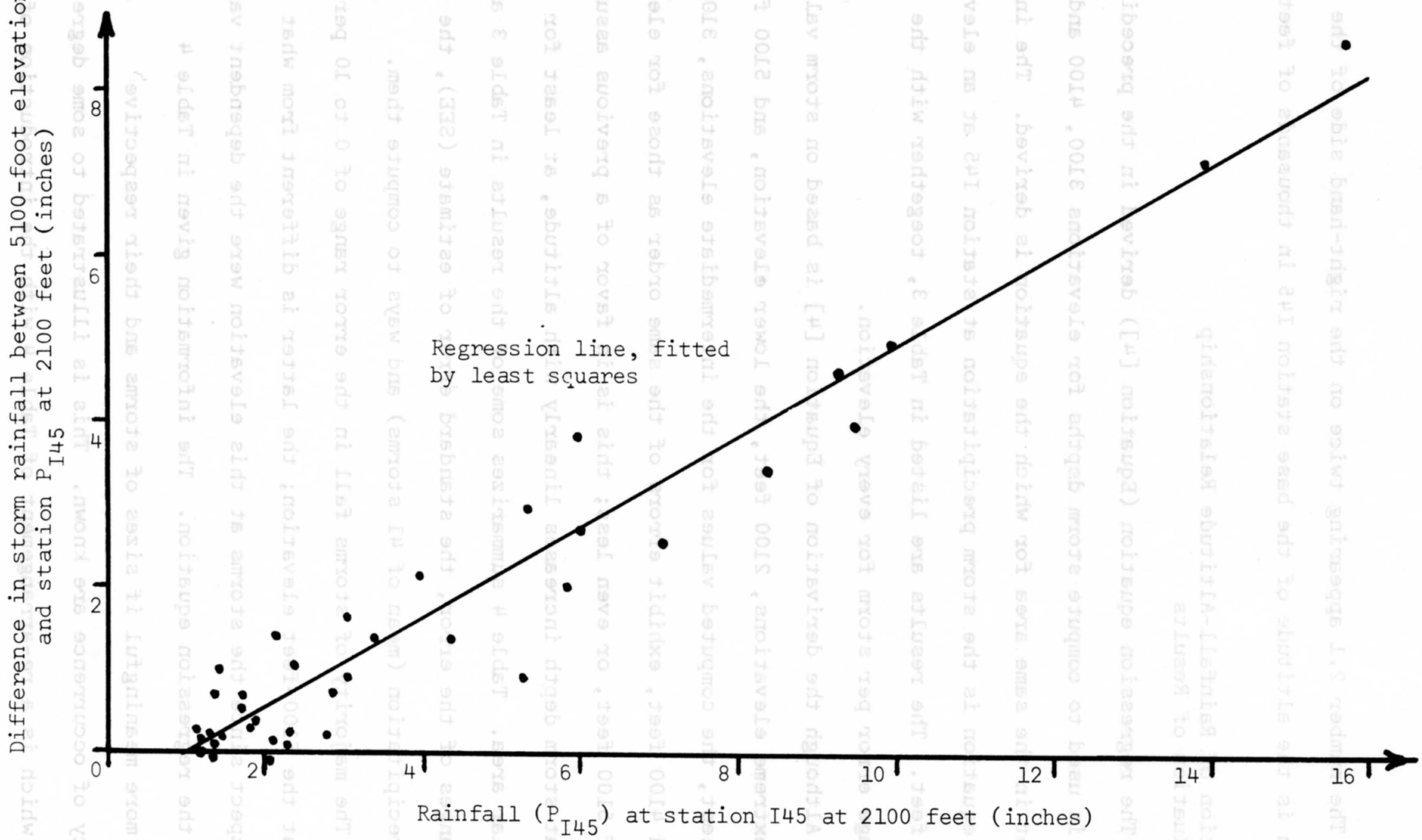


Figure 11. Storm precipitation difference between station I45 at 2100 feet and 5100-foot contour in combined Watersheds 2 and 4.

The number 2.1 appearing twice on the right-hand side of the equation is the altitude of the base station I45 in thousands of feet.

Application of Rainfall-Altitude Relationship and Evaluation of Results

The regression equation (Equation [4]) derived in the preceding section is used to compute storm depths for elevations 3100, 4100 and 5100 feet in the same area for which the equation is derived. The input to this equation is the storm precipitation at station I45 at an elevation of 2100 feet. The results are listed in Table 3, toegether with the percentage error per storm for every elevation.

Although the derivation of Equation [4] is based on storm values at two extreme elevations, 2100 feet, the lower elevation, and 5100 feet, the highest, the computed values for the intermediate elevations, 3100 feet and 4100 feet, exhibit errors of the same order as those for elevations of 5100 feet, or even less; this is in favor of a previous assumption that storm depth increases linearly with altitude, at least for this particular area. Table 4 summarizes some of the results in Table 3 and gives ranges of the error, the standard error of estimate (SEE), the mean storm precipitation (mean of 41 storms) and ways to compute them.

The majority of storms fall in the error range of 0 to 10 percent except at the 5100-feet elevation; the latter is different from what one would expect since the storms at this elevation were the dependent variable in the regression equation. The information given in Table 4 becomes more meaningful if sizes of storms and their respective frequency of occurrence are known. This is illustrated to some degree in Table 5 which is a re-arrangement of Table 4 with the introduction of

Table 3. Observed versus computed storm precipitation.

			Storm Precipitation in Inches								
			Av. for El. 3100'			Av. for El. 4100'			Av. for El. 5100'		
Storm No.	Duration (hrs)	I45* 2100' Obs.	Obs.	Comp.	% Error	Obs.	Comp.	% Error	Obs.	Comp.	% Error
271+272	42.7	2.38	3.11	2.61	16.1	3.05	2.84	6.9	3.44	3.07	10.8
276	23.3	1.89	2.17	2.04	6.0	2.14	2.19	- 2.3	2.28	2.33	- 2.2
279	8.5	1.06	1.03	1.14	-10.7	1.12	1.23	- 9.8	1.34	1.31	2.2
283	6.2	1.36	1.46	1.47	- 0.7	1.56	1.58	- 1.3	2.37	1.69	28.7
243	11.2	1.80	2.12	1.93	9.0	2.18	2.07	5.0	3.10	2.20	29.0
247	55.5	5.30	6.76	6.02	10.9	7.12	6.73	5.5	8.24	7.45	9.6
219	12.7	1.22	1.37	1.32	3.6	1.14	1.42	- 0.7	1.43	1.51	- 5.6
220-1	56.5	9.25	10.57	10.63	- 0.6	12.64	12.00	5.1	13.84	13.38	3.3
223	10.5	1.12	1.15	1.21	- 5.2	1.26	1.30	- 3.2	1.34	1.39	- 3.7
225	54.7	5.79	6.04	6.59	- 9.1	6.93	7.39	- 6.6	7.80	8.19	- 5.0
227	65.0	2.96	3.13	3.29	- 5.1	3.27	3.61	-10.4	3.85	3.94	- 2.3
195	12.0	1.28	1.31	1.38	- 5.3	1.36	1.48	- 8.8	1.24	1.59	-20.2
198	36.0	4.37	5.25	4.93	6.1	5.17	5.49	- 6.2	5.70	6.06	- 6.3
199	53.5	2.99	3.40	3.32	2.4	3.90	3.65	6.4	4.60	3.98	13.5
200	11.4	1.33	1.60	1.44	10.0	1.49	1.54	- 3.4	1.48	1.65	-11.5
209+210	86.6	9.49	11.66	10.91	6.4	12.58	12.32	2.1	13.47	13.74	- 2.0
215	19.2	1.42	1.54	1.53	0.7	1.61	1.65	- 2.5	1.64	1.76	- 7.3
179	11.8	1.25	1.24	1.35	- 8.9	1.36	1.45	- 6.6	1.24	1.55	-25.0
181	19.0	1.24	1.33	1.34	- 0.8	1.30	1.44	-10.8	1.38	1.54	-11.6
182	59.2	15.75	19.23	18.21	5.3	22.36	20.68	7.5	24.35	23.14	5.0
186	55.5	5.95	6.91	6.78	1.9	8.32	7.60	8.7	9.82	8.43	14.2
187	43.5	6.98	7.55	7.98	- 5.7	8.34	8.98	- 7.7	9.51	9.97	- 4.8
164	20.5	2.10	2.66	2.28	14.3	3.01	2.47	17.9	3.49	2.65	24.1
169	17.1	1.29	1.50	1.39	7.3	1.51	1.50	0.7	2.03	1.60	21.2
170	57.0	1.72	1.95	1.84	5.6	1.92	1.96	- 2.1	2.11	2.08	1.4
147	38.7	3.95	4.37	4.44	- 1.6	4.86	4.93	- 1.4	6.12	5.43	11.3
128	18.9	2.25	2.38	2.46	- 3.4	2.46	2.67	- 8.5	2.34	2.87	-22.6

126	16.3	1.71	2.35	1.83	22.1	2.60	1.95	25.0	2.44	2.06	15.6
94a,95,96	94.3	8.38	9.66	9.61	0.5	9.28	10.84	-16.8	11.78	12.08	- 2.5
97	11.2	1.86	1.77	2.00	-13.0	2.01	2.15	- 7.0	1.92	2.29	-19.3
98	19.5	1.47	2.17	1.59	26.7	2.12	1.71	15.3	1.52	1.82	-19.7
305	82.4	1.95	2.62	2.11	19.5	2.93	2.27	22.5	2.10	2.42	-15.2
307	24.2	3.37	3.65	3.76	- 3.0	4.41	4.16	5.7	4.72	4.55	3.6
309	40.0	5.25	5.26	5.96	-13.3	5.69	6.67	-17.2	6.12	7.38	-20.6
315	27.3	2.70	2.56	2.98	-16.4	2.82	3.27	-16.0	2.90	3.55	-22.4
317	11.3	2.05	1.92	2.22	-15.6	2.07	2.40	-15.9	2.20	2.57	-16.8
318	15.3	1.68	1.88	1.79	4.8	1.94	1.91	1.5	2.24	2.02	9.8
337	8.2	1.89	1.98	2.04	- 3.0	2.11	2.19	- 3.8	2.10	2.33	-11.0
338	62.1	2.81	3.06	3.11	- 1.6	3.30	3.41	- 3.3	3.57	3.71	- 3.9
339+340	18.0	2.29	1.88	2.50	-33.0	2.39	2.72	-13.8	2.54	2.93	-15.4
341	15.4	2.02	1.62	2.19	-35.2	1.82	2.36	-29.7	1.86	2.53	-36.0

*Base station

Av. = average
Obs. = observed
Comp. = computed

Table 4. Classification of the 41 storms by elevation and error range.

(1) Elevation (feet)	(2) 0-10 percent	Error Range (percent) (3) 10-20 percent	(4) 20-30 percent	(5) Over 30 percent	Storm mean (inches)	SEE (inches)
3100	28	9	2	2	3.785	0.3651
4100	29	9	3	0	4.137	0.4970
5100	18	13	9	1	4.575	0.5348

Table 5. Listing of storms by magnitude classes per error range for every elevation.

(1) Absolute Error Range Percent	(2) Storm Depth Range (inches)	(3) El. = 3100 No. of Storms	(4) El. = 4100 No. of Storms	(5) El. = 5100 No. of Storms	(6) Storms in Depth Range
0-10	1-2	12	11	4	22.0
	2-5	8	8	6	17.9
	5-10	5	5	4	11.4
	> 10	3	5	4	9.8
10-20	1-2	4	1	4	7.3
	2-5	4	6	7	13.8
	5-10	1	2	2	4.1
	> 10	-	-	-	0.0
20-30	1-2	-	1	2	2.4
	2-5	2	2	6	8.1
	5-10	-	-	1	0.8
	> 10	-	-	-	0.0
Over 30	1-2	2	-	1	2.4
	2-5	-	-	-	0.0
	5-10	-	-	-	0.0
	> 10	-	-	-	0.0

Note: El. stands for elevation which is given in feet.

classes or ranges of storm size for each error range. Column 6 in Table 5 gives the percentage of storms, averaged for the three elevations, falling within each size-range for every error-range.

Because the 41 storms used in the present study were selected in a biased manner for reasons explained earlier, the information in Table 5 is limited only to the distribution of error with storm size. Storm-size distribution, as shown in the same table, does not relate to reality; however, the real storm-size distribution can be found by an analysis of a random sample, or even all, of the storms for which records are available (Reimann and Hamilton, 1959).

In an attempt to investigate the possibility of regionalizing, to some extent, the precipitation-altitude relationship as described by Equation [4], Watersheds 2, 3, 4 and 5 are grouped together and examined as a single watershed. Using the same base station, I45, as before, storm depths are computed for points at different elevations in this area for the same 41 storms. In addition, mean areal precipitation for each storm is computed by using the Thiessen method and compared to areal storm precipitation as calculated from observed values and the Thiessen method. The results are shown in Table 6; the explanation of the variables appearing on this table follow.

Variable	Description
I45	Base rainfall station at 2100 feet.
I32	Rainfall station at 5200 feet.
I42	Rainfall station at 2600 feet.
I35	Rainfall station at 2700 feet (master gage).

Table 6. Results of computations for testing regionalization of the precipitation-altitude relationship (Equation [4]).

I45	I32		I42		I35		5-4100		4-4100		3-5100		PAVER	PCAVER
P	P	PC	P	PC	P	PC	P	PC	P	PC	P	PC	P	PC
2.38	3.39	3.11	2.85	2.50	3.06	2.52	3.46	2.85	2.76	2.85	3.70	3.09	3.11	2.75
1.59	2.28	2.34	1.86	1.96	1.98	1.98	1.80	2.18	2.09	2.18	2.07	2.33	1.98	2.12
1.06	1.34	1.04	1.10	1.06	.98	1.06	.97	1.05	1.00	1.05	1.36	1.04	1.10	1.05
1.36	2.50	1.51	1.69	1.38	1.65	1.39	1.52	1.46	1.26	1.46	2.37	1.51	1.75	1.44
1.80	3.25	2.20	2.23	1.85	2.15	1.88	2.05	2.06	2.10	2.06	3.00	2.19	2.35	2.00
5.30	8.04	7.59	6.97	5.68	6.45	5.76	8.08	5.84	6.67	6.84	7.82	7.61	7.11	6.50
1.22	1.40	1.29	1.20	1.23	1.28	1.23	1.27	1.27	1.33	1.27	1.30	1.29	1.28	1.26
9.25	13.98	13.88	9.74	10.00	9.73	10.15	11.20	12.24	12.32	12.24	13.00	13.73	11.20	11.57
1.12	1.21	1.14	1.12	1.12	1.37	1.12	1.33	1.13	1.18	1.13	1.36	1.13	1.25	1.13
5.79	8.01	8.45	5.60	6.22	6.18	6.31	6.37	7.51	6.83	7.51	6.88	8.37	6.46	7.13
2.96	3.81	4.02	2.77	3.13	2.58	3.17	2.62	3.64	3.20	3.64	3.38	3.99	2.98	3.49
1.28	1.16	1.39	1.15	1.30	1.26	1.30	1.23	1.35	1.34	1.35	1.14	1.38	1.22	1.33
4.37	5.32	6.23	5.19	4.67	5.09	4.73	5.65	5.57	4.93	5.57	5.94	6.17	5.26	5.30
2.99	4.65	4.07	2.88	3.16	3.51	3.20	3.48	3.68	3.61	3.68	4.45	4.03	3.62	3.53
1.33	1.37	1.46	1.31	1.35	1.61	1.36	1.23	1.42	1.53	1.42	1.42	1.46	1.41	1.40
9.49	17.05	14.25	10.01	10.26	11.74	10.41	12.38	12.56	10.56	12.56	12.48	14.10	11.88	11.88
1.42	1.56	1.61	1.37	1.45	1.49	1.46	1.49	1.54	1.58	1.54	1.66	1.60	1.51	1.51
1.25	1.26	1.34	1.13	1.26	.93	1.27	.86	1.31	1.22	1.31	1.23	1.34	1.10	1.29
1.24	1.24	1.32	1.26	1.25	1.45	1.26	.98	1.29	1.40	1.29	1.51	1.32	1.30	1.28
15.75	25.72	24.06	18.33	17.09	20.90	17.36	21.45	21.11	23.09	21.11	22.05	23.80	21.08	19.92
5.95	10.63	8.71	6.08	6.39	6.45	6.48	7.56	7.73	8.36	7.73	9.68	8.62	7.69	7.33
6.98	9.79	10.32	4.72	7.52	6.94	7.63	7.45	9.13	8.22	9.13	9.18	10.21	7.42	8.65
2.10	3.54	2.57	2.39	2.19	2.01	2.21	2.78	2.47	2.74	2.47	3.44	2.65	2.68	2.39
1.29	2.02	1.40	1.30	1.31	1.56	1.31	1.60	1.36	1.36	1.36	2.04	1.40	1.59	1.35
1.72	1.98	2.08	1.82	1.78	1.87	1.79	1.81	1.95	1.88	1.95	2.05	2.06	1.88	1.90
3.95	5.88	5.57	3.83	4.21	3.96	4.26	4.92	5.00	4.54	5.00	6.20	5.52	4.69	4.76
2.25	2.47	2.91	2.21	2.36	2.47	2.38	2.48	2.67	2.36	2.67	2.34	2.89	2.37	2.58
11.71	2.48	2.06	2.15	1.77	2.21	1.78	2.18	1.94	2.41	1.94	2.70	2.05	2.28	1.89

8.38	11.88	12.51	8.91	9.05	9.09	9.18	10.16	11.05	8.34	11.05	12.77	12.38	9.89	10.45
1.86	2.08	2.30	2.03	1.93	2.35	1.94	2.14	2.14	2.14	2.14	1.92	2.28	2.10	2.08
1.47	1.41	1.68	1.80	1.50	.93	1.51	1.13	1.61	1.88	1.61	1.38	1.68	1.41	1.58
1.95	1.70	2.44	2.54	2.03	1.79	2.04	2.46	2.26	3.42	2.26	2.05	2.42	2.29	2.19
3.37	4.71	4.66	3.73	3.58	3.87	3.62	4.22	4.20	4.68	4.20	4.02	4.62	4.08	4.02
5.25	6.42	7.61	5.34	5.63	5.42	5.71	5.13	6.77	5.62	6.77	5.24	7.53	5.46	6.43
2.70	2.92	3.61	2.64	2.85	2.56	2.88	2.56	3.29	2.90	3.29	2.60	3.58	2.68	3.16
2.05	2.15	2.59	1.80	2.14	1.74	2.16	1.74	2.40	2.03	2.40	1.85	2.58	1.88	2.32
1.68	2.25	2.01	1.80	1.73	1.87	1.74	1.89	1.89	1.78	1.89	2.38	2.00	1.94	1.85
1.89	2.10	2.34	1.72	1.96	1.80	1.98	1.71	2.18	2.16	2.18	2.28	2.33	1.93	2.12
2.81	3.69	3.78	2.87	2.97	2.73	3.00	2.67	3.44	3.34	3.44	3.16	3.75	3.00	3.30
2.29	2.56	2.97	2.38	2.40	2.72	2.42	2.66	2.73	2.44	2.73	2.54	2.95	2.53	2.63
2.02	1.86	2.55	1.88	2.10	1.82	2.12	1.72	2.36	1.78	2.36	1.84	2.53	1.83	2.28
0.0*	.7643*		.5768*		.6701*		.6079*		.7323*		.7704*		.4153*	

*SEE

5-4100	Average of precipitation at two neighboring points at 4100 feet elevation in Watershed 5.
4-4100	Same as above but for Watershed 4.
3-5100	Same, but for 5100 feet and Watershed 3.
P	Observed storm precipitation in inches.
PC	Computed storm precipitation inches using Equation [4].
PAVER	Mean areal storm precipitation (observed).
PCAVER	Mean areal storm precipitation (computed).
SEE	Standard error of estimate.

The locations of the points mentioned above and the Thiessen polygons for areas 2, 3, 4 and 5 grouped together are shown in Figure 12. An examination of Table 6 reveals that the erros are somewhat higher than those in Table 2 where the values refer to computed storm depths in the same area as the one for which the relationship is derived. However, the error in the mean areal precipitation (last two columns of Table 6) is much smaller than for individual points or elevations in the watershed, thus suggesting a smoothing of the erros arising at individual points. This si important in the case of using a watershed model with precipitation as a lumped input. Table 7 is a self-explanatory summary of the error distribution of the computed mean value of the areal storm precipitation. The absence of storms in the error range of 30 percent of higher is indication of the tendency of lumping to remove or rather attenuate high spatial variations.

In an effor to test the transferability of Equation [4], raingage I35 is used as a base station in this equation to compute mean areal storm

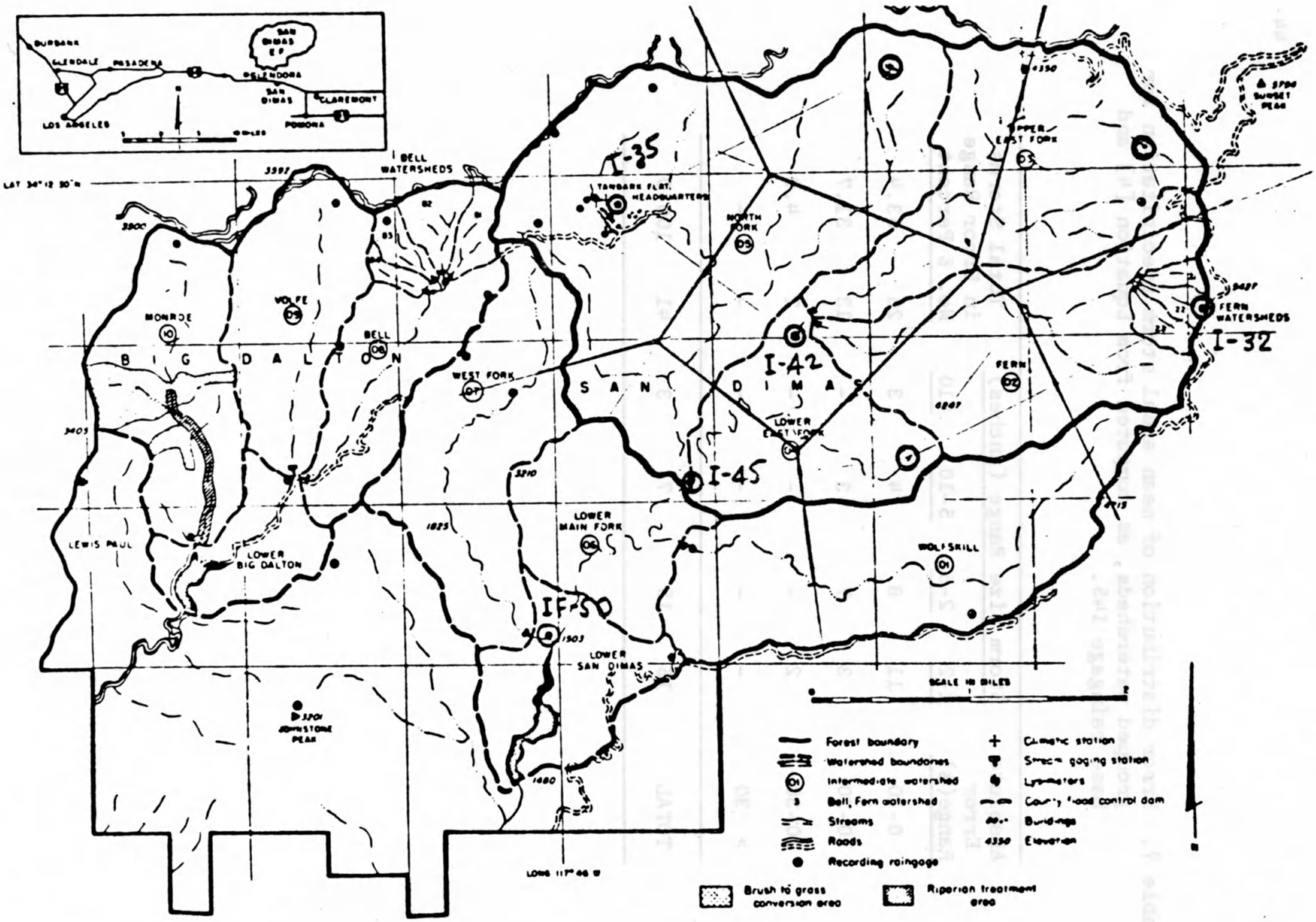


Figure 12. Thiessen polygons for calculating mean areal precipitation in grouped Watersheds 2, 3, 4 and 5.

Table 7. Error distribution of mean areal storm precipitation for grouped watersheds, as computed from Equation [4] and base raingage I45.

Absolute Error Range(%)	Storm Size Range (inches)				Total Storms in Error Range No. & Percent	
	1-2	2-5	5-10	>10		
0-10	11	8	4	3	26	63.4
10-20	3	7	3	-	13	31.7
20-30	2	-	-	-	2	4.9
> 30	-	-	-	-	-	-
TOTAL	16	15	7	3	41	100.0

precipitation for the same group of watersheds. Raingage I35 is in Watershed 5 at an elevaiton of 2700 feet, and Equation [4] is modified to read as follows:

$$P_H = P_{I35} + 0.183(H - 2.7)P_{I35} - 0.2(H - 2.7) \qquad [5]$$

where P_{I35} is the storm precipitation at station I35 at 2700 feet and the other variables are as defined earlier.

The results from Equation [5] are listed in Table 8, which is the counterpart of Table 6 obtained from Equation [4]. Comparing the results in these two tables, Equation [4] (Table 6) is found to give better results than Equation [5] (Table 8) for most individual points in the watershed. This is also indicated by the values of the standard error of estimate (SEE) appearing at the bottoms of the two tables. The computed mean areal storm precipitation, however, shown in the last column of each of the two tables, does not display much variation between the two, except for the two extreme storms (11.88 and 21.08 inches). These two storms were over-estimated by Equation [5] (Table 8), and thus, gave an SEE of 0.6542 as compared to 0.4153 (Table 6) for mean areal storm values. The observed areal precipitation (PAVER) average over the 41 storms is 3.916 inches. Thus, the two values of SEE are 20.4 and 13.0 percent of the mean, respectively.

The error-size distribution is better illustrated in Table 9 which is a summary of the absolute values of error between the last two columns in Table 8. In comparing Tables 7 and 9, no appreciable difference can be detected, except for the storm-size range greater than ten inches. The errors in these two tables also compare favorably with those in Table 5,

Table 8. Results of computation in testing the transferability of the precipitation-altitude relationship (Equations [4] and [5]).

I35	I32		I42		I45		5-4100		4-4100		3-5100		PAVER	PCAVER
P	P	PC	P	PC	P	PC	P	PC	P	PC	P	PC		
3.06	3.39	3.96	2.85	3.02	2.38	2.84	3.46	3.56	2.76	3.56	3.70	3.92	3.11	3.40
1.98	2.28	2.39	1.86	1.96	1.89	1.88	1.80	2.21	2.09	2.21	2.07	2.37	1.98	2.13
.98	1.34	.93	1.10	.98	1.06	.99	.97	.95	1.00	.95	1.36	.93	1.10	.96
1.65	2.50	1.90	1.69	1.64	1.36	1.59	1.52	1.79	1.26	1.79	2.37	1.89	1.75	1.75
2.15	3.25	2.63	2.23	2.13	1.80	2.03	2.05	2.42	2.10	2.42	3.00	2.61	2.35	2.33
6.45	8.04	8.90	6.97	6.35	5.30	5.86	8.08	7.82	6.67	7.82	7.82	8.80	7.11	7.38
1.28	1.40	1.37	1.20	1.28	1.22	1.26	1.27	1.33	1.33	1.33	1.30	1.36	1.28	1.31
9.73	13.98	13.68	9.74	9.57	9.25	8.78	11.20	11.94	12.32	11.94	13.00	13.52	11.20	11.24
1.37	1.21	1.50	1.12	1.36	1.12	1.34	1.33	1.44	1.18	1.44	1.36	1.49	1.25	1.42
6.18	8.01	8.51	5.60	6.09	5.79	5.62	6.37	7.48	6.83	7.48	6.88	8.41	6.46	7.07
2.58	3.81	3.26	2.77	2.55	2.96	2.42	2.62	2.96	3.20	2.96	3.38	3.23	2.98	2.84
1.26	1.16	1.34	1.15	1.26	1.28	1.24	1.23	1.30	1.34	1.30	1.14	1.33	1.22	1.29
5.09	5.32	6.92	5.19	5.02	4.37	4.65	5.65	6.11	4.93	6.11	5.94	6.85	5.26	5.79
3.51	4.65	4.62	2.88	3.47	2.99	3.24	3.48	4.13	3.61	4.13	4.45	4.57	3.62	3.93
1.61	1.37	1.85	1.31	1.60	1.33	1.55	1.23	1.74	1.53	1.74	1.42	1.84	1.41	1.70
11.74	17.05	16.61	10.01	11.55	9.49	10.57	12.38	14.47	10.56	14.47	12.48	16.42	11.88	13.60
1.49	1.56	1.67	1.37	1.48	1.42	1.45	1.49	1.59	1.58	1.59	1.66	1.66	1.51	1.56
.93	1.26	.86	1.13	.93	1.25	.95	.86	.89	1.22	.89	1.23	.86	1.10	.90
1.45	1.24	1.61	1.26	1.44	1.24	1.41	.98	1.54	1.40	1.54	1.51	1.61	1.30	1.51
20.90	25.72	29.96	18.33	20.54	15.75	18.73	21.45	25.97	23.09	25.97	22.05	29.60	21.08	24.36
6.45	10.63	8.90	6.08	6.35	5.95	5.86	7.56	7.82	8.36	7.82	9.68	8.80	7.69	7.38
6.94	9.79	9.62	4.72	6.83	6.98	6.30	7.45	8.44	8.22	8.44	9.18	9.51	7.42	7.96
2.01	3.54	2.43	2.39	1.99	2.10	1.91	2.78	2.24	2.74	2.24	3.44	2.41	2.68	2.17
1.56	2.02	1.77	1.30	1.55	1.29	1.51	1.60	1.68	1.36	1.68	2.04	1.77	1.59	1.64
1.87	1.98	2.23	1.82	1.86	1.72	1.78	1.81	2.07	1.88	2.07	2.05	2.21	1.88	2.01
3.96	5.88	5.27	3.83	3.91	3.95	3.65	4.92	4.69	4.54	4.69	6.20	5.22	4.69	4.46
2.47	2.47	3.10	2.21	2.44	2.25	2.32	2.48	2.82	2.36	2.82	2.34	3.07	2.37	2.71
2.21	2.48	2.72	2.15	2.19	1.71	2.09	2.18	2.50	2.41	2.50	2.70	2.70	2.28	2.40

9.09	11.88	12.75	8.91	8.94	8.38	8.21	10.16	11.14	8.34	11.14	12.77	12.60	9.89	10.49
2.35	2.08	2.93	2.03	2.33	1.86	2.21	2.14	2.67	2.14	2.67	1.92	2.90	2.10	2.57
.93	1.41	.86	1.80	.93	1.47	.95	1.13	.89	1.88	.89	1.38	.86	1.41	.90
1.79	1.70	2.11	2.54	1.78	1.95	1.71	2.46	1.97	3.42	1.97	2.05	2.10	2.29	1.91
3.87	4.71	5.14	3.73	3.82	3.37	3.57	4.22	4.58	4.68	4.58	4.02	5.09	4.08	4.35
5.42	6.42	7.40	5.34	5.34	5.25	4.94	5.13	6.53	5.62	6.53	5.24	7.32	5.46	6.18
2.56	2.92	3.23	2.64	2.53	2.70	2.40	2.56	2.94	2.90	2.94	2.60	3.20	2.68	2.82
1.74	2.15	2.04	1.80	1.73	2.05	1.67	1.74	1.91	2.03	1.91	1.85	2.02	1.88	1.85
1.87	2.25	2.23	1.80	1.86	1.68	1.78	1.89	2.07	1.78	2.07	2.38	2.21	1.94	2.01
1.80	2.10	2.12	1.72	1.79	1.89	1.72	1.71	1.98	2.16	1.98	2.28	2.11	1.93	1.92
2.73	3.69	3.48	2.87	2.70	2.81	2.55	2.67	3.15	3.34	3.15	3.16	3.45	3.00	3.02
2.72	2.56	3.46	2.38	2.69	2.29	2.54	2.66	3.14	2.44	3.14	2.54	3.43	2.53	3.00
1.82	1.86	2.15	1.88	1.81	2.02	1.74	1.72	2.01	1.78	2.01	1.84	2.14	1.83	1.95
0.0*	.9021*		.6053*		.5732*		.9180*		1.0148*		1.4793*		.6542*	

*SEE

Table 9. Error distribution of mean areal storm precipitation for the grouped watersheds, as computed from Equation [5] and base raingage I35.

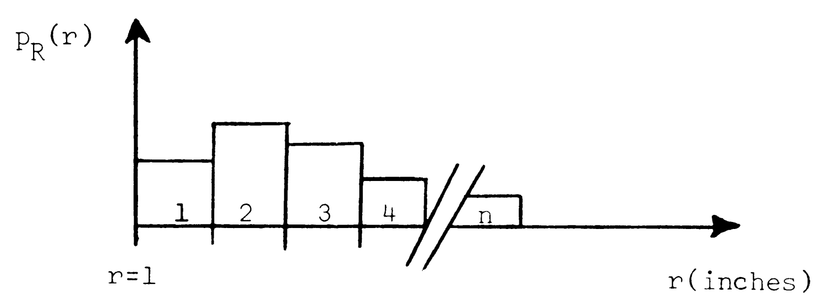
Absolute Error Range (%)	Storm Size Range (inches)				Total Storms in Error Range No. & Percent	
	1-2	2-5	5-10	>10		
0-10	12	9	5	1	27	65.8
10-20	3	4	2	2	11	26.8
20-30	1	1	-	-	2	4.9
> 30	1	-	-	-	1	2.5
TOTAL	17	14	7	3	41	100.0

which is a summary of the errors in computed storms over areas 2 and 4 for which Equation [4] was developed. This indicates that the error in estimating mean areal precipitation in the San Dimas area is smaller for larger areas, which is compatible with the regionality of storms in this area.

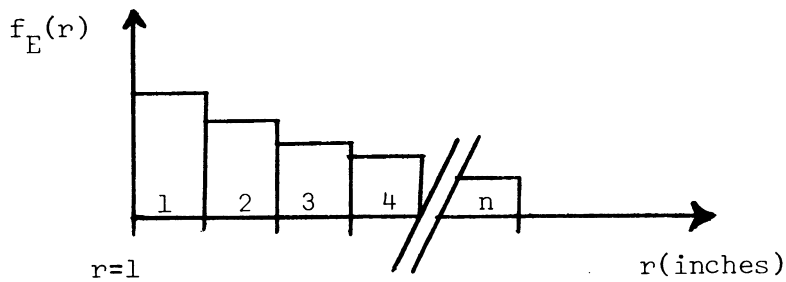
It may be concluded that, for more than 60 percent of the storms examined, the regionalizing and transferring of the precipitation-altitude relationship (Equations [4] and [5]) produce errors in mean areal storm precipitation less than ten percent. Storms in the error range of 10-20 percent account approximately for 24 percent of the 41 storms examined. The reader is reminded that these 41 storms are a biased sample of the storm population in this area.

A weighted error ($\overline{E}$) may be calculated for storms larger than one inch as follows:

1. Find the probability distribution $p_R(r)$ of all storms larger than one inch appearing in the record (arrange storms in n magnitude classes as shown in sketch below).



2. Find the distribution of error $f_E(r)$ with storm size by applying Equations [4] and [5] to the same number of storms (use same classes as in 1 above).



3. The weighted error $\overline{E}$ may be estimated as follows,

$$\overline{E} = \frac{\sum_{i=1}^{n} r_i p_R(r_i) f_E(r_i)}{\sum_{i=1}^{n} p_R(r_i) f_E(r_i)}$$

The importance of the previous findings, and in general any other similar information based on more detailed studies, depends on the purpose for which the precipitation data is required. It is shown in Chapter 4 how similar errors arising in the estimation of mean areal precipitation influence the "accuracy" of simulated streamflow, when the estimated precipitation is used as the input to a rainfall-runoff model.

CHAPTER 4

STREAMFLOW SIMULATION

General

As previously pointed out, the variability of precipitation, spatial, temporal or both, is of varying importance, depending on the purpose for which precipitation is to be used, and the magnitude of error that can be tolerated in the variable of interest. In real field applications the error magnitude, which is directly proportional to the degree of uncertainty with respect to knowledge of this variable, or even the system itself, is related to the economic feasibility of a particular application.

In streamflow simulation where precipitation is the main input to the rainfall-runoff model, the damping effect and the memory of a watershed system, the kind of output required and the tolerable error are factors which determine the adequacy of a rainfall sampling network. In existing raingage networks the problem of identifying a minimum network to suit a specific purpose is not difficult; existing data may be used to test different alternative networks before selecting the best one. The problem arises in watersheds with no data at all, or with only very few raingages, sometimes only one. In such cases transferring of information from neighboring, better gaged areas may prove beneficial; but how is one assured that transferring of information is good, and how good?

In the last part of Chapter 3 the spatial variability of precipitation in the San Dimas area was investigaged and an equation was derived describing the relationship between storm size and altitude for a small area within the San Dimas Experimental Forest. The same relationship was later applied to a raingage outside this small area for computing mean storm depths over a larger area. The errors in the results of these computations are tabulated in the same chapter. The importance of these errors and the effectiveness of transferring of information depends on the use to which precipitation values will be put.

The variable of interest, as stated earlier, is the monthly values of streamflow; thus, a simple rainfall-runoff model is used to simulate two sets of streamflow for the same period. In the first set the precipitation input is regarded as the actual or true precipitation on Watershed 5 and the values of simulated streamflow are assumed, for the sake of this study, to be the same as the observed values and are considered as the basis for comparison. The second set of precipitation input is computed by using Equation [5] (see Appendix 2); errors in this set of precipitation are expected to produce corresponding errors in the simulated streamflow. Dawdy and Bergman (1969) evaluated the errors in simulated flood peaks by using only one raingage on a basin (Chapter 5).

The Rainfall-Runoff Model

A full description of the rainfall-runoff model used is given in Appendix 1. The original version of this model which bears the name of Mero (Mero's watershed model), who was the first to develop it, was improved and modified by the writer to comply more with physical processes and the kind of input and output desired. The rainfall-runoff relationship which was derived elsewhere (see Appendix 1) was not changed in the program used here, due to time limitations forbidding the derivation of a new relationship for the San Dimas watersheds.

No detailed study of the model's sensitivity to the different parameters and inputs was undertaken. However, some qualitative remarks based on a preliminary sensitivity analysis and the experience of the author are given in Table 10 which describes the model parameter values after calibration. In addition, later findings (Chapter 4) indicate that the model is very sensitive to the precipitation input.

The Input to the Model

Two kinds of input are required by the rainfall-runoff model: 24-hour mean areal precipitation and 24-hour potential evaporation values.

Precipitation

For both sets of computations of simulated streamflow the same rainfall-runoff model is used and the same set of watershed parameters. In the first case, however, "actual" precipitation is used as the input to the model, the precipitation being that as measured at raingage I35. Appendix 2 gives a description of the work done and the reasoning for concluding on the use of point measurements at I35 as being representative of mean areal precipitation for Watershed 5.

For the second set of simulated streamflows the precipitation input is computed by using Equation [5] (same as Equation [4] but modified

Table 10. Parameter values for rainfall-runoff model estimated during calibration.--Definitions of these parameters are given in Appendix 1.

Parameter	Value & Units	Remarks
A1	2.0 km^2	Important in streamflow volumes
A2	3.0 km^2	
A3	3.0 km^2	
A4	11.1 km^2	
TO1	200 days	Important in shape of hydrograph and length of memory
TO2	20 days	Recall $Q(t) = Q_o e^{-t/T_{oi}}$ (i = 1, 2, 3, 4)
TO3	3 days	
TO4	1 day	
LIST	18 mm	Important in both hydrograph shape and total streamflow volume
LFC	80 mm	
LST	300 mm	
DM	60 days	
CL2	0.005*	
EVPC	0.9*	
EC	0.5*	
S1	100.0 mm°	Not important in the long run
L1	5.0 mm	
QO	0.020	Important in overland flow quantities and peaks
CT	0.300	

* Constants
° Controls initial baseflow

for the new base station I35), to calculate mean areal precipitation values. More details on this are given in Appendix 2.

Since the precipitation-altitude relationship (Equations [4] and [5]) was developed by using a number of storms of varying durations, it is necessary to allocate storm-precipitation to 24-hour periods in order to conform to the input specifications of the model. The manner in which this is accomplished is described in Appendix 2, together with the computer sub-routine developed.

Potential Evaporation

Historical data on 24-hour potential evaporation, which is the second input to the rainfall-runoff model, were obtained through a U.S. Weather Bureau pan at Tanbark Flat near raingage I35 in Watershed 5. In the model, the pan figures are multiplied by 0.9 in an attempt to approximate their values to overall watershed potential evaporation rates. The error introduced by this input is insignificant if one considers the low evaporation rates prevailing during the winter months which account for most of the year's streamflow. The same potential evaporation data are used for both sequences of simulated streamflow.

Streamflow Characteristics

The greatest part of the streamflow in Watershed 5 occurs during the winter months, 92 percent, on the average, taking place during the six-month period of December to May. During this period a continuous baseflow persists and super-imposed on this are sharp flood peaks followed by a steeply, and later mildly, receding limb; a few days after the flood the recession limb joins asymptotically with the baseflow which increases accordingly after flood events. Due to the sandy nature of the soil in this area, flood peaks are not as high as one would expect. From June onwards the baseflow decreases rapidly due to transpiration from groundwater and the lower soil horizon, and evapotranspiration from the stream itself. Following winters of low precipitation the stream may dry up completely during the late summer months.

The 13-year period of streamflow records used in the study starts in October 1941 and terminates in September 1954. No records are available for the water year 1946/47. Although the mean annual runoff for the 12-year period is approximately 22 million cubic feet, runoff volumes for individual years deviate tremendously from this value, ranging from less than one million cubic feet to approximately one hundred million cubic feet. It is interesting to note that the extremely wet years are clustered together and so are most of the dry years.

At this point the assumption that the watershed system is stationary is introduced. This implies that watershed properties as described by the set of parameters and equations in the rainfall-runoff model do not change with time, at least for the period 1941 to 1954.

Calibration

For the purpose of calibrating the rainfall-runoff model and applying it to predict streamflows, the period from October 1941 to September 1954 was chosen. For this period concurrent records of mean daily flows and 24-hour precipitation and pan evaporation are available, except for a gap in the streamflow record from October 1946 to September 1947.

The period from October 1943 to September 1944 is chosen to serve as a calibration period for estimating model parameter values. The reason for choosing this particular period (1943/44, water year) instead of the first and second years in the record is the uncertainty in the accuracy of the 1942/43 streamflow record, whereby a 20-inch storm is shown to produce insignificant streamflow, while much smaller storms, under similar conditions, usually give rise to higher streamflows. The first year, 1941/42, is used as a warm-up period, which is believed to help the evenning-out of errors in the initial conditions.

The calibration year 1943/44 was a very wet year, during which the volume of recorded streamflow was as high as 74 million cubic feet as compared to the 12-year annual mean of 22 million cubic feet. As previously explained, only the wet months of December through May are considered in the calibration and prediction due to failure of the rainfall-runoff model to simulate evapotranspiration from groundwater and channel losses; both become very important during the summer months.

A set of initial parameter values are chosen in a subjective manner and several computer simulations are made, every time changing certain parameter values and visually comparing the simulated and observed streamflows on a semi-logrithmic plot. The two hydrographs on Figure 13 are the results of the final calibration runs, the solid line referring to the observed mean daily streamflow and the dashed line to the simulated streamflow for the same period. The final set of parameter values, together with short remarks on their importance are given in Table 10. Definitions

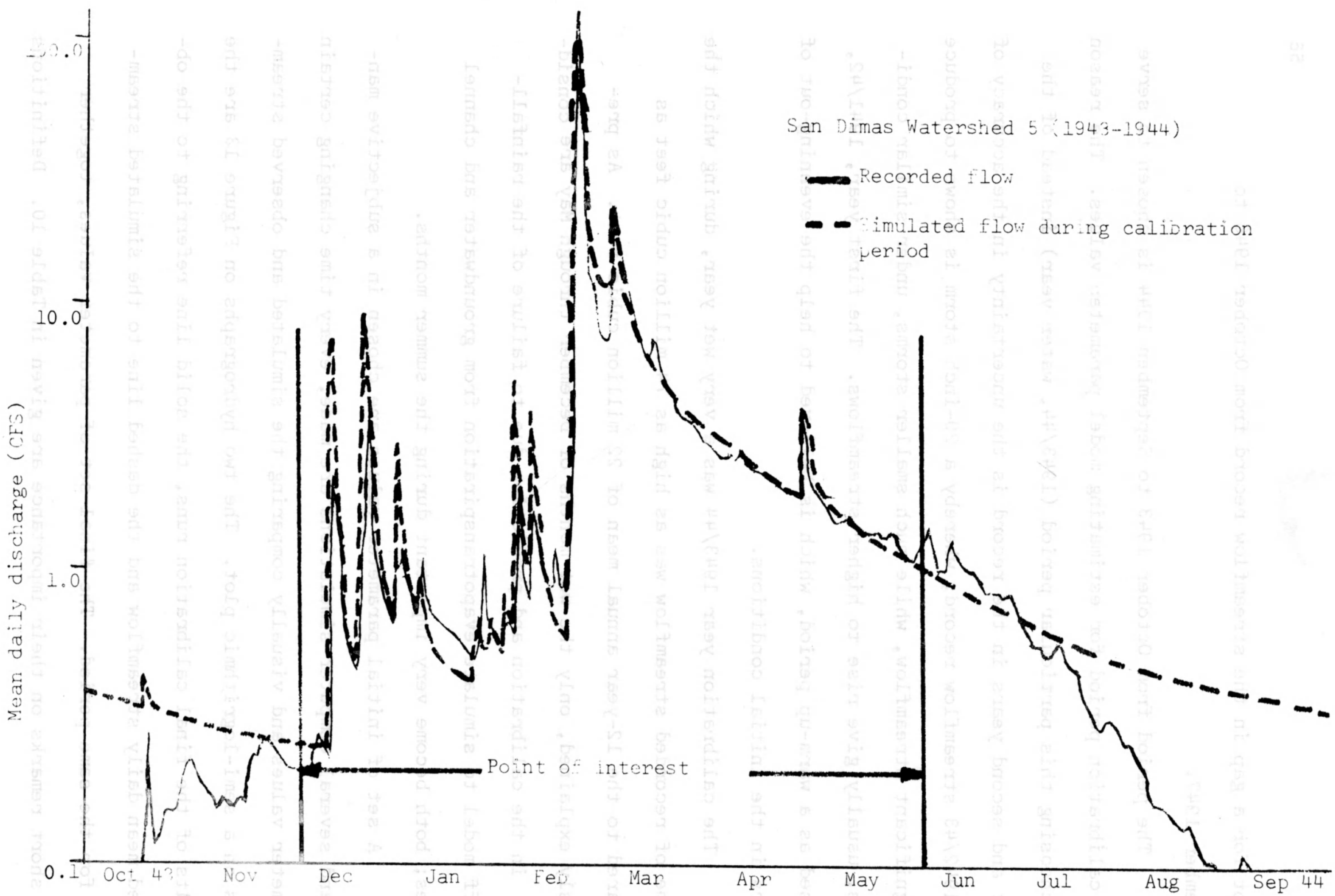


Figure 13. Observed and simulated mean daily flows for calibration period 1943-1944.

of the variables in this table are given in Appendix 1. The precipitation input to the rainfall-runoff model during calibration are the 24-hour precipitation values at raingage I35 (Appendix 2).

For the calibration period, monthly volumes of simulated streamflow are very close to the observed ones (Table 11) especially for the period of interest, December 1943 to May 1944.

Application of Model and Results

Using the same set of model parameter values, as decided upon during calibration, two streamflow sequences are generated by simulation for the entire 13-year period, beginning October 1941. Of this, only the period from October 1944 onwards is considered to be the prediction period.

First Sequence of Simulated Streamflows

The input for the first sequence is the 24-hour precipitation at raingage I35 which is also used in the calibration stage. The precipitation input for the second sequence of simulated streamflows are the 24-hour values as computed from Equation [5] (see also Appendix 2).

The results of the first sequence, together with the corresponding observed sequence, are plotted (Figure 14) in the form of monthly volumes for the period of interest, December to May. Monthly precipitation and the soil moisture at the end of every month as computed by the rainfall-runoff model are also shown in the same figure. For the first two years of prediction, immediately following the calibration period, the results are close to the observed ones, especially in the first

Table 11. Observed and simulated streamflow for the calibration period.--Monthly volumes, in million cubic feet.

	December 1943	January 1944	February 1944	March 1944	April 1944	May 1944
Observed	3.116	2.009	32.164	19.946	6.843	4.277
Simulated	5.906	1.950	32.202	20.370	7.438	4.258

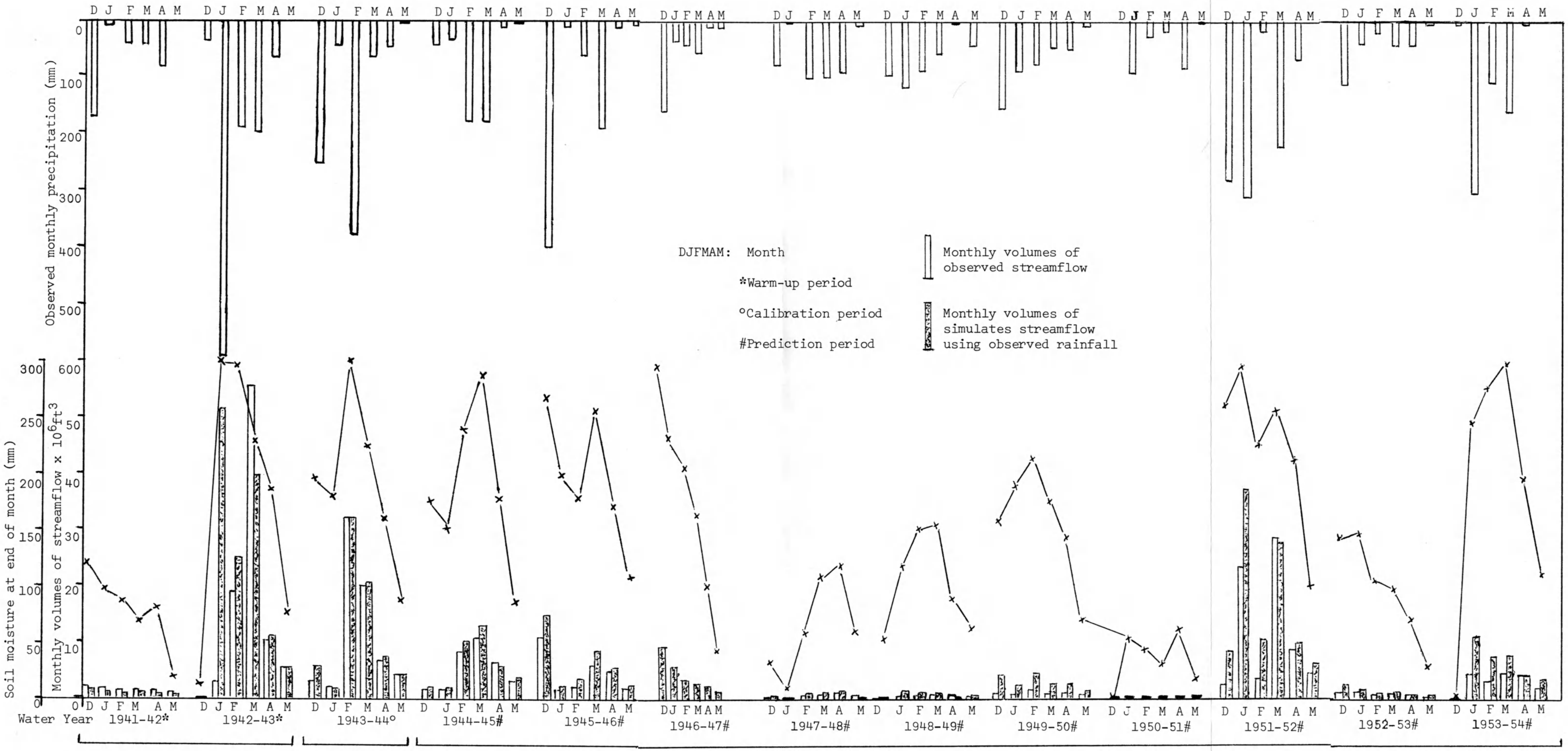


Figure 14. Monthly values of observed and simulated streamflow, precipitation and soil moisture for Watershed 5.--The precipitation is that recorded at raingage I35 and is the input to the rainfall-runoff model for obtaining the simulated streamflow. The soil moisture is in depth of water and is the current value at the end of each month as calculated by the model.

prediction year. It may be useful to point out that these two years are preceded by two extremely wet years. Following the first two years of the prediction period is a series of dry years up to, and including, 1950/51. The predicted monthly flows for these dry years are much higher than the observed, sometimes by 600 percent or more, as illustrated in Table 12. It appears that for years preceded by wet years, the prediction is better than that for years preceded by one or more dry years. Wet years, like 1951/52, following a series of dry years give the worst prediction. In almost all cases, predicted flows are on the high side. This may be partly due to the use of a very wet year for calibration.

The above observations probably suggest that the Mero model is not capable of describing the soil-water replenishment, storage and disposition in a realistic way. The model's complete depletion of soil moisture from both horizons, L_1 and L_2 (see Appendix 1), during summer, absolutely eliminates any effects on soil-water status that any particular year may have on the same parameter in following years, no matter how wet or dry this year may be. This absence of carryover effects in the soil-water status is unrealistic, especially in bare areas (Rowe and Colman, 1951), and indicates the need to improve the relevant parts of the model.

Furthermore, the net precipitation-overland flow relationship (Appendix 1), presently in the model, is the one derived for Cyprus watersheds, and is partly responsible for the high peaks and generally higher volumes in the simulated flows. The sequencing of the different processes in the overland flow and other parts of the model also appear to have some significance. For example, in the Mero model, overland

Table 12. Observed and simulated monthly flows, in million cubic
Observed precipitation at gage I35 was used for the si

	Water Year	December		January		February		
		Obs.	Sim.	Obs.	Sim.	Obs.	Sim.	Obs.
W	1941/42	2.127	2.008	1.595	1.039	1.212	0.793	1.387
W	1942/43	0.145	0.081	3.177	51.571	19.016	25.048	55.657
C	1943/44	3.116	5.906	2.009	1.950	32.164	32.202	19.946
P	1944/45	1.764	1.997	1.635	1.813	8.402	10.703	10.728
P	1945/46	10.858	14.948	1.742	2.200	2.017	3.513	5.988
P	1946/47	?	9.610	?	5.823	?	3.462	?
P	1947/48	0.157	0.460	0.122	0.108	0.671	0.919	0.883
P	1948/49	0.172	0.299	0.445	1.451	0.615	1.032	0.968
P	1949/50	0.626	4.120	0.742	2.292	1.592	4.437	0.755
P	1950/51	0.051	0.067	0.131	0.329	0.088	0.121	0.100
P	1951/52	2.149	8.160	23.106	37.414	3.694	11.083	28.479
P	1952/53	1.197	2.518	1.323	1.816	0.666	0.794	0.903
P	1953/54	0.059	0.083	4.129	10.952	2.887	7.207	4.171

Obs.: observed

Sim.: simulated

W: warm-up period

C: calibration period

P: prediction period

cubic feet, for San Dimas Watershed 5.--
he simulated series.

March	April		May	
Sim.	Obs.	Sim.	Obs.	Sim.
1.079	1.470	0.732	0.770	0.359
36.964	10.693	11.202	5.505	5.512
20.370	6.843	7.438	4.277	4.258
12.997	6.174	5.987	3.096	3.626
8.376	4.950	5.292	1.994	2.206
2.976	?	2.311	?	1.351
1.168	1.055	1.250	0.457	0.333
1.075	0.470	0.377	0.360	0.648
2.472	0.737	2.587	0.372	1.274
0.310	0.141	0.357	0.086	0.124
28.124	8.765	9.427	4.366	5.617
1.249	0.740	0.798	0.473	0.579
7.450	3.636	3.781	1.314	2.947

flow takes place first, as a function of the soil-moisture status of the previous day, and what is left is added to the soil-moisture as infiltration. Normally, in most watershed models, the reverse is the case; first infiltration losses are satisfied, and what is left is the overland flow or "rainfall excess."

In the original version of the Mero model, evapotranspiration takes place first from the lower of the two soil horizons (L_2), while the top horizon (L_1) maintains a constant soil-water status, at or below field capacity. Evapotranspiration from L_1 starts when the soil-moisture in L_2 is depleted. This sequence of soil moisture depletion, though unrealistic, gave best results for the Cyprus watersheds, probably due to fractures and joints in the rocky watersheds, aiding the roots of the coniferous forest to use water from greater depths. For the San Dimas area, however, best results are obtained when this sequence is reversed and evapotranspiration takes place first from the upper soil-horizon (L_1), while L_2 continues to be active in the transfer of soil water, if any is available, to the groundwater body. This more natural sequence is presently used in the model.

The present calibration and set of model parameter values is by no means the best that can be achieved. Additional work on small changes in the different components and relations in the model to comply more with conditions in the San Dimas area, and the use of more than one year, preferably a combination of wet, average and dry years, is expected to give a more optimum, in the long-run sense, set of model parameter values.

Nevertheless, irrespective of the large deviations from the observed values, the sequence of simulated streamflows just described is used as the basis for calculating differences in the second sequence resulting from errors in the precipitation input to the model. It should be understood that this particular rainfall-runoff model is used only as a tool to facilitate the evaluation of the watershed's response to precipitation errors. Any other watershed model may be used to serve the same purpose.

Second Sequence of Simulated Streamflows

The second sequence of simulated streamflow is, as before, obtained through the use of the rainfall-runoff model; however, the precipitation input used for this sequence contains errors when compared to the observed precipitation used to simulate the first sequence. The model parameter values are the same for both simulations and no attmept is made to recalibrate the model. In other words, what is done is similar to a sensitivity analysis of the rainfall-runoff model with respect to errors in the precipitation input (Dawdy and Bergmann, 1969). The response of the watershed to these errors is in the form of deviations of this sequence of simulated flows from corresponding values in the first sequence.

Comparison of Annual Values of the Two Sequences and Discussion of Findings

In order to save time in the evaluation of the results, the annual values of the two pairs (precipitation and simulated streamflow) of sequences are used. Although the period of interest was earlier specified to be the six-month winter period (December to May), which is important in comparing observed and predicted flows (Figure 14), the use of annual values for this comparison is logical since the comparison now is between simulated values. Furthermore, the streamflow for the summer months (June to November) accounts only for less than ten percent of the annual runoff. Table 13 lists the annual values of the two series of precipitation input and the corresponding series of simulated flows. The percent error for each of these two pairs is also given for every year. This error is positive for both the precipitation and simulated streamflow, due to the over-estimation of mean areal precipitation by the transferring of the precipitation-altitude relationship developed in another watershed.

The two error series (EP and EQ) given in Table 13 indicate that, on an annual basis, the percent error in simulated streamflow is much higher, sometimes by more than 700 percent, than the corresponding error in the precipitation input. Errors in annual values of the precipitation input range from 1.4 to 4.25 percent, while errors in simulated annual streamflows range from 3.44 to 12.48 percent. In general, the low error values in precipitation correspond to low error values in the simulated streamflow; similarly high errors of both variables are related.

A double mass plot of the two series of errors (EP and EQ), as listed in Table 13, is shown in Figure 15. The purpose of this plot is to illustrate the accumulation of error in simulated streamflow and the stabilization of this error after a long enough period (three to four years). The accumulation is attributed to the memory of the watershed system as described by the rainfall-runoff model; the stabilization of the error magnitude, after three or four years, is evidence of the

Table 13. Yearly values of the two sequences of precipitation input and the corresponding values of simulated streamflow, and the percent error for each pair of sequences.

Water Year	Precipitation Input (mm)		Percent Error	Simulated Streamflow x $10^6 ft^3$		Percent Error
	P_{I35}	$\overline{P}_c$	$\frac{\overline{P}_c - P_{I35}}{P_{I35}} \times 100$	Q_1 from P_{I35}	Q_2 from $\overline{P}_c$	$\frac{Q_2 - Q_1}{Q_1} \times 100$
1941/42	423	429	1.42	8.458	8.749	3.44
1942/43	1149	1200	4.44	139.82	150.506	8.14
1943/44	850	879	3.42	80.919	87.665	8.34
1944/45	753	779	3.46	51.897	57.584	10.97
1945/46	685	713	4.08	39.822	44.794	12.48
1946/47	700	722	3.14	39.081	43.714	11.85
1947/48	402	412	2.99	4.911	5.501	12.00
1948/49	430	438	1.86	5.336	5.847	9.60
1949/50	528	544	3.03	19.511	21.651	10.96
1950/51	291	294	1.03	1.602	1.695	5.80
1951/52	1044	1082	3.64	108.158	116.069	7.30
1952/53	393	397	1.02	11.071	11.874	7.26
1953/54	633	657	3.79	35.747	40.328	12.82

Note: P_{I35} is annual precipitation at gage I35 and is taken as representative of mean areal precipitation for Watershed 5 (Appendix 2).

$\overline{P}_c$ is mean annual areal precipitation computed from Equation [5] and data at gage I35.

Q_1 is annual simulated streamflow using P_{I35} as input to the model; Q_1 is taken as the reference for detecting errors in Q_2.

Q_2 is annual simulated streamflow using $\overline{P}_c$ as input.

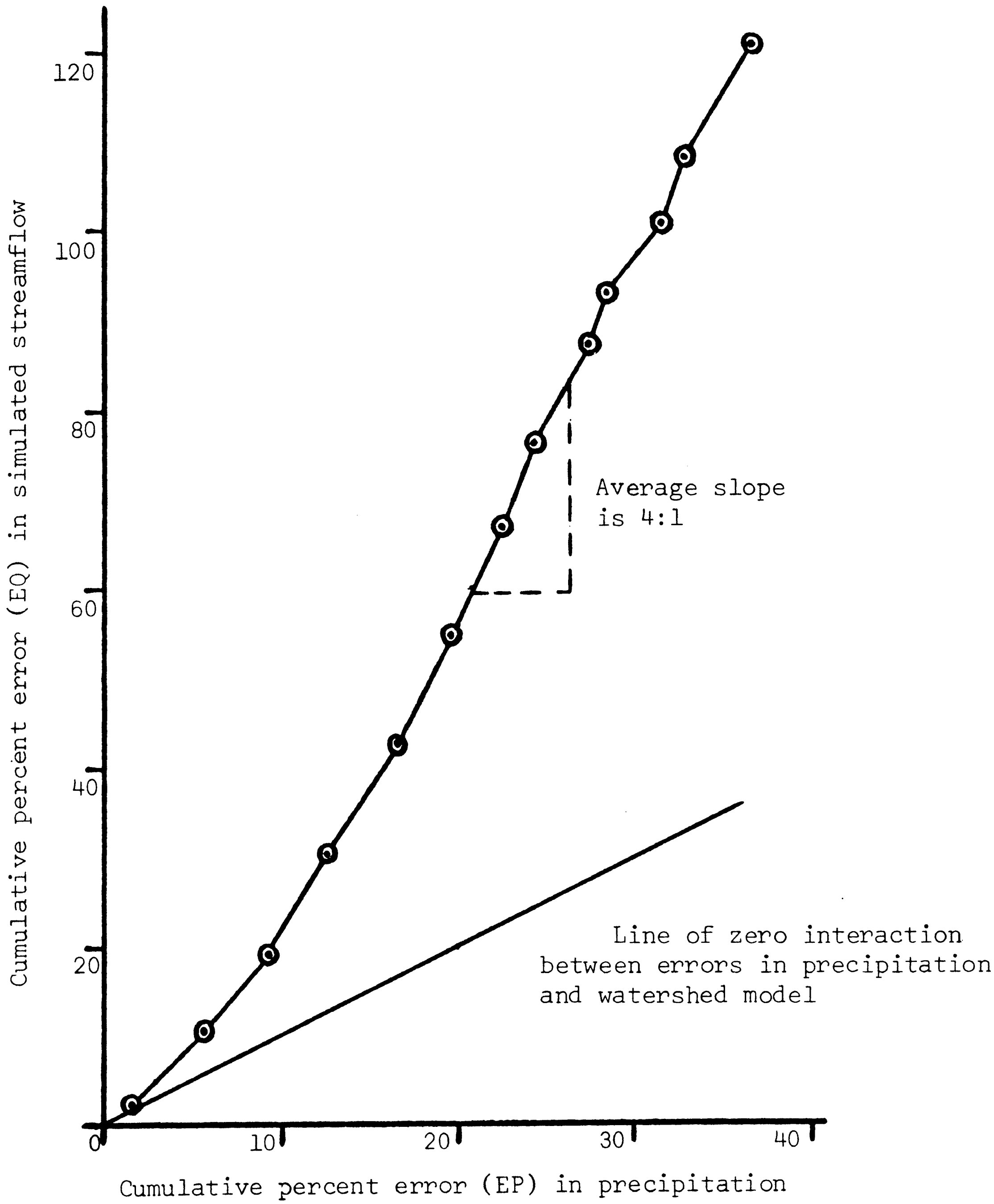


Figure 15. Double mass plot of errors in precipitation and the resulting errors in simulated streamflow.--Errors are based on annual values of precipitation and streamflow during the period 1941-1954 (water year). EP and EQ are defined in Figure 17.

effectively finite nature of this memory. In other words, the watershed system has an error propagation property with bounded cumulative effect.

In the long-run, points on the double mass curve of Figure 15 tend to deviate about a constant mean slope which has, approximately, a 400 percent gradient, or

$$\frac{\Delta(EQ)}{\Delta(EP)} \approx 4 \; . \qquad [6]$$

Between individual years (every point on Figure 15 represents a particular year), however, the error ratio, EQ/EP, varies, depending on the state (very wet, wet, dry, etc.) of the preceding year. A qualitative illustration of this concept is given in Figure 16, in which five states are identified. These states and the water years falling in each state are given in Table 14.

Figure 16(a) shows all state transitions, with their respective error ratio (EQ/EP), which are occurring in the 13-year period, 1941-1954; these are not all the possible state transitions. Figure 16(b) shows the transition from the two commonly occurring extreme states of very wet (VW) and dry (D), to a number of other states. It is interesting to note that in both cases the error ratio (EQ/EP) increases as the new state goes from very wet to dry, but it tends to increase more rapidly in going from very wet (VW) to drier states. Further work in the same direction, including the investigation of the dependence of error in any particular year on the states of two or more years, may be worth considering in a stochastic framework.

Another way of presenting the magnitude of the errors, EP and EQ, in precipitation and simulated streamflow, respectively, as related to

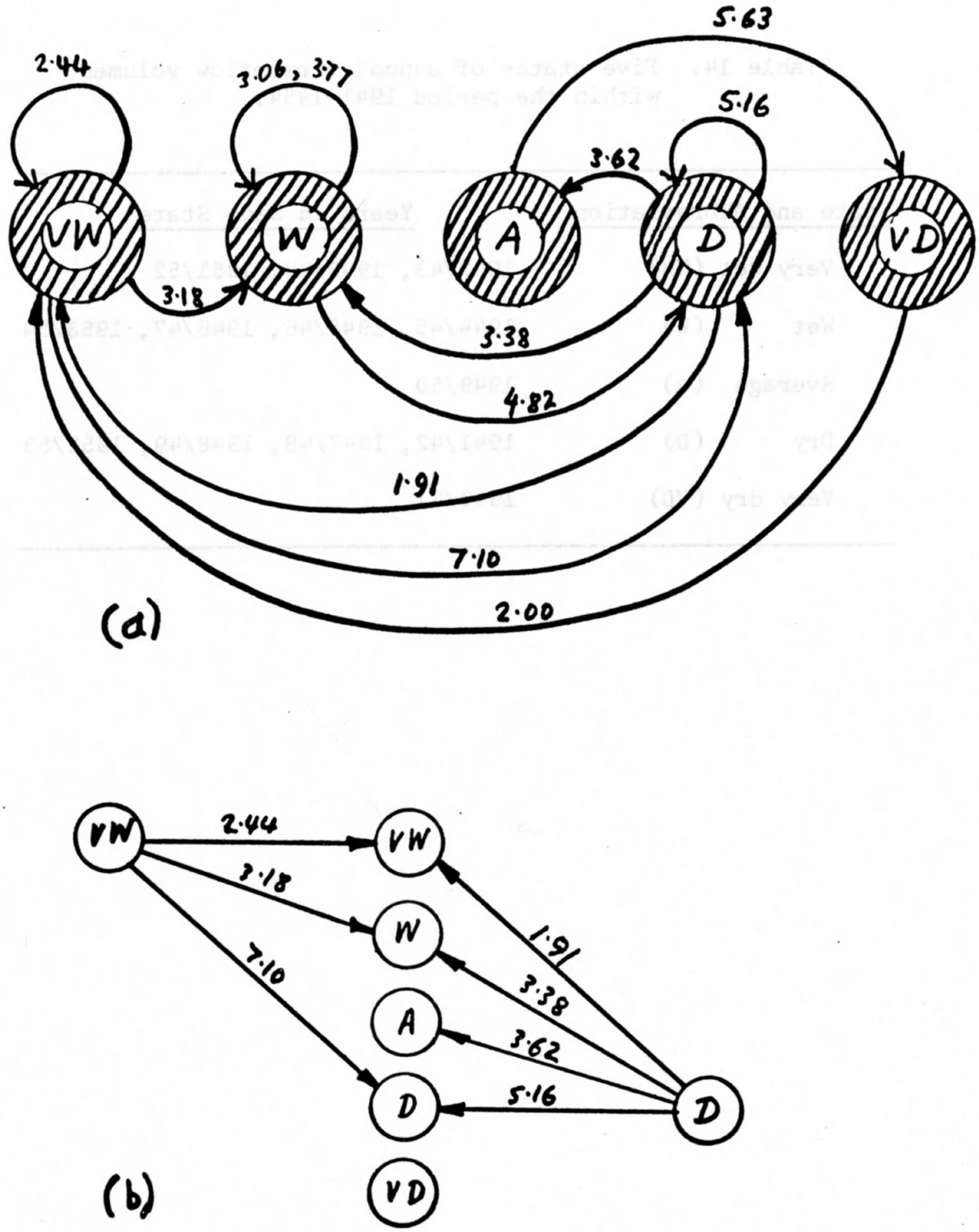


Figure 16. Schematic representation of the variation in error ratio (EQ/EP) in going from one state to a new state (see also Table 14).

Table 14. Five states of annual streamflow volumes within the period 1941-1954.

State and Abbreviation		Years in Same State
Very wet	(VW)	1942/43, 1943/44, 1951/52
Wet	(W)	1944/45, 1945/46, 1946/47, 1953/54
Average	(A)	1949/50
Dry	(D)	1941/42, 1947/48, 1948/49, 1952/53
Very dry	(VD)	1950/51

the magnitude of the event of interest, is the plot show in Figure 17. The event of interest is taken as the annual values of the first simulated streamflow series (Q_1), which is considered to be the reference streamflow sequence; these values are shown on the abscissa in million cubic feet, while values of EP and EQ are given on the ordinate.

For low values of annual streamflow (and hence precipitation), both errors EP and EQ are small and increase rapidly, particularly EQ, with increasing values of Q_1, the reference streamflow. EP reaches a maximum value of approximately 4 percent just above the mean value of Q_1, and tends to stay at this level except for the very high values of Q_1 when it again begins to climb very mildly. On the other hand, EQ increases far more rapidly, reaching its maximum value of about 13 percent around the mean value of Q_1; thence it decreases to a minimum of 7.3 percent from where it again starts climbing mildly as in the case of EP.

Obviously there appears to be some degree of a nonlinear relationship between the two errors; the large scatter of points, especially for small values of Q_1 is what one would expect considering the earlier discussion on how the magnitude of the error in any particular year is also affected by the state of the preceding year (see Figure 16).

Comparison of Monthly Simulated Flows and Discussion of Results

Following the previous discussion on the errors between pairs of annual values of the two sequences of simulated streamflow, and having in mind that the variable of interest was specified to be the monthly volumes of streamflow, a similar discussion of the errors on monthly values is presented. Table 15 lists the two series of simulated monthly

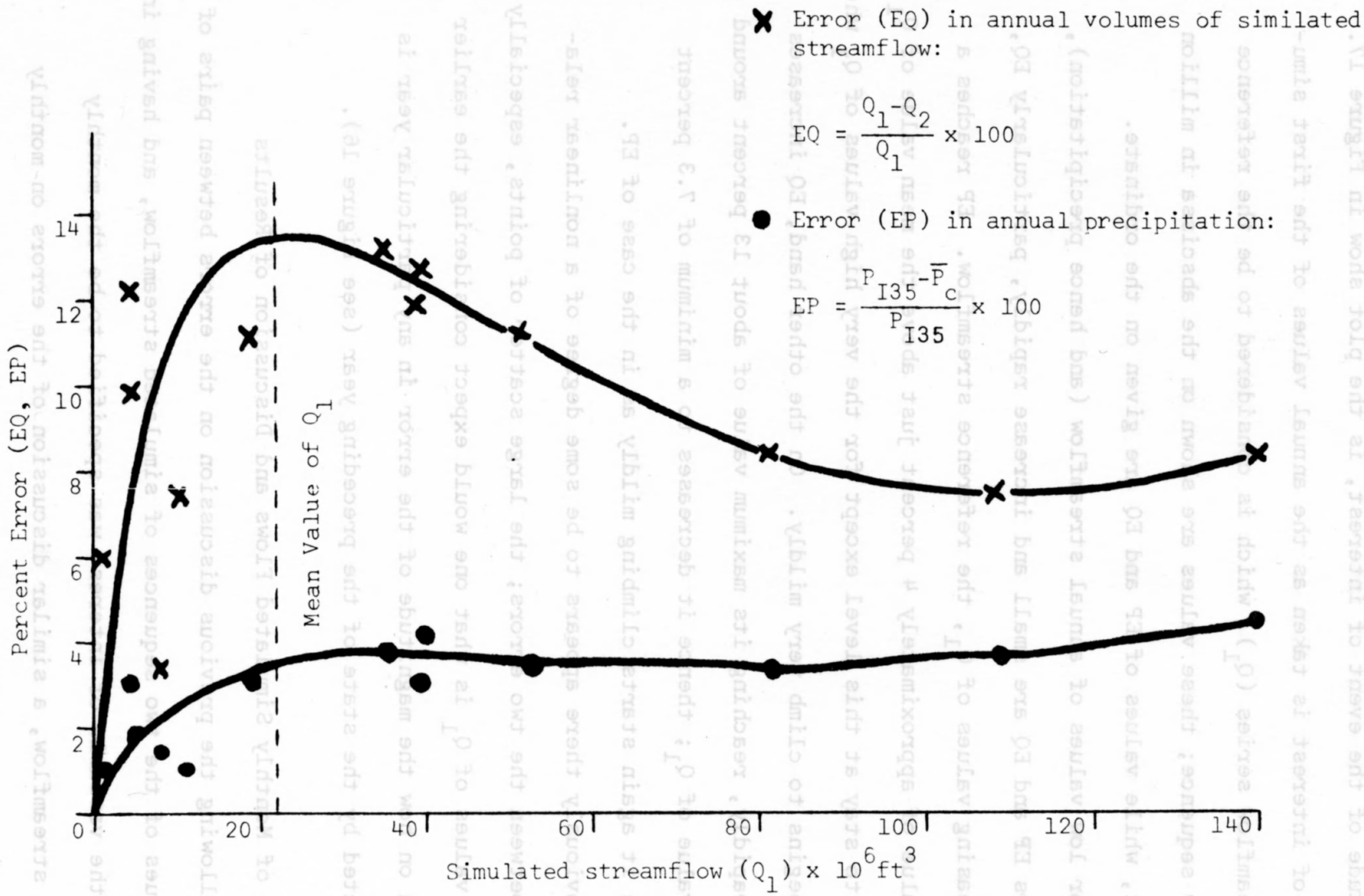


Figure 17. Distribution of input (precipitation) and output (simulated streamflow) errors.-- These errors are plotted versus Q_1 and are listed in Table 13 for the period 1941-1954. For the description of Q_1, Q_2, P_{I35} and $\overline{P}_c$ refer to Table 13.

Table 15. Monthly volumes of the two simulated series of streamflow, Q_{m1} and Q_{m2}, and percent error, EQ_m (see Equation [7]).--The precipitation inputs for simulating Q_{m1} and Q_{m2} are P_{I35} and $\overline{P}_c$, respectively; these are defined in Table 13.

	Simulated Flows				Simulated Flows		
Year/Month	Q_{m1}	Q_{m2}	EQ_m	Year/Month	Q_{m1}	Q_{m2}	EQ_m
1941/10	0.998	1.000	0.20	1945/01	1.813	2.051	13.12
1941/11	0.761	0.761	0.00	1945/02	10.073	11.156	10.78
1941/12	2.008	2.103	4.73	1945/03	12.997	14.465	11.30
1942/01	1.039	1.074	3.37	1945/04	5.987	6.600	10.25
1942/02	0.793	0.830	4.79	1945/05	3.626	3.881	7.03
1942/03	1.079	1.148	6.30	1945/06	1.764	1.913	8.45
1942/04	0.732	0.758	3.55	1945/07	1.009	1.125	11.50
1942/05	0.359	0.373	3.90	1945/08	0.787	0.885	12.44
1942/06	0.261	0.265	1.53	1945/09	0.591	0.670	13.36
1942/07	0.198	0.201	1.51	1945/10	0.445	0.511	14.80
1942/08	0.135	0.137	1.48	1945/11	0.338	0.390	15.40
1942/09	0.096	0.098	2.08	1945/12	14.948	16.861	9.83
1942/10	0.084	0.085	1.19	1946/01	2.200	2.554	16.10
1942/11	0.066	0.066	0.00	1946/02	3.512	3.973	13.10
1942/12	0.081	0.085	4.94	1946/03	8.376	9.305	11.10
1943/01	51.571	56.540	9.62	1946/04	5.292	5.850	10.55
1943/02	25.048	27.312	9.05	1946/05	2.206	2.481	12.45
1943/03	36.964	39.480	6.80	1946/06	1.219	1.381	13.30
1943/04	11.202	11.834	5.63	1946/07	0.581	0.670	15.30
1943/05	5.512	5.809	5.39	1946/08	0.411	0.475	15.56
1943/06	3.179	3.388	6.58	1946/09	0.293	0.342	16.70
1943/07	2.241	2.411	7.58	1946/10	0.369	0.416	12.72
1943/08	1.807	1.950	7.92	1946/11	11.290	12.487	10.60
1943/09	1.429	1.547	8.25	1946/12	9.610	10.925	13.70
1943/10	1.212	1.316	8.58	1947/01	5.823	6.949	19.35
1943/11	0.949	1.035	9.06	1947/02	3.462	3.779	9.15
1943/12	5.906	6.553	10.95	1947/03	2.976	3.152	5.92
1944/01	1.950	2.144	9.95	1947/04	2.311	2.431	5.20
1944/02	32.202	35.161	9.18	1947/05	1.351	1.449	7.25
1944/03	20.370	21.914	7.59	1947/06	0.709	0.787	11.00
1944/04	7.438	7.894	6.14	1947/07	0.533	0.601	12.76
1944/05	4.258	4.501	5.70	1947/08	0.384	0.438	14.05
1944/06	2.529	2.696	6.61	1947/09	0.261	0.301	15.30
1944/07	1.720	1.858	8.02	1947/10	0.187	0.216	15.50
1944/08	1.335	1.450	8.62	1947/11	0.142	0.163	14.80
1944/09	1.050	1.145	9.04	1947/12	0.460	0.518	12.60
1944/10	0.887	0.970	9.35	1948/01	0.108	0.124	14.80
1944/11	10.367	11.611	12.00	1948/02	0.919	1.061	15.47
1944/12	1.997	2.258	13.08	1948/03	1.168	1.268	5.95

Table 15--Continued

Year/Month	Q_{m1}	Q_{m2}	EQ_m	Year/Month	Q_{m1}	Q_{m2}	EQ_m
	Simulated Flows				Simulated Flows		
1948/04	1.250	1.371	9.70	1951/09	0.016	0.018	12.50
1948/05	0.333	0.382	14.70	1951/10	0.073	0.081	10.95
1948/06	0.163	0.192	17.76	1951/11	0.400	0.445	11.23
1948/07	0.084	0.095	13.10	1951/12	8.160	9.045	10.80
1948/08	0.053	0.061	15.10	1952/01	37.414	40.620	8.56
1948/09	0.043	0.048	11.06	1952/02	11.083	11.950	7.83
1948/10	0.034	0.042	13.50	1952/03	28.124	29.824	6.05
1948/11	0.031	0.034	10.65	1952/04	9.427	9.884	4.85
1948/12	0.299	0.317	6.02	1952/05	5.617	5.856	4.25
1949/01	1.451	1.589	9.52	1952/06	3.079	2.242	5.30
1949/02	1.032	1.126	9.10	1952/07	1.975	2.109	6.79
1949/03	1.075	1.161	7.93	1952/08	1.573	1.686	7.18
1949/04	0.377	0.433	14.85	1952/09	1.232	1.326	7.63
1949/05	0.648	0.721	11.25	1952/10	1.018	1.100	8.05
1949/06	0.195	0.215	10.25	1952/11	1.335	1.455	8.99
1949/07	0.087	0.094	8.05	1952/12	2.518	2.668	5.97
1949/08	0.058	0.064	10.32	1953/01	1.816	1.918	5.51
1949/09	0.047	0.051	8.52	1953/02	0.794	0.862	8.57
1949/10	0.041	0.044	7.33	1953/03	1.249	1.337	7.05
1949/11	1.125	1.265	12.45	1953/04	0.798	0.860	7.77
1949/12	4.120	4.666	13.26	1953/05	0.579	0.626	8.13
1950/01	2.292	2.500	9.08	1953/06	0.349	0.379	8.60
1950/02	4.437	4.866	9.67	1953/07	0.274	0.298	8.76
1950/03	2.472	2.746	11.10	1953/08	0.197	0.215	9.13
1950/04	2.587	2.850	10.15	1953/09	0.143	0.155	7.75
1950/05	1.274	1.408	10.50	1953/10	0.116	0.125	7.75
1950/06	0.596	0.666	11.73	1953/11	0.100	0.107	7.00
1950/07	0.260	0.294	13.08	1953/12	0.083	0.089	7.23
1950/08	0.182	0.205	12.62	1954/01	10.952	12.168	11.10
1950/09	0.125	0.139	11.20	1954/02	7.207	8.057	11.70
1950/10	0.094	0.104	10.63	1954/03	7.450	8.389	12.60
1950/11	0.116	0.124	6.90	1954/04	3.781	4.615	22.00
1950/12	0.067	0.073	8.95	1954/05	2.947	3.272	10.85
1951/01	0.329	0.348	5.78	1954/06	1.545	1.702	10.15
1951/02	0.121	0.128	5.78	1954/07	0.754	0.855	13.40
1951/03	0.310	0.319	2.90	1954/08	0.475	0.551	16.00
1951/04	0.357	0.375	5.05	1954/09	0.338	0.396	17.20
1951/05	0.124	0.131	5.65				
1951/06	0.026	0.028	7.70				
1951/07	0.023	0.025	8.70				
1951/08	0.020	0.021	5.00				

volumes of flow, Q_{m1} and Q_{m2}, corresponding to the previously discussed annual series, Q_1 and Q_2, respectively; the percent error between corresponding monthly values is also given in the same table.

The error between paired values of the two series is always positive, implying that Q_{m2} is always larger than Q_{m1}. The percentage error, EQ_m, is calculated as follows:

$$EQ_m = \frac{Q_{m2} - Q_{m1}}{Q_{m1}} \times 100 \qquad [7]$$

where Q_{m1} is the reference monthly volume of simulated flow,

and Q_{m2} is the corresponding volume of the second series of simulated monthly flow.

The reason why the error is one-sided (always positive) is the over-estimating of the precipitation input to the rainfall-runoff model which in turn imposes errors in the simulated streamflow (the output). The constant over-estimation of all storms (Table 12) greater than one inch, is a result of the transferring of the precipitation-altitude relationship, in the form of Equation [5], to Watershed 5 alone, and using raingage I35 as a base station. In comparison, better estimates (Tables 8 and 9, Chapter 3) were obtained for the grouped Watersheds 2, 3, 4 and 5 by applying the same equation and using the same base station. This is in support of the previous argument that errors in estimating mean areal precipitation for this type of cyclonic storms are attenuated by an increase in the area under consideration. Nothing, however, is known about the size and sign of the error in the observed "actual" mean areal precipitation.

Errors in monthly precipitation (EP_m) range from zero to approximately 5.5 percent of the observed values; these errors are not substantially different from those of the annual precipitation values in which the maximum error is approximately 4.0 percent. Contrary to the proximity of the two precipitation errors just described, errors in simulated monthly streamflows are as high as 22.0 percent, as compared to the maximum error for the annual volumes which is only 12.8 percent. The maximum monthly error of 22.0 percent is for the month of April 1954. No consistent relation seems to exist between the size of error and the size of the event (monthly volumes of streamflow, Q_{m1}). This is evident of the strong dependence of the streamflow volume for any particular month on the combination of the states ("state" is used in the same context as for the annual flows discussed earlier) of several preceding months; this dependence appears to be so high that for most cases it completely obscures the direct influence of precipitation errors on the corresponding simulated streamflow. In addition, the propagation of error in monthly simulated streamflow is more obvious and well-defined than in the case of annual values. The plot in Figure 18(a) illustrates the distribution of error for the monthly volumes in the 13-year period of simulated streamflow, Q_{m2}. The nine non-zero-error points in Figure 18 approximate a straight line on normal probability paper.

The plot in Figure 18(b) is a cumulative plot of the error shown in Figure 18(a). For 56 percent of the months in the 13-year period the error is less than 10 percent and it is less than 15 percent for more than 90 percent of the time. It should be emphasized again that the

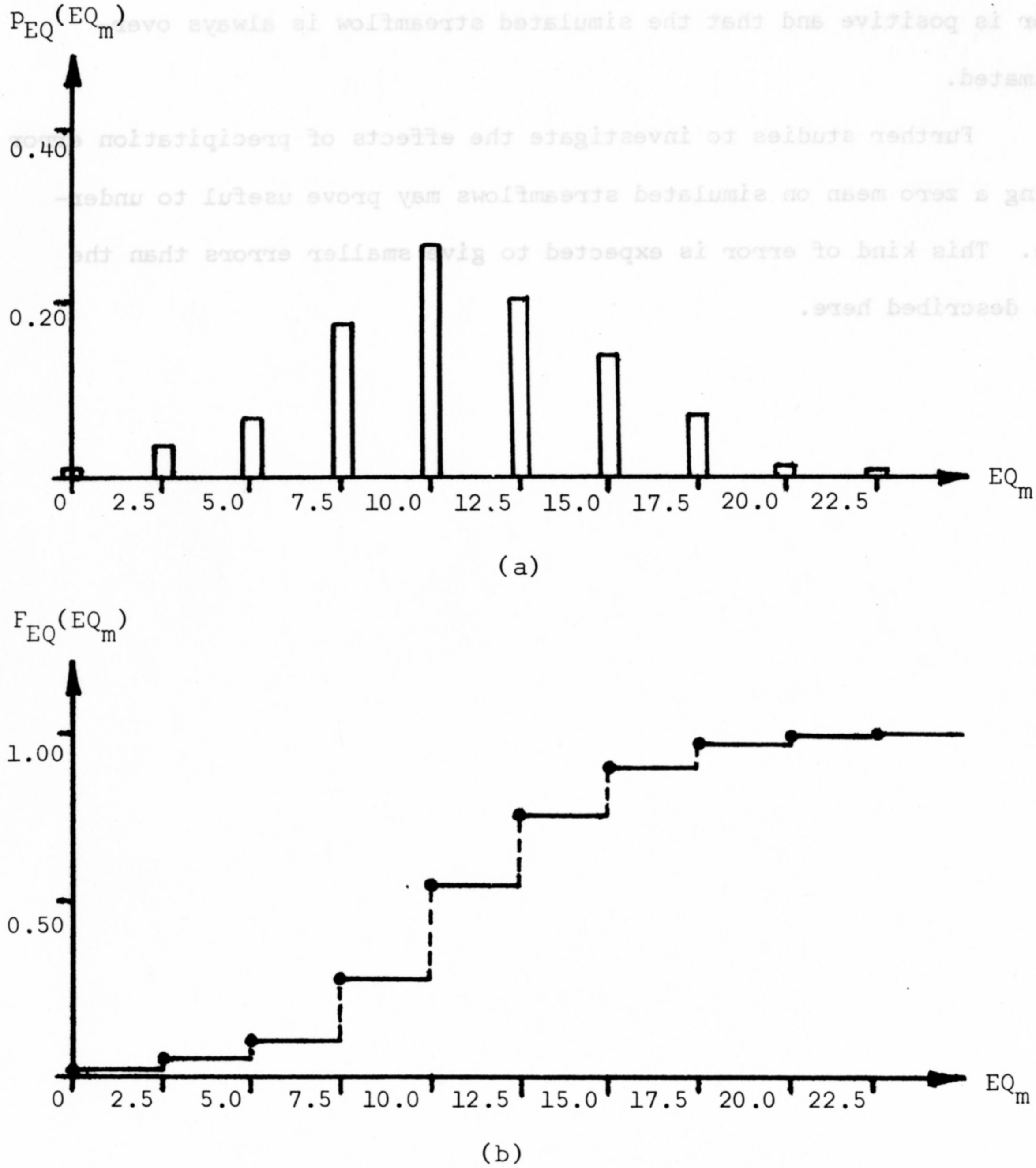


Figure 18. Distribution of error (EQ_m) for the simulated monthly streamflow series Q_{m2}.--(a) Probability mass function; (b) Cumulative distribution function.

error is positive and that the simulated streamflow is always overestimated.

Further studies to investigate the effects of precipitation error having a zero mean on simulated streamflows may prove useful to undertake. This kind of error is expected to give smaller errors than the ones described here.

CHAPTER 5

CONCLUSIONS

The influence of altitude on precipitation amounts in the San Dimas area is very obvious. This influence is more well-defined for storms exceeding one inch of rainfall; for storms smaller than one inch, the influence, if any, is obscured by the larger spatial variability of small storms and the errors of measurement.

In the present study, a precipitation-altitude relationship for storms greater than one inch is derived for two small watersheds combined. Under the assumption of linear increase of precipitation with altitude, this relationship is generalized to apply to any altitude in the area. Further generalized, the same relationship is found to describe storm precipitation in the San Dimas area with different degrees of accuracy, depending on the area of the watershed, the base raingage used and the degree of spatial lumping of the precipitation to be computed.

Three different applications of the precipitation-altitude relationship are examined and the errors evaluated.

In the first case this relationship is applied to the 41 storms used for its derivation to compute corresponding values of average elevational storm precipitation for the elevations of 3100, 4100 and 5100 feet in the two combined watersheds. The standard error of estimate (SEE) for these three elevations is 9.6, 12.0 and 11.7 percent of the mean, respectively (see Table 4). Although the precipitation-altitude relationship is based on storm data for the elevation of 5100 feet, the proximity of the three values of SEE for the three elevations is in support of the assumed linearity in the precipitation-altitude relationship.

In the second case, an attempt is made to regionalize the relationship by grouping together four watersheds and using the same base raingage as in the first case. The standard error of estimate for the same 41 storms in computing mean areal storm precipitation is 10.6 percent of the mean. For individual points in the group of watersheds, however, the SEE is much higher (up to 17 percent). This is indication of the tendency of spatial lumping to have an attenuation effect on the error.

In the third case, the transferability of the precipitation-altitude relationship is tested by applying it to the data from another raingage which is used as the base station. For the same 41 storms and the same group of watersheds as in the second case, the standard error of estimate in computing mean areal storm precipitation is 16.7 percent. For individual points, values of SEE are much higher. In using the same relationship and base station to compute mean areal precipitation for a smaller watershed, precipitation is consistently over-estimated (see Table 13). This indicates that raingage I35, at 2700 feet, which is used as base station, receives higher precipitation than the average precipitation for the elevation of 2700 feet over the grouped watersheds. Assuming that measurement and sampling errors are insignificant, this phenomenon may be attributed to the different watershed topography and orientation. When used as an input to a rainfall-runoff model, this precipitation causes consistent over-estimation of streamflows and induces a cumulative propagation of error.

For evaluating the effect of errors in precipitation, when the latter constitutes the input to a watershed model, a modified version of Mero's watershed model (Appendix 1) is used, and the error in the simulated streamflow is examined, both on an annual and a monthly basis. The following conclusions, with respect to these effects, are reached:

1. In considering annual volumes of precipitation and the respective simulated streamflow, errors in the latter are, in the long-run, four times as high as the long-term annual precipitation error (Figure 15). Errors in annual precipitation range from 1.02 to 4.44 percent, while errorrs in simulated annual streamflow range between 3.44 and 12.82 percent. The standard error of estimate for the yearly values is 12 percent of the mean, as compared to 3.7 percent for the respective annual precipitation values. On a monthly basis, though the errors in precipitation depths are not substantially higher than those for the annual values, the errors in simulated streamflow are as high as 22.0 percent, compared to the highest of 12.8 percent for the annual runoff. This is an indication of the strong dependence that exists in the streamflow volumes of consecutive months, and the high sensitivity of the watershed model to errors in the precipitation input. All the errors mentioned above are on the high side as a result of the consistently over-estimated precipitation input to the model.

2. Because of the consistently over-estimated precipitation, a propagation of error in the simulated streamflow is evident; this error

propagation is characterized by a bounded cumulative property which causes the propagation to attain stability after three or four years (Figure 15). This stabilization period represents the practically effective memory of the watershed model. The use of other watershed models may give different results; in a qualitative sense, however, all models are anticipated to give similar results.

3. The effect of precipitation error during a particular year on errors in simulated streamflow for the same year depends on the state of the preceding year. This effect becomes more pronounced if the year for which the simulation is performed is dry (low precipitation year) and less pronounced if it is a wet year (Figure 16). The same is true but in a much stronger sense with monthly simulated streamflow; in this case the combination of states of several preceding months becomes very important in influencing the error in simulated streamflow for a particular month. In fact, this is so important that, in most cases, it obscures completely the direct effect of precipitation on simulated streamflow errors for the same month.

4. The error in the simulated monthly streamflow, which is the variable of interest, is less than 10 percent for 56 percent of the time, and is less than 15 percent for over 90 percent of the time. As stated earlier, this error is consistently on the high side. Dawdy and Bergmann (1971), found a standard error of estimate of approximately 20 percent in the prediction of isolated peak discharges by using precipitation containing errors; they do not give any information as to the differences between the different precipitation inputs describing the

same event. The Mero watershed model gives results comparable to those obtained by Shanholtz and Lillard (1971), by applying the Stanford model on agricultural watersheds in Virginia. For these two watersheds, the absolute error in simulated monthly streamflow was 50 percent or less for 45 and 86 percent of the time, respectively. For the period of interest (December to May), in the San Dimas area, the Mero model gave an absolute error of 50 percent or less (Table 12) for approximately 60 percent of the time over the prediction period 1944 to 1954.

APPENDIX 1

THE RAINFALL-RUNOFF MODEL

The Mero watershed model was first developed in Cyprus during 1967 by F. Mero, hydrologist with "TAHAL" Israel, in co-operation with hydrologists from the Cyprus Water Planning Project and the Department of Water Development, Cyprus. The Cyprus Water Planning Project which was a United Nations Special Fund Project with FAO experts and Cypriot counterparts had undertaken the assessment of the land and water resources of Cyprus for an overall island-wide water development plan. Following the over-exploitation of groundwater, greater emphasis was placed on surface water resources, that is riverflows, which exhibit a higher variability both with time and space. The then existing streamflow records were very short, up to six years, and of disputable quality, and were far from adequate to describe the long-term behavior of streamflows.

Fortunately, records of daily precipitation were available for several stations in the area of interest, since 1916. Since the main objective was to make an inventory of the water resources of the island, especially in the areas of high precipitation and runoff, it was decided that knowledge of the distribution of 10-day volumes of flow would be adequate. Considering the objective and data and other limitations it was decided to construct a watershed model which would be capable to accept 24-day precipitation as input and compute the mean flow for a similar period. Volumes of computed 10-day flows were required to be within say 15% of actual, even though daily flows would not meet this requirement for some of the time.

Thus, the adoption of the Mero watershed model for extending the record of 10-day flows in some watersheds in the island of Cyprus may be attributed mainly to the following:

1. Data available - 24-hour precipitation and potential evaporation did not permit the use of a more sophisticated model, operating on time intervals shorter than one day.
2. Objectives of study - 10-day volumes of flow were adequate. No flood studies necessary at this stage.
3. Fund limitations - could not afford to hire larger and more expensive computer in another country, say Rome, Italy.
4. Time limitations - the job had to be completed within a specific period of time.
5. Computer limitations - the available computer was of too small a capacity to accept a long program of a sophisticated model.

Description of Model

Mero's hydrometeorological water balance model computes mean daily streamflow given inputs of daily precipitation, evaporation, initial soil moisture (two zones), initial springflow or groundwater flow storage. The various components of the model will be described with appropriate equations and a summary set of tables to present the use of the model.

There are several assumptions required to understand and simplify the model operation for the given input data:

1. Physical
    a. All parts of the watershed contribute to storm runoff.
    b. No evaporation-transpiration occurs from the groundwater body.
    c. Evaporation from interception storage occurs at the potential rate.
    d. The watershed(s) is(are) small enough to allow concentration of storm runoff at the outlet on the same day.
    e. All precipitation falls in the form of rainfall.
2. System
    a. The watershed is stationary in time.
    b. Linearity with respect to L/LST

In addition, certain initial conditions are assumed to be present. These include:

U (interception storage moisture) = 0

L_2 (soil moisture between field capacity and sat.) = 0

S_2 (initial flow from shallow springs) = 0

STPR (runoff from net precipitation, routed to overland flow) = 0

INMM (interflow storage) = 0

Components of Model

Interception

Before any infiltration or overland flow takes place the interception storage, U, must be fully satisfied to its maximum value, UMAX.

Interception storage includes interception by vegetation and that part of precipitation which is retained by depressions and the land surface, and is available for evaporation at the prevailing potential rates.

UMAX can be estimated by trial and error and checked by changing the values of UMAX until the stream flow satisfactorily fits the observed hydrograph, especially for rainfalls occurring after dry periods of several days. The assumption is made that UMAX does not vary throughout the year. This assumption, while valid for conifer forests, will not hold for deciduous forests. Minor modification of the program will enable use of a variable UMAX for the months desired. Agricultural watersheds should be handled on a variable UMAX, especially if different crops are cultivated or cultivation does not extend throughout the year. Values of UMAX used in Cyprus ranged from 10 to 25 mm and in the San Dimas Watershed in California from 12 to 18 mm.

Net Precipitation (PN)

Net precipitation refers to the precipitation remaining after both daily potential evaporation (EVP) and the interception storage (U) are satisfied,

$$PN = P - EVP - (UMAX - U) \tag{1.1}$$

where PN = net precipitation

P = actual precipitation

EVP = potential evaporation

UMAX = maximum value of interception storage (same as UST)

U = current value of interception storage at the beginning of the observation period (previous day).

From the equation given above it is obvious that PN can be negative, which means that the evaporation demand for the day is not met by P and/or U. In this case evaporation takes place at the expense of soil moisture, as will be explained below.

Soil Moisture Storage (L)

The "active" soil layer has been divided into two soil-water storage zones. The upper soil zone L_1 which may be thought of as the root zone and in which soil-moisture can reach a maximum value up to field capacity LFC, from which only evapotranspiration can take place, and the lower zone L_2 which receives moisture from above when field capacity is exceeded in L_1. Both evapotranspiration and groundwater recharge are assumed possible from L_2. The model evaporates and transpires water from L_1 first and then from L_2 if evapotranspiration is not satisfied by the interception storage.

The soil moisture capacity of both zones, L_1 and L_2, is assigned a maximum value, LMAX, and when this is attained, any moisture added to the soil through infiltration is considered direct recharge to groundwater.

Addition to soil moisture is, although in an indirect way, a function of current soil moisture conditions, LMAX, and net precipitation PN. In the case of PN > 0, the amount of water added to the soil-moisture storage is what is left after overland flow and interflow are subtracted from PN, these being functions of PN, soil moisture and LMAX. For details refer to the section on overland flow and interflow.

Summarizing the notation used in soil moisture storage:

L_1 = soil moisture in upper soil zone, $WP \leq L_1 \leq LFC$

WP = wilting point

LFC = field capacity

L_2 = soil moisture in lower soil zone

LMAX = maximum soil moisture in both upper and lower soil zones

$$L = L_1 + L_2. \qquad [1.2]$$

The boundary between the two soil zones has no physical significance, nor are the values of LFC and LMAX based on actual field measurements. However estimates of the soil parameters based on actual data might be desirable. The two-soil-zone system is rather a conceptual model for the sake of flexibility in dealing with allocation and disposition of soil water, and a large degree of subjective judgement is involved in the estimation of soil-parameter values.

Overland Flow and Interflow

These two components of streamflow are considered as functions of PN and L/LMAX and are computed first, before any additions to L are effected. Overland flow itself is also a function of the infiltration characteristics of the soil.

In the absence of infiltration data, and due to the use of a time period of 24 hours which makes infiltration calculations over this interval almost meaningless, an empirical relationship may be developed between net precipitation (PN) and overland flow by plotting the 24-hour PN versus the resulting observed overland flow for the same period in the form of a scattergram (Figure 1.1). It may be more advantageous to use

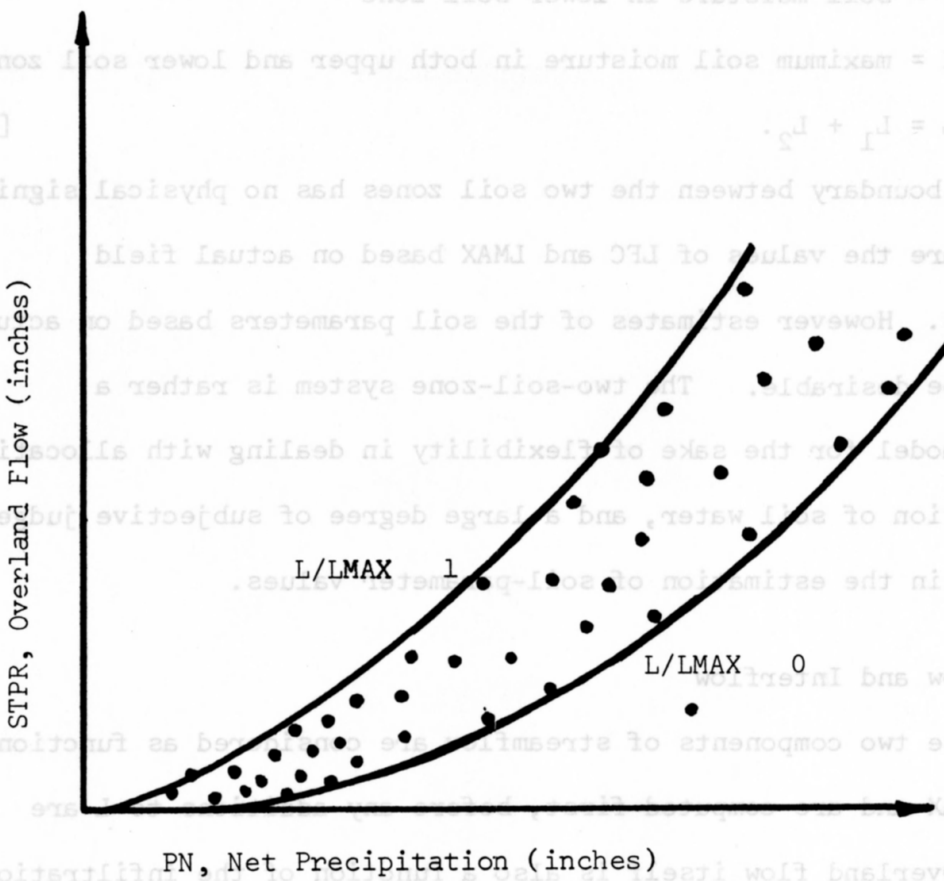


Figure 1.1 Scattergram for PN vs. STPR.

well-defined storms and the corresponding overland flow irrespective of duration time; in this way errors in determining quantity of flow due to a certain storm are avoided. Storms should be chosen to cover a wide spectrum of soil moisture conditions, ranging from storms following long dry periods with minimum soil moisture to storms during the rainy season preceded by long wet periods when soil moisture has reached its maximum. Two smooth enveloping curves may be drawn to include, if possible, all points in the scattergram, and the assumption follows that these curves relate to minimum and maximum soil-moisture conditions respectively. Intermediate soil-moisture conditions are given by drawing curves between the two extremes by linear or non-linear interpolation, thus covering all possible cases. This family of curves may be expressed in a mathematical form for use with a computer.

For demoting soil moisture status the dimensionless quantity L/LMAX is employed. Thus, for the watershed

$$STPR = f(PN, L/LMAX) \qquad 0 < L/LMAX \leq 1.0 \qquad [1.3]$$

where STPR = that part of PN which will be routed to overland flow.

Data from Cyprus showed that the following relationships for STPR worked well:

For $PN \leq 10$ mm,

$$STPR = (-0.32 + 0.07PN + 0.05PN(PRST - 0.5) - Q_o)\ CT \qquad [1.4a]$$

For $10 < PN \leq 40$ mm,

$$STPR = (-0.45 + 0.06PN + 0.0025PN^2 - 0.00001PN^3 + 0.25(PN - 8.0)(PRST - 0.5) - Q_o)\ CT \qquad [1.4b]$$

For $PN > 40$ mm,

$$STPR = (1.43 - 0.039PN + 0.0032PN^2 - 0.000003PN^3 + 0.25(PN - 8.0)(PRST - 0.5) - Q_o)\ CT \qquad [1.4c]$$

where PN = net precipitation

$STPR$ = that part of PN which will be routed to overland flow

$$PRST = L/LMAX \qquad [1.5]$$

CT = "tilt" coefficient in PN-STPR relationship.

The value of STPR, which is in depth of water in mm is added to STMM, the surface storage which is routed into overland flow.

Interflow is that component of streamflow which occurs during and shortly after overland flow. It may be defined as flow reaching the stream through loose surface soil and talus material, through cracks and joints, under mulch and as outflow from the top soil at the contact with a less permeable underlying layer where this contact is exposed.

As in the case of overland flow, interflow is assumed to be a function of L/LMAX and (PN - STPR). The interflow part of PN is added to the interflow storage (INMM) before being routed to contribute to the total hydrograph.

Addition to the interflow storage, INMM, occurs in the form of GPR1, which is described by the following equation:

$$GPR1 = (PN - STPR)(0.01 + 0.05\ L/LMAX) \qquad [1.6]$$

This relation is different from the one given in the original Mero model.

Groundwater

Groundwater is conceived as a storage reservoir the recharge of which is affected in two possible ways operating simultaneously at maximum soil-moisture.

In the first case when the soil moisture in the two soil zones is less than LMAX, the only contribution to groundwater, GPR2, comes from L_2 when $L_2 > 0$, and is taken as a function of $L_2/(LMAX - LFC)$, i.e.,

$$GPR2 = (L_2)\ L_2/(LMAX - LFC)\ CL_2\ ND/DM \qquad [1.7]$$

where $(CL_2)(ND/DM)$ = arbitrary delay function controlling the recharge rate from L_2.

$1.0 \leq CL_2 \leq 0.001$ (arbitrary)

ND = number of days since LFC is exceeded

DM = maximum delay time in days.

<u>and</u> when $L \geq LMAX$, then $ND/DM = 1.0$

Since the model requires that L_2 continues to contribute to groundwater as long as $L_2 > 0$, this implies that L_2 is a conceptual storage, its values ranging between field capacity and saturation.

In the second case, when L = LMAX, any further addition of moisture to L is automatically transferred to the groundwater storage without any delay, i.e., ND/DM = 1.0. The groundwater storage is in mm of water as in the case of surface storage, STMM, and interflow, INMM.

The model does not predict groundwater levels, but the groundwater storage reservoir is useful in routing of springflow contributions to the total hydrograph where such exist.

Contribution of the Different Storages to Streamflow

The storages considered are those for:

1. Overland flow (STMM)
2. Interflow (INMM)
3. Persistent springs (S_1MM) } Groundwater
4. Temporary springs (S_2MM)

The outflow from a storage reservoir is described by an equation of the form

$$Q(t) = Q_o e^{-t/t_o} \qquad [1.8]$$

where $Q(t)$ = outflow as a function of time

Q_o = initial outflow at $t = 0$

t_o = time constant (same units as t) characteristic of the depletion of the reservoir

t = time.

It follows from the above equation that

$$\ln Q(t)/Q_o = -t/t_o \qquad [1.9]$$

which implies that a straight line is obtained when plotting ln Q(t) versus t.

Semi-log plots of the recession part of observed streamflow hydrographs, however, rarely occur as straight lines, except at the very extreme part of the recession several days or weeks after recharge of any kind ceases to take place. The recession parts of hydrographs can be analyzed and the total recession hydrographs may each be described as follows by the sum of a number of straight lines on a semi-log plot, each representing outflow from a single reservoir.

$$Q(t) = \sum_{i=1}^{n} Q_{oi}\, e^{-t/t_{oi}} \qquad [1.10]$$

In this case t = 1 and n = 4 which means that the total hydrograph, Q(t), is made up of four components as mentioned above.

It is preferable that parts of the hydrograph chosen for the above analysis be free of any addition to groundwater from soil moisture or elsewhere.

Each day $Q_i(t)$ is subtracted from the respective storage and

$$Q_i\,(t+1) = Q_i\,(t)\, e^{-1/t_{oi}} \qquad [1.11]$$

The area of the watershed, all of which is normally assumed to contribute to overland flow storage, is denoted by A_4. Areas contributing to the other storages are:

A_3 = interflow storage area

A_1 = persistent springflow area

A_2 = temporary springflow area.

If all the water infiltrating and feeding these three storages eventually appears as flow in the stream, then

$A_1 + A_2 + A_3 = A_4$.

In the case of no base flow or interflow the respective A_i may be set equal to zero. Further, when

$A_1 + A_2 + A_3 < A_4$, then there exists a "losing" watershed,

and when

$A_1 + A_2 + A_3 > A_4$, then there exists a "gaining" watershed.

Each Q_i is converted to mean daily flow in cubic meters per second (cms) when multiplied by A_i and a time conversion factor.

The Total Hydrograph

As mentioned previously the total hydrograph, $Q(t)$, is the sum of the four components referred to above.

No stream routing is included in the model, as the routing is assumed to be automatically taken care of in the routing of the four storages and in the attempt to fit computed to observed hydrographs at the outflow point of the watershed.

Evaporation and Evapotranspiration

No provision is made in the model for the calculation of potential evaporation rates. These rates, in parallel to the precipitation, are part of the input and may be previously obtained by any acceptable method to suit the climatic conditions of the area under study. In using pan evaporation data, monthly coefficients must be supplied for converting pan to field potential evaporation.

Evaporation from interception and precipitation is assumed to take place at potential rates and the precipitation is added to the current interception storage before subtracting the potential evaporation for the day.

Evaporation from soil moisture commences when the current interception storage is zero, i.e., when $U = 0$, and is a function of the current moisture in the soil. Transpiration is not treated separately but is included in the evaporation term.

Evaporation occurs at the expense of the total soil moisture in storage, L, with L_1 being depleted first.

$$E = EVP\ (L/LMAX)^{EC} \tag{1.12}$$

where E = actual evaporation

EVP = potential evaporation

EC = constant governing evaporation rates with L/LMAX.

Figure 1.2 shows the relationship of L/LMAX to $(L/LMAX)^{EC}$. When EC = 1.0, E is linearly related to L.

In choosing the value of EC, the presence, type, and density of vegetation may be subjectively taken into consideration.

Since evaporation from the groundwater storage is not considered, the fit of the computed hydrograph during periods of dry months is poor. This is perhaps explained by the fact that vegetation transpires from the storages of S_1MM and S_2MM.

Output

The output from the model, as previously stated, is on a daily basis, but monthly and yearly totals may also be computed. The following computed data is printed every time period, the time period being equal to 24 hours in the present program:

1. Mean daily flow per sub-watershed (maximum of four sub-watersheds)
2. Total of all sub-watersheds
3. S_1, S_2, and the interflow per sub-watershed

where S_1 = persistent springflow in cms

S_2 = temporary springflow in cms.

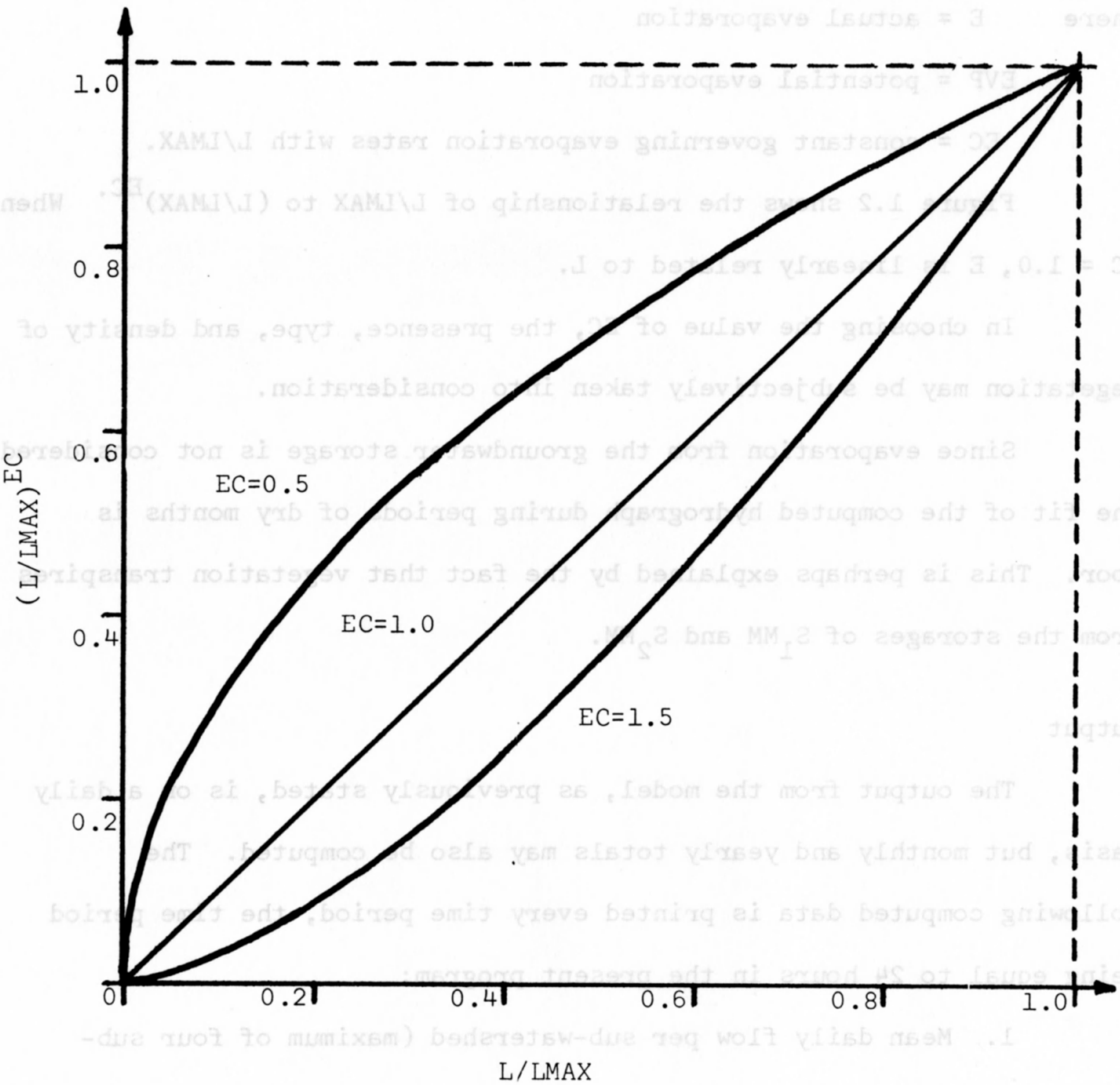


Figure 1.2 L/LMAX vs. $(L/LMAX)^{EC}$.

On a monthly and yearly basis the following are printed:

1. Totals of the above three daily printouts
2. Precipitation
3. Soil moisture in L_1 and L_2 at the end of every month and year.

Computer Program and List of Variables

A listing of the computer program for the rainfall-runoff model, followed by a list of variables and their definitions is given in the following pages.

Modified Mero Watershed Model

```
      PROGRAM CHRIS(INPUT,OUTPUT,TAPE2=INPUT,TAPE5=OUTPUT)
      INTEGER Z,ZC,YR,DM(4),SC,YRM,RUN1,RUN2,RUN3,YRMO(4),
     1RUN4,CODE1(4),CODE2(4),CODE3(4),CODE4(4),STAT(4),RSTAT(4)
      REAL L1(4),L2(4),L,LFC(4),LST(4),IN(4),INMM(4)
      DIMENSION NDM(12),U(4),UST(4),CL2(4),TSMM(4),RO4(4),CO4(4),RO3(4),
     1RO2(4),CO2(4),S2MM(4),RO1(4),CO3(4),CO1(4),S1MM(4),DA(4),ND(4),
     2VTY(4),VT(4),CT(4),QO(4),STMM(4),TSTMM(4),TINMM(4),Q(4),TO3(4),
     3TO4(4),A1(4),TO1(4),TO2(4),A2(4),A3(4),A4(4),ST(4),S2(4),S1(4),
     4STIN(4),EVPC(4),PVTOT(12),EC(4),EVPD(4,31),P(4,31),
     5TOTPM(4),TOTPY(4),PC(4,31),A(5),H(5),PESTO(5),PPOLY(5)
      ICARD=2
      IPR=5
      IPUN=1
C-----NO OF DAYS IN MONTH (NDM)
      NDM(1)=31
      NDM(2)=28
      NDM(3)=31
      NDM(4)=30
      NDM(5)=31
      NDM(6)=30
      NDM(7)=31
      NDM(8)=31
      NDM(9)=30
      NDM(10)=31
      NDM(11)=30
      NDM(12)=31
C-----INPUT CARD WITH NO OF ZONES AND IDENTIFICATION
      READ(ICARD,1000) ZC,RUN1,CODE1,RUN2,CODE2,RUN3,CODE3,RUN4,CODE4,
     1SC,YR,MON,NOYRS,VEGET
 1000 FORMAT(I1,4(I2,4A2),I1,3I2,F3.1)
C-----INITIALIZATION
      DO 2 Z=1,ZC
      TOTPM(Z)=0.
      STMM(Z)=0.
      INMM(Z)=0.
      TSTMM(Z)=0.
      TINMM(Z)=0.
      TSMM(Z)=0.
      S2MM(Z)=0.
      VT(Z)=0.
      VTY(Z)=0.
      TOTPY(Z)=0.
      L2(Z)=0.
      ND(Z)=0
      U(Z)=0.
    2 CONTINUE
      QTOT=0.
      VTOT=0.
      VTOTY=0.
      KFI=0
C-----PARAMETER INPUT CARD
      DO 1 Z=1,ZC
      READ(ICARD,1005)EC(Z),A1(Z),TO1(Z),A2(Z),TO2(Z),A3(Z),TO3(Z),A4(Z)
     1,TO4(Z),DM(Z),UST(Z),LFC(Z),LST(Z),CT(Z),QO(Z),S1MM(Z),L1(Z),CL2(Z
     2),EVPC(Z)
 1005 FORMAT(6X,F2.1,F5.1,F4.0,F5.1,F3.0,F5.1,F3.1,F5.1,F3.1,I3,F3.1,F4.
     11,F4.1,F4.3,F4.3,F6.3,F4.1,F4.3,F3.2)
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```
      RO1(Z)=(1.-EXP(-1./TO1(Z)))
      RO2(Z)=(1.-EXP(-1./TO2(Z)))
      RO3(Z)=(1.-EXP(-1./TO3(Z)))
      RO4(Z)=(1.-EXP(-1./TO4(Z)))
      CO1(Z)=A1(Z)/86.4
      CO2(Z)=A2(Z)/86.4
      CO3(Z)=A3(Z)/86.4
      CO4(Z)=A4(Z)/86.4
      DA(Z)=A4(Z)-A3(Z)-A2(Z)-A1(Z)
      WRITE(IPR,2000)Z,A1(Z),A2(Z),A3(Z),A4(Z),DA(Z),DM(Z),UST(Z),LFC(Z)
     1,CT(Z),S1MM(Z),TO1(Z),TO2(Z),TO3(Z),TO4(Z),EC(Z),CL2(Z),EVPC(Z),
     2LST(Z),Q0(Z),L1(Z)
 2000 FORMAT(7H  ZONE ,I1,4H A1=,F6.1,4H A2=,F6.1,4H A3=,F6.1,4H A4=,F6.
     11,4H DA=,F6.2,4H DM=,I5,5H UST=,F5.1,5H LFC=,F5.1,4H CT=,F6.3,5H
     2S1=,F7.2/ 8X,4HTO1=,F5.0,5H TO2=,F5.0,5H TO3=,F5.1,5H TO4=,F5.1
     3,4H EC=,F2.1,4HCL2=,F5.3,6H EVPC=,F4.2,5H LST=,F5.1,4H Q0=,F6.3,
     44H L1=,F6.1/////)
    1 CONTINUE
      WRITE(IPR,1001)CODE1,RUN1,CODE2,RUN2,CODE3,RUN3,CODE4,RUN4,SC
 1001 FORMAT(1X,*ZONE 1=*,4A2,*(*,I2,*),ZONE 2=*,4A2,*(*,I2,*),ZONE 3=*,
     14A2,*(*,I2,*),ZONE 4=*,4A2,*(*,I2,*)   SC=*,I2)
C-----DAILY EVAPORATION AND RAINFALL INPUT IN INCHES
    5 DO 10 Z=1,ZC
      READ(ICARD,1010)STAT(Z),YRMO(Z),(EVPD(Z,I),I=1,31)
 1010 FORMAT (A8,I4,31F2.2)
      READ(ICARD,1015)RSTAT(Z),YRMO(Z),(P(Z,I),I=1,10)
      READ(ICARD,1015)RSTAT(Z),YRMO(Z),(P(Z,I),I=11,20)
   10 READ(ICARD,1016)RSTAT(Z),YRMO(Z),(P(Z,I),I=21,31)
 1015 FORMAT (A11,I4,1X,10F5.2)
 1016 FORMAT (A11,I4,1X,11F5.2)
      KFI=KFI+1
      LINE=50
C-----COMPUTE NO OF DAYS IN MONTH
      MON=MON+1
      IF(MON-13)109,101,109
  101 YR=YR+1
      MON=1
  109 K=NDM(MON)
      IF(MON-2)110,105,110
  105 IF((YR*100)/4-(YR/4)*100)110,106,110
  106 K=K+1
  110 DO 500 I=1,K
      DO 30 Z=1,ZC
      L=L1(Z)+L2(Z)
      PRST=L/LST(Z)
      GPR1=0.
      GPR2=0.
      PN=(P(Z,I)-EVPD(Z,I)*EVPC(Z))*25.4+U(Z)
      TOTPM(Z)=TOTPM(Z)+P(Z,I)*25.4
      P(Z,I)=P(Z,I)*25.4
      IF(PN)31,32,32
   31 U(Z)=0.
      DL=PN*(PRST**EC(Z))
      GO TO 33
   32 IF(PN-UST(Z))42,43,43
   42 U(Z)=PN
      DL=0.
      GO TO 33
```

```
43 PN=PN-UST(Z)
   U(Z)=UST(Z)
   IF(PN-10.)46,46,44
44 IF(PN-40.)57,57,58
46 STPR=(-0.32+0.071*PN+0.05*PN*(PRST-0.5)-Q0(Z))*CT(Z)
   GO TO 59
57 STPR=(-0.45+0.06*PN+0.0025*(PN**2)-0.00001*(PN**3)+0.25*(PN-8.)*
  1(PRST-0.5)-Q0(Z))*CT(Z)
   GO TO 59
58 STPR=(1.43-0.0309*PN+0.00332*(PN**2)-0.000003*(PN**3)+0.25*(PN-8.)
  1*(PRST-0.5)-Q0(Z))*CT(Z)
59 IF(STPR)60,60,61
60 STPR=0.
61 PN=PN-STPR
   GPR1=PN*(0.01+0.5*PRST)
   IF(GPR1)67,68,68
67 GPR1=0.
68 DL=PN-(GPR1*A3/A4)
   STMM(Z)=STMM(Z)+STPR
   INMM(Z)=INMM(Z)+GPR1
   TSTMM(Z)=TSTMM(Z)+STPR
   TINMM(Z)=TINMM(Z)+GPR1
33 L1(Z)=L1(Z)+DL
   IF(L1(Z)-LFC(Z)) 133,133,143
133 IF(L1(Z)) 153,153,144
153 L2(Z)=L2(Z)+L1(Z)
   L1(Z)=0.
   GO TO 144
143 L2(Z)=L2(Z)+L1(Z)-LFC(Z)
   L1(Z)=LFC(Z)
144 IF(L2(Z)+LFC(Z)-LST(Z)) 193,39,39
193 IF(L2(Z)) 163,163,173
163 L2(Z)=0.
   ND(Z)=0
   GO TO 40
173 IF(L2(Z)-10.) 164,164,175
164 ND(Z)=0
   GO TO 40
175 ND(Z)=ND(Z)+1
   IF(ND(Z)-DM(Z))40,39,39
39 ND(Z)=DM(Z)-1
40 GPR2=L2(Z)*CL2(Z)*0.58*(EXP(L2(Z)/(LST(Z)-LFC(Z)))-1.)
   L2(Z)=L2(Z)-GPR2
   IF(L2(Z)+LFC(Z)-LST(Z))45,45,34
34 GPR2=GPR2+L2(Z)+LFC(Z)-LST(Z)
   L2(Z)=LST(Z)-LFC(Z)
45 DG=GPR2
   TSMM(Z)=TSMM(Z)+DG
   S2MM(Z)=S2MM(Z)+DG
   S1MM(Z)=S1MM(Z)+DG
   S1S2=S1MM(Z)+S2MM(Z)
   IF(MON-5) 303,304,304
304 IF(MON-10) 305,305,303
305 TRGR=EVPD(Z)*VEGET*(1.-PRST)*EXP(-20./S1S2)
   S2MM(Z)=S2MM(Z)-0.5*TRGR
   IF(S2MM(Z)) 404,404,405
404 A=1.
   S1MM(Z)=S1MM(Z)+S2MM(Z)
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```
      S2MM(Z)=0.
  405 S1MM(Z)=S1MM(Z)-0.5*TRGR
      IF(S1MM(Z)) 406,406,303
  406 S1MM(Z)=0.
  303 ST(Z)=STMM(Z)*R04(Z)
      STMM(Z)=STMM(Z)-ST(Z)
      ST(Z)=ST(Z)*C04(Z)
      IN(Z)=INMM(Z)*R03(Z)
      INMM(Z)=INMM(Z)-IN(Z)
      IN(Z)=IN(Z)*C03(Z)
      S2(Z)=S2MM(Z)*R02(Z)
      S2MM(Z)=S2MM(Z)-S2(Z)
      S2(Z)=S2(Z)*C02(Z)*35.315
      S1(Z)=S1MM(Z)*R01(Z)
      S1MM(Z)=S1MM(Z)-S1(Z)
      S1(Z)=S1(Z)*C01(Z)*35.315
      STIN(Z)=ST(Z)+IN(Z)
      STIN(Z)=STIN(Z)*35.315
      Q(Z)=STIN(Z)+S1(Z)+S2(Z)
      VT(Z)=VT(Z)+Q(Z)
   30 CONTINUE
      DO 48 J=1,SC
   48 QTOT=QTOT+Q(J)
      QTOT=0.
  500 CONTINUE
      DO 52 Z=1,ZC
      VT(Z)=VT(Z)*0.0864
      VTY(Z)=VTY(Z)+VT(Z)
   52 TOTPY(Z)=(TOTPY(Z)+TOTPM(Z))
      DO 53 Z=1,SC
   53 VTOT=VTOT+VT(Z)
      PVTOT(MON)=VTOT
      VTOTY=VTOTY+VTOT
C-----PRINT MONTHLY OUTPUT
      WRITE(IPR,2015)MON,YR,VTOT,(VT(Z),TOTPM(Z),L2(Z),L1(Z),Z=1,ZC)
 2015 FORMAT(*0*,I2,*/19*,I2//5X,*VTOT*,3X,*V(MCM)*,4X,*P*,3X,*L2*,3X,
     1*L1(MM)*,2X,*V(MCM)*,4X,*P*,3X,*L2*,3X,*L1(MM)*,2X,*V(MCM)*,4X,*P*
     2,3X,*L2*,3X,*L1(MM)*,2X,*V(MCM)*,4X,*P*,3X,*L2*,3X,*L1(MM)*/2X,2F7
     3.3,F7.1,2F6.1,F8.3,F7.1,2F6.1,F8.3,F7.1,2F6.1,F8.3,F7.1,2F6.1)
      DO 54 Z=1,ZC
      VT(Z)=0.
   54 TOTPM(Z)=0.
      YRM=YR-1
      VTOT=0.
      IF(MON-9)5,56,5
   56 WRITE(IPR,2020)YRM,YR,VTOTY,(VTY(Z),TSTMM(Z),TINMM(Z),TSMM(Z),Z=1,
     1ZC)
 2020 FORMAT(*0YEARLY TOTALS FOR 19*,I2,*/*,I2//5X,*VTOT*,4X,*V(MCM)*,3X
     1,*ST*,4X,*IN*,2X,*S(MM)*,3X,*V(MCM)*,3X,*ST*,4X,*IN*,2X,*S(MM)*,3X
     2,*V(MCM)*,3X,*ST*,4X,*IN*,2X,*S(MM)*,3X,*V(MCM)*,3X,*ST*,4X,*IN*
     3,2X,*S(MM)*/4X,F7.3,F8.3,3F6.1,F9.3,3F6.1,F9.3,3F6.1,F9.3,3F6.1)
      WRITE(IPR,2025) (TOTPY(Z),Z=1,ZC)
 2025 FORMAT(* *,2X,*P(MM)*,4X,F4.0,23X,F4.0,23X,F4.0,23X,F4.0)
      VTOTY=0.
      DO 99 Z=1,ZC
      VTY(Z)=0.
      TOTPY(Z)=0.
      TSTMM(Z)=0.
```

```
    TINMM(Z)=0.
 99 TSMM(Z)=0.
    IF(KFI/12-NOYRS)520,521,520
520 GO TO 5
521 STOP
    END
```

Definition of Variables in Mero's Watershed Model

Variable	Units	Description
A_1	km^2	Area allocated to persistent springs
A_2	km^2	Area allocated to temporary springs
A_3	km^2	Area allocated to interflow
A_4	km^2	Area allocated to overland flow (same as total watershed area)
CL2	1	Constant in groundwater recharge from L2
CØDE1		
CØDE2		Identification of each of the four zones (Z) or sub-watersheds
CØDE3		
CØDE4		
CØ1	$10^3 Tm^2/s$	Conversion coefficient for persistent springs
CØ2	$10^3 Tm^2/s$	Conversion coefficient for temporary springs
CØ3	1	Ratio of interflow area
CØ4	$10^3 Tm^2/s$	Conversion coefficient for interflow and overland flow
CT	1	"Tilt" coefficient in PN-STPR relationship
DA	km^2	Area active in a "losing" or "gaining" stream
DG	mm/T	Recharge to groundwater (same as GPR2)
DL	mm/T	Addition to soil-moisture storage (+ or -)
DM	1	Maximum number of days in time delay function
EC	1	Exponent in $(L/LST)^{EC}$
EVPC	1	Coefficient to modify pan evaporation

EVPD	in/T	Potential evaporation per day (converted to mm/T in program)
GPR1	mm/T	Addition to interflow storage
GPR2	mm/T	Addition to DG from soil moisture
I	Integer	Used in matrices or DØ statements in program
ICARD		Set equal to 2, and used in READ statement
IN	mm/T, ft^3/s	Interflow contribution to daily streamflow
INMM	mm	Current depth of water in interflow storage
IPR		Set equal to 5 and used in WRITE statement
IPUN		Set = 1, and used in PUNCH statement (not used in this program)
J	Integer	Same as I above
K	Integer	Same as I above
KFI		Index used during program
L	mm	Total moisture in soil
LFC	mm	Field capacity $\geq$ L1
LINE		Index used in program
LST(same as LMAX)	mm	Maximum possible soil moisture in both horizons
L1	mm	Current moisture in upper soil horizon, $\leq$ LFC
L2	mm	Current moisture in lower soil horizon, $\leq$ LMAX-LFC
MØN	1	Initial month (set equal to 9 in input card)
ND	1	Number of days elapsed since L2 > 0
NDM	1	Number of days in month
NØYRS	1	Number of years simulation is required (first input card)

P	in/T	Daily precipitation input (converted to mm/T in program)
PN	mm/T	Net precipitation (has multiples uses in program, Caution!)
PRST	1	L/LMAX or L/LST
Q	ft^3/s	Average discharge over period (in this case mean daily discharge)
QO	mm/T	Threshhold level of PN in order to produce overland flow
RØ1	1	Routing coefficient for persistent springs (computed in program)
RØ2	1	Routing coefficient for temporary springs (computed in program)
RØ3	1	Routing coefficient for interflow (computed in program)
RØ4	1	Routing coefficient for overland flow (computed in program)
RSTAT		Rainfall station identification
RUN1	1	
RUN2	1	Number of runs made per sub-watershed in calibration stage (for identification purposes only)
RUN3	1	
RUN4	1	
SC	1	Number of sub-watersheds to be grouped in outputting discharge (maximum of 4)
ST	mm/T, ft^3/s	Overland flow (after routed from surface storage
STAT		Evaporation station identification
STIN	mm/T, ft^3/s	= ST + IN
STMM	mm	Current storage in overland flow "reservoir"
STPR	mm/T	Addition to STMM (see Equation [1.4])

S1	mm/T, ft^3/s	Flow from persistent springs
S1MM	mm	Current depth of storage in persistent spring "reservoir"
S1S2	mm	= S1MM + S2MM
S2	mm/T, ft^3/s	Flow from temporary springs
S2MM	mm	Current depth of storage in temporary spring "reservoir"
TJNMM	mm	Yearly addition to interflow storage
TØTPM	mm	Total monthly rainfall over watershed
TØTPY	mm	Total yearly rainfall over watershed
TRGR*	mm	Transpiration from groundwater (S1MM and S2MM) effective from May to October
TSMM	mm	Yearly addition to groundwater storage
TSTMM	mm	Yearly addition to surface storage
TØ1	T	Time constant in routing from the persistent springs reservoir
TØ2	T	Time constant in routing from the shallow springs reservoir
TØ3	T	Time constant in routing from interflow storage
TØ4	T	Time constant in routing from surface storage
U	mm	Current moisture in interception storage, $0 \leq U \leq UST$
UST(same as UMAX)	mm	Maximum value of interception storage (input parameter)
VEGET*	1	Portion of watershed covered by vegetation (input parameter)
VT	ft^3/s, $10^6 ft^3$	Total monthly volume fo flow for sub-watershed
VTØT		Total monthly volume of flow for all sub-watersheds

VTØTY		Total annual volume of flow for all sub-watersheds
VTY	$10^6 ft^3$	Total yearly volume of flow for sub-watershed
YR	1	Initial year, current year
YRM	1	Previous year = YR - 1
YRMØ	1	Current year and month
Z	Integer	Used in DØ statements to denote number of sub-watersheds within main watershed
ZC	Integer	Denotes number of sub-watersheds for which the flow is summed together in the output

APPENDIX 2

PRECIPITATION INPUT TO THE RAINFALL-RUNOFF MODEL

Two sets of precipitation inputs are used in the simulation of streamflow as described in Chapter 4 and Appendix 1:

1. Values of 24-hour precipitation representing "actual" mean areal precipitation on Watershed 5.
2. Computed values of 24-hour precipitation for the same period as computed by the use of Equation [5], thus enabling a study of how the differences in the input influence output in the rainfall-runoff model.

In the first case 24-hour precipitation records, which are readily available in published form (Reimann and Hamilton, 1959), for station I35 are used as representing mean areal precipitation; but before concluding on their use, these point values are compared to mean areal values by using a regression analysis; the same 41 storms as the ones mentioned in Chapter 3 are used in this analysis. The regression equation as calculated by the least square method is as follows:

$$\overline{P}_R = 0.0543 + 0.972P_{I35} \quad . \ . \ . \qquad [2.1]$$

where $\overline{P}_R$ = computed mean areal storm precipitation for Watershed 5 (see Equation [2.2] for observed $\overline{P}$).

and P_{I35} = storm precipitation at raingage I35.

The results of the analysis are given in Table 2.1. Column 2 in this table is a listing of the values of the 41 storms as measured at

Table 2.1. Table showing the results of the regression analysis (Equation [2.1] for 41 storms.

Storm No.	Storm Size (inches) P_{I35}	Storm Size (inches) $\overline{P}$	Error = $P_{I35}-\overline{P}$ (inches)	5 Percent Error = $\frac{P_{I35}-\overline{P}}{\overline{P}} \times 100$	6 $\overline{P}_R$ (inches)	7 Error = $\overline{P}-\overline{P}_R$	8 Percent Error = $\frac{\overline{P}-\overline{P}_R}{\overline{P}} \times 100$
271+272	3.06	2.93	-.13	- 4.2	3.05	-.12	- 4.1
276	1.98	1.90	-.08	- 3.9	1.99	-.09	- 4.8
279	.98	1.02	.04	4.1	1.01	.01	.6
283	1.65	1.62	-.03	- 1.7	1.67	-.05	- 3.0
243	2.15	2.10	-.05	- 2.1	2.16	-.05	- 2.6
247	6.45	6.61	.16	2.5	6.37	.24	3.6
219	1.28	1.27	-.01	- .5	1.31	-.03	- 2.7
220-1	9.73	9.96	.23	2.4	9.58	.38	3.8
223	1.37	1.36	-.01	- 1.1	1.40	-.04	- 3.0
225	6.18	6.06	-.12	- 2.0	6.11	-.05	- .8
227	2.58	2.60	.02	.8	2.58	.02	.8
195	1.26	1.19	-.07	- 5.5	1.29	-.10	- 8.2
198	5.09	5.24	.15	3.0	5.04	.21	3.9
199	3.51	3.31	-.20	- 5.8	3.49	-.19	- 5.6
200	1.61	1.25	-.36	-22.1	1.63	-.38	-30.0
209+210	11.74	11.49	-.25	- 2.1	11.55	-.06	- .5
215	1.49	1.45	-.04	- 2.6	1.51	-.06	- 4.2
179	.93	.99	.06	6.8	.97	.03	2.8
181	1.45	1.28	-.17	-11.7	1.47	-.19	-15.1
182	20.90	20.17	-.73	- 3.5	20.52	-.35	- 1.7
186	6.45	6.65	.20	3.1	6.37	.28	4.2
187	6.94	7.04	.10	1.4	6.85	.19	2.7
164	2.01	2.21	.20	9.8	2.02	.18	8.4
169	1.56	1.48	-.08	- 5.3	1.58	-.11	- 7.1
170	1.87	1.80	-.07	- 3.6	1.89	-.08	- 4.6
147	3.96	4.18	.22	5.7	3.93	.25	6.0
128	2.47	2.40	-.07	- 3.0	2.47	-.08	- 3.2

126	2.21	2.20	-.01	- .5	2.22	-.02	- .9
94a,95,96	9.09	9.05	-.04	- .4	8.96	.10	1.1
97	2.35	2.15	-.20	- 8.5	2.36	-.21	- 9.5
98	.93	1.16	.23	24.8	.97	.20	16.8
305	1.79	2.16	.37	20.8	1.81	.35	16.4
307	3.87	3.87	-.00	- .1	3.84	.02	.6
309	5.42	5.28	-.14	- 2.5	5.36	-.08	- 1.5
315	2.56	2.59	.03	1.2	2.56	.03	1.1
317	1.74	1.78	.04	2.3	1.76	.02	1.2
318	1.87	1.83	-.04	- 2.0	1.89	-.05	- 2.9
337	1.80	1.83	.03	1.6	1.82	.01	.6
338	2.73	2.66	-.07	- 2.6	2.73	-.07	- 2.6
339+340	2.72	2.66	-.06	- 2.1	2.72	-.05	- 2.1
341	1.82	1.83	.01	.3	1.84	-.01	- .6

Note: P_{I35} is storm precipitation as measured at gage I35 in Watershed 5

$\overline{P}_R$ is observed mean areal precipitation (Equation [2.2]) for Watershed 5

$\overline{P}$ is computed mean areal precipitation (Equation [2.1]) for Watershed 5 station I35, and column 3 is a list of mean areal precipitation for the same storms. Columns 4 and 5 give the absolute and percentage difference (error) between values in columns 2 and 3, respectively. The values in column 3 are computed from observed altitudinal average storm values, available for altitudes of 2100, 2600, 3100 and 4100 feet. To each altitude an area is assigned, the upper and lower boundaries of which run parallel to and halfway between two adjacent elevation contours for which rainfall observations are given. Equation [2.2] is used for weighting the altitudinal storm values on an area basis in order to obtain the mean areal precipitation.

$$\overline{P} = \frac{\sum_{i=1}^{4} A_i P_i}{\sum_{i=1}^{4} A_i} \qquad [2.1]$$

where $\overline{P}$ = mean areal storm precipitation

i = number of altitudinal area zones

A_i = area of zone i

P_i = observed storm precipitation, average over the contour at the mid-elevation of zone i.

As indicated in column 5 of Table 2.1 the error is less than 5 percent for three-fourths of the storms; in the three cases for which the error exceeds 20 percent the storm depth is small and so is the absolute error. Values in columns 6, 7 and 8 in Table 2.1 refer to computed mean areal storm depths (by use of Equation [2.1]), and absolute and percent deviation (error) from corresponding values in column 3, respectively.

Values of the standard error of estimate are computed for the two pairs of values: columns 2 and 3 and columns 6 and 7. These values are 0.1826 and 0.1640 inch, respectively, and their proximity is an indication of the close agreement between corresponding values of $\overline{P}$ and $\overline{P}_R$.

Without further examination, the precipitation values as recorded at gage I35 are considered representative of the mean areal precipitation on Watershed 5, at least for the purpose of this study. Thus, the precipitation at gage I35 is considered the first of the previously mentioned two precipitation sets or sequences.

For the second set of precipitation input, areal weighting is employed to obtain mean areal precipitation values; as in case 1, Equation [2.2] is used in the weighting. This time, however, P_i refers to elevational precipitation as computed from corresponding values at station I35 and Equation [5] (see Chapter 3). A small sub-routine had to be introduced in the rainfall-runoff program to allocate storm precipitation to 24-hour periods to comply with input specifications. It may be useful at this point to repeat that all storms used in rainfall relationships and analysis so far had a duration from a few hours to about 100 hours, contrary to the 24-hour precipitation values required by the model. Only for station I35 are 24-hour precipitation depths available, and these have provided the basis for allocating storm to 24-hour precipitation, under the assumption of zero temporal variability of this variable, at least in Watershed 5. A listing of the sub-routine and explanation of the variables used follows.

Listing of sub-routine used to compute storm precipitation from storm data at I35 using Equation 4, and to distribute computed storm values to the respective 24-hour periods.

```
        IF(MON-2)110,105,110
105     IF((YR*100)/4-(YR/4)*100)110,106,110
106     K=K+1

110     DO 600 Z=1, ZC
        H(1)=2.7
        H(2)=2.1
        H(3)=2.6
        H(4)=3.1
        H(5)=4.1
        A(2)=0.64
        A(3)=4.10
        A(4)=4.40
        A(5)=1.96
        ATOT=11.10
        I=1
601     IF(P(Z,I)) 605,605,610
605     PC(Z,I)=P(Z,I)
        I=I+1
        IF(I-K) 606,606,600
610     PSTO=P(Z,I)
        ISTRT=I
        I=I+1
        IF(I-K) 608,608,621
608     IF(P(Z,I)) 621,621,622
622     PSTO=PSTO+P(Z,I)
        I=I+1
        IF(I-K) 609,609,621
609     GO TO 608
621     NSD=I-ISTRT
        M=ISTRT+NSD-1
        IF(PSTO-1.) 840,840,818
818     PCSUM=0.
        DO 819 J=2,5
        PESTO(J)=PSTO+0.183*(H(J)-H(1))*PSTO-0.2*(H(J)-H(1))
        PPOLY(J)=A(J)*PESTO(J)
819     PCSUM=PCSUM+PPOLY(J)
        PCSTO=PCSUM/ATOT
        DO 830 J=ISTRT,M
830     PC(Z,J)=P(Z,J)*PCSTO/PSTO
        GO TO 845
840     DO 841 J=ISTRT,M
```

```
841     PC(Z,J)=P(Z,J)
845     IF(I-K) 606,606,600
606     GO TO 601
600     CONTINUE

        DO 500 I=1,K
```

Explanation of variables in sub-routine used to compute mean areal precipitation for 24-hour periods from storm precipitation at station I35 and Equation [5]. Storms of one inch or less are not recomputed, but the values at I35 are used.

Z	Variable, defined as an integer in rainfall-runoff program
ZC	Number of zones in watershed (integer)
H(I)	Altitudes for which mean altitudinal storm precipitation data is available (in thousands of feet)
A(I)	Area corresponding to H(I) in square km
ATØT	Total area in Watershed 5 (ATOT = $\sum_{I=2}$ A(I))
P(Z,I)	24-hour precipitation values at station I35 read in during main program. Z is as defined above and I is the date in a particular month and ranges from 1 to 31.
PC(Z,I)	24-hour precipitation as computed by Equation [4], mean for Watershed 5. For storms of 1 inch or less Equation [4] was not applied and PC(Z,I)=P(Z,I)
PSTØ	Storm precipitation, being the sum of one or more 24-hour values of P(Z,I)
ISTRT	Date storm starts
NSD	Number of storm days in one storm
PESTØ	Estimated storm depth from Equation [4]
PPØLY	A(J)*PESTØ(J), J = number of area zones
PCSUM	Accumulator of PPØLY

PCSTØ <u>Computed storm</u> precipitation. Computed storm values are distributed to 24-hour periods of computed precipitation as follows:

PC(Z,J)=P(Z,J)*PCSTØ/PSTØ, where J values range between the beginning and terminating date of the storm at hand.

SELECTED BIBLIOGRAPHY

Bean, Robert T. 1944. Geology of watersheds V, VII and lower VI, San Dimas Experimental Forest Progress Report. California Forest and Range Experimental Forest, Berkeley, California. U.S. Dept. of Agriculture Forest Service.

Bean, Robert T. 1946. Geology of watersheds I, II, III and IVA. California Forest and Range Experimental Forest, Berkeley, California. U.S. Dept. of Agriculture Forest Service.

Bentley, Jay R. 1961. Fitting brush conversion to San Gabriel watersheds, Miscellaneous Paper No. 61. Pacific Southwest Forest and Range Experiment Station, Berkeley, California. U.S. Dept. of Agriculture Forest Service.

Crawford, James M., Jr. 1962. Soils of the San Dimas Experimental Forest, Miscellaneous Paper No. 76. Pacific Southwest Forest and Range Experiment Station, Berkeley, California. U.S. Dept. of Agriculture Forest Service.

Cyprus Water Planning Project. 1971. Mero's hydrometeorological water balance computations (unpublished). Dept. of Water Development, Nicosia, Cyprus.

Davis, D.R., Duckstein, L. and Kisiel, C.C. 1972. Optimum design of mountainous raingage networks using Bayesian decision theory. To be presented at the International Symposium on Distribution of Precipitation in Mountainous Areas (Oslo, August).

Dawdy, D.R. and Bergmann, J.M. 1969. Effect of rainfall variability on streamflow simulation. Water Resources Research, Vol. 5, No. 5, pp. 958-966.

Dawdy, D.R., Lichty, R.W. and Bergmann, J.M. 1970. A rainfall-runoff simulation model for estimation of flood peaks for small drainage basins -- a Progress Report. U.S. Geological Survey open file report.

Eagleson, P.S. 1967. Optimum density of rainfall networks. Water Resources Research, Vol. 3, No. 4, pp. 1021-1034.

Foucarde, H.G. 1942. Some notes on the effect of the incidence of rain on the distribution of rainfall over the surface of unlevel ground. Royal Soc. South Africa, Cape Town Trans. 29: pp. 235-254.

Hamilton, E.L. 1954. Rainfall sampling on rugged terrain. Technical Bulletin No. 1096. U.S. Dept. of Agriculture, Washington, D.C.

Hamilton, E.L. and Reimann, L.F. 1958. Simplified method of sampling rainfall on the San Dimas Experimental Forest, Technical Paper No. 26, California Forest and Range Experiment Station, Berkeley, California. U.S. Dept. of Agriculture Forest Service.

Hill, Lawrence W. 1963. The San Dimas Experimental Forest. Pacific Southwest Forest and Range Experiment Station, Berkeley, California. U.S. Dept. of Agriculture Forest Service.

Kisiel, C.C. and Duckstein, L. 1972. Economics of hydrologic modelling. University of Arizona, Tucson. To be presented at the International Symposium on Mathematical Modelling Techniques in Water Resources, Ottawa, Canada, May 9-12, 1972.

Krammer, J.S., Lent, J.D. and Clarke, J.W. 1965. Streamflow records from the San Dimas Experimental Forest, 1939-1959. Pacific Southwest Forest and Range Experiment Station, Berkeley, California. U.S. Dept. of Agriculture Forest Service, Research Note PSW-79.

Mero, F. Application of the groundwater depletion curves in analyzing and forecasting spring discharges influenced by well fields. IASH publ. No. 63, Symposium on Surface Water, pp. 107-117.

Parl, B. 1967. Basic Statistics. Doubleday & Company, Inc. Garden City, New York.

Reimann, L.F. 1959. Mountain evaporation -- data from the San Dimas Experimental Forest, Miscellaneous Report No. 35. Pacific Southwest Forest and Range Experiment Station, Berkeley, California. U.S. Dept. of Agriculture Forest Service.

Reimann, L.F. and Hamilton, E.L. 1959. Four hundred sixty storms -- data from the San Dimas Experimental Forest, Miscellaneous Paper No. 37. Pacific Southwest Forest and Range Experiment Station, Berkeley, California. U.S. Dept. of Agriculture Forest Service.

Rowe, P.B. and Colman, E.A. 1951. Disposition of rainfall in two mountain areas of California, Technical Bulletin No. 1048. U.S. Dept. of Agriculture, Washington, D. C.

Shanholtz, V.O. and Lillard, J.H. 1971. Simulations of watershed hydrology on agricultural watersheds in Virginia with the Stanford model. Dept. of Agricultural Engineering Research Division, Virginia Polytechnic Institute and State University, Blacksburg, Virginia.

Storey, H.C. 1939. Topographic influences on precipitation. Proc. of the Sixth Pacific Science Congress, Vol. IV, Berkeley, Stanford and San Francisco, California.

Storey, H.C. and Hamilton, E.L. 1943. A comparative study of raingages. Trans. AGU: 1943(3), pp. 133-141.

Storey, H.C. and Wilm, H.G. 1944. A comparison of vertical and tilted raingages in estimating precipitation in mountain watersheds. Trans. AGU: 1944(3), pp. 518-523.

Wilm, H.G., Nelson, A.Z. and Storey, H.C. 1937. An analysis of precipitation measurements on mountain watersheds. California Forest and Range Experiment Station, Berkeley, California. U.S. Dept. of Agriculture Forest Service.