THE UNIVERSITY OF NEW SOUTH WALES

water research laboratory

Manly Vale, N.S.W., Australia

Report No. 91

IMPROVED TECHNIQUES FOR ESTIMATING RUNOFF WITH BRIEF RECORDS

by

F. C. Bell

June, 1966

The University of New South Wales

WATER RESEARCH LABORATORY

Report No. 91

Improved Techniques for Estimating Runoff with Brief Records

by

F.C. Bell, B.Sc. (Sydney)

June, 1966.

https://doi.org/10.4225/53/5797fd99839ed

SUMMARY

A selective literature survey and general analysis of runoff statistics are presented. From these it is concluded that single, brief records give inaccurate estimates of means and quite unreliable estimates of extremes. Improved estimates are possible if single records are supplemented with other relevant information and it is shown that spatial correlation or regional consistencies can be utilised for this purpose.

The recently developed "complete" rainfall-runoff models provide a means of extending or synthesizing records but some of the components of these models are inconsistent with current knowledge of the physical processes of hydrology. Other deficiencies in the models are caused by their attempts to integrate fragmentary hydrological concepts that have evolved without reference to the rainfall-runoff cycle as a whole.

A more comprehensive approach is suggested by replacing the traditional "infiltration theory" with a general "retention theory" from which an improved rainfall-runoff model is developed.

The improved model is demonstrated and tested with data from a 60 acre watershed.

Acknowledgements

This investigation was undertaken by the author in 1964-66 as part fulfilment of the requirements for the degree of Master of Science in the School of Civil Engineering.

Hydrologic data for the research were obtained from a number of experimental catchments in the Lidsdale State Forest. The experimental catchments were set up under a grant from the Water Research Foundation of Australia for research into the effects of exotic softwood afforestation on water yield. Preliminary results of the studies of water yield investigation are to be published by the Foundation as its Bulletin No. 15.

The research described in the present report is of a fundamental nature, and has general application. The work formed part of the programme of hydrologic research of the Department of Water Engineering and the author expresses appreciation of the encouragement given by Professor C. H. Munro, Foundation Professor of Civil Engineering and the helpful advice by his supervisor, Associate Professor E. M. Laurenson.

CONTENTS

		Page
1.	INTRODUCTION	
1.01	Purpose and Scope of Study	1
1.02	The Limitations of Brief Records	1
1.03	Terminology	2
1.04	Statistical and Deterministic Estimates	2
2.	THE DIRECT ESTIMATION OF RUNOFF STATISTICS	
2.01	Runoff Statistics	4
2.02	Annual, Seasonal and Monthly Runoff	5
2.03	Frequency Distribution of Annual, Seasonal and Monthly Runoff	5
2.04	Extreme Low Flows	7
2.05	Frequency Distribution of Extreme Low Flows	8
2.06	Flood Runoff	9
2.07	Frequency Distribution of Floods	10
2.08	Flow Duration Curves	13
3.	THE ESTIMATION OF RUNOFF STATISTICS FROM RAINFALL DATA	
3.01	Rainfall Statistics	15
3.02	A General Rainfall-Runoff Relationship	16
3.03	Annual and Seasonal Estimates	17
3.04	Monthly Estimates	18
3.05	Storm and Short-Period Estimates	18
4.	ESTIMATES WHEN RUNOFF AND RAINFALL DATA	
4.01	The General Problems	22
4.02	Normal Runoff and Rainfall	22
4.03	Period Bias	24
4.04	Extreme Runoff and Rainfall	26

Page

4.05 Frequency Ratios 26

4.06 Hydrogeography 28

4.07 Climatic and Hydrologic Regions 29

4.08 Hydrogeography and Data Collection 32

5. EXAMPLES OF IMPROVED TECHNIQUES FOR STATISTICAL ESTIMATES

5.01 Mean Annual Rainfall in the Central Tablelands of N.S.W. 33

5.02 Mean Annual Runoff in the Central Tablelands of N.S.W. 34

5.03 Regional Estimates with one or two Watersheds 35

5.04 The Synthesis of Long Records 36

6. COMPLETE RAINFALL-RUNOFF MODELS

6.01 Introduction 43

6.02 The General Rainfall-Runoff Cycle 44

7. UNITED STATES WEATHER BUREAU MODEL

7.01 General Features 46

7.02 Specification of Moisture Status by U.S.W.B.Model 46

7.03 Simulation of Drying Phase by U.S.W.B. Model 48

7.04 Simulation of Wetting Phase by U.S.W.B. Model 48

7.05 Conversion of Rainfall-excess to Streamflow by U.S.W.B. Model 51

8. STANFORD UNIVERSITY MODEL

8.01 General Features 52

8.02 Specification of Moisture Status by Stanford Model 57

8.03 Simulation of Drying Phase by Stanford Model 57

8.04 Simulation of Wetting Phase by Stanford Model 58

8.05 Conversion of Rainfall-excess to Streamflow by Stanford Model 61

Page.

9. COMMONWEALTH BUREAU OF METEOROLOGY MODEL

9.01 General Features 62

9.02 Specification of Moisture Status by C.B.M. Model 62

9.03 Simulation of Drying Phase by C.B.M. Model 62

9.04 Simulation of Wetting Phase by C.B.M. Model 63

9.05 Conversion of Rainfall-excess to Streamflow by C.B.M. Model 64

10. OTHER RAINFALL-RUNOFF MODELS

10.01 Multiple Regression and Multivariate Analysis 65

10.02 T.G. Chapman 67

10.03 R.P. Betson 69

10.04 W.C. Boughton 71

11. SOME RELEVANT STUDIES OF INDIVIDUAL PROCESSES

11.01 Evaporation and Transpiration 75

11.02 Infiltration, Soil Water Retention and Drainage 79

11.03 Conversion of Rainfall-excess to Streamflow 82

12. PRESENT DEFICIENCIES IN RAINFALL-RUNOFF MODELS

12.01 State of Knowledge of Hydrologic Phenomena 84

12.02 Specification of Moisture Status 84

12.03 Simulation of Drying Phase 85

12.04 Simulation of Wetting Phase 87

12.05 General Conclusions on Current Rainfall-Runoff Models 88

13. THE DEVELOPMENT OF RETENTION THEORY

13.01 Infiltration Theory 92

13.02 Runoff Separation 94

13.03 Time Delay between Rainfall and Runoff 96

13.04 Retention Theory 104

Page

14. APPLICATIONS OF RETENTION THEORY

14.01 A Complete Rainfall-Runoff Model 106

14.02 The Runoff Depletion Function 109

14.03 The Streamflow Function 114

14.04 The Runoff Accretion Function 116

14.05 The Retention Depletion Function 121

14.06 The Retention Accretion Function 124

14.07 The Recovery Functions 129

14.08 The Basic Retention Function 131

14.09 Limited Wetting Periods and Initial Loss 136

14.10 Rainfall-Runoff Estimation by the Complete Model 142

14.11 Introduction to Watershed Analysis 165

15. CONCLUSIONS

15.01 Estimates of Runoff Statistics 169

15.02 Complete Rainfall-Runoff Models 169

15.03 An Improved Rainfall-Runoff Model 170

REFERENCES 171

APPENDIX. GLOSSARY OF SPECIAL TECHNOLOGY 182

1. INTRODUCTION

1.01 Purpose and Scope of Study

The objective of the study is to develop improvements in existing techniques of runoff estimation mainly from a thorough consideration of physical processes such as evaporation and infiltration.

For most Australian conditions the volume of runoff for a particular period is largely dependent on the floods within or preceding the period and it is apparent that many aspects of flood hydrology must therefore be dealt with.

Because of the broad scope of the study, the effects of snow and other freezing phenomena have been omitted. It is realized that these are of great significance in many circumstances and necessitate major modifications to some of the techniques described.

1.02 The Limitations of Brief Records

From the point of view of engineering design the most important hydrological estimates are those of extreme events, perhaps with recurrence intervals of 50 to 10,000 years. It is generally recognized that such estimates cannot be obtained with a high degree of reliability when only 30 or 40 years of records are available, and these circumstances are quite common.

The position is even worse (and the situation perhaps even more common) when only 5 or 10 years of records are available and in these circumstances it is also a dubious business to directly estimate non-extreme events such as the mean annual runoff or the 2 year flood peak.

In many cases improvements are possible, however, if additional hydrological information can be utilized. This may be in the form of data from similar watersheds with longer records or it may consist of special knowledge of the hydrological phenomena involved. The problems of using such data systematically and efficiently will be the concern of this study.

1.03 Terminology

The original title of the thesis involved the word "yield" which was understood to mean "volume of streamflow during a specific time period of one month or longer duration or as a result of a specific storm event." (Ayers, ref. 5 p.1). However, an examination of the text books showed several different meanings for the term (e.g. ref. 3, p.15 and ref. 1, p.360) and it was therefore abandoned in favour of the more general "runoff" which is defined as "any volume of streamflow having its origin in rainfall on the watershed concerned."

The very unsatisfactory state of hydrological terminology has also hampered the study (and no doubt many other studies) by the absence of recognised terms to describe certain relatively simple concepts that should be used commonly in hydrological thinking. It has therefore been necessary to invent or adopt various expressions that may be unfamiliar to other hydrologists. These are listed and defined in an Appendix.

1.04 Statistical and Deterministic Estimates

Runoff estimates are of two main types, viz. statistical or deterministic. Statistical estimates are concerned with representative events during periods of time, for example the maximum value in 100 years. A major objective with this type of estimate is the assessment of frequency distributions rather than chronological or precise time distributions.

On the other hand, deterministic estimates are concerned with the actual times of particular events and therefore often give detailed attention to cause and effect, for example the estimation of an unknown runoff resulting from a known set of storm conditions.

In practice statistical and deterministic estimates tend to be interdependent and their differences are somewhat obscured. It should be recognized that most statistical techniques are inherently empirical and it is inappropriate to criticise them on the grounds that they are "not rational." Rational or deterministic considerations may often be used, however, to improve or develop statistical techniques.

Sections 2 to 5 of this report deal largely with statistical aspects of runoff estimation and are considered necessary for a proper perspective of the subject. They consist of a literature survey accompanied by a number of essentially untested ideas for improved techniques.

The chief interest of the author has been in the deterministic aspects of runoff estimation and these are submitted in sections 6 to 14 as the main contribution of the study.

2. THE DIRECT ESTIMATION OF RUNOFF STATISTICS

2.01 Runoff Statistics

In general the frequency distributions of statistical populations are analysed by techniques involving measures of central tendency (e.g. the mean and median), dispersion (e.g. the standard deviation and various percentiles), skewness (e.g. Pearson's coefficients) and Kurtosis (e.g. moment coefficient). Associated with all of these measures are the various moments of distribution. Unfortunately the usefulness of such techniques is limited in runoff studies because:

(a) Simple distributions based on statistical theories are not completely appropriate due to different ranges of values being caused by quite different physical factors. For example the extreme floods at a particular place may be caused by rare visitations from a climatic regime that has little influence on relatively common floods.

(b) In hydrological problems extreme events are of prime importance but the usual frequency distribution parameters are derived largely from the characteristics of common events. This is closely associated with (a) above.

(c) The samples available for estimates of runoff distributions are usually small and their information content is further reduced by autocorrelation. (Ref. 72).

The processes of rainfall and runoff generation are so complex that no satisfactory physical theories have been developed to explain and predict the general forms of runoff frequency distributions, although the techniques of storm maximization and transposition may possibly be regarded as steps in this direction.

It is advantageous to consider runoff statistics in three groups, viz:

(a) Annual, seasonal and monthly runoff.

(b) Extreme low flows.

(c) Flood runoff.

The required estimates are of a different nature and for a different purpose within each of the above groups.

2.02 Annual, Seasonal and Monthly Runoff

Studies of annual, seasonal and monthly runoff are usually concerned with "average" values which are useful for the general assessment of water resources and preliminary comparisons between alternative development schemes.

Estimates of storage behaviour etc. may also be concerned with extreme values and for these purposes it is necessary either to adopt some form of frequency distribution or to assume that future events will occur of the same magnitude and in the same sequence as the past recorded events. The latter assumptions simplify the calculations but are not very satisfying from a theoretical point of view, as suggested in several references (e.g. ref. 6 and ref. 20).

Extreme runoff estimates for design purposes, such as the critical low flow or design flood, are not necessarily associated with annual, seasonal or monthly periods of time. These are discussed in sections 2.04 and 2.07.

2.03 Frequency Distribution of Annual, Seasonal and Monthly Runoff

The frequency distributions of runoff for most Australian and U.S. streams are positively skewed. Studies by Standish Hall (ref. 14) and Waitt (ref. 15) suggest that the skewness tends to be greater in drier regions, which is evidently a consequence of the general form of the relationships between losses and rainfalls, (see 3.02). Skewness also tends to be greater for monthly values than for seasonal or annual values.

Some European streams, however, appear to have symmetrical distributions for both annual and monthly runoff. (ref. 17).

Alexander (ref. 12) and Blokhinov (ref. 18) have proposed the general use of the two-parameter gamma distribution which allows for any degree of skewness. Applications of this have been suggested for advanced statistical analyses but they may sometimes prove unsatisfactory for the reasons outlined in 2.01.

Various theoretical distributions for the complete range of values may be derived with the Markov techniques described by Federov (ref. 19) and Fiering (ref. 20). These are mathematically elegant but are probably no more reliable than other simpler approaches because of the tacit assumptions involved and the large sampling errors of the basic data.

Logarithmic transformations of runoff values are often successful in "normalizing" the data. This technique was used in the previously mentioned study by Waitt (ref. 15) for a number of Australian streams.

Gerny (ref. 21) has suggested a general empirical form of distribution expressed by:

$$Q_f = C \log_{10} f + 5 D \log_{10} \left(\frac{f}{f - 1}\right) \qquad \text{............(2a)}$$

where Q_f = annual runoff that may be expected once in f years.

Equation (2a) should provide a good fit to most data but the constants C and D have no direct statistical significance.

Because of skewness in runoff frequency distributions it is often preferable to use the median rather than the arithmetic mean as a measure of central tendency, particularly with logarithmic transformations, (see ref. 15). The relative merits of means and medians are discussed in some detail by Foster (ref. 6).

When lower values are of primary significance the geometric mean is a useful measure of central tendency as it is is not unduly affected by irrelevant high values, (Beard,ref. 22). Similarly the root mean square should be suitable for studies concerned mainly with high values although no applications of this have been found in the literature.

2.04 Extreme Low Flows

Estimates of extreme low flows are required as design criteria for such purposes as water supply, hydro-electricity and navigational development. Early techniques of analysing low flows involved mass curves and behaviour diagrams. Some of the deficiencies in these were outlined by Waitt, (ref. 16), who suggested an improved method based on the "drought characteristic curve."

Some modern methods of reservoir storage analysis (e.g. Gould, ref. 23) utilize frequency estimates of runoff during convenient time units such as years or months but the complete specification of low flows required runoff-duration-frequency data of the type reported by Stall and Neill (ref. 24) and Gannon (ref. 30) and Balek and Holecek (ref. 31).

Statistical analyses are commonly carried out with the annual series, i.e the minimum values occurring in each year. It may be preferable in some circumstances to consider the partial duration series, i.e. all occurrences of very low flows whether they are the minima for particular years or not. This question is discussed in some detail by Stall and Neill (ref. 24), Langbein (ref. 25) and Chow (ref. 9). The distributions of the two series differ considerably for recurrence intervals of one or two years but they tend to converge for recurrence intervals greater than about ten years.

Partial duration distributions are closely related to flow duration curves, as outlined in 2.08.

2.05 Frequency Distribution of Extreme Low Flows

Fisher and Tippett (ref. 26) showed that if the smallest values of many large samples of a particular population are considered together, their distribution is almost (but not completely) independent of the population distribution and approaches one of three limiting functions, called the 1st, 2nd and 3rd asymptotic distributions of smallest values.

This theory was examined by Gumbel (refs. 27, 28 and 29) who suggested that extreme low flows conform to the 3rd asymptotic distribution expressed by:

$$\pi(q) = \exp\left[-\left(\frac{q - \varepsilon}{\theta - \varepsilon}\right)^{k}\right] \qquad \text{..........(2b)}$$

where $\pi(q)$ = probability of an extreme low flow greater than or equal to q during the specified period.

ε = minimum possible flow during the specified period (usually zero), i.e. the limiting asymptotic value.

θ = a characteristic (Normal) low flow, i.e. flow with

k = a constant between 0 and 1 expressing the curvature of the distribution

In the above the return period Y is given by:

$$Y = \frac{1}{1 - \pi(q)} \qquad \text{..........(2c)}$$

Many recent studies of extreme low flows have been based on Gumbel's approach, for example Bernier (ref. 32), (Gannon ref. 30), and Kaczmarck (ref. 33). With this method, unfortunately, the equations indicate an infinite return period for the minimum flow which is obviously inappropriate for arid regions where zero flows may occur quite frequently.

Other possible distributions have been suggested (see Foster, ref. 6), but these are theoretically less satisfactory than Gumbel's. In practice, however, all distributions fitting the available data give approximately the same estimates (statistically) if the extrapolation is relatively small. When large extrapolations are necessary the estimates depend considerably on the adopted or implied minimum flow and its frequency.

In ref. 29, Gumbel has indicated a statistical method of estimating the minimum flow based only on the observed values but it appears that more satisfactory estimates may be possible by utilizing additional information from geohydrological considerations. These would involve estimates of the groundwater recharge and transmission characteristics which impose limits on the possible extreme conditions. The concepts of the "storage flow ratio" (Chapman, ref. 34), "response time" and "recharge ratio" (Langbein, refs. 14 and 35) may be very useful for such studies which would be essentially an extension of the deterministic approach discussed later in sections 6 to 14.

2.06 Flood Runoff.

Engineering structures associated with most water development projects are designed to withstand flood conditions and the various methods of assessing these conditions form a very extensive subject. Only the most relevant aspects can be mentioned.

Flood estimates differ from low flow estimates in that the frequencies of instantaneous peak rates are usually more important than the frequencies of flow volumes for particular durations.

Although the statistical approach to flood estimation has been used for many years, (see Hazen, ref. 2) in the past it has suffered a certain amount of unjustified disrepute, as discussed very thoroughly by Alexander (ref. 11). The development of the meteorological technique of flood maximization was not particularly helpful in this regard, in fact, it tended to divide hydrologists into two strongly opposed schools of thought. One group considered that maximization techniques should completely supercede frequency studies while the other group ridiculed this suggestion. (see for example, the discussion on "Wyangala Dam Inflow Flood Estimates", ref. 37 and also ref. 11).

Much of the controversy is emotional and unnecessary as the two approaches are not mutually exclusive. Each provides some extra, independent information which should all be utilized for the best possible estimates.

2.07 Frequency Distribution of Floods

Fisher and Tippett's asymptotic distributions of smallest values were mentioned in 2.05 with regard to extreme low flows. There are similar distributions for largest values and Gumbell (ref. 28) has suggested the 1st asymptotic distribution of these for floods, i.e.

$$\pi(q) = \exp\left[-e^{k(q-\theta)}\right] \qquad \ldots\ldots\ldots\ldots (2d)$$

where $\pi(q)$ = probability of a flood less than or equal to q during the specified period.

θ = characteristic flood, i.e. with $\pi(\theta) = \frac{1}{e} = .368$

k = a constant between 0 and 1 expressing the curvature of the distribution.

In (2d) the return period Y is given by equations (2c), i.e.

$$Y = \frac{1}{1 - \pi(q)}$$

As the above is a two-parameter distribution it can be plotted linearly on graph paper with appropriately spaced ordinates, (see refs. 27, 28, 38, 39, 40) which provides a convenient means of extrapolating from the observed data to more extreme values. Modifications to this approach have been suggested by Jenkinson (ref. 38) who advocated the 2nd or 3rd asymptotic distributions if the plot is not linear, and by Lieblein (ref. 38) who weighted the observed values according to their frequency to derive the "most probable" linear distributions.

The estimation of confidence limits or "control curves" for Gumbel's method has been examined in some detail by Alexnader (ref. 11) and Kaczmarck (ref. 33).

Several other forms of distribution for floods have been advocated, for example, the Gamma distribution (refs. 12 and 18), log probability (ref. 10), log normal (refs. 11) and cube root normal, (ref. 38). Studies by Hershfield suggest that these all tend to underestimate extrapolated values and there is little theoretical justification for their use under most circumstances, (ref. 38).

In practice, when the sample distribution is not extrapolated very far, the precise mathematical form is relatively unimportant because all methods fitting the observed values give approximately the same result. When a large extrapolation is necessary, however, different forms of distribution give different estimates but in many cases the confidence limits are so wide that the variations are not very significant statistically.

Some of the hydrologists favouring probability methods are very critical of the concept of a maximum flood. There is no comparable criticism of the concept of a minimum non-zero flow, in fact, Gumbel's approach to this subject actually depends on an estimate of this flow.

The postulation of a maximum flood is surely as feasible as the postulation of a minimum (non-zero flow);yet the former is rejected by some hydrologists who readily accept the latter.

There is no strong reason apparent to the author why the maximization method should not be combined with the third asymptotic distribution of largest values, i.e.

$$\pi\ (q) \ = \ \exp -\left[\left(\frac{\omega - q}{\omega - \theta}\right)^{k}\right] \qquad ....2(e)$$

where $\pi(q)$, θ and k are the same as for equation (2d) and ω is the maximum flood estimated by the most reliable means available.

It should be realized, however, that any estimate of ω is subject to various errors and the selection of this value for design purposes does not ensure 100% security. By introducing ω into the distribution it should be possible to reduce the confidence limits but not eliminate them completely.

Hershfield (ref. 37) has developed a probability approach to the estimation of ω and this may also provide a reasonable method of determining the likely values at the extremity of the distribution.

However, a valid criticism of using equation (2e) is that conditions can be envisaged in which ω has a finite recurrence interval, perhaps as low as 200 or 300 years. This possibility is unlikely to be thoroughly investigated until more knowledge and data have accrued on the relevant physical processes. In the meantime, the selection, of suitable values of K in the equation should give close approximations to most feasible distributions.

As equation (2e) involves three parameters it does not necessarily result in a linear plot on Gumbel probability paper which therefore has no special advantages. One possible disadvantage with this paper is that the ultimate maximum value cannot be properly plotted. The same applies if the abscissa represents the logarithm of the recurrence interval as suggested by Alexander (ref. 11). A simple probability scale overcomes the above difficulty and should be satisfactory if only the rarer events are required on the plot, for example events with recurrence intervals exceeding ten years ($\pi (q) < .10$).

2.08 Flow Duration Curves

There are several different forms of flow duration curve, the commonest showing the percentages of time during which various flows are exceeded, (ref. 3). This is somewhat similar to a frequency distribution but normal flows and relatively frequent extremes are given more emphasis than rare extremes. Flow duration curves are particularly useful for the assessment of hydro-electric potential and water supply schemes requiring little artificial storage, (See Foster, ref. 6). They have been used in a number of recent studies of streamflow behaviour (e.g. refs. 31, 41, 42).

Kunkle (ref. 41) separated the baseflow and surface flow of various U.S. streams and derived "baseflow duration" curves which were found to be related to hydrogeologic factors such as channel and bank characteristics, gravel deposits etc. Notable consistencies were observed over wide physiographic regions.

Balek and Holecek (ref. 31) postulated the "unit duration curve" for direct comparisons between different periods of record and also between different catchments. This enabled analyses of drought characteristics in relation to climate and catchment size.

Extensions of streamflow data from correlations of flows of equal percent duration have been carried out by Langbein and Hardison (ref. 13). Similar ideas were used by Reinhart and Eschner (ref. 42) in studying the effects of different forestry practices on streamflow.

Wisler and Brater show that flow duration curves vary considerably when derived from different periods of records as they strongly reflect the effects of wet and dry cycles, (ref. 3). The same authors also deal with the detailed interpretation of the curves and suggest that this is assisted by separating them into homogeneous segments such as "late season flows" and so on.

3. THE ESTIMATION OF RUNOFF STATISTICS FROM RAINFALL DATA

3.01 Rainfall Statistics

Much of the previous discussion concerning the distribution and variability of runoff also applies to the distribution and variability of rainfall, although there are some significant differences that will be mentioned below.

Rainfall data represents samples at particular points while runoff data refers to comparatively large watershed areas. In general the frequencies of point values of rainfall are lower than the frequencies of corresponding values over areas exceeding 10 square miles, the difference being greater for short duration or thunderstorm rainfall than for prolonged wet spells. Empirical methods of allowing for this are given by refs. 3, 43 and 44.

The frequency distribution of annual rainfalls in Australia do not appear to be as skewed as those for annual runoff, in fact they are almost normally distributed in many cases. Maps of median annual rainfall for Australia are practically identical with maps of average annual rainfall (refs.45,46), indicating quite low coefficients of skewness for the whole continent.

A study by Gibbs (ref.47) suggests that the square root of the hourly, daily and monthly rainfall are all normally distributed. The logarithm of daily and monthly rainfall is also assumed to be normally distributed in some circumstances, (ref.6).

The most suitable distribution for extreme low rainfalls appears to be the Fisher-Tippett 3rd asymptotic distribution of lowest values as suggested by Gumbel for streamflow (see equation (2b) and ref. 27).

Extreme high rainfall is usually given in the form of depth-duration-frequency data, typical examples being ref. 48 for the U.S.A. and ref. 49 for Australia. In the U.S.A. work, the extreme values were assumed to conform to the 1st asymptotic distribution (equation (2d) for streamflow) while those for Australia were assumed to have a log probability distribution (see ref. 50). The durations in these studies range from 5 minutes up to 3 days.

Extreme short duration rainfalls up to about 2 hours are usually considered to be caused by thunderstorms and throughout the U.S.A. the 5 minute, 10 minute, 15 minute and 30 minute values are approximately 29%, 45%, 57% and 79% (respectively), of the 1 hour rainfall of the same frequency, (refs. 48, 51). Reich has suggested that these "Hershfield ratios" may be universal and has shown that they also apply to South Africa, (ref. 52). This hypothesis appears to be supported by the data from several Australian cities as demonstrated by the author in the appendix of ref. 49.

Similar ratios for long-duration rainfall are consistent within fairly broad climatic regions but differ considerably between such regions, (see ref. 53). The "frequency ratios," expressing relationships between rainfalls of the same duration but different frequencies, should also be consistent within broad climatic regions for both short and long duration rainfall, as will be discussed in section 4.05. These are directly relevant to the problem of specifying frequency distributions of rainfall.

3.02 A General Rainfall-Runoff Relationship

Ignoring catchment leakage, a very general relationship between rainfall and runoff is expressed by:

$$Q = P - \Delta S - W \qquad \text{..............(3a)}$$

where P = volume of rainfall in a specified time

Q = volume of runoff in the same time

ΔS = change in catchment water storage during the specified time.

W = actual evaporation from catchment during the specified time

In most cases P is known and the other terms are dealt with according to the length of the specified time and the nature of the problem.

3.03 Annual and Seasonal Estimates

Equation (3a) is used to give estimates of mean annual runoff from mean annual rainfall if the appropriate terms represent these quantities. Under such conditions ΔS is omitted because it should average zero over along period of time and mean annual W may be estimated from only a few years of records because it is less variable than either P or Q.

The latter estimate should take into account the tendency for W to be correlated with P, and a fairly obvious method of doing this is demonstrated in 5.02 where an improved annual rainfall-runoff study is presented. Other studies of this type have been made by McCutchan (ref. 54) McArthur and Cheney (ref. 55), Alexander and Sutcliffe (ref. 15), Turc (ref. 56), Thornthwaite (ref. 57), and the U.S. Geological Survey, (ref. 58).

In individual years, equation (3a) can be used to estimate the annual runoff from a given annual rainfall as shown by Cordery, (ref. 59). This technique also ignores ΔS as it is usually small compared with the other values particularly when a "water year" is selected. Some allowance should be made, however, for the distribution of rainfall within the year because a few heavy storms tend to produce more runoff than a number of lighter storms with the same total rainfall. For this purpose Cordery suggests the additional parameter of "maximum percentage of rain in one month." The "maximum rain in one month" or the total number of wet days in the year may also be suitable.

Similar comments apply when estimates are required on a seasonal basis, although in some cases it may be necessary to adopt a seasonal value of ΔS rather than assume an average value of zero.

3.04 Monthly Estimates

In estimating monthly runoff from monthly rainfall, ΔS is more important and more difficult to allow for. Any statistical techniques of this type, however, would probably not be directly concerned with the separate values of ΔS and W.

In some techniques for estimating the monthly runoff from particular values of monthly rainfall, ΔS is virtually lumped together with W and their combined effects are related to variables such as rain in previous month, number of wet days, initial groundwater flow, temperature and so on. Examples of this type of study have been reported by Sharp et alia (ref. 109) and Samra (ref. 60).

Somewhat more rational approaches have been used by McDonald (ref. 61) and Lewis and Burgy (ref. 62) who endeavoured to calculate complete water budgets on a monthly basis. In these studies W was estimated mainly from meteorological data and ΔS was related to rainfall and watershed storage characteristics.

3.05 Storm and Short-Period Estimates

Statistical estimates of storm runoff are usually concerned with representative extreme values rather than mean values. The peak rate of runoff is of prime importance and this may be related to rainfall-duration-frequency data by the so-called "rational method", expressed as follows:

$$q_f = CAp_f \qquad \text{..........(3b)}$$

where q_f = peak discharge in cfs with frequency f

p_f = average rainfall rate in ins/hour with frequency f for duration t_c.

A = area of watershed in acres.

C = coefficient of runoff, usually between 0.3 and 0.9.

t_c = critical duration i.e. the rainfall period that is most effective in producing a high peak discharge for the given watershed.

The coefficient of runoff allows for the combined effects of loss rates and attenuation due to storage. The rational method is generally only advocated for design floods on smaller watersheds because in these the storage attenuation is less important and C therefore tends to be independent of A.

Design flood estimates are also made by the unitgraph-loss rate method which may be expressed by:

$$q_f = h_m (p_f - r_f) \qquad \text{..........(3c)}$$

where q_f, p_f and t_c are the same as for (3b)

h_m = peak ordinate for unitgraph of critical duration t_c.

r_f = an appropriate or representative loss rate in ins/hr.

For conservative estimates, r_f is sometimes neglected, in which case the unitgraph-loss rate method, as expressed above, becomes equivalent to the rational method, h_m being equal to CA.

There are possible refinements of (3b) and (3c) to allow for the effects of varying rainfall intensities within t_c, but it is doubtful that the extra work involved can be convincingly justified (see Coulter, ref.63). Various other modifications of these two methods have been proposed recently, for example by Reich (ref.64), Wood (ref.65), and Harold (ref.66)

The estimation of C and t_c are challenging problems that cannot be examined in this report. Various approaches are suggested in the above references and by Turner (ref.67), Cook (ref.68) and "Australian Rainfall and Runoff" (ref.43). Most text books also present methods of estimating these parameters for different watershed and storm conditions, (e.g. Refs. 1 and 4).

The unitgraph-loss rate method may be used for constructing complete design hydrographs, if required. In such cases equation (3c) is modified slightly by substituting the hydrograph ordinates q_1, q_2 .....etc. for q_f and the corresponding unitgraph ordinates h_1, h_2 .....etc. for h_m.

Rigorous deterministic techniques for estimating short-term runoff from rainfall data are complex because of the inter-relations and variability of the physical processes involved. These techniques may be regarded as "complete rainfall-runoff models" and their detailed examination forms the main part of the report in sections 6 to 14.

Complete rainfall-runoff models may be used to synthesize long periods of runoff data if sufficient information is available to enable the derivation of the required parameters and relationships. Once a long synthetic record has been obtained, it provides a basis for general estimates of runoff statistics, i.e. annual means, monthly means, extreme floods or other required values.

Techniques have also been developed for generating hypothetical runoff records directly from rainfall statistics if there is sufficient data on the inter-correlations between the statistics themselves. These are the "Markov techniques" described by Fiering (ref. 20) and Federov (ref. 19) as mentioned previously in 2.03. They differ considerably from the complete rainfall-runoff models as they are highly empirical and not directly concerned with the detailed physical processes.

4. ESTIMATES WHEN RUNOFF AND RAINFALL DATA ARE BRIEF

4.01 The General Problems

The previous sections have been primarily concerned with circumstances in which either runoff or rainfall records are available over a reasonably long period, say 40 years or more. Similar techniques can be used when the records are much shorter than this but the sampling errors are sometimes so great that the results are misleading. It is often possible, however, to reduce these errors, as will be shown in 4.03 and 4.05.

When no records exist at all in an area it is usually necessary to estimate runoff by comparison with similar or neighbouring areas for which data is available. The selection of such data and its modification require decisions that are unavoidably subjective. Some ideas for a more systematic approach to this problem will be presented in 4.06.

4.02 Normal Runoff and Rainfall

The oft-quoted studies of Binnie (ref.66) have led to the general belief that about 30 years of rainfall record are sufficient to give a mean annual value within one or two percent of the "true mean annual value", i.e. as would be obtained from an indefinitely long period of records. More recent studies by Hidore (ref.70) suggest that this is reasonable in some climatic regions but in other parts of the world the 50 year mean can differ from the 100 year mean by as much as 10%.

The importance of the length of record on estimates of rainfall statistics is also dealt with extensively by McDonald and Green (ref.69), and briefly by Chow (ref.9, under "Normalcy Tests.")

As runoff values are relatively more variable than rainfall values it appears that the length of record is even more important in the estimation of runoff statistics. In this regard Wisler and Brater (ref. 3) suggest that it is not uncommon for the ten year mean to differ from the fifty year mean by 30%. Deviations of a similar magnitude evidently occur when statistics such as the one or two year flood are required, (see Hayes, ref. 71).

The errors in estimates from brief records are essentially sampling errors which may be regarded as arising from two sources viz. random fluctuations and temporal correlations.

Random fluctuations are sometimes of a highly localized nature, for example the unsurpassed 20 inches of rainfall in 24 hours at South Head, Sydney, in October 1844. In other circumstances they may apply to thousands of square miles, for example the record floods occurring simultaneously in a number of streams as a result of the February 1956 storm over most of N.S.W.

Temporal correlation, or the tendency for similar events to occur closely together in time, generally applies with some consistency to wide climatic regions. Thus the flood records for most of N.S.W. streams have an extraordinary number of large values within the period 1949 to 1956 inclusive.

Matalas and Langbein (ref. 72), Langbein and Harbeck (ref. 70), Alexander (ref. 74), Urban (ref. 75) and Whitmore (ref. 76) all deal with aspects of temporal correlation which is also referred to as "serial correlation", "autocorrelation," "persistence" and "coherence." In general these studies show that the greater the degree of temporal correlation the less effective is the length of the records.

The factors responsible for temporal correlation are theoretically grouped under the terms "periodicity" and "trend" (see O'Mahony, ref. 77) although these cannot usually be readily separated. There is an extensive literature on periodicity in relation to rainfall and runoff estimation, some relevant studies including those of Marvin (ref. 78), Laszloffy (ref. 79), Leeper (ref. 80), Somov (ref. 94), Rodier (ref. 81), Yevdjevich (ref. 82) and Dixley (ref. 83).

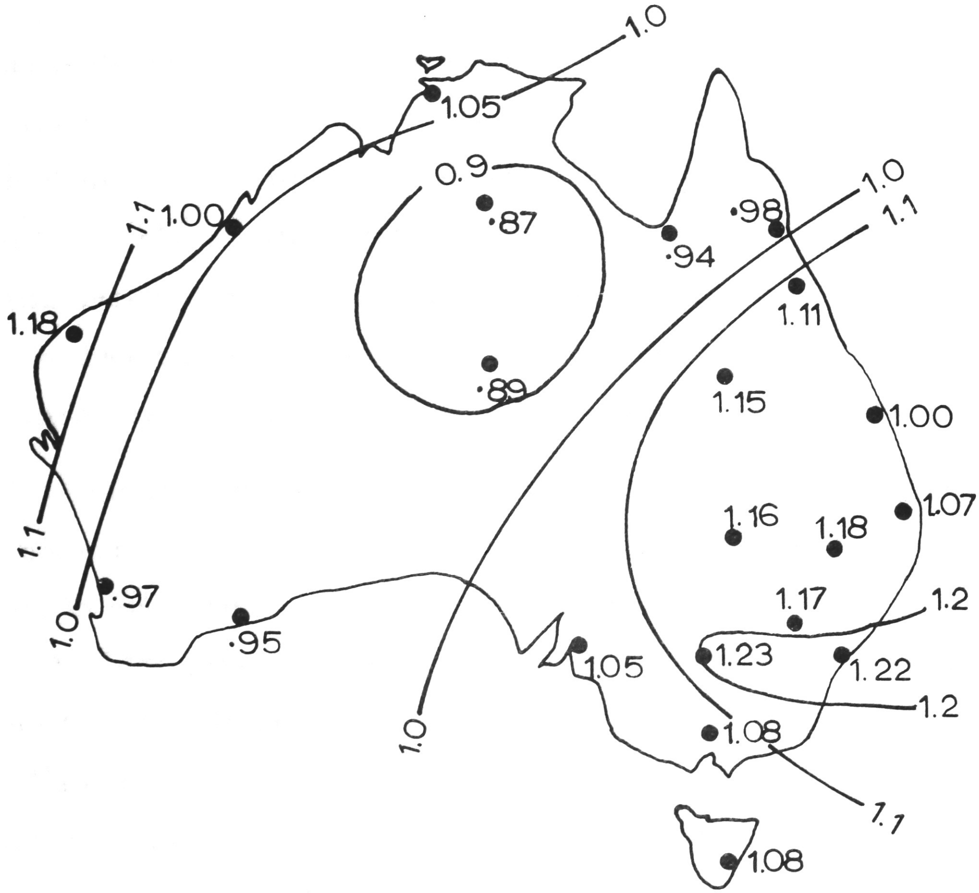
4.03 Period Bias

The ratio of a statistic estimated from a particular period of data to the value obtained from a standard (or infinitely long) period of data is conveniently called the "period bias." For a particular statistic and period this bias is expected to be reasonably consistent over wide areas unless local random fluctuations are of prime significance. Therefore the unknown "error" in an estimate from a brief record at a particular place may often be assessed from the comparable known "error" at another place with a complete record over the standard period.

The above is demonstrated in fig. 1 which shows the mean annual rainfall period bias for 1954-63 for a number of Australian stations. The adopted standard period is 1924 to 1963 inclusive (40 years).

It is evident from fig. 1 that the bias tends towards similar values over wide areas. Regional averages or interpolated values can be assumed at any place with a 1954-63 record and these should provide a basis for improved estimates if longer records are not available. A specific example of this technique will be given in 5.01.

A similar idea is implied by the double mass curve technique of testing and adjusting data (ref. 8). Examples of this type of adjustment have been reported by Chapman (ref. 114), and in the design estimates for the Snowy Mountains Hydro-Electric scheme (ref. 8).



1954-63 PERIOD BIAS FOR
MEAN ANNUAL RAINFALL
Standard period: 1924-63

FIG. 1

4.04 Extreme Runoff and Rainfall

The period bias concept could be applied to the estimation of extremes such as the 50 year flood from brief records but it would probably be unsatisfactory for the following reasons:

(a) The estimation of extremes is largely dependent on one or two extraordinary events that may be highly localized.

(b) The estimation of extremes from brief records requires the extrapolation of frequency distributions ~~which implies~~ doubtful assumptions concerning the forms of the distributions. Extra errors are therefore introduced by this procedure and greater variability of the calculated bias may be expected.

As indicated in earlier sections, the assumed form of the frequency distribution is one of the major factors in estimating extremes and it is of greater importance when records are shorter. The form of the distribution reflects the characteristics of the phenomena causing the extremes and it seems reasonable to expect that these characteristics do not change abruptly from place to place unless there are abrupt changes in topography or other relevant factors. This "spatial correlation" is conveniently utilized by means of the "frequency ratio" which was mentioned in 3.01.

4.05 Frequency Ratios

The ratio of the 100 year value of rainfall or runoff to the 2 year value may be called the 100:2 year frequency ratio. This is a concise way of expressing part of the frequency distribution and simplifies the analysis and comparison of extremes. Other parts of the distribution are expressed by ratios such as the 40:2, 50:2, 200:2, and so on.

The 1, 5 or 10 year values can be used as a base instead of the 2 year but the latter has the following advantages:

(a) It is estimated with greater accuracy than the 5 and 10 year values.

(b) It is not subject to the graphical plotting difficulties of the 1 year value (for example on Gumbel Paper). The latter is also more remotefrom, and less likely to be related to the relevant extreme values.

A period bias adjustment should be used, if necessary, for reliable estimates of the 2 year event from brief records.

McIllwraith (ref. 50) used a type of frequency ratio in his analysis of rainfall data for N.S.W. Following some preceding U.S. studies and based on rather scanty data he derived relationships between these ratios, the rainfall duration and the "standard deviations of the logarithm of the primary 24 hour rainfall." Later studies by Hershfield (ref. 48), Stewart (ref. 84), and Reich (ref. 52) suggest that the complexity of McIllwraith's approach is hardly justified in view of the sampling errors and assumptions involved. These researchers used frequency ratios that varied with rainfall duration but areal variations were accounted for on a broad regional basis.

A recent runoff study involving frequency ratios was reported by Benson (ref. 85) who examined flood peaks over a large part of the U.S.A. The results of this work indicated that rare floods on any single watershed can be estimated more confidently from appropriate average frequency ratios of several comparable watersheds. When obtaining such averages, however, it is important that the individual values are reasonably independent and not unduly influenced by a particular storm common to several watersheds, otherwise the estimate is little better than from a single record.

4.06 Hydrogeography

The previously suggested approaches for reducing the deficiencies in brief records depend largely on the tendency for hydrological phenomena to have consistent and therefore predictable characteristics within certain geographical localities. This is also referred to as "spatial", "areal" and "inter-station" correlation and the associated field of study might be designated "Hydrogeography".

"Hydrogeography" is hereby defined as the study of the distribution of hydrological phenomena over the earth's surface. (c.f. "Geography" ref. 86). It is concerned with the general location, description and classification of phenomena, all of which may require statistical or numerical specification. It is particularly concerned, however, with the qualitative integration of whole complexes of variables so that they may be analysed on a relatively broad scale.

Hydrology differs from other natural sciences in that it is elaborately quantitative with a weak qualitative structure. Systematic description and classification are fundamental to any true science and it is hoped that the naming of this aspect of hydrology will give it more prestige and encourage its recognition as a unified field of study.

A basic problem in hydrogeography is to ascertain the extent to which hydrologic information from one locality can be used in other localities. This may involve the areal extrapolation of data within countries or the transference of data from continent to continent with appropriate modifications for local conditions.

This type of problem may be approached through the systematic analysis of climatic and hydrologic regions.

4.07 Climatic and Hydrologic Regions

The term "region" is used rather loosely in everyday language but in geography it has a special meaning, viz "an area of the earth that possesses within its boundaries a comparatively high degree of uniformity in a particular characteristic" (ref. 86 and ref. 87). Thus, there are "cultural regions," "political regions" and so on.

Various types of climatic regions have been studied by geographers and an excellent survey of these is given by Gentili (ref. 88). A system developed by Miller (ref. 89) seems to be particularly relevant for hydrologic work as it is based on the genetic factors of climatic variations and shows the general areas affected by the characteristic moist air masses. These air masses may be regarded as determining the "potential" rainfall characteristics over wide regions.

The above "potential rainfall" is modified considerably by local factors such as topography and proximity to moisture source which result in typical patterns of distribution that may be sub-divided into "rainfall regions" (see the Appendix at end of this report for complete definitions).

Natural features are frequently inter-related to a high degree, particularly climate, vegetation, soils, topography and potential land use. Geographers therefore commonly refer to "physiographic" or "natural" regions in which there are consistent recurring patterns of features. Such patterns are utilized by the C.S.I.R.O. method of land classification which subdivides areas into "land systems" and "land units" (see refs. 90, 91, 92). The land unit is a basic "micro-region" that is characterized by a particular soil-vegetation complex or some other dominant feature relevant to land use.

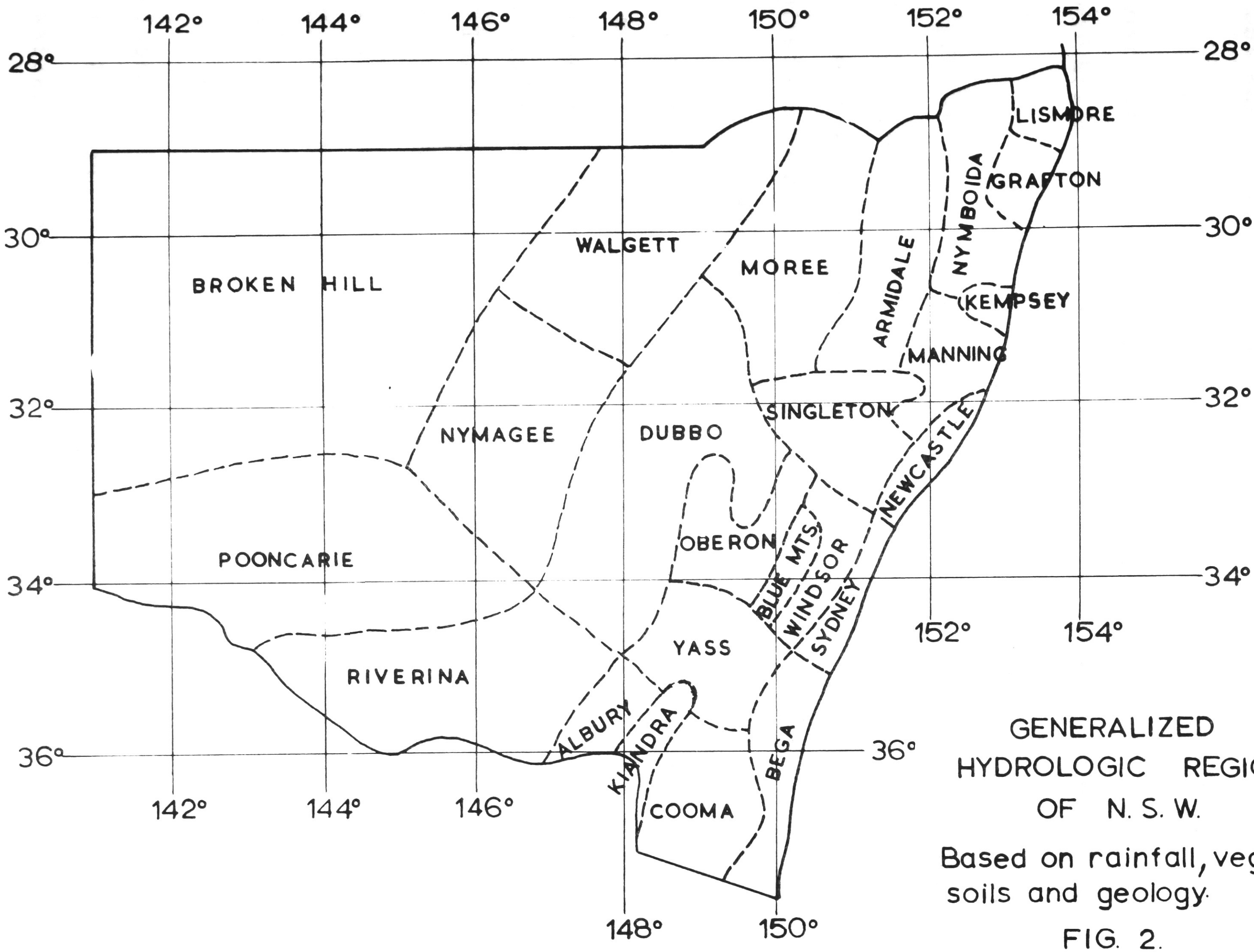
The above suggests the adoption of basic "hydrologic units" which may be defined as areas of land generally less than 200 square miles and having comparatively unitform hydrological characteristics that distinguish them from adjoining units.

The boundaries of hydrologic units indicate approximate limits of factors that have special effects on rainfall-runoff relations. Such factors may include urban areas with extensive impervious surfaces, particular vegetation-soil complexes, particular geological formations or types of land modification. Considerable knowledge of the significance of all of these factors in the hydrologic cycle would be necessary for the proper definition of units.

In general, rainfall regions are larger than hydrologic units and it is therefore suggested that the term "hydrologic region" be applied to a group of units corresponding approximately with a rainfall region. Hydrologic regions would therefore tend to have fairly consistent rainfall characteristics but the variability in runoff characteristics within them would largely depend on the differences between the constituent units.

Fig. 2 shows a system of hydrologic regions for N.S.W. based on rainfall, vegetation, soils and geology. The definition of the boundaries was not as subjective as it may appear because significant changes in these factors tended to coincide in most areas.

The above indicates, therefore, that hydrogeography should be concerned with broad climatic regions as major groupings of hydrological factors. It is also concerned with the subdivision of these into hydrologic regions to account for important variations in rainfall distribution and the further subdivision of regions into units to account for variations in watershed surface characteristics.



GENERALIZED
HYDROLOGIC REGIONS
OF N. S. W.

Based on rainfall, vegetation, soils and geology.

FIG. 2.

The a priori definition of these regions in the absence of extensive rainfall-runoff records should provide a more systematic basis for areal extrapolation of data, for example by means of the period bias and frequency ratio techniques described previously.

Reasonable estimates for ungauged watersheds are also possible with this type of approach, as will be demonstrated in 5.02 and 5.03.

4.08 Hydrogeography and Data Collection

The ideas and principles of hydrogeography are particularly relevant to the International Hydrological Decade which is assessing water resources on a world-wide scale. They should be very useful in undeveloped countries where maximum use must be made of sparse data and a sound basis is required for the economical expansion of data networks.

The extreme variability of hydrological phenomena presents many difficulties in obtaining adequate measurements at reasonable cost. Authorities responsible for data collection are usually restricted in their resources which therefore tend to be concentrated in areas where specific projects have been planned. Other areas are invariably given a low priority with the consequence that most water resource projects are ultimately designed on inadequate data. This wasteful policy is not confined to smaller countries, as pointed out recently by Hidore (ref. 70) who deplored the lack of streamflow records for the extensive southern and Great Plains regions of the U.S.A. (see also McCall, Ref. 93).

The above position may be improved by a carefully designed network of "representative basins", as advocated for the International Hydrological Decade. It is a hydrogeographical problem to assess the main types of physical environment to be represented by the basins.

5. EXAMPLES OF IMPROVED TECHNIQUES FOR STATISTICAL ESTIMATES

5.01 Mean Annual Rainfall in the Central Tablelands of N.S.W.

The distribution of mean annual rainfall throughout the Central Tablelands was required as part of a comprehensive Rainfall-Runoff study for watersheds comparable with those at Lidsdale State Forest (ref. 189).

Detailed maps of mean annual rainfall were available (refs. 95 and 96) but an examination of these suggested that they were based on either:

(a) A small number of stations with means calculated for a standard period of data

Or

(b) A large number of stations with unadjusted means calculated for the full period of records at each station.

Neither of these was considered suitable because detailed, accurate values were required and it was necessary that they be strictly comparable from place to place. The data from 44 stations were therefore analysed in three groups:

(a) The 40 year period from 1924 to 1963 for 29 stations.

(b) The 19 year period from 1945 to 1963 for an additional 15 stations, all of which had records that commenced after 1935.

(c) The 19 year period from 1945 to 1963 for the 29 stations of (a) above.

The mean annual rainfalls were calculated for each of these groups and the 1945-63 values of period bias for (c) were plotted in fig. 3, adopting 1924-63 as the standard period. It should be noted that the bias was fairly consistent within regions.

The average values of bias within each region were used to make appropriate adjustments to the calculated mean annual rainfalls of group (b) and the final values of mean annual rainfall were plotted on fig. 4. It should be noted that the resulting isohyetal pattern differs from that of the Weather Bureau maps, particularly in the vicinity of Helensburgh, Kurrajong and Oberon.

5.02 Mean Annual Runoff in the Central Tablelands of N.S.W.

There are 23 gauged watersheds in the regions concerned, varying in size from 6 square miles to about 3000 square miles. Their usable records vary in length from 5 up to 40 years.

The mean annual runoff and corresponding mean annual rainfall were calculated for the full period of streamflow data available at each gauging station. These values are set out in table 1 and plotted on fig. 5. It may be observed that the plotted points of fig. 5 form three distinct groups viz:

(a) The high rainfall Sydney region.

(b) The high rainfall Blue Mountains region.

(c) The moderate to low rainfall regions, i.e. Windsor Yass and Oberon.

Average curves drawn through each of the above groups should therefore enable reasonable estimates of runoff for any ungauged stream in the region if the mean average rainfall can be calculated e.g from fig. 4.

The curves also enable the estimated means to be adjusted to correspond with a standard period of records rather than the actual period of records. For example the 7 year record for the Fish River at Oberon showed a mean annual rainfall of 34.6 ins. and a mean annual runoff of 5.7 ins. which is 5 percent lower than the runoff given by the regional curve for the same rainfall. The 40 year mean annual rainfall for the Fish River at Oberon is 33 ins. and the corresponding regional runoff is 5.5 ins. The required 40 year estimated mean is therefore assumed to be 5 percent lower than 5.5 ins., i.e. 5.3 ins.

The estimated 1924-63 mean annual runoff was calculated as above for each watershed and these values, together with corresponding estimates for the ungauged areas, were used to construct the runoff map of fig. 6.

5.03 Regional Estimates with One or Two Watersheds

The regional curves of fig. 5 can be estimated from the records of one or two representative watersheds if they are of reasonable length.

This technique is demonstrated in fig. 7 where the plotted points represent mean values calculated from groups of six or seven years for the Avon River, Burralow Creek and the Macquarie River, as shown in table 1. The years were grouped in chronological order which gave adequate ranges of values in these examples. In some cases, however, it may be necessary to group the annual values in order of magnitude otherwise only a small segment of the curve is defined.

It should be appreciated that regional runoff relations such as the above should give reasonable estimates if the hydrologic units do not differ greatly throughout the region. In cases where the units are highly variable it may be necessary to derive separate curves for similar groups of units rather than for entire regions.

5.04 The Synthesis of Long Records

As mentioned previously in 3.05 complete rainfall-runoff models have been developed to enable the synthesis of long runoff records from rainfall records and these may then be used for improved statistical estimates.

Such techniques are still relatively new and are undoubtedly amenable to many further improvements.

Several rainfall-runoff models will therefore be outlined in sections 6 to 10 and a critical analysis will be attempted in sections 11 and 12. Following this analysis an improved model will be developed and tested in sections 13 and 14.

TABLE 1

REGION	CATCHMENT	PERIOD OF DATA	ANNUAL AVERAGES FOR PERIOD P	Q	EST. AVERAGES 1924-63 (incl.) from fig. 5 P	Q
			ins.	ins.	ins.	ins.
SYDNEY	Avon R. at Avon Dam	1924-63	53.7	21.9	54	22
	" " "	1924-33	47.8	15.4	"	"
	" " "	1934-43	43.5	13.9	"	"
	" " "	1944-53	57.2	27.2	"	"
	" " "	1954-63	66.4	31.0	"	"
	Cordeaux R. at Cordeaux Dam	1924-50	54.6	21.0	60	26
	Cataract R. at Cataract Dam	1924-63	57.5	25.6	58	26
	Woronora R. at Woronora Dam	1931-50	49.9	18.0	55	23
	O'Hares Ck. at Wedderburn	1931-51	48.3	15.8	51	19
	Nepean R. at Nepean Dam	1924-63	46.1	14.9	46	15
	Waratah Rivulet at Waterfall	1926-41	50.5	15.7	57	21

TABLE 1 (Cont'd)

REGION	CATCHMENT	PERIOD OF DATA	ANNUAL AVERAGES FOR PERIOD		EST. AVERAGES 1924-63 (incl.)	
			P	Q	P	Q
			ins.	ins.	ins.	ins.
BLUE MTS.	Burralow Ck. at Kurrajong	1927-61	48.9	12.0	49	12
	" " "	1927-35	47.2	10.4	49	12
	" " "	1936-43	42.7	7.9	49	12
	" " "	1944-51	55.7	17.7	49	12
	" " "	1952-61	50.0	12.4	49	12
	Grose R. at Recorder	1946-58	54.2	14.6	48	11
	Cox's R. at McMahon's	1943-50	42.7	8.4	39	6
	Colo R. at Upper Colo	1924-63	38.6	5.5	36	5
OBERON	Fish R. at Oberon	1944-50	34.6	5.7	33	5
	Fish R. at Tarana	1955-63	41.5	10.6	34	6
	Slippery Creek at Damsite	1955-62	38.2	6.1	35	5
	Middle R. at Marangaroo	1925-29	33.6	6.3	33	6
	Cox's R. at Bathurst Rd.	1951-63	30.2	5.6	28	5
	Turon R. at Sofala	1948-63	32.2	6.2	28	6
	Macquarie R. at Bathurst	1933-63	32.0	4.5	32	5
	" " "	1933-39	26.8	2.1	32	5
	" " "	1940-47	26.3	1.7	32	5
	" " "	1948-55	38.6	7.8	32	5
	" " "	1956-63	36.7	6.1	32	5
OTHER REGIONS	Warragamba R. at Damsite	1924-50	31.9	4.3	33	5
	Abercrombie R. at Caves	1931-63	32.4	4.4	32	4
	Wollondilly R. at Pomeroy	1947-56	38.7	8.0	33	6
	South Ck. at Mulgoa Rd.	1956-61	35.0	6.5	30	5
	Shoalhaven R. at Welcome Reef	1939-53	28.8	9.4	-	-
	Hunter R. at Singleton	1943-54	27.1	2.9	-	-
	Richmond R. at Casino	1952-60	49.0	15.5		-

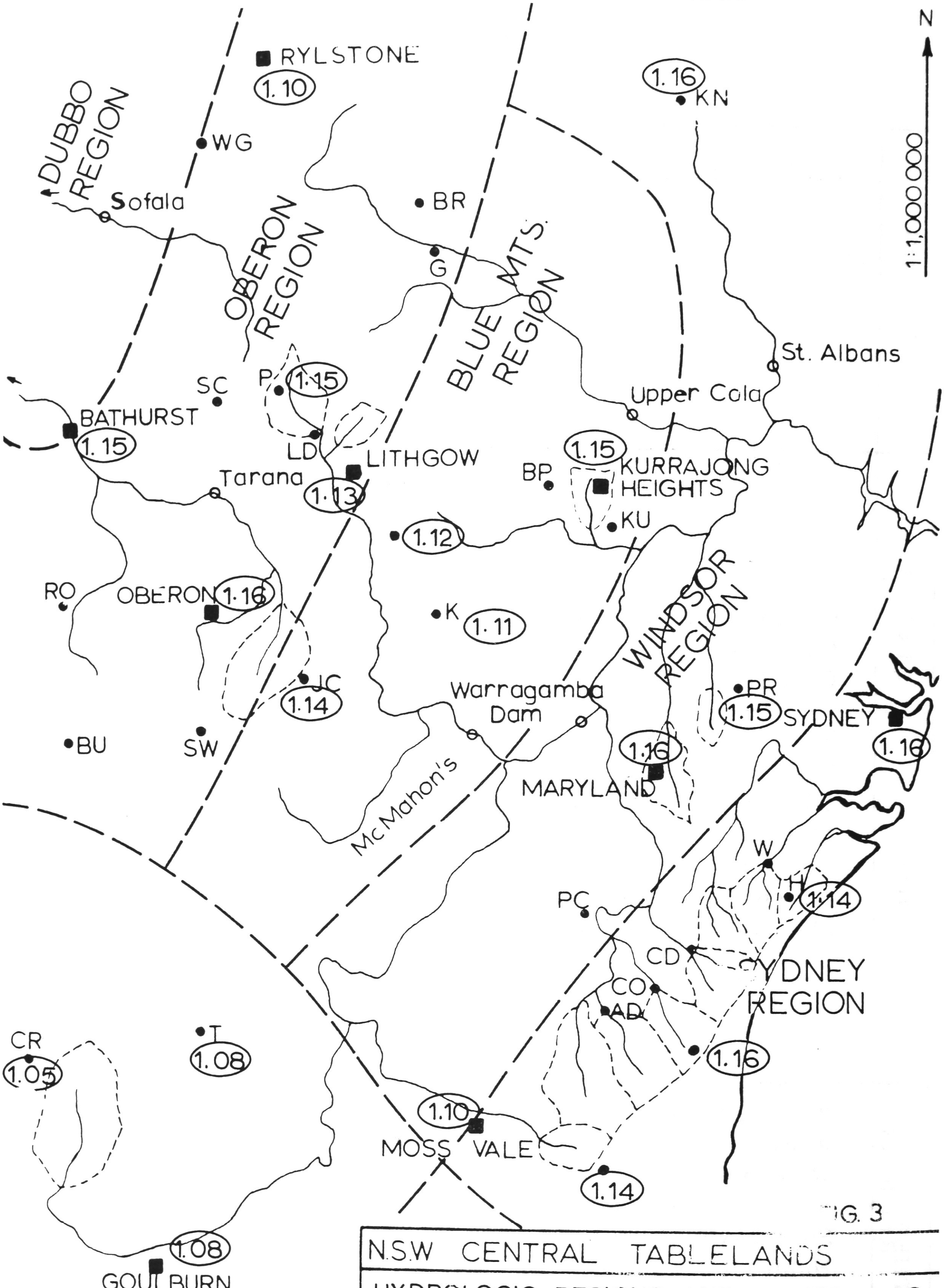


FIG. 3

N.S.W CENTRAL TABLELANDS

HYDROLOGIC REGIONS AND PERIOD BIAS FOR RAINFALL STATIONS WITH LONG RECORDS

Numbers indicate 1945-63 period bias for mean annual rainfall.

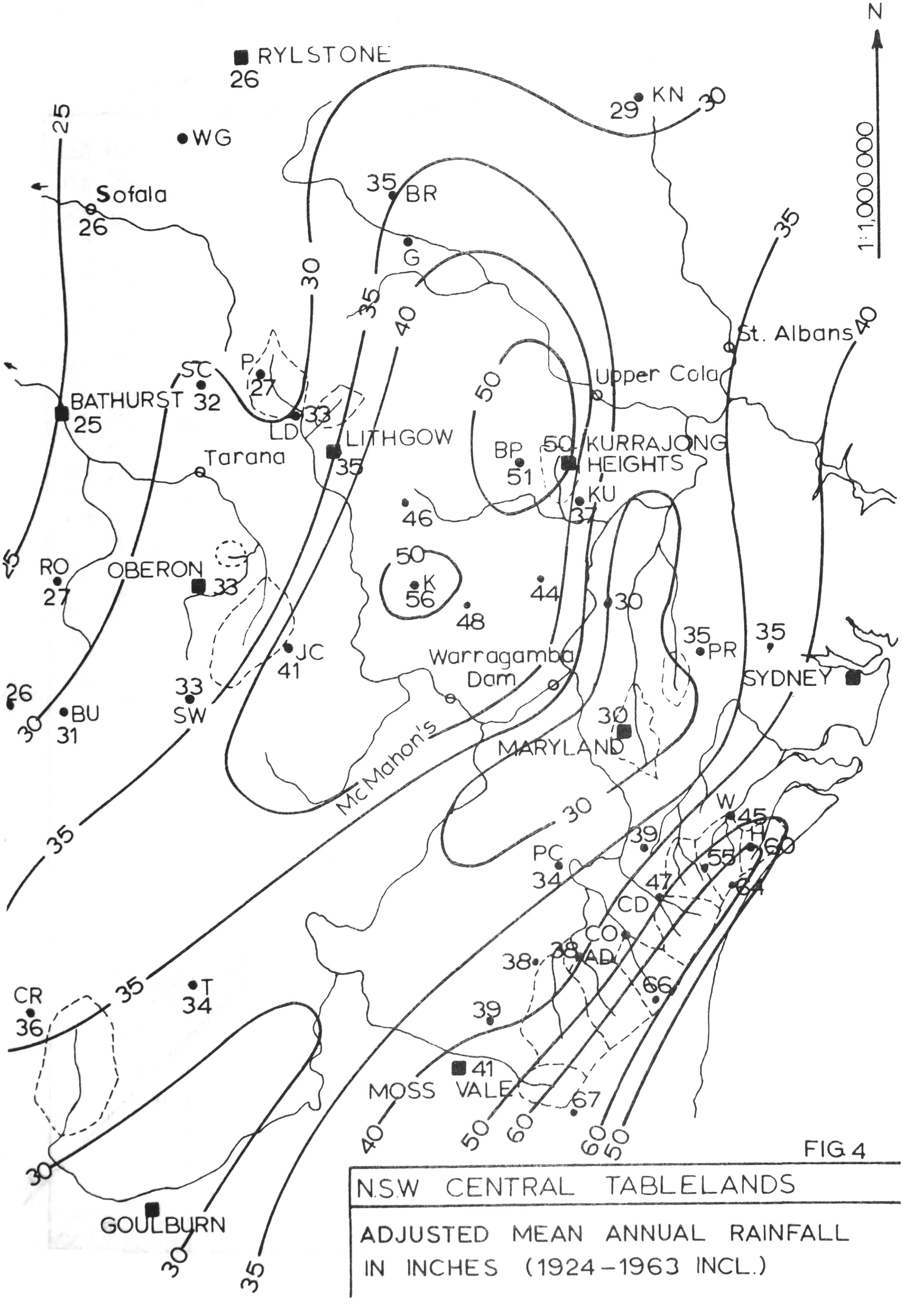
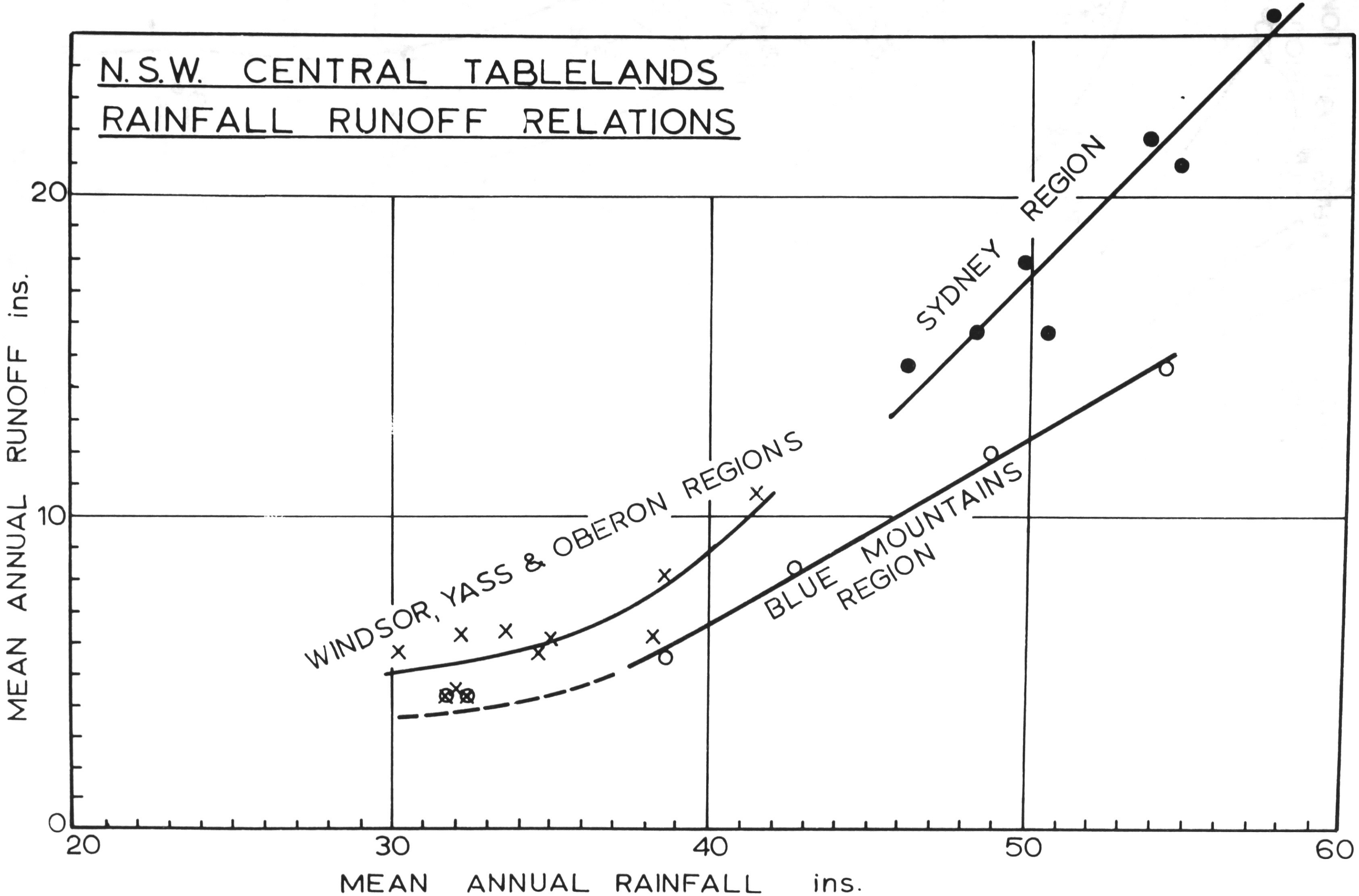


FIG 4

N.S.W CENTRAL TABLELANDS

ADJUSTED MEAN ANNUAL RAINFALL IN INCHES (1924-1963 INCL.)



N.S.W. CENTRAL TABLELANDS
RAINFALL RUNOFF RELATIONS
SYDNEY REGION
WINDSOR, YASS & OBERON REGIONS
BLUE MOUNTAINS REGION
MEAN ANNUAL RUNOFF ins.
MEAN ANNUAL RAINFALL ins.
0
10
20
20
30
40
50
60

FIG. 5

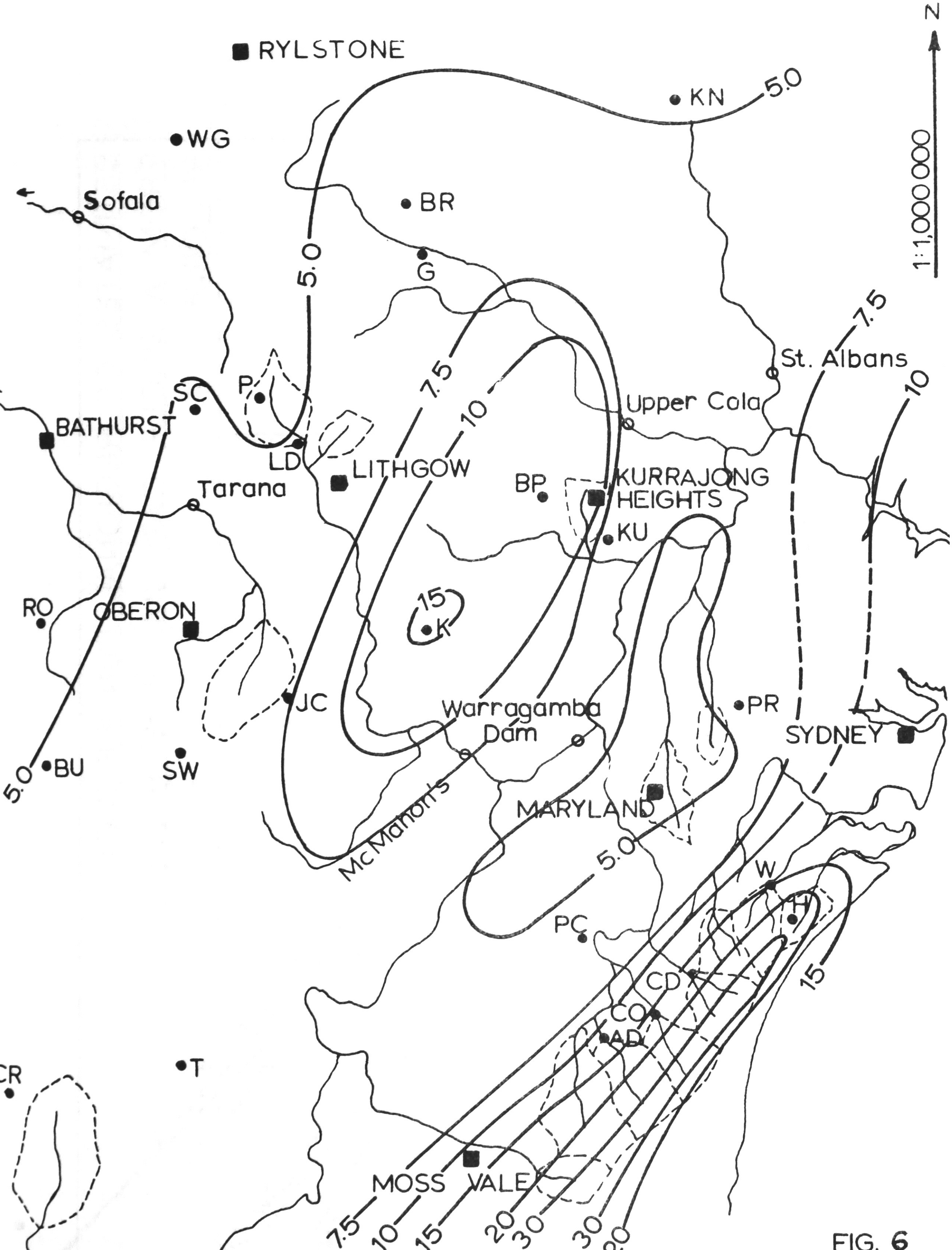


FIG. 6

N.S.W CENTRAL TABLELANDS

ESTIMATED MEAN ANNUAL RUNOFF IN INCHES. (1924–63 INCL.)

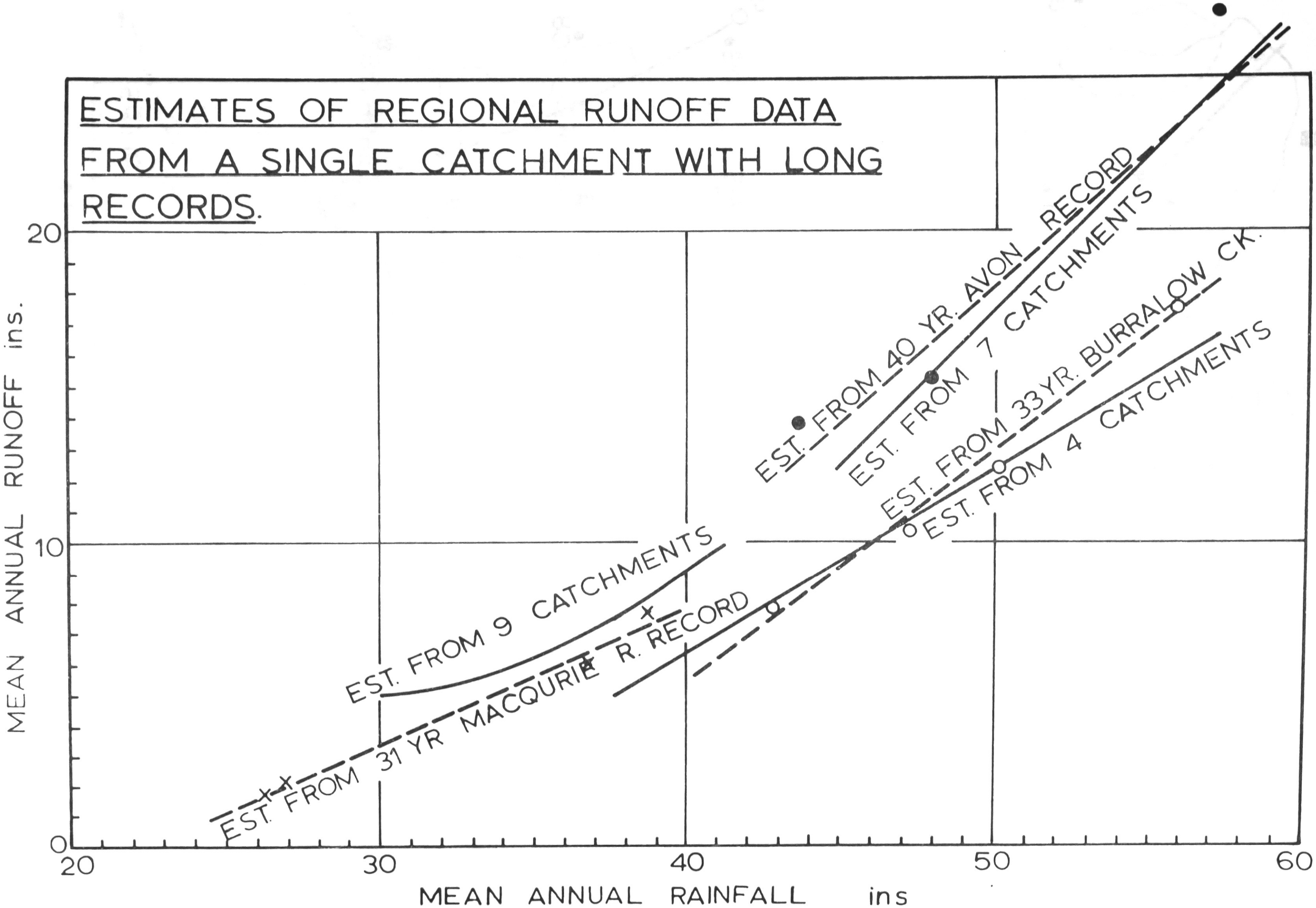

ESTIMATES OF REGIONAL RUNOFF DATA
FROM A SINGLE CATCHMENT WITH LONG
RECORDS.
EST. FROM 40 YR. AVON RECORD
EST. FROM 7 CATCHMENTS
EST. FROM 33 YR. BURRALOW CK.
EST. FROM 4 CATCHMENTS
EST. FROM 9 CATCHMENTS
EST. FROM 31 YR MACQUARIE R. RECORD
MEAN ANNUAL RUNOFF ins.
MEAN ANNUAL RAINFALL ins
0
10
20
20
30
40
50
60

FIG. 7

6. COMPLETE RAINFALL-RUNOFF MODELS

6.1 Introduction

The estimation or prediction of runoff from rainfall may be divided into two parts:

(a) The estimation of the rainfall-excess which is essentially the residual when abstractions are made from the gross rainfall to allow for infiltration and other so-called "lossess."

(b) The conversion of rainfall-excess to hydrographs of streamflow at the catchment outlet.

Although all of the errors in (a) are carried through to (b), engineering hydrologists in the past have given more attention to the latter for which several satisfactory techniques are now available. (e.g. unit hydrographs and storage routing.) No techniques of comparable precision are yet available for estimating rainfall-excess.

In relatively recent years however, a certain amount of progress has been made with this problem through the application of high-speed computers to flood forecasting and the extension of streamflow records. These methods virtually simulate mathematically the entire rainfall-runoff cycle as most of the processes of the cycle have some effect on the rainfall-excess.

The general features of simulation techniques have been reviewed by Amorocho and Hart (ref. 97) who suggest that such techniques are profoundly influenced by our knowledge and interpretation of the relevant physical phenomena. They also suggest that the present state of knowledge of hydrologic phenomena is so inadequate that numerous emotional controversies are apt to arise regarding the relative merits of different approaches. While this may be true to a certain extent, some of the current controversies appear to have been resolved by recent developments in soil and plant physics, as will be discussed later.

Greater difficulties are due to the extreme variability of hydrologic factors in time and space, necessitating a vast amount of information for their complete specification. It appears that approximations and simplifying assumptions will always be required because the amount of basic data that can be collected and analysed is invariably restricted, even with the advent of high-speed computers.

Some of the new rainfall-runoff models will therefore be outlined with special emphasis on the estimation of rainfall-excess. The main underlying assumptions will be examined in the light of relevant studies of individual processes such as infiltration and transpiration.

6.2 The General Rainfall-Runoff Cycle

It is convenient to consider hydrologic processes within two phases:

(a) The drying phase when there is no rainfall and the dominant process is evaporation.

(b) The wetting phase when rainfall is the dominant process.

The term "moisture status" may be used for the dryness or wetness of a catchment with particular regard to its capacity for absorbing rainfall and preventing this rainfall from becoming runoff. The specification and accounting of the moisture status are the fundamental problems in the procedures to be described.

The concept of "potential evaporation" is necessary for modern hydrological studies (ref. 98 to 101). Briefly, it is intended to be independent of factors associated with the water being evaporated, i.e. the water temperature, rate of supply and degree of exposure. In this paper no distinction is made between the terms "potential evaporation" and "potential evapotranspiration" for reasons outlined in 11.01.

Potential evaporation corresponds closely with the evaporation from large lakes or short, well-watered grass and may be calculated fairly rigorously from meteorological data (refs. 100 and 101). Reasonable relative values are obtained from standard pan readings which may be converted to approximate absolute values by applying correction factors that vary with the type of pan and season of year. The actual rate of moisture loss from a catchment by evaporation is usually less than the potential rate, particularly when the catchment is in a dry condition.

During the wetting phase, evaporation diminishes in importance and the catchment gains water from rainfall through the processes of interception, depression storage and infiltration.

Most hydrology text-books still present the "infiltration theory" which asserts that the rate of generation of surface runoff is equal to the excess of rainfall intensity over infiltration capacity. This may be reasonable for an isolated point on a catchment but the theory has serious deficiencies when applied to a typically heterogeneous area, as will be discussed in section 13. Concepts such as "retention," "recharge," "absorption" and "initial loss" are now frequently used in rainfall-runoff studies to describe processes that largely involve infiltration but imply significant departures from the traditional infiltration theory.

7. UNITED STATES WEATHER BUREAU MODEL

7.01 General Features

Kohler (refs.163 and 167) has outlined investigations by the U.S. Weather Bureau in order to improve forecasts of river stage and discharge. He describes a very promising model that may be represented diagrammatically by fig.11.

The computations are made for sub-areas surrounding individual rainfall stations rather than for whole catchments. This enables better allowances to be made for the wide variations in conditions that may be expected over large areas, particularly in the distribution of rainfall.

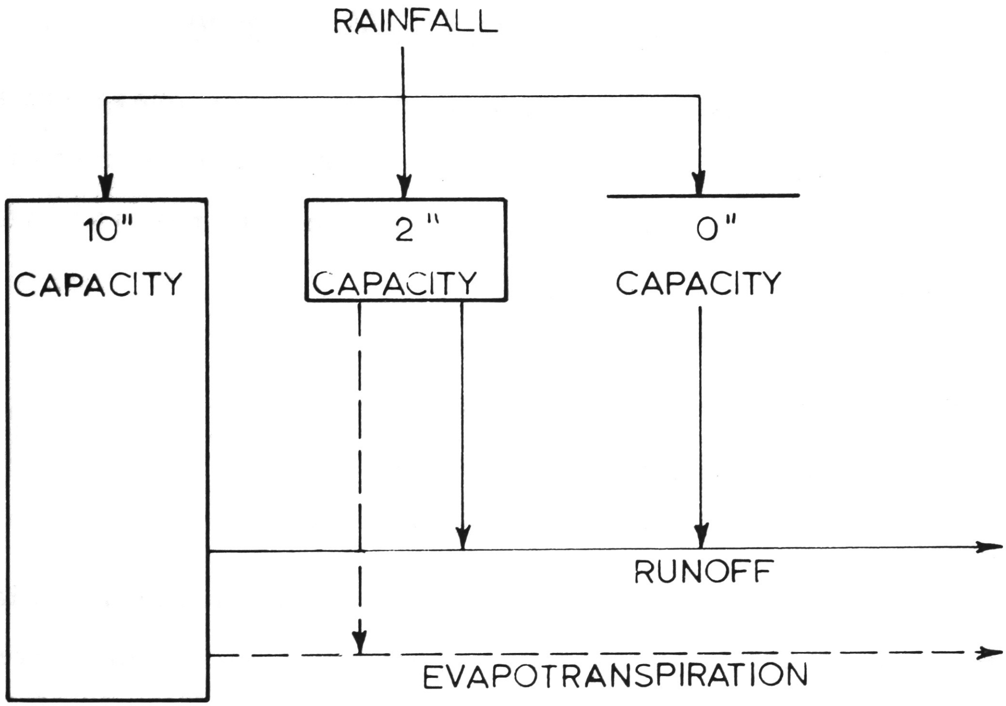
As shown in fig.11 a sub-area is represented by several storage units with fixed capacities such as 0, 2 and 10 inches. The water stored in these units is the water that is absorbed by the catchment and therefore does not contribute to the runoff. The effective areas or weightings to be applied to the capacities are characteristic constants of a particular sub-area that determine its wetting and drying behaviour.

7.02 Specification of Moisture Status by U.S.W.B. Model

The moisture status of a sub-area is specified by the deficiencies of its storage units, i.e. by the volumes of water required to fill the units to capacity.

The U.S.W.B. has tried as many as five storage units in each sub-area but Kohler states that little is gained by using more than three, including 0 inches. As the 0 inch capacity does not change, the moisture status of a sub-area at any time may be adequately specified by only two parameters representing the deficiencies of the 2 and 10 inch capacities.

FIGS. 8 -10 in the original print copy.



COMPONENTS OF U.S.W.B. MODEL.

FIG. 11.

The deficiency of the 2 inch capacity is essentially an index of the "upper zone" or surface condition while the deficiency of the 10 inch capacity is an index of the total moisture condition. Capacities other than 2 and 10 inches may be selected, of course, and a certain amount of experimenting may suggest more suitable values for a particular catchment.

7.03 Simulation of Drying Phase by U.S.W.B. Model

Changes in moisture status during the drying phase are attributed entirely to evapotranspiration and no direct provision is made for the effects of drainage to lower levels. As explained in ref. 163, values of potential evaporation are obtained from:

(1) Mean daily air temperature.

(2) Total daily solar radiation.

(3) Total daily run of wind.

(4) Mean daily dew point temperature.

The water storage in each unit is depleted at the potential evaporation rate until none remains. The low capacity units are therefore usually emptied early and, as the drying phase progresses, later losses occur only from the high capacity storages. The curve of moisture depletion for a whole sub-area therefore becomes less steep with increasing time, the actual shape of the curve depending on the effective areas of the storage units.

The accounting is done on a daily basis and is carried forward independently for each of the selected storage capacities. There is no need to be concerned with the effective areas until the final stages of the runoff calculations.

7.04 Simulation of Wetting Phase by U.S.W.B. Model

During the wetting phase the storage units are recharged, the volume of recharge being equal to the change in moisture deficiency. The difference between the recharge and the rainfall in any period is equal to the runoff generated in the period. The equations used for the computations are as follows:

$$f_{os} = f_m \left(\frac{d_{02}}{2}\right)^{\frac{2f_{os}}{f_m d_{os}}} \qquad \text{..........(7a)}$$

$$f_{ts} = f_{os} \exp\left(\frac{-f_{os}t}{d_{os}}\right) \qquad \text{..........(7b)}$$

$$Q_s = P - d_{os}\left[1 - \exp\left(\frac{-f_{os}T}{d_{os}}\right)\right] \qquad \text{..........(7c)}$$

$$Q = \sum a_s Q_s \qquad \text{..........(7d)}$$

$$\sum a_s = 1.00 \qquad \text{..........(7e)}$$

where s = moisture capacity, i.e. the maximum amount of water that can be held in a storage unit, excluding the amount that will become runoff.

Q = runoff from a subcatchment, i.e. an area represented by a precipitation station.

Q_s = runoff from a storage unit with effective moisture capacity s within a subcatchment.

P = precipitation on a catchment.

d_{ts} = moisture deficiency at time t of unit with moisture capacity s i.e. the amount of water required to make the storage equal to s.

d_{02} = moisture deficiency at $t = 0$ of unit with capacity of 2 inches.

f_{ts} = capacity rate of absorption at time t of unit with moisture capacity s i.e. the potential time rate of decrease of d_{ts} if adequate water is available.

f_{os} = f_{ts} at $t = 0$.

f_m = maximum capacity rate of absorption.

T = effective storm duration in period considered

a_s = characteristic weighting of units with moisture capacity s.

In the above, a_s and f_m are characteristic constants of the sub-catchment. For a rainfall of depth P with effective duration T and initial moisture status d_{02} and d_{os}, the value of Q may be obtained as follows:

(a) Calculate f_{os} from (7a). (As this is not in explicit form a graphical method is suggested by Kohler, ref. 163).

(b) Calculate Q_s from (7c).

(c) Calculate Q from (7d).

It may be shown that the equations imply the following assumptions (ref. 168):

(1) The rate of recharge is limited by the "absorption capacity" which is directly proportional to the moisture deficiency. This means that the absorption capacity decreases exponentially with time if sufficient water is available, which is roughly in line with Horton's approach to infiltration. (refs. 103 and 104).

(2) The maximum absorption capacity is the same for all storage units and applies when the surface is completed dry i.e. when there is no water in the 2 inch capacity storage unit.

(3) When there is water in the 2 inch capacity unit, the wetting phase proceeds as if the existing conditions had been caused by preceding absorption from a state of surface dryness. (in reality, however, the existing conditions may be due to preceding evaporation also).

The "effective duration" is an equivalent period in which the rainfall intensity continuously exceeds the absorption capacity. In practice there are usually periods when the absorption capacity cannot be satisfied by the available rainfall and in such cases an effective duration can only be estimated by subjective or trial and error methods. At the time of publication of the references, investigation was proceeding to obtain more satisfactory methods of dealing with these conditions. Kohler suggested that some function of both time and volume of rainfall may eventually replace effective duration.

Before the model can be operated the following characteristic constants must be evaluated for a particular catchment:

(1) The effective area of each subcatchment.

(2) The effective area or weighting to be applied to each storage capacity within each subcatchment. (a_s).

(3) The maximum absorption capacity. (f_m).

These are derived by a computer to give the best fit to past records of rainfall, runoff and potential evaporation.

The residual errors of the model are not random but show a seasonal bias. Small seasonal variations of the weightings are therefore made and these probably account for such factors as vegetation changes, agricultural practices and heat storage within the soil.

7.05 Conversion of Rainfall-excess to Streamflow by U.S.W.B. Model

Kohler does not include this aspect in the given references but presumably it may be dealt with by either a unitgraph or routing technique.

The term "rainfall-excess" strictly applies to surface runoff only. In the U.S.W.B. model it may also include interflow and, in some cases, base flow, depending on whether these components are readily separated in the basic runoff data.

STANFORD UNIVERSITY MODEL

8.01 General Features

Another fairly complex rainfall-runoff model has been developed at Stanford University, California, by Crawford and Linsley (refs. 105 and 106). It produces hourly streamflow estimates from hourly rainfall and daily values of potential evaporation and may be used for purposes such as the extension of runoff records and flood estimation.

The mean hourly rainfall is calculated from all appropriate gauges on the catchment and areal variation are ignored. The model is therefore probably suitable for catchments between about 20 and 100 square miles, depending on the rainfall characteristics. (It has been tried on an 80 sq. mile catchment).

The daily potential evaporation is calculated from Meteorological data by the U.S.W.B. method, as described previously.

Figure 12 shows how the Stanford model represents a catchment by four main units, viz:

(a) An upper zone storage which provides for the effects of the catchment surface and upper soil layers.

(b) A lower zone storage which provides for the effects of the major part of the soil profile above the water table.

(c) A groundwater storage which controls base flows.

(d) A unit for impervious areas from which all rainfall becomes runoff.

Some of the detailed relationships of the model have evolved by a process of trial, and error, arbitrary adjustments having been made when the reproduced data differed from actual observations. The equations given in the references may be summarized as follows:

Evaporation

$$C_u = c_1 V + \exp(-c_2 R_L) \quad \text{.........(8a)}$$

$$V = \sum_{i=1}^{\infty} (0.9)^i W_{i-1} \quad \text{.........(8b)}$$

where i = number of days before specified day.

$$\left.\begin{aligned} W &= E \quad \text{when } S_u > 0 \\ &= E - \frac{E^2}{2rR_L} \quad \text{when } S_u = 0 \text{ and } E < rR_L \\ &= \frac{rR_L}{2} \quad \text{when } S_u = 0 \text{ and } E \geq rR_L \end{aligned}\right) \quad \text{(8c)}$$

Surface Runoff

$$Q_i = P\left[a + b\sqrt[4]{0}\right] \quad \text{.........(8d)}$$

$$\left.\begin{aligned} Q_m &= \frac{P_L^2}{2M} \quad \text{when } P_L \leq M \\ &= P_L - \frac{M}{2} \quad \text{when } P_L > M \end{aligned}\right\} \quad \text{.........(8e)}$$

Infiltration

$$\left.\begin{aligned} I_u &= P_u\left[1.0 - 0.89 R_u^3\right] \quad \text{when } R_u \leq 0.5 \\ &= P_u\left[1.0 - \left(\frac{R_u}{R_u + 1}\right)^{1/R_u}\right] \quad \text{when } R_u > 0.5 \end{aligned}\right\} \quad \text{.........(8f)}$$

$$\left.\begin{aligned} I_m &= P_L - \frac{P_L^2}{2M} \quad \text{when } P_L \leq M \\ &= \frac{M}{2} \quad \text{when } P_L > M \end{aligned}\right) \quad \text{.........(8g)}$$

$$\left.\begin{aligned} I_n &= P_L - \frac{P_L^2}{2N} \quad \text{when } P_L \leq N \\ &= \frac{N}{2} \quad \text{when } P_L > N \end{aligned}\right) \quad \text{.........(8h)}$$

$$I = G\, I_u \quad \text{.........(8i)}$$

Drainage from Upper to Lower Levels

$$O_u = f(1-R_L)\,S_u \qquad \text{........(8j)}$$

$$\left.\begin{aligned} O_L &= G R_L O_u && \text{when } R_L \leq 1.0 \\ &= G O_u && \text{when } R_L > 1.0 \end{aligned}\right\} \text{.......(8k)}$$

$$\left.\begin{aligned} G &= h_1 + h_2\,(R_L)^3 && \text{when } R_L \leq 0.5 \\ &= (1-R_L)^{1/R_L} && \text{when } R_L > 0.5 \end{aligned}\right\} \text{.......(8l)}$$

Interflow and Groundwater Flow

$$Q_n = I_m - I_n \qquad \text{......(8m)}$$

$$\left.\begin{aligned} O_g &= k_4 S_g && \text{when } S_g \leq C_g \\ &= k_4 S_g + k_3\,(S_g - C_g) && \text{when } S_g > C_g \end{aligned}\right\} \text{......(8n)}$$

where

P = rainfall on catchment in each hour

P_u = rainfall available to upper zone in each hour

P_L = rainfall available to lower zone in each hour

Q_i = runoff generated from impervious areas in each hour

Q_m = runoff generated from upper zone in each hour

Q_n = interflow generated each hour

Q_s = surface runoff generated each hour = $Q_i + Q_m$

S_u = storage in upper zone

S_L = storage in lower zone

S_g = groundwater storage

C_u = nominal storage capacity of upper zone

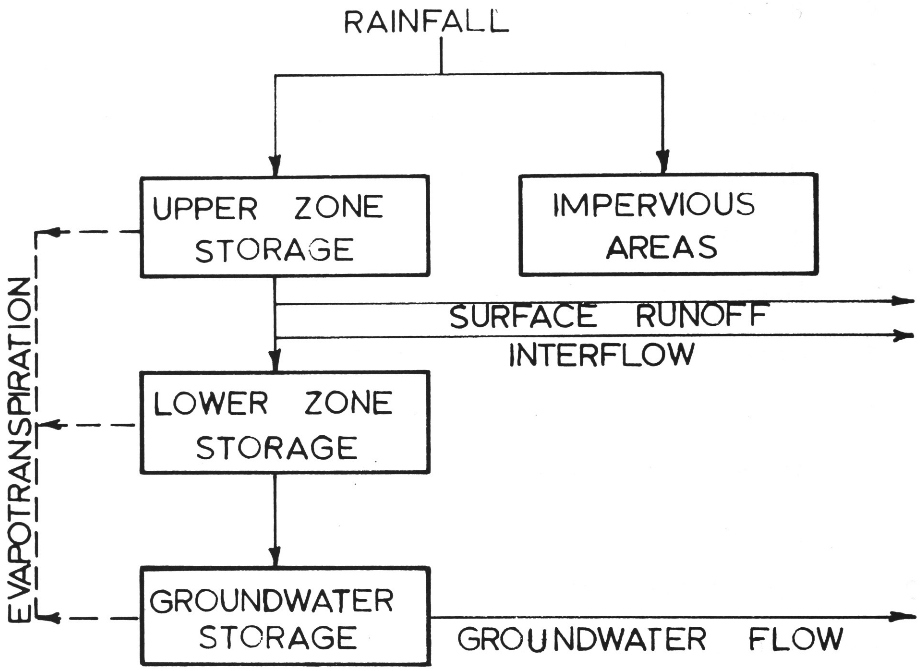
C_L = maximum probable storage of lower zone

C_g = nominal groundwater storage capacity

R_u = upper zone moisture ratio = S_u/C_u

R_L = lower zone moisture ratio = S_L/C_L

O_u	=	drainage from upper zone to lower zone
O_L	=	drainage from lower zone to groundwater
O_g	=	groundwater flow at catchment outlet
O	=	total discharge at catchment outlet
I_u	=	direct hourly infiltration to upper zone
I_m	=	direct hourly infiltration to lower zone, including interflow.
I_n	=	direct hourly infiltration to lower zone excluding interflow
I_g	=	hourly infiltration to groundwater storage (i.e. part of I_u)
E	=	daily potential evaporation
G	=	groundwater accretion index
M & N	=	lower zone infiltration indices
V	=	antecedant evaporation index
W	=	actual daily evaporation
a	=	impervious proportion of catchment
b	=	parameter representing water surface area
c_1 & c_2	=	upper zone storage constants
f	=	upper zone drainage constant
h_1 & h_2	=	groundwater storage constants
k_1 & k_2	=	surface and interflow storage-outflow ratios
k_3 & k_4	=	storage-groundwater flow ratios
r	=	lower zone evaporation constant



COMPONENTS OF STANFORD MODEL

FIG. 12.

8.02 Specification of Moisture Status by Stanford Model

Three parameters are required to specify the moisture status of a catchment, i.e.

Interception-depression storage deficiency.

Upper zone moisture ratio.

Lower zone moisture ratio.

The maximum deficiency of the interception-depression storage is a fixed amount between .05 and .50 inches depending on the catchment concerned. This maximum value applies at the end of most drying phases because the storage is rapidly depleted. After the commencement of rainfall, however, the storage is rapidly filled and a minimum deficiency therefore applies during most of the wetting phase to account for evaporation etc. The interception-depression storage is only a minor part of the total upper zone storage which also includes most of the water involved in early infiltration.

The ratio of the total upper zone storage to the "nominal capacity" is the upper zone moisture ratio. Although the nominal capacity is limited it varies to a certain extent with the lower zone storage and becomes progressively larger as the catchment dries out (equation 8a).

The lower zone moisture ratio is given by the ratio of the lower zone storage to the maximum lower zone storage. This largely controls long-term infiltration and evaporation.

8.03 Simulation of Drying

The drying phase is simulated on a daily basis from the calculated potential evaporation and the appropriate parameters of moisture status.

Evaporation from the upper zone is assumed to occur at the potential rate and while any water remains in this storage there is no evaporation from the lower zone. When the upper zone storage is empty evaporation commences from the lower zone at rates that vary with the moisture ratio and the potential evaporation as given by equation (8c). This accounts for the effects of the limited availability of water for evaporation as the soil moisture is depleted.

A progressive adjustment is made to allow for the relatively small amount of evaporation from the groundwater storage. Presumably it is varied on a seasonal basis and may be estimated from available records.

To provide for drainage from upper to lower levels a percentage reduction of the upper zone storage is made each day, depending on the lower zone moisture ratio (equation (8j). The portion of this drainage entering the groundwater storage is also expressed as a function of the lower zone moisture ratio (equation (8k)).

8.04 Simulation of Wetting Phase by Stanford Model

Computations of the wetting phase are carried out on an hourly basis during periods of rainfall. The runoff generated in each hour consists of:

The surface runoff from impervious areas (Qi)
The surface runoff from pervious areas (Qm)
Interflow. (Qn)
Groundwater flow (Og).

The surface runoff from impervious areas is the gross rainfall on a fixed percentage of the catchment representing sealed surfaces plus a variable percentage representing water surfaces (equation (8d)).

The surface runoff from pervious areas is equal to the remaining rainfall after abstractions have been made for:-

Interception-depression storage

Infiltration to the upper zone

Infiltration to the lower zone

If the hourly rainfall is less than, or equal to the deficiency of the interception-depression storage, it all enters this storage. Rainfall in excess of the interception-depression storage deficiency becomes available for infiltration. The percentage infiltrating to the upper zone is a function of the upper zone moisture ratio and decreases as the ratio increases (equation 8f).

Infiltration to the lower zone is similarly regulated by the lower zone moisture ratio, but in this case the model assumes that infiltration capacities vary throughout the catchment from a minimum value of 0 to a maximum value of M. The distribution curve is assumed to be linear and is represented by OM in fig.13. The runoff from pervious areas is therefore largely determined by M which is a function of the lower zone moisture ratio. The curve relating these is unique to each catchment and its parameters are derived to fit the data. Equations (8e) and (8g) may be readily obtained from fig.13 to express both surface runoff and infiltration in terms of the available rainfall (inflow to lower zone) and the parameter M.

Interflow is generated from part of the infiltration to the lower zone and is represented in fig.13 by the area between lines OM and ON. Parameter N may be regarded as the maximum "net infiltration capacity" and is related to the moisture ratio by a unique curve similar to that for parameter M.

Part of the net infiltration in each hour enters the groundwater storage, (equation (8i)). The groundwater flow is a direct function of the storage as given by equation (8n).

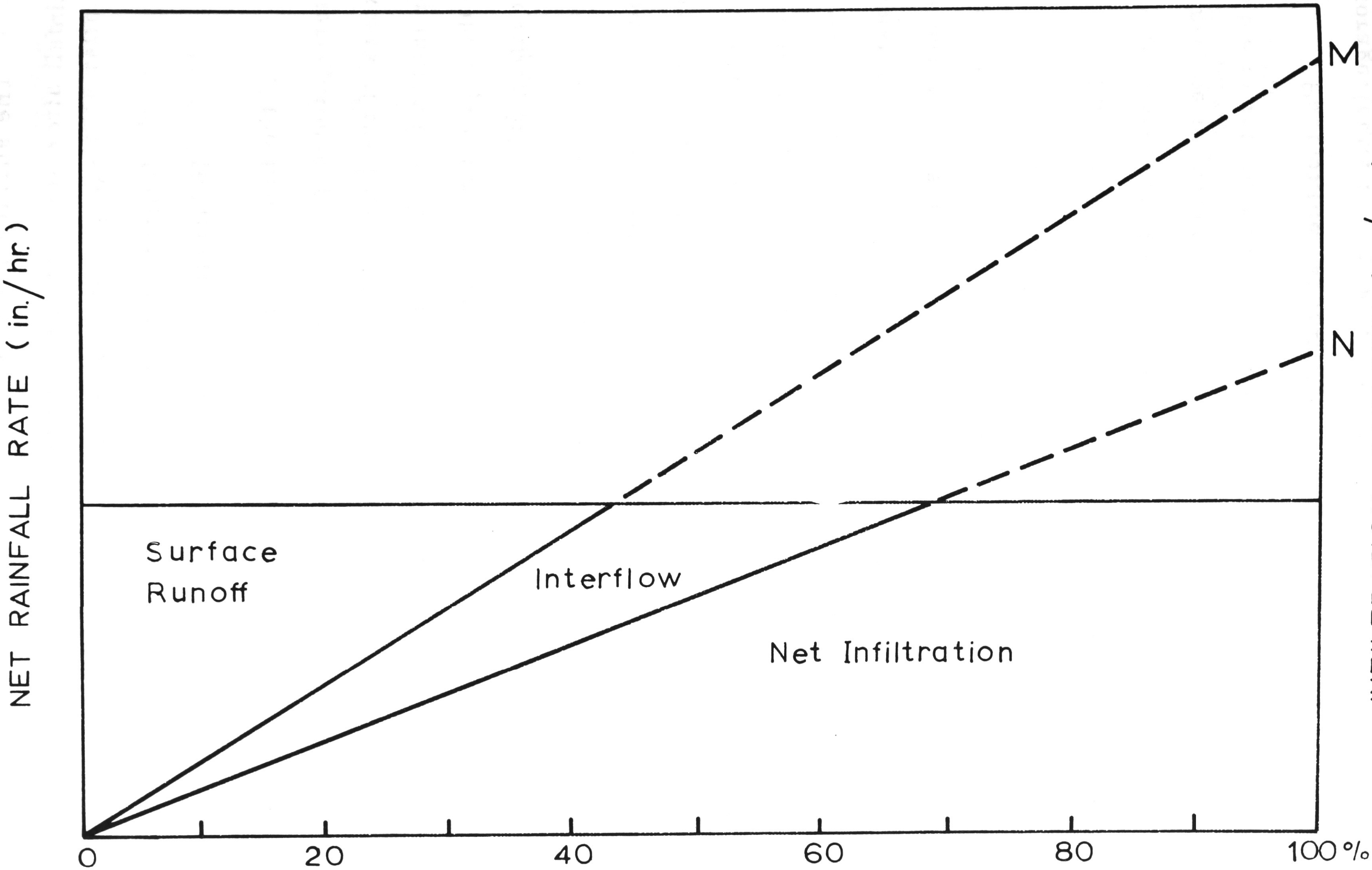


FIG. 13. % OF AREA WITH INFILTRATION CAPACITIES EQUAL TO OR LESS THAN INDICATED VALUES.
(After Crawford and Linsley 1960)

Throughout the wetting phase all storages are affected by drainage from upper to lower levels (equations (8j) and (8k)) and the interception-depression storage is depleted by a constant hourly rate. These processes continue into the drying phase for limited periods.

8.05 Conversion of Rainfall-excess to Streamflow by Stanford Model

The computed hourly increments of surface runoff and interflow are translated by characteristic constant values of time determined from the records. The resulting histograms are then routed through separate linear storages to form surface and interflow hydrographs at the catchment outlet. These hydrographs and the groundwater flow given by equation (8n) are added together to produce the total hydrograph of streamflow.

9. COMMONWEALTH BUREAU OF METEOROLOGY MODEL

9.01 General Features

A relatively simple model has been successfully used by the Commonwealth Bureau of Meteorology to predict flood flows from isolated storm events on several Australian rivers (refs. 107 and 164). The basic assumption of this approach is that no significant flood flows occur during a storm until the rainfall has satisfied an "initial loss" that depends on the moisture status. It is further assumed that the loss rate after the commencement of flood runoff is a relatively small, constant value.

An electronic computer is used to derive unit hydrographs, loss rates and initial losses from past pluviograph and streamflow data.

9.02 Specification of Moisture Status by C.B.M. Model

The moisture status is specified by a single parameter that endeavours to account for the effects of preceding rainfall on the catchment. This parameter is the "antecedent precipitation index" and is calculated from the following equation as suggested by Linsley, Kohler and Paulhus (ref. 2):

$$API = 0.9\ API' + P \qquad (9a)$$

where
- API = Antecedent precipitation index for specified day
- API' = Antecedent precipitation index for previous day
- P = Rainfall since previous day

A daily account of API is maintained throughout the drying phase.

9.03 Simulation of Drying Phase by C.B.M. Model

Although the above equation does not directly simulate the physical processes of the drying phase it may give reasonable predictions of the effects of these processes. Under some conditions the value of API may be interpreted as representing the actual amount of moisture storage within an appropriate layer of the catchment and the constant of 0.9 in the equation implies that during the drying phase this moisture is depleted daily by ten percent of the existing storage.

Variable values, rather than a constant of 0.9 would provide the most logical allowance for seasonal and regional differences but this complicates the derivation and operation of the model. The Bureau of Meteorology claims that runoff predictions are just as reliable if the API is computed with 0.9 and a seasonal parameter is used in the correlation between API and initial loss.

The value of API never becomes zero in a finite time but it is generally insignificant after the drying phase has exceeded about 30 days in length. Therefore, in computing the API it is usually not necessary to consider rainfalls preceding the specified day by more than a month.

9.04 Simulation of Wetting Phase by C.B.M. Model

The wetting phase of the C.B.M model represents the physical processes in a highly simplified manner. Interception, depression storage and early infiltration are all acounted for by initial loss which is dependant on the API and season of year, as mentioned previously. The correlation between initial loss, API and season is obtained from past records and is unique to each catchment.

It is assumed that during the early part of a storm a catchment can absorb rainfall at unlimited rates and, no matter how high the intensities may be, runoff does not occur. As soon as the total rainfall equals the initial loss, however, the absorption capacity of the catchment drops abruptly and most of the ensuing rainfall becomes rainfall-excess.

The small constant loss rate during this stage is usually between .02 and .10 inches per hour which accounts for continuing evaporation and deep infiltration.

The value of API is increased each day by the amount of rainfall entering the catchment during the previous 24 hours. It would be more logical to increase the API by rainfall minus corresponding runoff but this will not necessarily improve the accuracy of runoff predictions.

9.05 Conversion of Rainfall-excess to Streamflow by C.B.M. Model

Rainfall-excess is converted to streamflow by the unit hydrograph method with several refinements. On some streams the peaks of unit hydrographs derived from large storms tend to be higher than the peaks derived from smaller storms (ref. 108) and the use of an average unit hydrograph can therefore result in the under-estimation of extreme floods. The C.B.M. ananlysis by digital computer enables these tendencies to be quantitatively specified and appropriate adjustments are made when necessary.

Similar adjustments are made for large variations in areal distributions of rainfall-excess. On the Macleay River, N.S.W. for example, storms concentrated near the headwaters were separated from storms concentrated near the outlet and appropriate corrections were applied to estimates based on the average unit hydrograph.

10. OTHER RAINFALL-RUNOFF MODELS

10.01 Multiple Regression and Multivariate Analysis

The rainfall-runoff cycle may be regarded as the mathematical operation of a number of variables which ultimately produce runoff as the major dependent variable. Multiple regression is commonly used to analyse such systems and can evaluate the combined and individual effects of many independent variables such as rainfall and evapotranspiration.

"Best fit" equations for predicting runoff have been derived with linear multiple regression programmes which are readily available for most computers. Logarithmic transformations or polynomials are usually easily introduced, if necessary, to allow for curvilinear relations.

Unfortunately these methods have certain limitations and do not always give satisfactory rainfall-runoff estimation as discussed in detail by Sharp, Gibbs, Owen and Harris (ref. 109). The main difficulties are due to the inter-relationships between many of the variables which make some of the basic assumptions of the methods untenable. The high coefficients of correlation and tests of significance that are obtained with some multiple regression studies may be quite misleading because of the skewness in much hydrological data.

Harris, Sharp, Gibbs and Owen have developed an improved statistical model similar to multiple regression analysis but avoiding some of the above difficulties. In this approach the most important independent variable is first identified and its effects on all other variables are removed. The next most important independent variable is then identified and the procedure is repeated until there are no further significant reductions in the unexplained variance of the dependent variable. Although the predicting equation thus obtained may not give a "best fit" to the data it ensures a minimum number of significant variables, shows the relative importance of each variable and is less likely to give unrealistic predictions than the ordinary multiple regression methods.

Some aspects of the preceding method are approximated by the multiple graphical correlation method of deviation which may be used when there are only three or four independent variables. This has the advantage that the forms of the relationship are not restricted by specific mathematical functions but it is unsuitable if there is joint action or inter-dependence between the variables, (e.g. rainfall total and rainfall duration).

Co-axial graphical correlation, outlined by Linsley, Kohler and Paulhus (ref.1) has been widely used for estimating runoff with varying degrees of success. (Some Australian examples are in ref.60 and ref.111). It has certain advantages over the method of deviations but is still of limited value when there is joint action between a number of variables.

It is suggested by Snyder (ref.112) that multiple regression may frequently be satisfactory for predicting purposes but it should not be used for testing or evaluating an assumed rainfall-runoff model for the following reasons:

(1) Multiple regression curve-fitting (by least squares) associates errors with the dependent variable only. This is inappropriate as most of the independent variables are also subject to error in rainfall-runoff estimation.

(2) Multiple regression assumes that there are no inter-relationships between the independent variables. This aspect has been dealt with previously.

In the above reference Snyder briefly describes the techniques of multivariate analysis and advocates the use of these for some rainfall-runoff problems. Multivariate analysis enables the association of errors with any variables and also permits the separation of the truly independent components of the computational system. These features are very desirable if good physical interpretations of the system are required.

Techniques using multiple regression have been developed by Chapman (refs.113 and 114) and techniques using multivariate analysis have been developed by Betson (ref.115). Both of these will be described below.

10.02 T.G. Chapman

Chapman specified the moisture status of a catchment with a single parameter called the "catchment dryness index", designated by D (ref.113). He also defined a parameter K which is the ratio of the actual evapotranspiration rate to the potential evaporation rate, i.e.

$$K = \frac{dW/dt}{dE/dt} = \frac{dW}{dE} = \frac{dD}{dE} \qquad \ldots\ldots\ldots\ldots(10a)$$

where E = potential evaporation

W = actual evaporation

It is assumed that K is a function of D which may be derived graphically from a long period of rainfall-runoff records as follows:

(a) Select periods that commence and finish immediately after moderate or large floods with little or no intermediate rain.

(b) Assume that the catchment dryness is O at the beginning and end of the above periods so that the actual water loss by evaporation within each period is approximately equal to the "moisture recharge" at the end, i.e.

$$W = P - Q = D_1$$

where W = evaporation during period.

D_1 = dryness index immediately before flood at end of period

P = rainfall in period (mostly from storm at end)

Q = runoff due to P

(c) Plot values of W calculated in (b) against corresponding values of E and draw a smooth curve through the scatter of points. The slopes of this curve give a first estimate of K.

(d) Plot K as estimated above against D_1, (=W). This relationship can then be improved by trial and error to give the best fit to the available data.

After a satisfactory relationship betweek K and D has been obtained a daily accounting of D can be carried out from estimates of the potential evaporation. Any method of estimating potential evaporation may be used, including unadjusted pan readings, providing the same method is used consistently in the complete derivation and in all applications of a particular relationship.

Rainfall in minor storms is evaporated at rates corresponding with the maximum value of K, in the accounting procedure, and no further reduction in D is made until all of the inflow from the minor storm has been evaporated.

In an application of the above to the Upper Goulburn River, N.S.W., Chapman found that there were few periods with no significant rain between flood-producing storms. It was therefore assumed that an estimated maximum rate of K applied to the intermediate storms and appropriate values were subtracted from W and E in step (c) above.

To enable the estimation of storm runoff, some form of relationship between D, storm rainfall total and rainfall intensity must be derived from the available data.

In this study Chapman obtained good predictions of storm runoff by multiple regression with D, storm rainfall total and maximum daily rainfall as the (assumed) independent variables. The maximum daily rainfall was found to be a better index of the effects of rainfall intensity than the storm duration. Co-axial graphical correlation was also tried but the predictions of runoff were not as good as by multiple regression, apparently because of joint action between the variables.

10.03 <u>R.P. Betson</u>

R.P. Betson (ref. 115) used a multivariate technique to develop a single equation for predicting storm runoff from storm rainfall, storm duration and a soil moisture index. The soil moisture index was computed each day from:

$$S = S' + P - Q - E \qquad \text{.............(10b)}$$

where S = soil moisture index for the given day

S' = soil moisture index for the previous day

P = rainfall during 24 hours prior to 9 a.m. on given day

Q = runoff during 24 hours prior to 9 a.m. on given day

E = actual evaporation from catchment during 24 hours prior to 9 a.m.

The value of E is estimated from daily pan data which is adjusted so that the calculated average annual catchment evaporation is equal to the average annual difference between rainfall and runoff.

In deriving a relationship between storm runoff, storm rainfall and S, Betson commenced with Horton's infiltration equation (ref. 103), evaluated various parameters based on this, and made progressive adjustments suggested by the fit of the equation to the data and the physical rationality of the parameter values. The final equation is as follows:

$$Q_t = P_t\,(1-h) - \left[a + CT + \frac{b}{n}\exp(-nmS) - \frac{b}{n}\exp(-nmS-nT)\right] \quad \ldots\ldots(10c)$$

where Q_t = storm runoff

Q = runoff for day preceding time of S.

P_t = storm rainfall

P = rainfall for day preceding time of S

E = estimated evaporation from catchment for day preceding time of S

T = storm duration

a = constant representing interception storage

b = constant equal to the difference between maximum and minimum infiltration capacities

c = minimum infiltration capacity

$(1-h)$ = runoff-producing portion of catchment

m = constant relating infiltration and soil moisture

n = infiltration depletion constant

This equation was found to have an excellent adjustment to the data for all watersheds. The standard error of .02 is about the same order of accuracy as the rainfall data.

It was shown that the parameter (1-h) represents the effective run-off-producing area which Betson claims is constant for a particular water-shed. This area was surprisingly small, varying between watersheds from

from 4.6% to 86% with an average of only 25%.

Betson's approach appears to have developed from a series of studies published by the Tennessee Valley Authority (refs.116, 117 and 118). These deal extensively with the adaptation of traditional hydrological techniques to digital computer analysis.

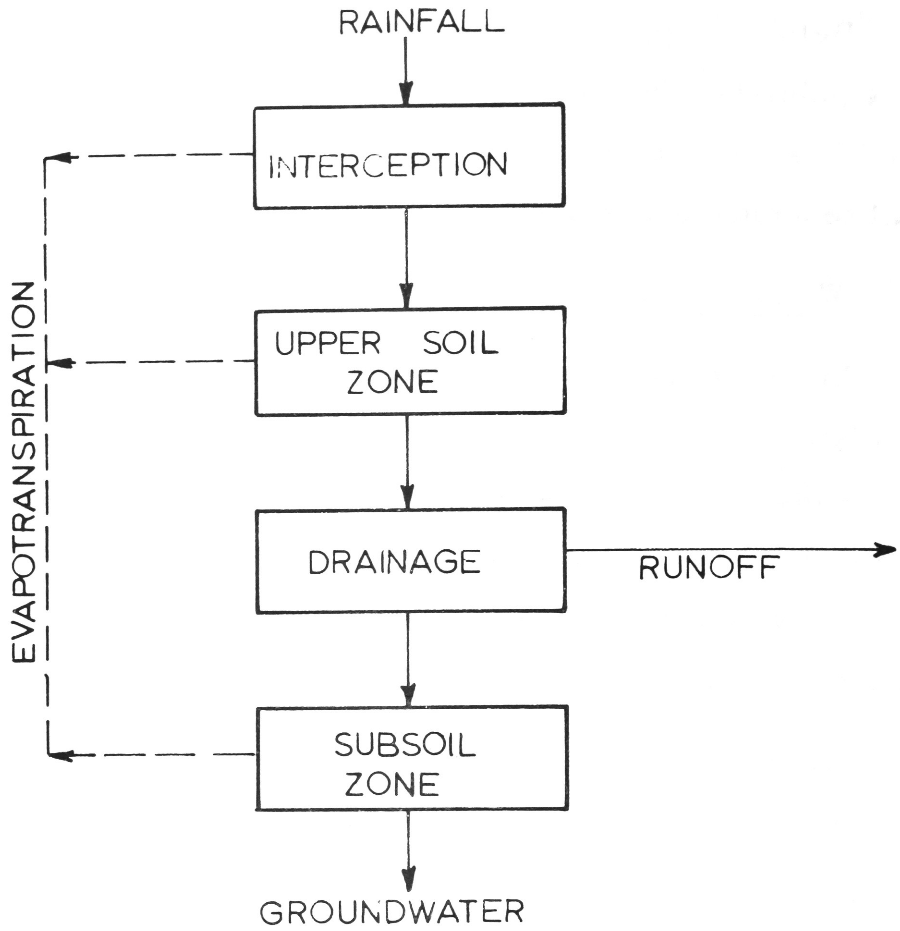
10.04 W.C. Boughton

In general, the preceding models involve either short periods of rainfall or some measure of rainfall intensity such as storm duration. Because pluviograph data is brief and deficient under many conditions, Boughton (refs.131 and 169) has developed a model requiring only daily values of rainfall and potential evaporation. Its components and their inter-relationships are shown diagrammatically in fig.14.

There are three surface storages in this model, viz.the interception, upper soil and drainage storages. The amount of water required to completely fill these at any particular time is the "potential initial loss" which must be satisfied during a wet spell before runoff commences.

The upper soil storage represents the moisture content between wilting point and field capacity of a very porous top soil layer that has no limiting infiltration capacity. The drainage storage represents the water between field capacity and saturation in this layer, and allows for a rapid recovery of potential initial loss between storms.

After the surface storages are satisfied rainfall in excess of the subsoil infiltration becomes runoff. The subsoil storage represents a much denser layer having a relatively low infiltration capacity that varies with its moisture content in an inverse exponential manner. At



COMPONENTS OF BOUGHTON'S MODEL

FIG. 14

maximum moisture content the infiltration capacity is a small constant accounting for deep seepage etc.

It is assumed that subsoil infiltration continues at capacity rates during and after rainfall while any water remains in the drainage storage. Some of the latter is also depleted as base flow, the rate being directly proportional to the amount of storage.

During the drying phase the interception storage is evaporated first at the potential rate. When the potential demand cannot be met from this, the evaporation from the upper soil and subsoil storages is given by:

$$\left.\begin{aligned} V_u &= \frac{Vmu\ Wu}{Mu} && \text{when } Vu < \frac{E}{A} \\ &= \frac{E}{A} && \text{when } Vu \nless \frac{E}{A} \end{aligned}\right) \quad \ldots\ldots(10a)$$

$$\left.\begin{aligned} Vs &= \frac{Vms\ Ws}{Ms} && \text{when } Vs < \frac{E}{1-A} \\ &= \frac{E}{1-A} && \text{when } Vs \nless \frac{E}{1-A} \end{aligned}\right) \quad \ldots\ldots(10b)$$

where

Vu, Vs	=	actual daily evaporation from upper soil and subsoil storages respectively
Vmu, Vms	=	maximum possible daily evaporation from upper soil and subsoil storages respectively
Mu, Ms	=	total capacities of upper soil and subsoil storages respectively
Wu, Ws	=	actual storage levels
A	=	proportion of total evaporation from upper zone (usually about 0.5)
E	=	daily pan evaporation

There is no evaporation from the drainage storage.

The operation of the model requires the derivation of the following characteristic constants for each watershed:

(a) Capacities of each of the four storages

(b) Maximum possible daily evaporation from the upper soil and subsoil storages.

(c) Two constants expressing relationship between subsoil moisture level and infiltration.

(d) Depletion factor expressing rate of base flow from drainage storage.

Boughton has applied the model to six gauged watersheds in N.S.W. using an IBM 620 computer to derive the constants and perform the daily water balance calculations. Reasonably good reproductions of the recorded runoff data were obtained in all cases.

An interesting by-product of this study was the significantly higher surface storage capacities derived for larger watersheds. Boughton attributed this to the effects of channel transmission losses which he claimed were relatively much lower in the smaller watersheds.

11. SOME RELEVANT STUDIES OF INDIVIDUAL PROCESSES

11.01 Evaporation and Transpiration

The term "evapotranspiration" is commonly used for the combined effects of transpiration and direct evaporation. This cumbersome expression appears to be unnecessary as the general term "evaporation" includes the process of transpiration and can be correctly substituted in all cases for "evapotranspiration", as has been done in this report. It should be appreciated, however, that the major evaporation losses from vegetated catchments are due to transpiration and it is surprising that this process has not been considered more thoroughly in rainfall-runoff studies.

It was stated earlier that the actual rate of moisture loss by evaporation is usually less than the potential rate, particularly when the catchment is in a dry condition, but this has long been a controversial question.

For example, Veihmeyer and Hendrickson (ref.119) have exerted considerable influence with their claims that transpiration rates are approximately equal to the potential evaporation rates and are not affected by differences in soil moisture unless conditions are so dry that wilting occurs. On the other hand, there has been wide support for Thornthwaite and Mather (ref.120) who postulated that relative transpiration (ratio of actual to potential) decreases linearly with soil moisture below a critical value of soil moisture near field capacity.

Between these two extreme theories there have been various other proposals such as those due to Penman (refs. 137 and 121) and Van Bavel (ref.122).

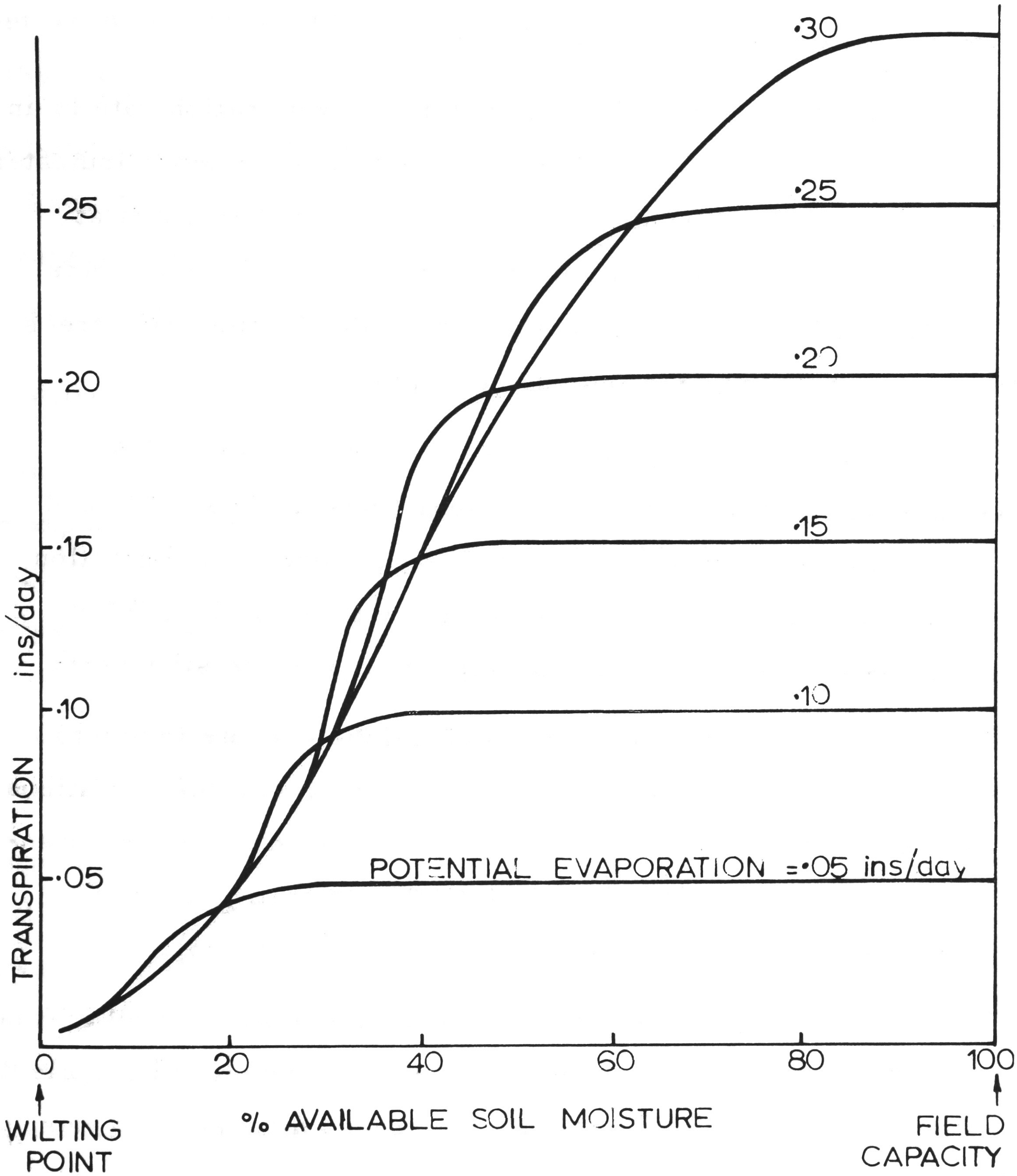
The controversy has almost been resolved by a number of recent studies that provide a better insight to the physical relationships involved (see refs. 123 to 130 inclusive). These suggest that for a particular plant-soil complex the transpiration rate should be expressed as a function of both soil moisture and potential evaporation, as shown in fig. 15. Marked decreases in transpiration at various critical values of soil moisture occur when the rates of water absorption by the roots can no longer satisfy the potential evaporation from the foliage, and moisture stress increases rapidly within the plant. The ensuing transpiration characteristics may be determined, in many cases, by the partial closure of the stomata or by other responses of the plant to moisture stress.

Some scatter about the curves of fig. 15 may be expected because of the plant responses to other conditions such as light and temperature and also because of the time lag between changes in potential evaporation and the resulting changes in plant moisture stress.

Fig. 15 shows that Veihmeyer and Hendrickson's theory is tenable under any of the following circumstances:

(a) Low rates of potential evaporation
(b) Vegetation able to maintain high rates of absorption at lower moisture contents, perhaps because of an extensive rooting system.
(c) Coarse soil that retains relatively small volumes of water between the so-called "field capacity" and "wilting point".

In other circumstances Thornthwaite and Mather's theory is closer to reality but the general approach of fig. 15 seems preferable to either



TRANSPIRATION - SOIL MOISTURE RELATIONS FOR A TYPICAL PLANT SOIL COMPLEX.

FIG. 15

of the above theories if sufficient information is available for the graphical derivation of the relationships.

It is frequently assumed that the potential evaporation rate is an upper limit to the actual evaporation rate but some evidence indicates that advection, wind and other effects may cause actual rates at least 10% greater than the potential based on free water etc. (refs. 140 and 141). This applies particularly to transpiration from trees and the evaporation of water intercepted by foliage.

Rainfall interception by foliage has been studied extensively and its physical processes are fairly clear (refs. 1, 100, 132, and 134). The maximum amount of water that can adhere to vegetation surfaces is usually less than 10 points at any particular time but continuing losses due to evaporation appear to be quite significant.

The evaporation component of interception accounts for up to 3 points per hour in some areas, despite the unfavourable conditions for this process during periods of rainfall. Some interception losses also include water absorbed directly into the vegetation but the quantities involved are generally small (refs. 129 and 135).

Current knowledge of direct evaporation from soil is well advanced, relevant contributions having been reported by Penman and Schofield (ref. 136), Penman (ref. 137), Philip (ref. 125), Philip and deVries (ref. 138), Gardner and Hillel (ref. 139).

This evaporation takes place at rates approximately equal to the potential while ever the transfer processes within the soil can maintain an adequate supply of water to the surface. When the supply of water becomes restricted the rate of evaporation decreases rapidly and becomes dependent on the moisture distribution and temperature gradients rather than on the potential evaporation.

The latter stage would not be very important in rainfall-runoff studies of vegetated catchments where most of all the soil moisture is lost by transpiration. It may be significant, however, in denuded areas, particularly under arid conditions.

11.02 Infiltration, Soil Water Retention and Drainage

During the past fifteen years a great deal of progress has been made in explaining and quantitatively predicting the physical phenomena of the infiltration process, (see for example, Watson, ref.142). Difficulties in applying this knowledge to real situations are now due more to the complexities of measuring, specifying and mathematically manipulating the variable conditions rather than to a lack of understanding of the fundamentals.

Because of such difficulties the old exponential equation of Horton's (ref.103) is still of practical value in describing the decrease in infiltration capacity with time during a storm. It enables the infiltration characteristics of a catchment to be specified by three empirically derived constants but these are not readily related to the effects of varying soil moisture and most applications of the equation are gross approximations of the actual conditions.

The following equation, due to Philip (ref.144) has greater physical significance:

$$f = \frac{1}{2} St^{-\frac{1}{2}} + B \qquad \text{........(11a)}$$

where f = infiltration rate at time t if ample water is available at the surface

S = the "sorptivity", i.e. the contribution due to capillarity

B = a parameter expressing the saturated permeability i.e. the contribution due to gravity

Philip has shown that each soil has characteristic moisture content-conductivity relationships which largely govern its hydrological behaviour. These relationships are conveniently expressed in a graphical form and may be obtained by appropriate measurements. They determine the parameters S and B in the above equation which enables a proper allowance to be made for the effects of varying moisture contents on infiltration.

Philip's equation was derived for a homogeneous soil colum but it is evidently suitable for small catchment areas with uniform soil profiles (ref. 143). For more general cases it should be possible to adapt the equation so that S and V vary with the depth of the wetting front as well as with the moisture content, but no applications of this are known to the author.

The above has been largely concerned with continuous infiltration while adequate water is available at the surface. When the water supply becomes restricted, for example after rainfall ceases, infiltration decreases to zero but draining or redistribution of water continues within the soil profile. Physical analysis of this problem has been attempted but it is complicated by the fact that hysteresis occurs in the moisture content-tension curve, i.e. the relationship for increasing soil moisture (as in continuous infiltration) differs from the relationship for decreasing soil moisture (as in draining or drying). Further research on the factors causing these differences will undoubtedly yield a satisfactory solution in the near future.

Studies to date suggest that when infiltration ceases the velocity and moisture content of the wetting front both fall abruptly to values that depend, to a certain extent, on the head of water above the wetting front (ref.145).

For a particular soil the"field capacity" is generally assumed to be a constant moisture content reached after it has freely drained for two or three days from a saturated state. The remaining water is said to be retained by "molecular forces" and consequently cannot contribute to streamflow. Unfortunately, this very useful concept is not strongly supported by the previously mentioned studies which show that drainage may continue for indefinite periods and any approximate equilbrium moisture content is not necessarily a constant for all conditions (ref.166).

Recent field observations have also shown that soil moisture contents below the normal field capacity cause significant contributions to ground-water flows (ref.146 and 147).

11.03 Conversion of Rainfall-excess to Streamflow

The unit hydrograph method has been widely adopted for the conversion of rainfall-excess to hydrographs of streamflow and it is adequately described in several standard text books (e.g. ref. 1 and ref.4).

Before the advent of high-speed computers the proper derivation of unit hydrographs from complex storms was a very laborious procedure. Several computer techniques are now available, one being in current use by the Commonwealth Meteorological Bureau, as mentioned previously. Another technique, using harmonic analysis, has been reported by O'Donnell (ref.148).

Despite its popularity the unit hydrograph method is a gross approximation and has several significant deficiencies (refs.108, 149 and 150). That is does not represent a correct physical relationship between rainfall-excess and runoff may be readily demonstrated by elementary hydraulic principles (ref.3).

For some watersheds there is no real loss of precision when unit hydrographs are derived by the simple S curve method which assumes a constant rate of rainfall-excess (ref.4). This was done by the author for nine storms on the Macleay River, N.S.W. and, although the differences in the resulting unit hydrographs were considerable, they were no greater than those obtained by the C.M.B. computer technique. Similar findings were reported by Coulter (ref.63) from studies on a number of other streams.

Ishihara (ref. 151) and Henderson and Wooding (ref. 152) have developed equations more closely related to the physical mechanisms by simulating a stream and its catchment with inclined planes. Henderson and Wooding's equations may be adapted to inclined flow through porous media, suggesting that their approach is suitable for interflow and groundwater flow, in addition to surface runoff.

Sugawara (ref. 153) studied the rainfall-runoff characteristics of various Japanese rivers and concluded that unit hydrographs are unreliable for high intensity rainfall. He therefore represented catchments with systems of interconnected cylindrical storages and derived a general equation for converting rainfall-excess to streamflow. It may be shown that this equation includes the unit hydrograph as a special, linear case.

The routing of rainfall-excess through a catchment by using storage principles has been thoroughly reviewed by Laurenson (refs. 154 and 170) who developed a very general approach that allows for many variable conditions. Relatively simple storage routing techniques have also been successfully used, one example having been described previously in the Stanford University model.

Further aspects of this topic will be examined in 13.03.

12. PRESENT DEFICIENCIES IN RAINFALL-RUNOFF MODELS

12.01 State of Knowledge of Hydrologic Phenomena

The present state of knowledge of hydrologic phenomena appears to be more advanced than is suggested by Amorocho and Hart (ref.97). The factors influencing the various processes are generally understood and may be assessed at least qualitatively for most conditions.

The greatest limitations in the development of rainfall-runoff models are evidently due to the difficulties of efficiently measuring and specifying the phenomena because of their extreme variability in time and space.

Nevertheless it seems that some of the current specialized hydrological knowledge could be used to more advantage in the rainfall-runoff models available at present, as will be suggested by considering their components in relation to the previously mentioned studies of individual processes.

12.02 Specification of Moisture Status

Studies of infiltration show that the speed of the wetting front and soil moisture content above it, depend on the initial dryness of the soil and the supply of water (refs.144, 155 and 166). It may take a number of days for the front to travel through the entire profile and it is therefore evident that the amount of retention from a storm is influenced by a limited depth of the profile that varies with the storm duration and antecedent moisture status.

Single parameters of moisture status refer essentially to one "average" effective depth of profile which is not necessarily a good index of other effective depths. Single parameters may often be representative of a large part of the profile, however, because of the tendency of moisture distributions to assume regular patterns, after a period, due to continuing drainage and differential root abstraction (refs.156, 161 and 162). Inaccurate estimates would be expected when the moisture patterns are irregular or unusual, for example, after very short or very long drying phases.

The C.B.M., Chapman and Betson methods are all limited as above but the U.S.W.B. and Stanford University use two and three parameters respectively, which can represent a wider range of moisture conditions.

An ideal model would probably maintain a continuous account of the moisture status of each significant layer within the catchment profile and this would generally involve at least four parameters, as has been used by Boughton. In practice at present, however, the characteristics of most models are derived from rainfall, runoff and evaporation data only, which probably would not be expected to yield more than two significant parameters of moisture status.

In both the Stanford and Boughton models the interception storage deficiency is determined with little reference to the available data and could probably be included in the upper zone moisture deficiency without detriment.

12.03 Simulation of Drying Phase

The Stanford and Boughton models use forms of relationship between evaporation, potential evaporation and moisture status that are consistent with the recent studies of 11.01. Most of the other models still appear to be unduly influenced by the earlier ideas of Veihmeyer, Thornthwaite, etc.

Consideration of fig.15 shows that relative evaporation (ratio of actual to potential) is a function of both the moisture status and the potential evaporation, increasing with wetter conditions and decreasing with higher values of potential evaporation. The U.S. Weather Bureau and Chapman assume that relative evaporation is a function of the moisture status only and their estimates would therefore tend to be too low in winter and too high in summer. These errors are probably a more important source of the reported seasonal bias in the U.S. Weather Bureau estimates than the other factors suggested by the reference.

Betson's model has similar deficiencies and would also tend to overestimate evaporation under dry conditions and underestimate it for wet conditions due to the further implied assumption that relative evaporation is constant throughout the year.

The A.P.I. technique used by the C.B.M. was discussed in 9.03. It would be restricted in representing many conditions, for example long drying phases in cool climates when the moisture status may remain close to field capacity for several months.

The effects of drainage from upper to lower levels do not appear to be recognised in some of the models. The U.S.W.B. may partly allow for drainage with relatively high weightings of the low capacity storage units but the total rate of removal of moisture in the model cannot exceed the potential evaporation rate. The method would therefore be suspect for storms following very short drying phases.

In his study of the Upper Goulburn River (ref. 114) Chapman derived actual evaporation rates much greater than the potential, for wet conditions. Some of these high rates would be due to the rapid evaporation of intercepted water, as discussed previously, but they could also reflect the effects of drainage from upper to lower levels.

12.04 Simulation of Wetting Phase

Runoff is influenced by areal variation in rainfall distribution and the only models that deal with this factor are those of the U.S.W.B. and C.B.M. The watersheds of the other models would therefore usually be limited to about 100 square miles or less.

It is rather surprising that Philip's analysis of infiltration has not been found more useful to date for rainfall-runoff models. Concepts such as "layer sorptivity" and "layer permeability" should provide a practical and rational approach if careful adaptions are made to allow for heterogeneity and similar factors.

The Stanford model is very versatile in its treatment of the wetting processes but expressions such as equation (8f) and equation (8g) are very cumbersome and should be amenable to simplification without detriment.

On the other hand the traditional, over-simplified exponential approach of Betson's suffers from some of the deficiencies of the old infiltration theory, as discussed more fully in 13.01. In particular, the use of different segments of a single infiltration curve for different values of moisture status is quite inconsistent with Philip's analysis (see ref.166 fig.4).

. Betson's conclusions concerning the constant runoff-producing areas of watersheds may be acceptable for some arid, flat regions but they are evidently inappropriate for more general circumstances. Various studies of loss rates (e.g. Laurenson and Pilgrim, ref.158, McCutchan, ref.54) suggest that the runoff-producing areas of many watersheds approach

100% in large floods. The storms analysed by Betson did not exceed $2\frac{1}{4}$ inches and therefore represent a very limited sample.

The drying behaviour of a watershed is determined by the storage capacity weightings in the U.S.W.B. model. The same constants also determine the wetting behaviour which implies that similar factors control both phases of the rainfall-runoff cycle. Although this may be reasonably satisfactory for bare soil, in most natural watersheds the drying behaviour is determined by the moisture passing upwards through the vegetation while the wetting behaviour is determined by the water passing downwards through the soil profile. It is rather optimistic to expect the same weightings to account for these distinctly different factors.

The initial loss concept, as used in the C.M.B. and Boughton models, is particularly relevant to some aspects of a later section (14.09) where it will be examined in detail. Its apparent relationships with watershed characteristics and rainfall intensity suggest that the allocation of losses by these models may sometimes be questionable especially for smaller watersheds.

12.05 General Conclusions on Current Rainfall-Runoff Models

Some of the components of the above models are poor representations of the factors they are supposed to simulate. However, their parameters are mathematically selected to reproduce the observed runoff data as closely as possible, which ensures that significant errors in the individual components are mutually compensating. Satisfactory predictions may therefore be expected for most conditions but poor results may sometimes occur in the following:

(a) The estimation of runoff from extreme and unusual conditions, particularly if these did not occur in the period of data used for the derivation of the parameters.

(b) Predicting the effects on runoff of major changes in watershed characteristics.

Both of the above are relevant to this report which may be regarded as part of the extensive "Lidsdale Project" aimed at estimating the effects of pine afforestation on runoff.

In the complete Lidsdale study, some of the most critical conditions are unlikely to be sampled during the very limited data collection period and a reliable rainfall-runoff model will therefore be sought to synthesize long records that include such critical periods. For this purpose the Stanford Model is probably the most suitable of those examined as it attends the most thoroughly to the individual processes.

On the other hand, there appear to be too many components in the Stanford Model for the data that is normally available and the mere reproduction of the observed values of runoff does not prove the validity of the arbitrarily selected parameters. It should be possible to check some of the parameters for the Lidsdale watersheds because detailed soil moisture and interception are being made but in more general cases this data is not available.

Tacit, qualitative extrapolations of the results of hydrologic studies are common but often invalid and improvements have been suggested in section 3 with the recognition of hydrogeography as a specialized field of study.

Reliable quantitative extrapolation now seems feasible in the near future if rainfall-runoff models can be developed with components that are closely related to static watershed characteristics such as soil type and depth, vegetation type and surface slope. It is envisaged that direct measurements or surveys of these factors should enable complete hydrological assessments without the necessity to wait for many years of runoff records.

Some of the deficiencies of current models are due to their attempts to combine fragmentary concepts that have evolved for specialized studies without reference to the rainfall-runoff cycle as a whole. Efforts are therefore being directed towards the development of a basic theory of rainfall-runoff relationships with the following attributes:

(a) It should be sufficiently simple and generalized to enable practical, comprehensive treatments of the entire cycle.

(b) It should be consistent with the individual physical processes and allow for detailed, complex analyses of these processes.

(c) The parameters involved in analyses based on the theory, should have some recognizable physical significance and be realistically related to measurable characteristics of watersheds.

It is believed that such a theory would lead rapidly to many major improvements in existing techniques and one possible approach towards this will be described in the remainder of the report.

The initial idea of the approach is to replace the traditional "infiltration theory" and its "working assumptions" with a more general, but somewhat similar, "retention theory." Some of the difficulties of the infiltration concept are therefore outlined in section 13, together with a brief analysis of the associated problem of runoff separation. An attempt is made to eliminate the need for runoff separation by viewing the time delay aspect "in toto". This appears to be a necessary pre-requisite to the development of a retention theory with attributes (a) and (b) above.

The importance of attribute (c) above is stressed in 14.10 and 14.11.

The major part of section 14, however, deals with a relatively simple rainfall-runoff model that follows readily from retention theory. Its complete operation is illustrated and tested with data from a small watershed at Lidsdale State Forest.

13. THE DEVELOPMENT OF RETENTION THEORY

13.01 Infiltration Theory

The great significance of infiltration in rainfall-runoff relationships was pointed out by Horton in 1933 (ref. 103) and the general ideas arising from this have been very influential in modern hydrology.

Essentially, the so-called "infiltration theory" asserts that the rate of generation of surface runoff is equal to the excess of the rainfall intensity over the "infiltration capacity," (ref. 3). Horton defined infiltration capacity as the maximum rate at which a given soil in a given condition can absorb rain as it falls, (ref. 103).

Unfortunately the application of these simple basic concepts to practical rainfall-runoff studies requires a number of questionable assumptions; most of which are discussed in a very thorough analysis by Cook (ref. 102).

The main difficulties are summarized as follows:

(a) Infiltration capacity decreases rapidly in the early stages of a storm and it is generally impossible to determine the actual value at any particular time from rainfall-runoff data alone

(b) It is also difficult to derive true average values of infiltration capacity during a storm because the length of period in which rainfall intensity exceeds infiltration capacity is usually uncertain.

(c) In applications of the theory it is necessary to estimate the infiltration capacities corresponding to particular values of moisture status but the derivation of relationships for this purpose is hindered by the difficulties mentioned in (a) and (b) above.

(d) Infiltration does not include abstractions of rainfall by interception and depression storage, which are sometimes significant factors in the generation of surface runoff.

(e) Instantaneous rainfall intensities are not completely relevant because much infiltration occurs from overland flow and surface storage. The total rainfall during periods of up to half an hour is probably a better index of the available water.

(f) It is readily demonstrated that the effective infiltration capacity of a heterogeneous watershed varies significantly with the rainfall intensity, (ref. 102, p. 738).

(g) Infiltration theory is only applicable to surface runoff but the various methods of separating this from total runoff are all of doubtful validity.

Because of (a), (b), (c), (d) and (e), it has become common practice to deal with approximate average infiltration rates for whole storms, these being variously called "loss rates," ϕ indices or W indices, according to the method of computation (ref. 1). A considerable amount of data of this type has been assembled for Australian watersheds by Laurenson and Pilgrim (ref. 158) and compared with similar data from U.S. watersheds.

"Standard" infiltration curves have also been advocated with the following equation (ref. 3):

$$f = f_c + (f_o - f_c)e^{-kt} \qquad \text{..........(13a)}$$

where f = infiltration capacity at time t during a storm

f_c = the ultimate infiltration capacity

f_o = infiltration capacity at start of storm

k = a constant

In general f_c and k are assumed to be constants for a particular watershed or soil-vegetation complex and f_o is varied with the antecedent moisture status. As indicated earlier (sections 11.02 and 12.04). Philip's more rational analysis for an ideal soil suggests that both f_c and k should also vary with the antecedent moisture status.

Infiltration theory is concerned with "surface runoff" only and herein lies the chief objection to its use in complete runoff studies. It is not readily integrated with other hydrological factors and attempts to do this have been a major cause of unnecessary complexities in more comprehensive studies.

At this stage it is advantageous to consider the various forms of runoff in more detail.

13.02 Runoff Separation

Three forms of runoff are widely recognised, viz. surface runoff, interflow and groundwater flow, all of which are adequately described in the standard text books. These represent the flows from different types of storage media and in an ideal watershed with abrupt changes between the media it should be possible to ascertain the relative contributions from each by the characteristics of total flow at the outlet.

In real watersheds, however, there may be a number of different storage media gradually merging into each other and closely interconnected. Attempts at estimating the relative contributions under such circumstances, with runoff data alone, are rather speculative. Nevertheless for flood studies there are advantages in separating the relatively steady "base flows" from the more transient "direct runoff" and many arbitrary methods for doing this have been proposed, (ref. 1, 149 and 114).

Most of these methods have little physical significance. Base flows are supposed to correspond approximately with groundwater flows but during flood periods the latter may be either impeded to a considerable extent or virtually negative because of direct recharge from the flood runoff.

Coulter (ref. 172) and Rangana (ref. 171) have shown that the calculated flood runoff can vary by 30% or more and the calculated average infiltration rate can vary by as much as .20 ins/hr. with different methods of base flow separation. This is particularly unsatisfactory for the analysis of infiltration from rainfall-runoff data and detracts considerably from the physical significance of infiltration theory even if "standard" or consistent methods of computation are extensively adopted.

Gross discrepancies have been observed between calculated average watershed infiltration rates and those measured directly with infiltrometers etc., (ref. 158). Some authorities claim that these discrepancies are mainly due to the effects of interflow which tends to be part of surface runoff in watershed calculations but is included in infiltration with infiltrometer measurements. Other studies suggest that interflow is particularly important in forested areas where it can account for up to 90% of the total annual runoff, (see Hertzler, ref. 173). This appears to be supported by some of the results obtained with the Stanford University model.

It should be apparent from the above that applications of infiltration theory depend considerably on the dubious and inconsistent methods available for separating runoff. In reality the source of flow cannot usually be determined and it is misleading to interpret precise manipulations of the data in terms of precise physical processes such as infiltration. From a scientific point of view it would be preferable to perform analyses in terms of less definite concepts that are consistent with the usual data but may be correctly related to specific concepts for special purposes.

A more generalised view of the runoff components will therefore be attempted.

13.03 Time Delay between Rainfall and Runoff

The first attempts of the author to develop a more general approach to this problem were along the rather obvious lines of the traditional unit hydrograph theory. It was thought that the separation of the total hydrograph due to each storm or significant rainfall burst would reveal a consistent shape throughout both the flood and base flow sections when reduced to unit volume. This was found to be a satisfactory approximation for some circumstances but in other circumstances it became apparent that the base flow proportion was significantly larger for larger storms. It was also considered that some attention should be given to the tendencies for peaks to increase (ref.108) and times of rise to decrease (ref.154) with large storms.

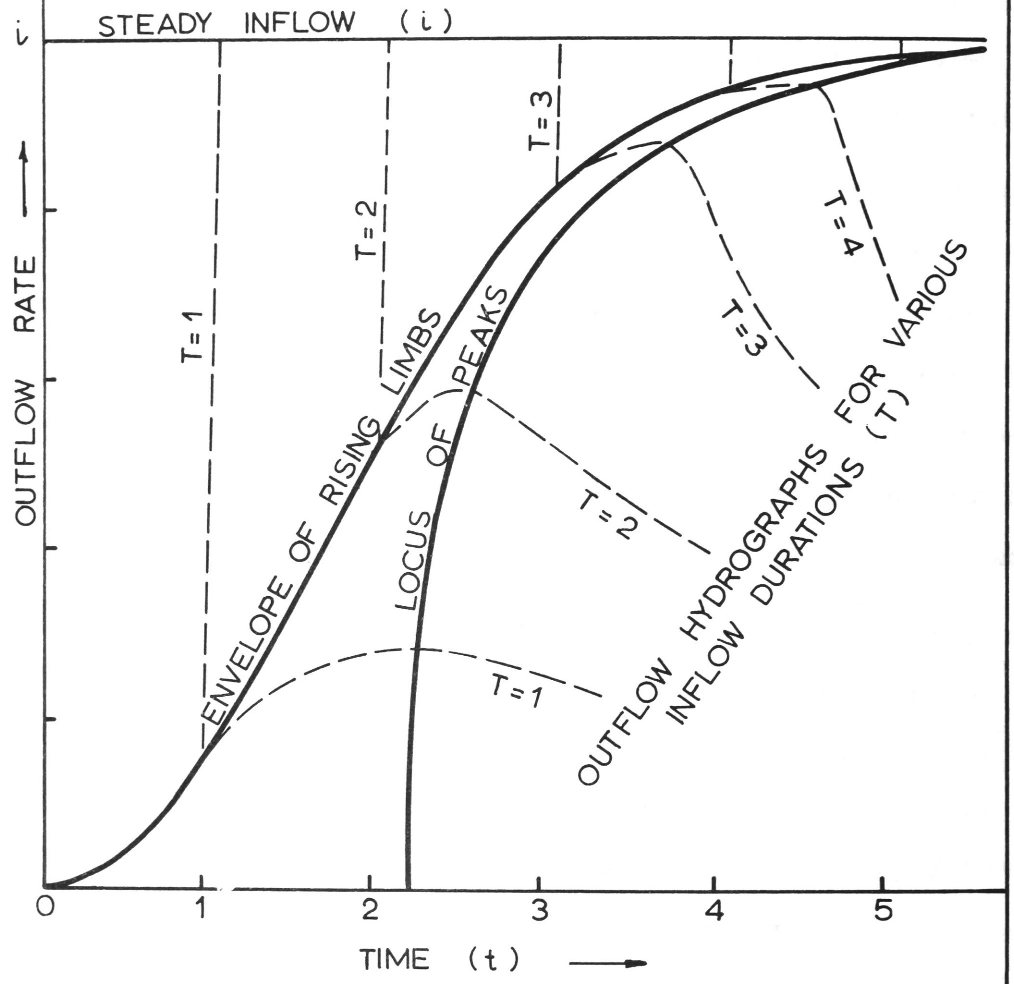
Efforts were therefore made to derive general "response functions" expressing the watershed outflow and rate of change of outflow. These were intended to be in three-variable graphical form (with the variables of watershed inflow, time and watershed outflow) so that any degree of watershed variability could be readily allowed for and examined. Possible approximations by Specific mathematical functions were also investigated. For this purpose previous studies by Zoch (ref.174), Sugawara (ref.153), Levi and Valdes (ref.175), Holten and Overton (refs.176 and 177), Court (ref.178), Velikanov (ref.179) and Wooding (ref.180) were all relevant and some of these provided considerable insight into the physical mechanisms and their mathematical expression.

At this stage there was serious concern about the way in which the studies had spread so broadly and it was decided not to pursue the aspect of time distribution of runoff in further detail. However, it was concluded that many of the relevant factors may be synthesized diagrammatically as shown in figures 16 to 19.

Figure 16 gives the general form of the outflow resulting from a steady inflow i for various periods T (ignoring losses). The locus of the peaks, the envelope of rising limbs, and shape of recession limbs all vary with i and T, and may be regarded as characteristic functions of the particular watershed. Figure 17 shows how the locus of peaks varies, in general, with i and T. Similar graphical functions may be used for the rising limbs and recession limbs.

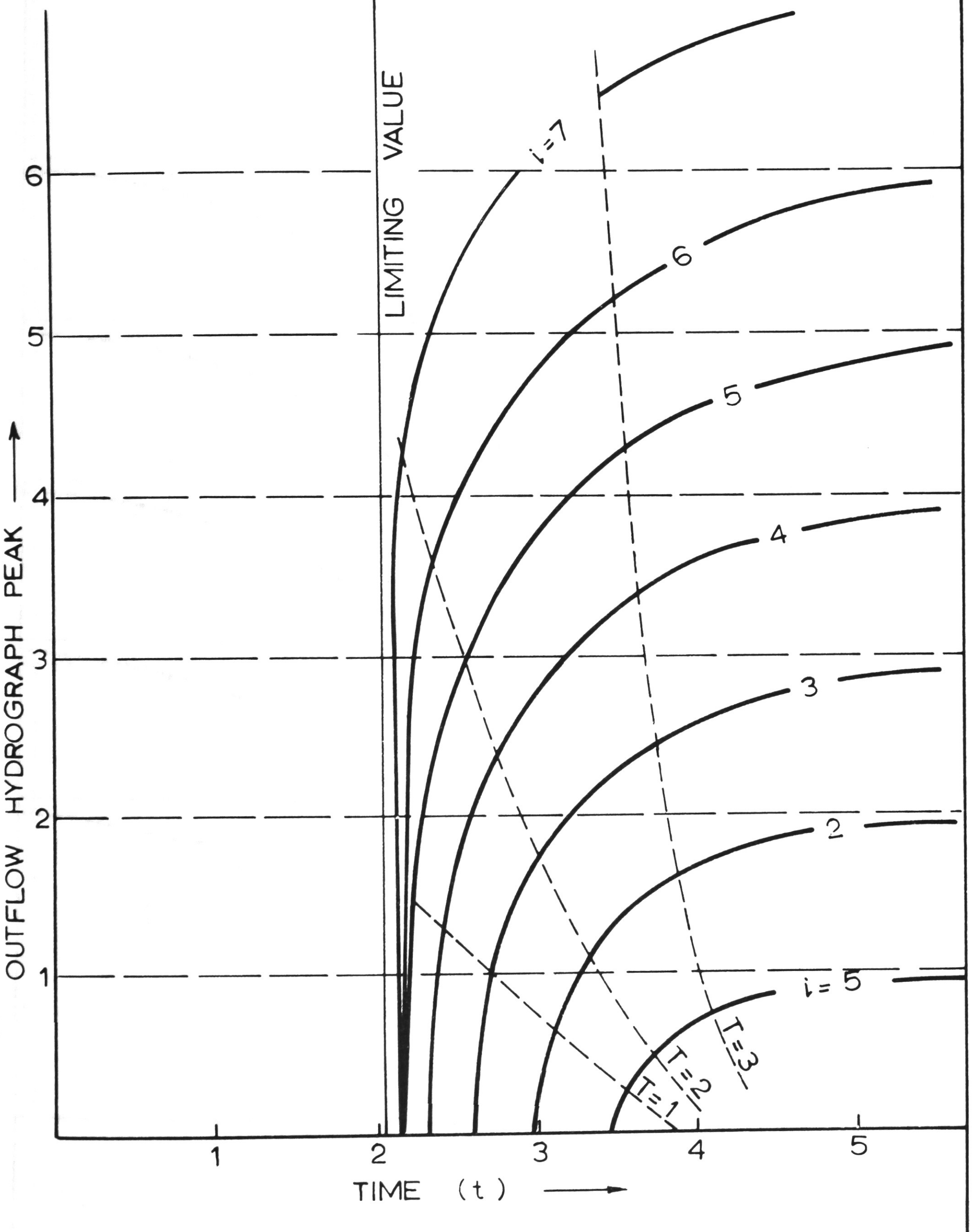
It is interesting to note that unit hydrograph theory and the so called "rational method" are special cases of fig.17, as shown in figs.18 and 19. The envelope of rising of fig.16 also corresponds approximately with the S curve of unit hydrograph theory and the time-area diagram of the rational method. With these diagrams interesting relationships can also be seen between the critical duration, time of rise and "constant discharge diagram," (see ref.43,Appendix A.)

A less cumbersome approach appeared possible with storage and routing concepts along the lines described by Laurenson (refs.154 and 170). It was thought that Laurenson's model for direct runoff could be adapted for total runoff by adding appropriate storages for base flow etc. Suggestions of this type have been made by Dooge (ref.181) who recommended the use of



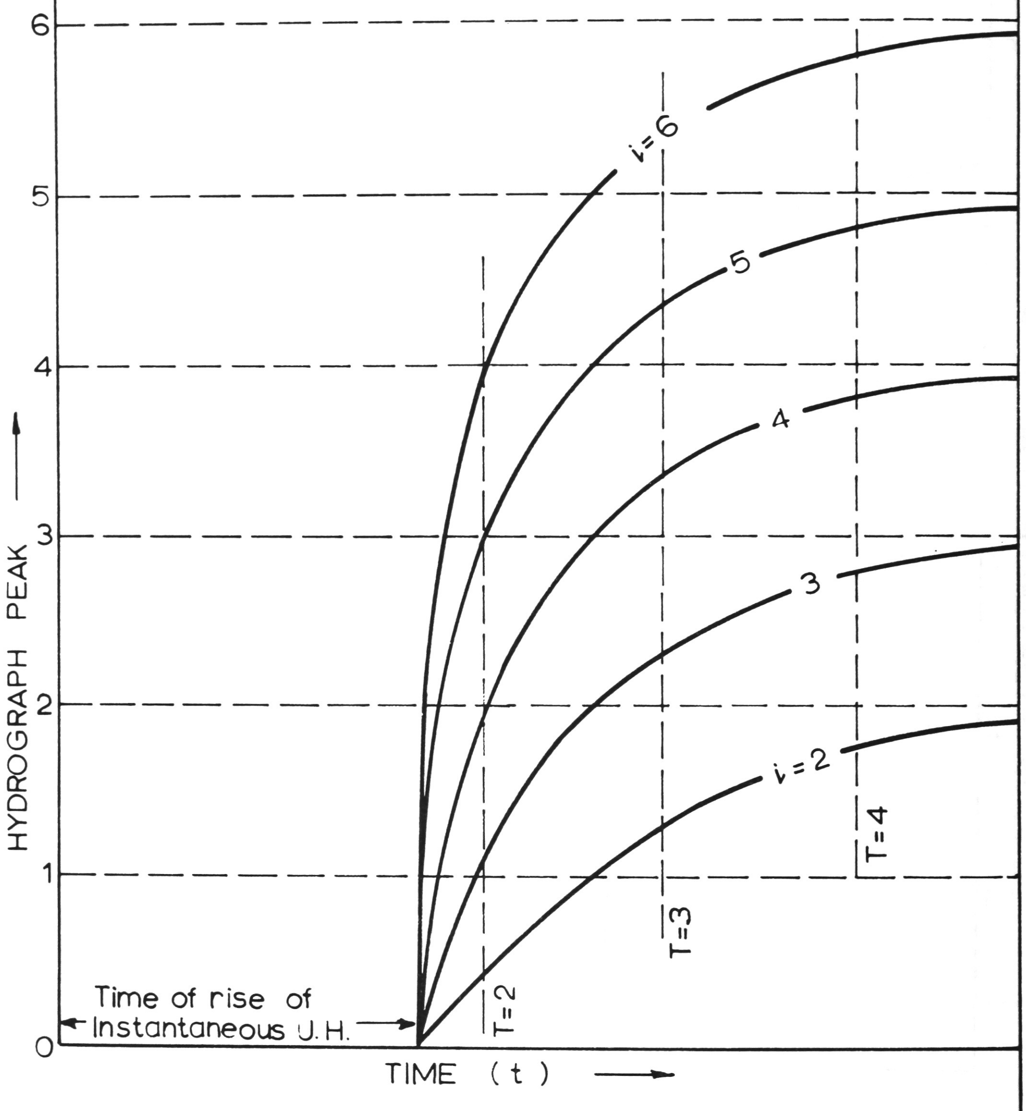
GENERAL FORM OF OUTFLOW HYDROGRAPHS FROM A STEADY INFLOW OF VARIOUS DURATIONS

FIG. 16.



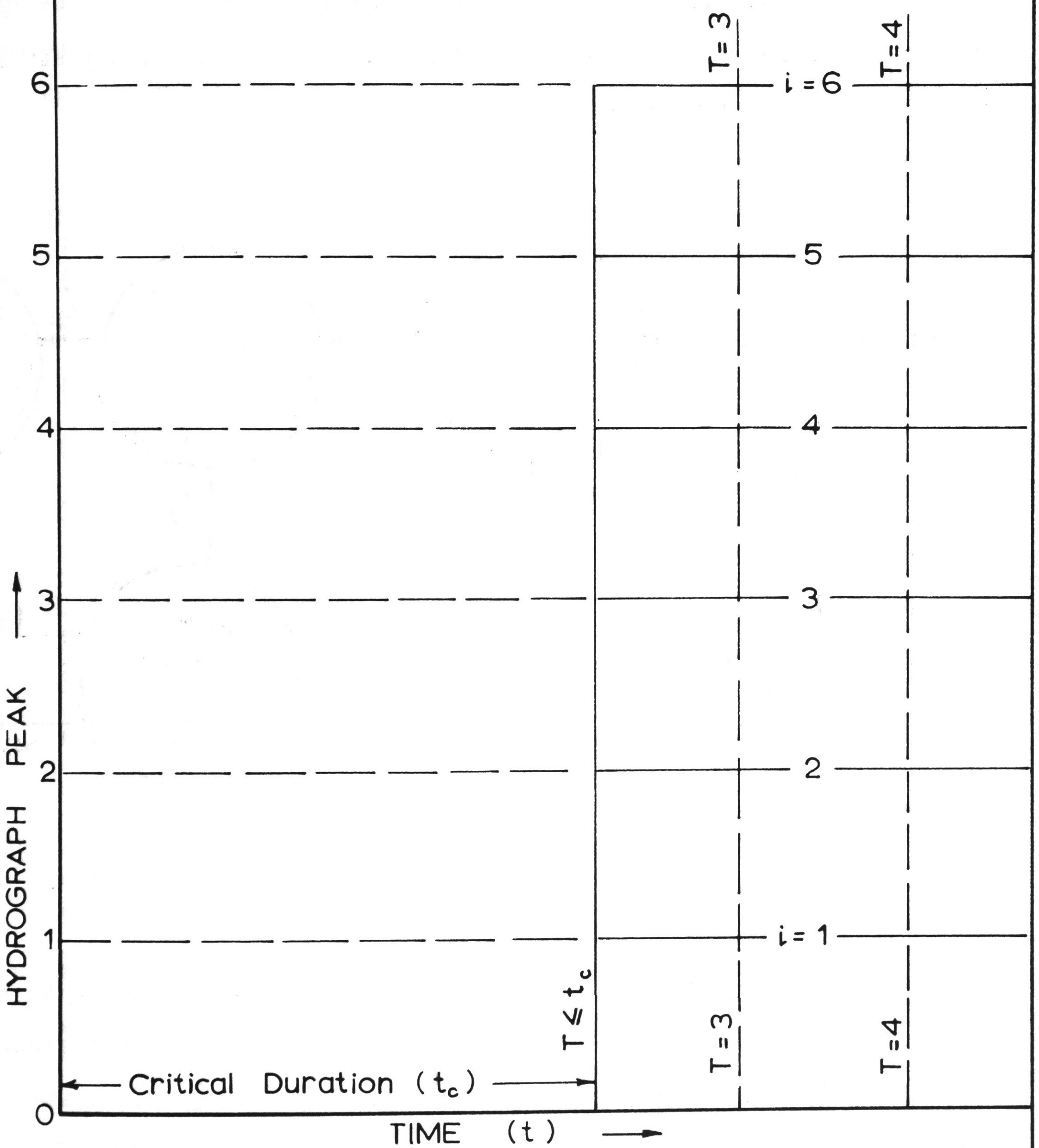
GENERAL FORM OF HYDROGRAPH PEAK LOCI FROM STEADY INFLOWS OF VARIOUS DURATIONS

FIG. 17.



HYDROGRAPH PEAK LOCI FROM STEADY INFLOWS BY UNIT HYDROGRAPH THEORY.

FIG. 18.

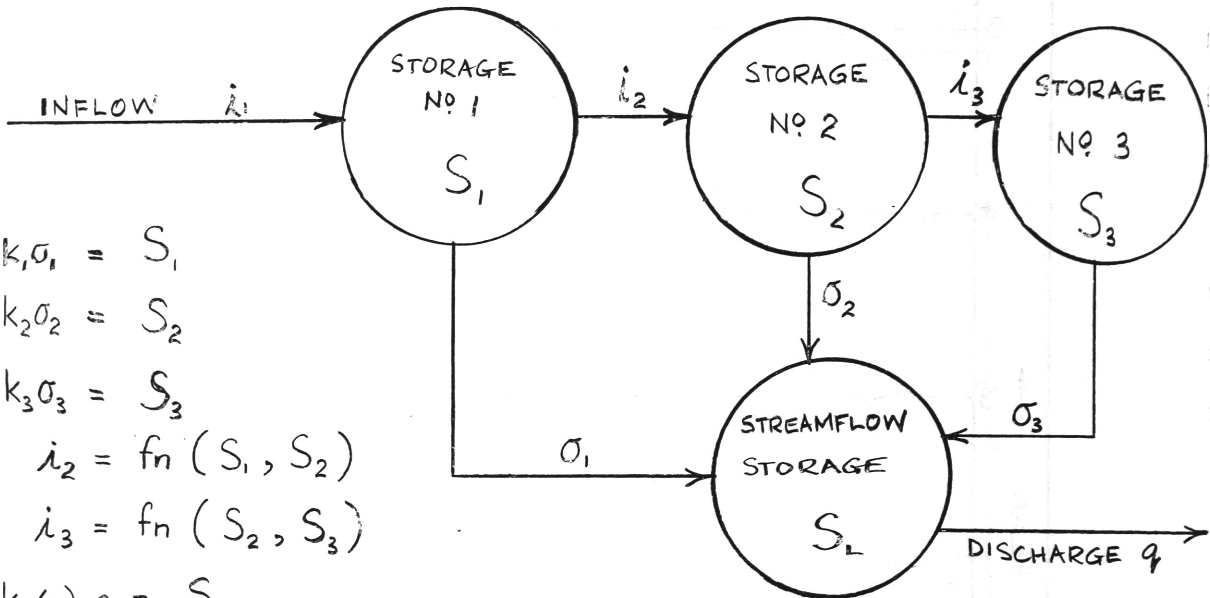


HYDROGRAPH PEAK LOCI FROM STEADY INFLOWS BY RATIONAL METHOD.

FIG. 19.

linear groundwater storage elements in parallel with similar surface storage elements to completely simulate a watershed.

Assuming, initially, that areal variations in rainfall could be ignored, some grouping of elements with similar characteristics seemed justified and the following model was therefore examined:



$$k_1\sigma_1 = S_1$$
$$k_2\sigma_2 = S_2$$
$$k_3\sigma_3 = S_3$$
$$i_2 = fn\,(S_1, S_2)$$
$$i_3 = fn\,(S_2, S_3)$$
$$k_2(q)\, q = S_L$$

FIG. 20

S_1, S_2 and S_3 are linear storage elements with delay times k_1,k_2 and k_3 (in increasing order of magnitude) and corresponding approximately with the watershed surface storage, the interflow storage and groundwater storage respectively. The streamflow storage is intended to simulate the effects of the main channel system of the watershed and its delay time is a non-linear function of the total discharge. The streamflow delay time is approximately equal to the lag between peak inflow to streamflow storage and the resulting peak outflow, (Laurenson, ref.154). This is easily obtained from the data as it is essentially the lag between the end of a distinct rainfall burst and the resulting hydrograph peak.

Although the above model would probably give good results if appropriate parameters were derived by electronic computer, it is subject to some of the objections outlined in 13.02. In particular the derivation of inflow functions to S_2 and S_3 were troublesome with ordinary computational methods and another approach was therefore sought.

If a single non-linear storage replaces S_1, S_2 and S_3, its outflow may be regarded as a joint function of the total storage and the distribution of the water within this storage, (i.e. the relative amounts in S_1, S_2 and S_3).

It is thus implied that the storage is completely specified by the two parameters of outflow and total storage. A small outflow with a large storage indicates a high proportion of groundwater runoff and conversely, a large outflow with a relatively small storage indicates a high proportion of surface runoff.

When there is no inflow the rate of change of outflow is a joint function of the outflow and total storage. This may be called the "runoff depletion function" and is described further in 13.05.

With this system the effect of an inflow on a storage must also be expressed as two parameters, viz. the increase in outflow and the increase in total storage. These are specified by a "runoff accretion function" which will be described in 13.06.

The model based on the above considerations is represented diagrammatically as follows:-

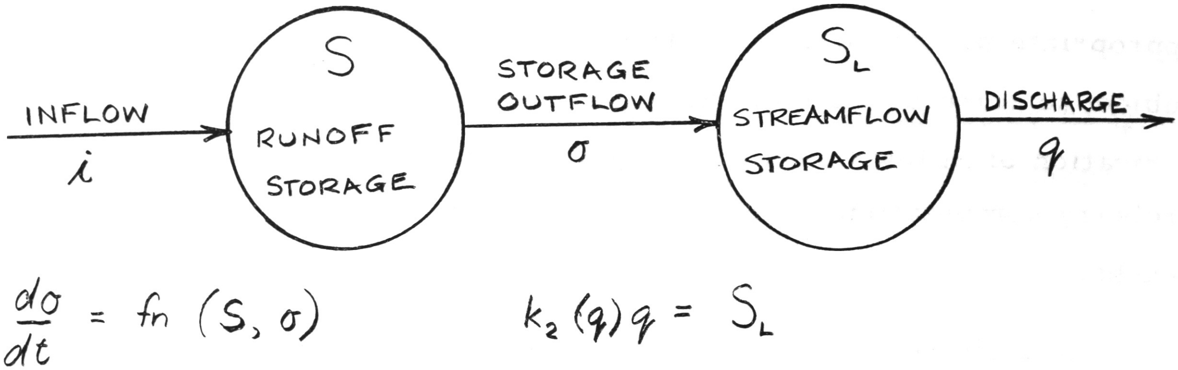


FIG. 21

The two-parameter, non-linear "runoff storage" represents relatively slow-moving, reservoir type storage such as overland "sheet flow," large pools and groundwater. The single parameter, non-linear "streamflow storage" simulates the effects of rapidly flowing water in channels etc.

The logical structure of the above model is not ideal but it seems to be as sound as that of other models and it has the following further advantages:

(a) Computational simplicity

(b) Readily integrated with other components of the rainfall-runoff cycle

(c) Consistent with the objective of developing a general approach

The model was therefore adopted for the remainder of the study and was subsequently found to be highly satisfactory, (see 14.10).

13.04 Retention Theory

"Retention storage" is defined as the volume of water in a watershed that is unlikely to become runoff. "Retention rate" is the rate of increase of retention storage through rainfall. This terminology has been used by various hydrologists in the past, (e.g. refs. 158, 182 and 183) and although it is not used by the recognized text books it is quite suitable for these important concepts.

Retention storage and runoff storage are mutually exclusive and any rainfall entering a watershed should become initially either retention storage or runoff storage.

The "retentivity" of a watershed in a given condition is the retention rate that would occur with rainfall of high intensity. It varies with the condition of the watershed and may be regarded as a parameter of moisture status.

Retention theory asserts that when rainfall occurs the retention rate is a joint function of the retentivity and the rainfall intensity. The specific form of this "basic retention function" is unique for each watershed and depends on watershed characteristics. Retention theory corresponds with infiltration theory if the watershed is homogeneous and interception, depression storage and baseflow are all insignificant.

For practical applications it may sometimes be necessary to assume the simplified form of basic retention function implied by infiltration theory, as shown in fig. 22.

Although it is deliberately indefinite for purposes of logic and generality, retention theory enables the development of simple versatile relationships between watershed rainfall, runoff and evaporation, as outlined in the next few sections.

There are some minor difficulties with the above definitions, for example the theoretical retentivity can sometimes differ seasonally for the same moisture status. This could probably be easily allowed for if necessary, but it has not hindered the studies to date.

14. APPLICATIONS OF RETENTION THEORY

14.01 A Complete Rainfall-Runoff Model

Retention theory and the final model of 13.03 lead naturally to the following complete watershed model:

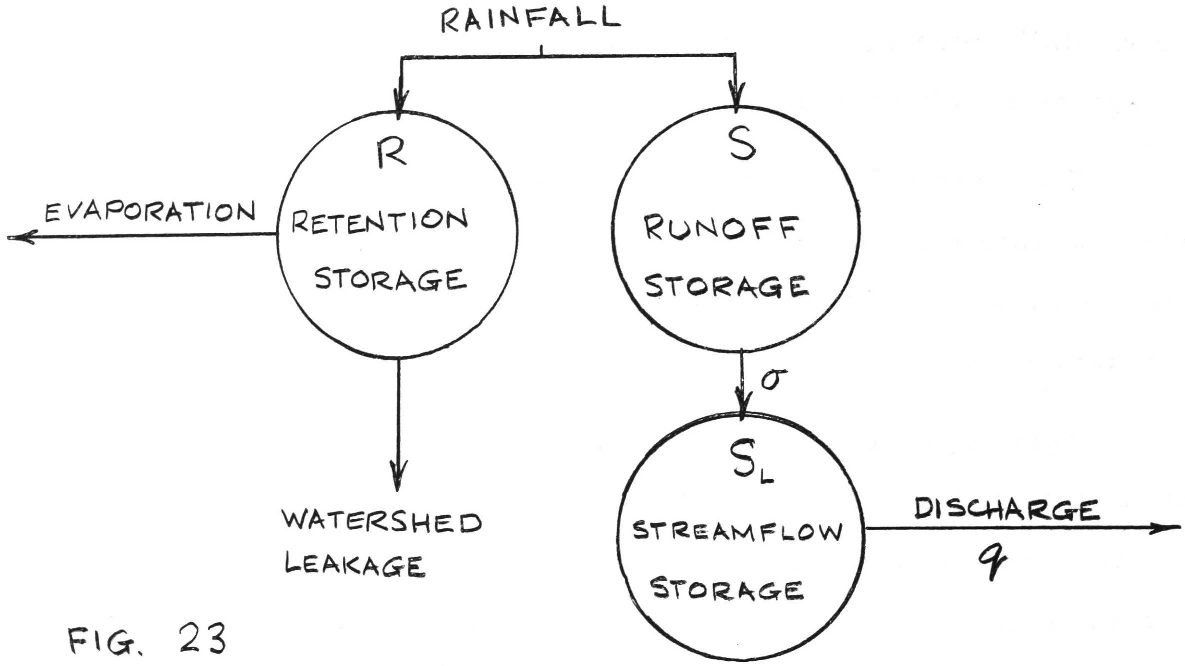


FIG. 23

Retention storage is specified essentially by two parameters viz. the total storage (R) and the retentivity (r). This is somewhat analagous to the method of specifying runoff storage, the retentivity being regarded as an index of the storage distribution in the same way as the discharge is an index of the runoff storage distribution. Low values of r indicate retention storage near the watershed surface while high values of r suggest a dry surface.

It is advantageous to use another parameter of retention storage which may be called the "evaporativity," (w). The evaporativity of a watershed in a given condition is the evaporation rate that would occur with extremely high potential evaporation. It is the limiting rate controlled by the availability of water from the watershed and varies with the moisture status as suggested by the studies in 11.01.

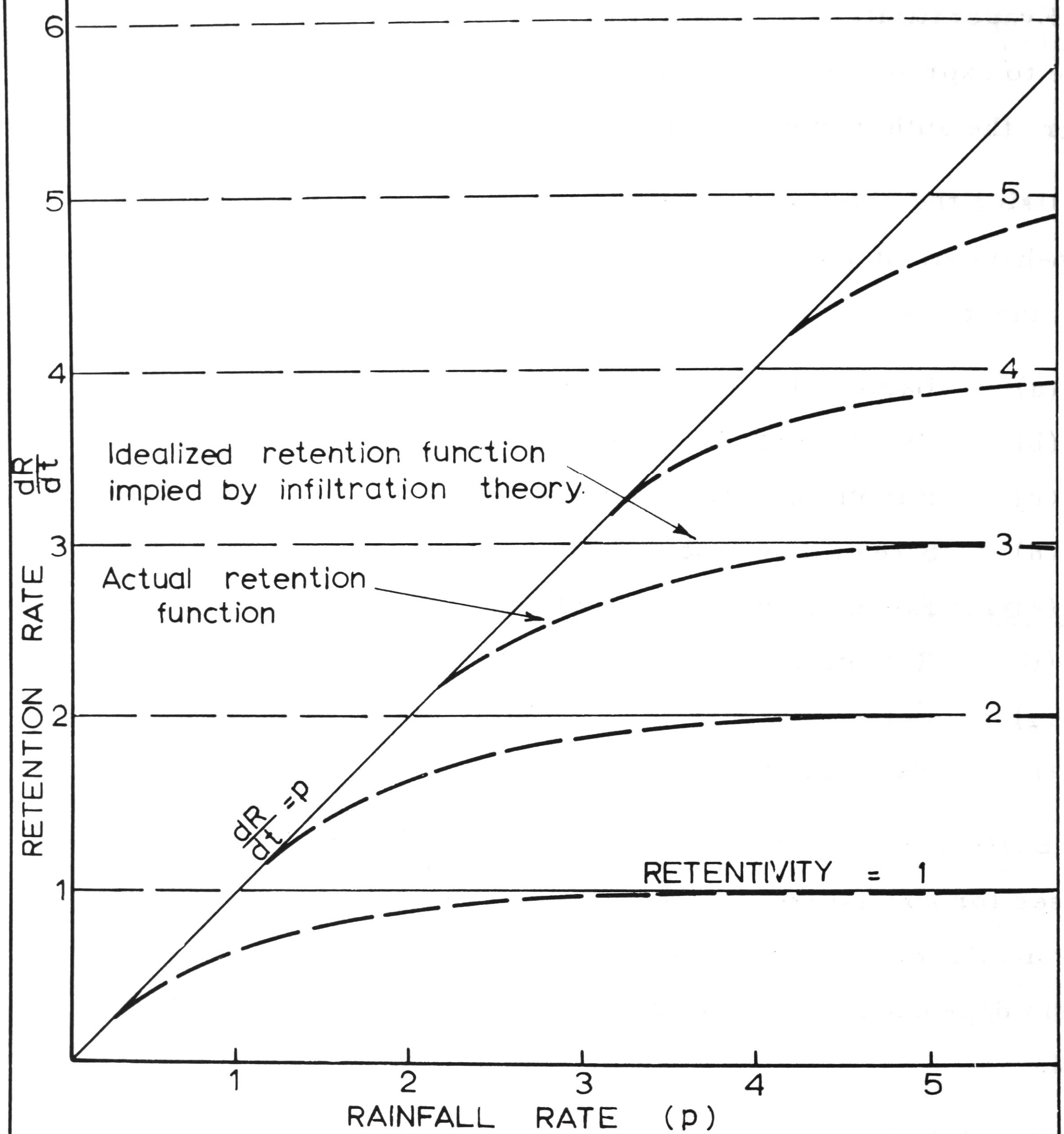
Evaporativity and retentivity are inversely related and it may be possible to express one in terms of the other for general conditions. However, the author has not endeavoured to do this to date.

Using the above concepts and the model of fig. 23 the rainfall-runoff behaviour of a watershed is completely specified and characterized by eight functions viz:

(a) Basic retention function
(b) Runoff depletion function
(c) Runoff accretion function
(d) Streamflow function
(e) Retention depletion function
(f) Retention accretion function
(g) Retentivity recovery function
(h) Evaporativity recovery function

Each of the above can be highly refined and related to individual processes for specialized studies but each can also be approximated by a few physically significant constants when detail is unjustified. The particular emphasis depends on the objectives of the study and the nature of the data available.

Only (b), (c) and (d) can be derived directly from rainfall and runoff data alone. The others may be obtained from these by indirect methods but they are greatly affected by small, inevitable errors in (b), (c) and (d). Under such circumstances it is desirable to assume specific mathematical forms for some of the functions so that complexities of doubtful significance are avoided. Unfortunately, this applies to the Lidsdale data as the soil



IDEALIZED AND ACTUAL RETENTION FUNCTIONS FOR A TYPICAL WATERSHED

FIG. 22.

moisture and interception measurements are not yet adequate for the complete direct derivation of the above functions.

For some studies the idealized form of the basic retention function, as shown in fig. 22, is probably quite adequate. In other cases it is necessary to initially adopt the idealized form and then adjust this to eliminate the bias (if any) in the reproduced data. This technique is described in 14.08.

Each of the functions will now be discussed in more detail.

14.02 The Runoff Depletion Function

The basic form of the runoff depletion function is:

$$\frac{d\sigma}{dt} = fn_1(S, \sigma)$$

where σ = outflow from runoff storage (volume per unit time)

S = runoff storage (volume) (see fig. 26)

In practice it has been found that the function can be derived from the recession limbs of the discharge hydrographs without allowing for the effects of the streamflow storage. This is because in the recession limb σ closely approximate the watershed discharge (q) when the latter is lagged by the relatively constant streamflow delay time (K_L).

In actual analyses and applications the instantaneous rates of change of outflow are not easy to work with and a more practical form of the runoff depletion function is:

$$\sigma_{t+1} = fn_2(S_t, \sigma_t)$$

where σ_t and S_t are outflow and storage at time t.

σ_{t+1} is the outflow at time t + 1 (hours or days)

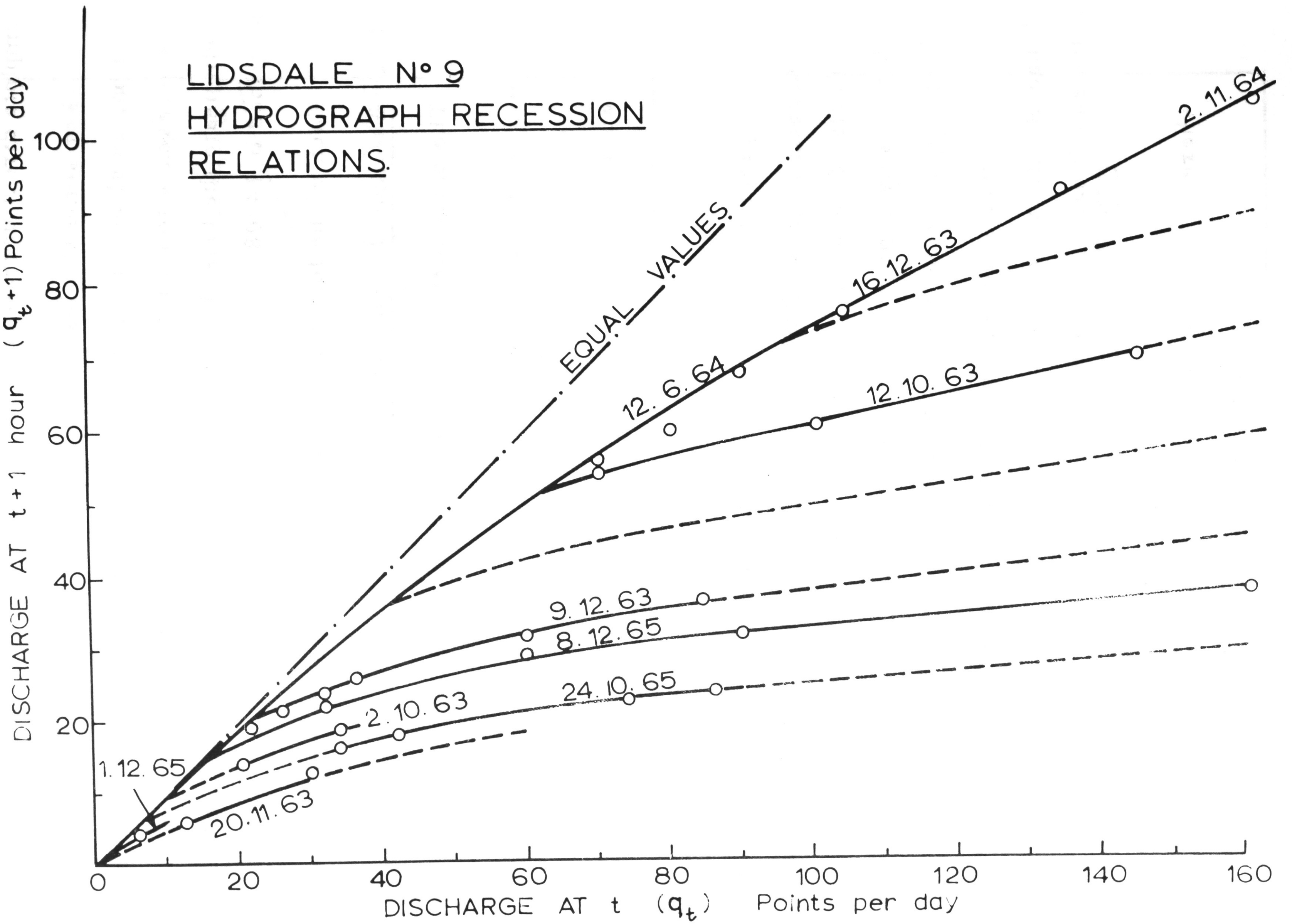

LIDSDALE N° 9
HYDROGRAPH RECESSION
RELATIONS.
DISCHARGE AT t + 1 hour (q_t +1) Points per day
DISCHARGE AT t (q_t) Points per day
EQUAL VALUES.
2.11.64
16.12.63
12.6.64
12.10.63
9.12.63
8.12.65
24.10.65
2.10.63
1.12.65
20.11.63
0
20
40
60
80
100
120
140
160

FIG. 24.

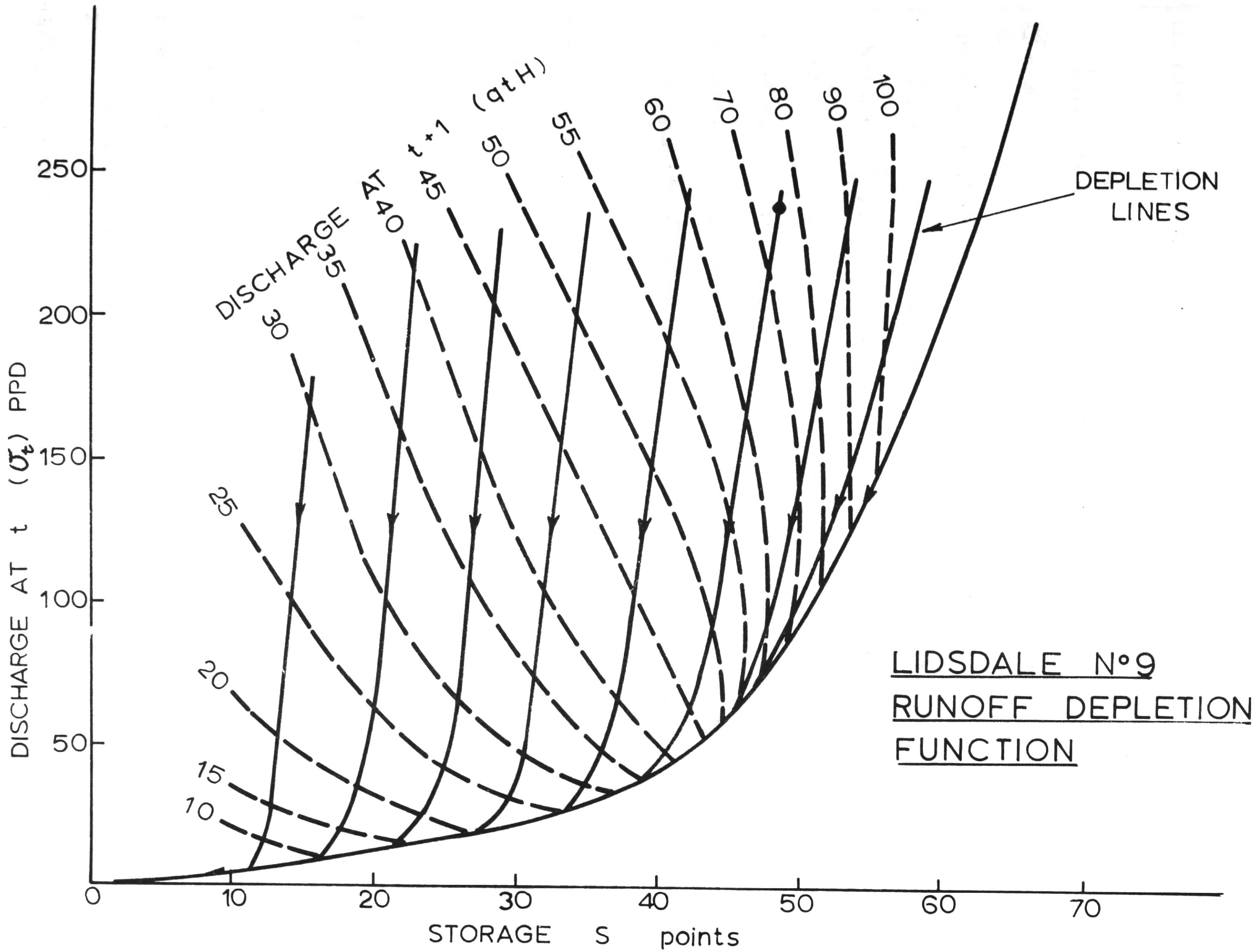

DISCHARGE AT t+1 (qtH)
10
15
20
25
30
35
40
45
50
55
60
70
80
90
100
DEPLETION LINES
LIDSDALE N°9
RUNOFF DEPLETION FUNCTION
DISCHARGE AT t (σ_t) PPD
STORAGE S points
0
10
20
30
40
50
60
70
50
100
150
200
250

FIG. 25

A convenient method of derivation is as follows:

(a) Obtain a wide range of segments of hydrograph recessions for the given watershed.

(b) Reject any segments coinciding with significant rainfall, or of doubtful accuracy.

(c) Plot the remaining segments as in fig. 24, i.e. with q_t versus q_{t+1}.

(d) If necessary, interpolate between segments, as shown by the dashed lines in fig. 24, so that the continuation of each segment to storage exhaustion may be estimated.

(e) Assume that $\sigma_t = q_t$ etc. and for various values of σ_t and σ_{t+1} calculate the corresponding values of S_t by integrating along the lines of fig. 24. These results are plotted in fig. 25 which is the required depletion function.

The above has been tried on a number of watersheds and in all cases the recession segments showed consistent trends similar to fig. 25. The function can be interpreted in terms of surface runoff, interflow and groundwater flow as in fig. 26.

The lower limiting curve of the depletion function represents a major watershed characteristic that considerably influences the rainfall-runoff behaviour. The flow and storages corresponding to this should be given special designations and the terms "saturated flow" and "saturated storage" are therefore suggested. At the lower stages they correspond with the groundwater flow and storage.

In some regions the groundwater storages are extremely high and conditions approaching exhaustion never occur. In such cases it would probably be desirable to consider only the storage above a selected datum level. This would correspond to a "datum flow."

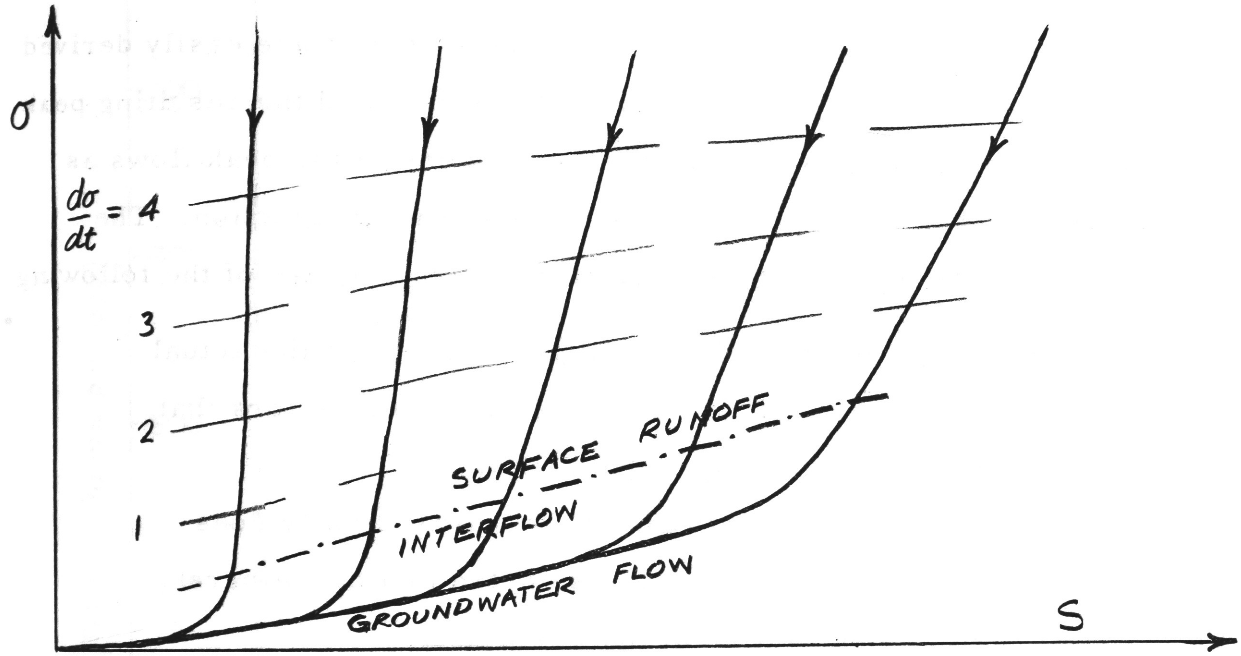


FIG. 26

If it is desirable to assume a specific mathematical form of the function, fig. 26 can be linearized as shown in fig. 27. This implies that the hydrograph recessions are exponential, which is similar to the conventional approach.

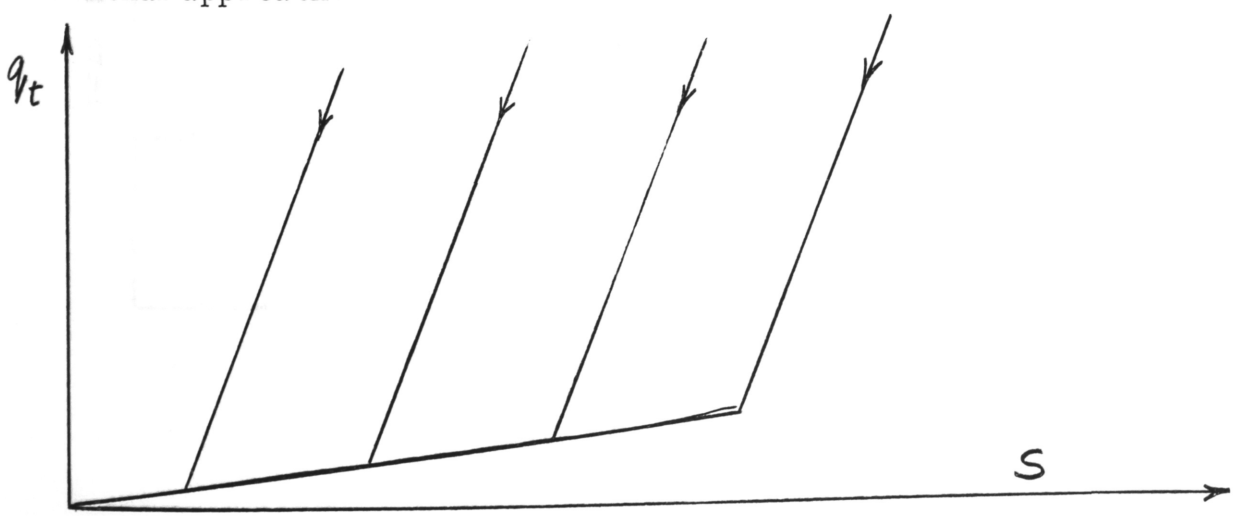


FIG. 27

14.03 The Streamflow Function

As mentioned earlier, streamflow delay times are easily derived from the lags between the ends of rainfall bursts and the resulting peak flows. They vary somewhat with the magnitudes of the peak flows as shown in fig. 28 which may be called the streamflow function. The scatter of points in this figure evidently occurs because of the following:

(a) The implied model is an approximation of the actual physical processes and there are other factors that affect the lag, (e.g. initial flows)

(b) The rainfall contributing to the peak usually does not cease abruptly at the selected end of the burst.

(c) Small time errors in the basic data.

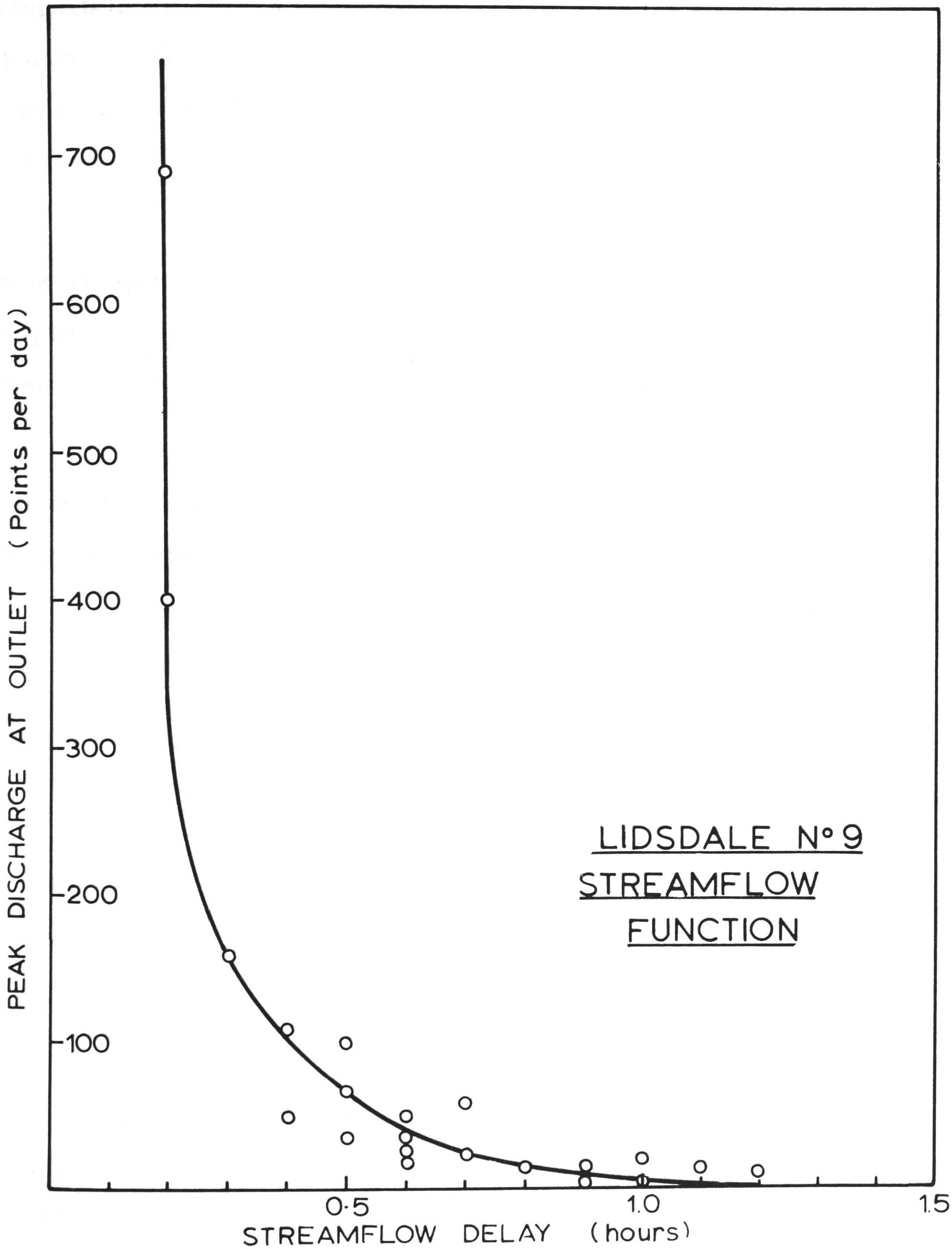

LIDSDALE N°9
STREAMFLOW
FUNCTION
PEAK DISCHARGE AT OUTLET (Points per day)
700
600
500
400
300
200
100
0·5
1.0
1.5
STREAMFLOW DELAY (hours)

FIG. 28

Where necessary the hydrograph of flow at the watershed outlet (q) will be designated by the term "streamflow hydrograph" to distinguish it from the "storage outflow hydrograph" which refers to the direct outflow from the runoff storage (σ). The storage outflow hydrograph may be calculated from the streamflow hydrograph, if required, by reverse routing.

There is usually no need, however, to obtain the complete storage outflow hydrograph because only the peak and start of rise are required for the author's procedures. A relatively simple method may be used for deriving these, as outlined in the next section.

In some cases a constant average (or minimum) streamflow delay time can be adopted without detriment, and the streamflow storage then becomes linear.

14.04 The Runoff Accretion Function

The runoff accretion function specifies the response of the runoff storage to an increment of inflow. It may be expressed by:

$$\Delta S = fn_3 (\Delta I \; \Delta T)$$

$$\Delta \sigma = fn_4 (\Delta I \; \Delta T)$$

where ΔS = increase in runoff storage due to ΔI

ΔI = increment of inflow to runoff storage

ΔT = time from start to end of ΔI

In the above the start and end of ΔI are selected so that the inflow rate is approximately constant during each increment. The length of ΔT is therefore not fixed in any particular storm. The function is derived from runoff records as shown in fig. 29 and as described below:

(a) Select a range of reliable hydrographs with simple clearly defined rising limbs. Reject any with erratic or unusual shapes.

(b) Estimate the recession limbs of the corresponding storage outflow hydrographs by using the previously derived streamflow delay time, as shown in fig. 29.

(c) Estimate the peak of each storage outflow hydrograph by extending the recession limb of (b) beyond the inflexion point back to the appropriate time. This enables the determination of $\Delta\sigma$ (see fig. 29).

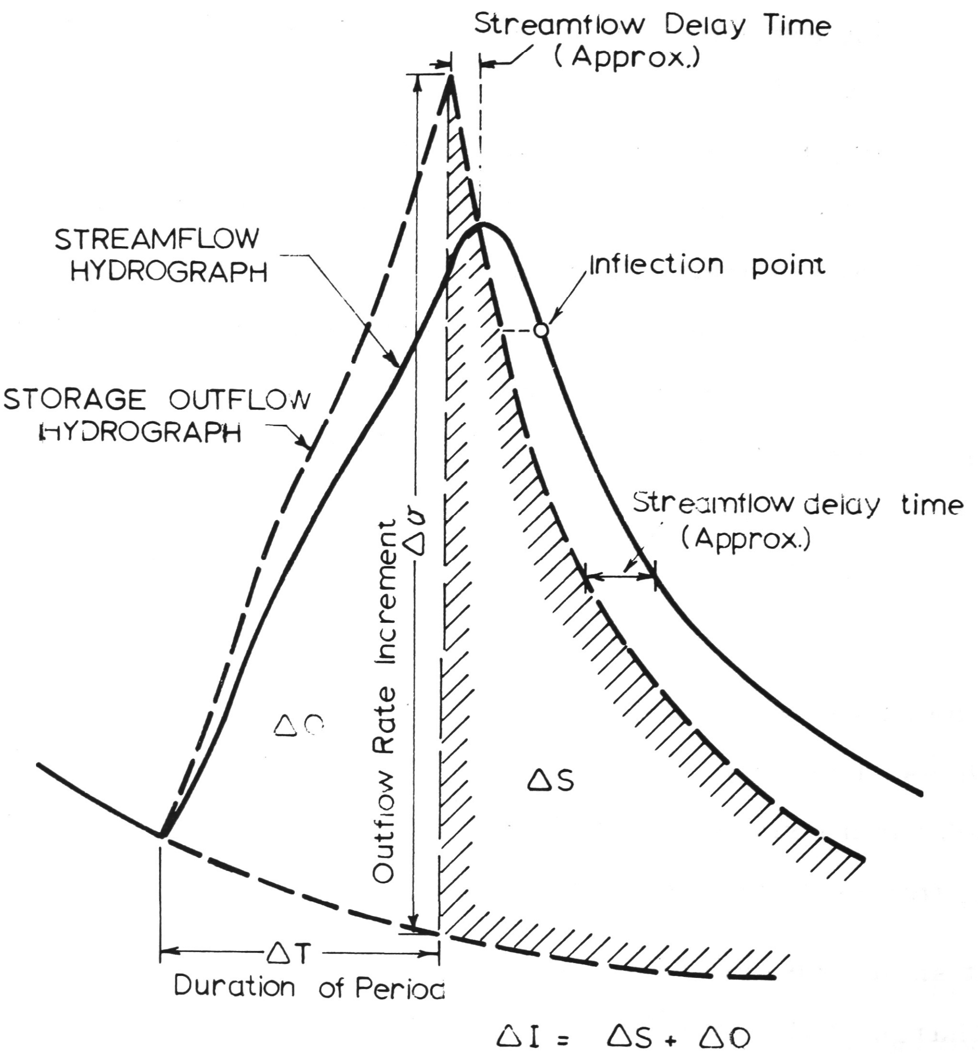
(d) Assume the start of rise of the storage outflow hydrograph coincides with the start of rise of the streamflow hydrograph, enabling the calculation of ΔT from this and (c).

(e) Estimate ΔI and ΔS as shown in fig. 29.

(f) Plot the derived values of ΔI, $\Delta\sigma$ etc. to give the Runoff Accretion Function. A typical example is given in fig. 30 and table 4.

The above was found to be very suitable for the small Lidsdale watersheds. For large watersheds it may be possible to derive similar functions for hourly or other fixed time units but the author's attempts at doing this were not very rewarding.

It should be apparent that the runoff accretion function allows for recharge of the interflow and groundwater storages without assuming specific forms for these processes.



DATA REQUIRED FOR DERIVATION OF RUNOFF ACCRETION FUNCTION.

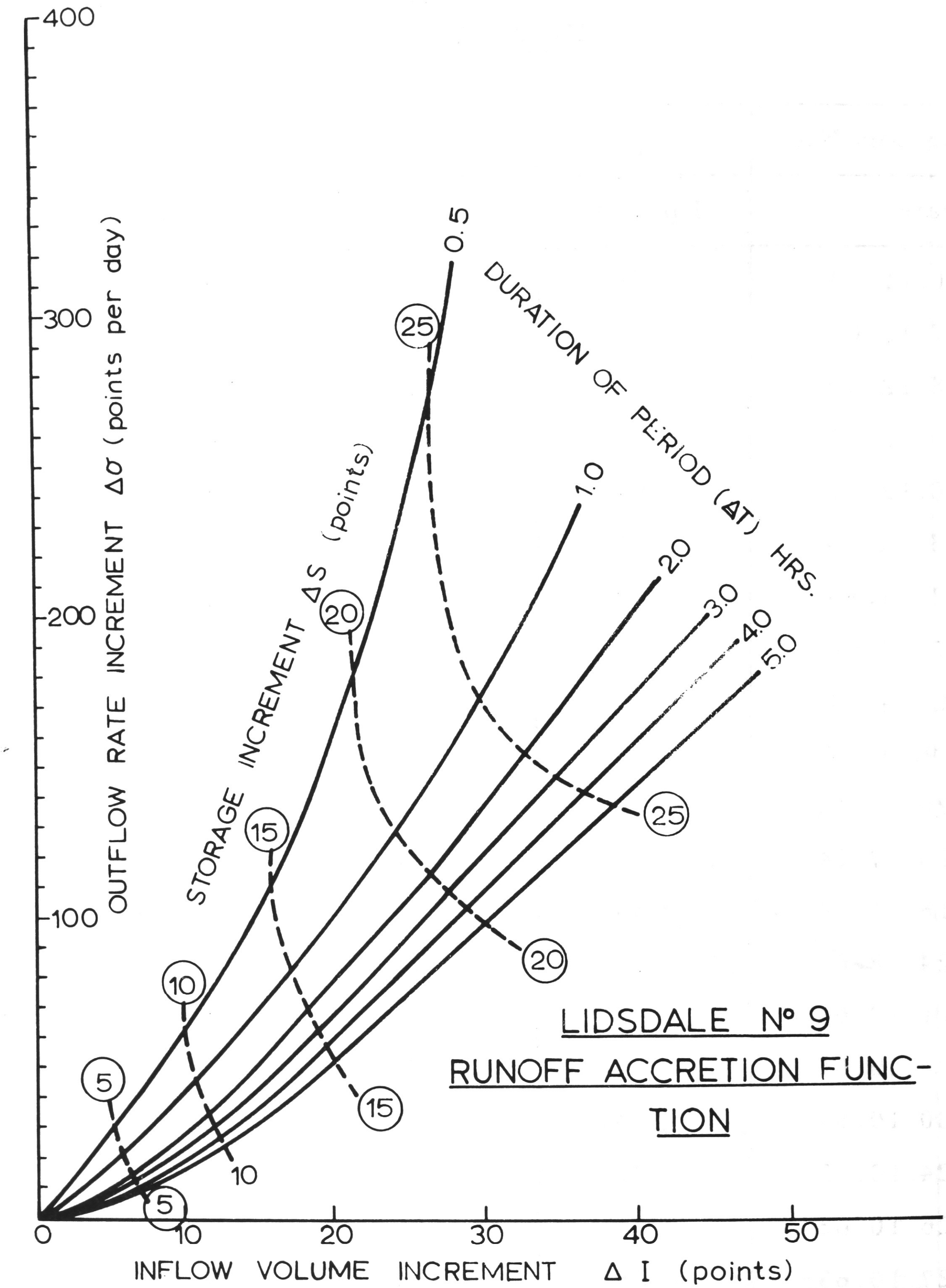
FIG. 29

TABLE 2.

DERIVATION OF RUNOFF ACCRETION FUNCTION

LIDSDALE No. 9

Column No.	2	3	4	5
Date	Δ I points	$\Delta\sigma$ p.p.d	Δ S points	Δ T hours
20.11.63	5	38	5	0.4
7.12.63	6	22	5	1.4
8.12.63	29	100	18	5.0
9.12.63	19	50	14	4.5
10.12.63	17	30	12	6.8
13. 1.64	4	3	4	6.5
22. 4.64	2	6	2	1.0
22. 4.64	3	14	3	0.9
22. 4.64	3	3	3	5.2
19. 6.64	6	13	5	1.1
20. 6.64	8	15	7	2.2
16. 7.64	7	13	6	1.8
16. 7.64	36	115	21	5.8
24. 8.64	12	45	11	1.3
11.10.64	8	15	7	3.2
30.10.64	10	20	8	5.2
30.10.64	3	8	3	1.0
24.10.65	24	140	22	0.7
26.10.65	7	16	6	1.5
27.10.65	21	80	16	2.9
28.10.65	11	40	10	1.3
5.12.65	7	17	6	1.5
8.12.65	31	410	27	0.5
12. 6.64	18	90	16	1.0



OUTFLOW RATE INCREMENT Δσ (points per day)
400
300
200
100
0
10
20
30
40
50
INFLOW VOLUME INCREMENT Δ I (points)
STORAGE INCREMENT ΔS (points)
DURATION OF PERIOD (ΔT) HRS.
0.5
1.0
2.0
3.0
4.0
5.0
5
10
15
20
25
LIDSDALE N° 9
RUNOFF ACCRETION FUNCTION

FIG. 30

14.05 The Retention Depletion Function

The general form of the retention depletion function is:

$$\frac{dw}{dR} = fn_5 \; (w, R)$$

where w is the evaporativity and R is the total retention storage.

This allows for most of the factors mentioned in sections 11 and 12, and can be represented diagrammatically as in fig. 31.

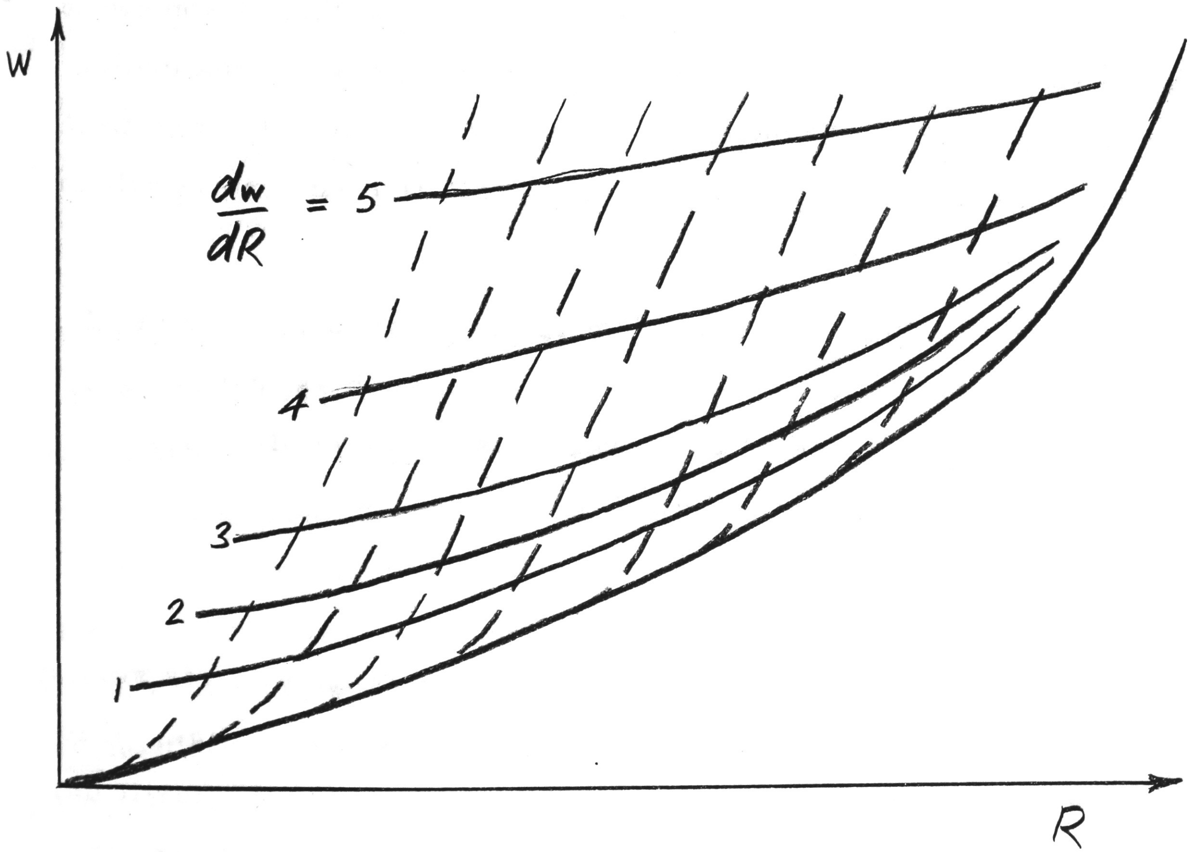


FIG. 31

The previously mentioned studies suggest that the drying phase should be divided into periods called "free drying" and "limited drying" periods.

In a free drying period the evaporation from a watershed is not restricted by the supply of energy and its rate is therefore equal to the evaporativity. This occurs commonly under warm dry climatic conditions.

In a limited drying period the evaporation is restricted by the supply of energy rather than the supply of moisture and it is therefore equal to the potential rate. This would be expected frequently in cool moist climates.

The above is essentially a "threshold" concept that is somewhat analagous to infiltration theory. It is a similar type of approximation and subject to similar difficulties but it is very useful in the practical derivation and application of the retention depletion function as will be demonstrated in later sections.

Since this function cannot be completely derived with the usual data, it is necessary to assume a specific form such as that of fig. 37. The conditions represented by the lower "base" line of slope 1/m in fig. 37 are expressed by:

$$w = \frac{R}{m}$$

Where m is a constant with the dimensions of time and is similar to the storage delay time relating discharge and runoff storage. It is a representative or characteristic "depletion delay" for the particular vegetation-soil complex of the watershed. In the example shown for Lidsdale No. 9, m was estimated initally from gypsum block observations during appropriate periods, (warm weather, advanced drying phase), assuming that most of the retention storage was in the form of soil moisture. Some small adjustments to the initial estimate of m were made later to give a better fit to the rainfall-runoff data. The more transient conditions represented by the family of parallel lines of slope $1/m$, in fig. 37 were estimated similarly, with some arbitrary allowances being made for surface moisture and interception in the early parts of the drying phase. (See also section 14.07).

The above retention depletion function could have been derived by trial and error methods from the rainfall-runoff data alone with the use of a computer but the formulation of a suitable programme and the preliminary testing would have involved more time than was available.

Once derived, the retention depletion function provides a very simple method for accounting the watershed moisture losses during a drying phase. Given the initial values of w and R (from previous accounting), the progress of the drying phase is followed along the depletion lines of the function. As an example, the complete accounting for Lidsdale No. 9 is given in 14.10 where fig. 37 is used for the calculation of changes in retention storage during drying periods.

In free drying periods the orthogonal time lines of fig. 37 show the values of w and R for successive days, and allow the conditions at the end of each period to be read off directly.

During limited drying periods the same depletion lines are followed but the time lines are ignored. The changes in R are equal to the potential evaporation for these conditions and corresponding changes in w are read directly from the figure. (See complete example in section 14.10).

In practice, drying periods tend to be either free or limited for extended durations and a very simple calculation gives the end or start of each period. Most drying phases consist of only one or two periods. The accounting can be done on a daily basis, if desired, but in the Lidsdale computations each drying period was treated as a single time unit and this reduced the volume of work considerably.

14.06 The Retention Accretion Function

Some aspects of the retention accretion function are analagous to the retention depletion function. Its general form may be expressed by:

$$\frac{dr}{dR} = fn_6 \quad (r, R)$$

This is represented in fig. 32.

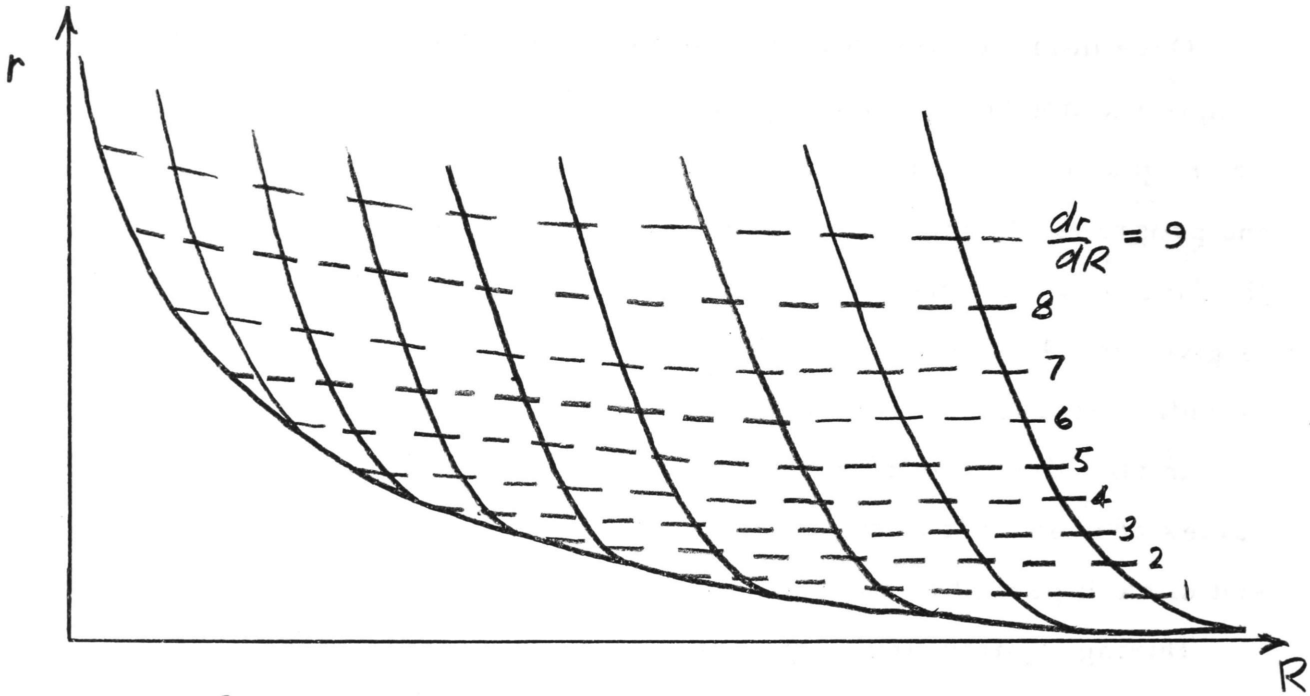


FIG. 32

The wetting phase is divided into "free wetting" and "limited wetting" periods. During free wetting periods the retention is not restricted by the supply of rainfall and it is therefore equal to the retentivity. This may be assumed to occur when a significant percentage of the rainfall becomes runoff inflow.

In limited wetting periods the actual retention rate is less than the retentivity because the supply of rainfall is restricted. When this occurs the runoff inflow is relatively small and the actual retention rate is approximately equal to the rainfall rate. The assumption of exact equivalence in such cases implies the idealized retention function which is discussed in 14.08.

To derive retention accretion functions for the Lidsdale watersheds it was necessary to assume the linear form given in fig. 39. The complete derivation required the estimation of the following:

(a) The slopes of the "base line" ($1/n$) and transient lines ($1/n'$)

(b) The maximum retention storage

(c) The minimum retentivity

n and n' are "accretion delay times" and may be obtained from the ratios of retention storage increments (ΔR) to corresponding changes in retention rates during continuous free wetting periods.

This procedure is demonstrated in table 3 for Lidsdale No. 9 which shows the derivation of the parameters from a number of wetting phases. These particular phases were selected because:

(a) They contained wetting periods that correspond with distinctly separable hydrographs.

(b) The retention rates could be assumed equal to the retentivities as the runoff rates were relatively high.

Because of (a) it was possible to calculate the average retention rate and change in retention storage for each period. These calculations are not shown in table 3.

The difference between the average retention rates for two consecutive periods (column 3 of table 3) therefore represents a decrement of retentivity in the retention accretion function (see fig. 32). The corresponding increment of retention storage (column 4 of table 3) is approximately equal to the mean of the two consecutive retention increments minus a small allowance for evaporation.

The ratios of the retention increments to the retentivity decrements are shown in column 5 of table 3 and these are estimates of the accretion delay time. They tend to occur in two groups of values with averages of .380 and .098 which are assumed to be the base and transient accretion delay times respectively (n and n').

Two independent methods were used to estimate the maximum retention storage for Lidsdale No. 9. The first consisted of adding together the maximum total retention of the observed consecutive limited wetting periods and the maximum total retention of the observed consecutive free wetting periods. The underlying logic of this procedure is not altogether convincing but it is intended to simulate a wetting phase in which the retention storage is recharged completely from exhaustion to maximum.

In the second method, the maximum retention storage was obtained from the estimated total soil moisture storage between field capacity and wilting point (based on soil measurements) plus an additional 20 points to account for interception and surface storage (ref. 132). The assumptions implied by this were those adopted for the derivation of the depletion function as mentioned in 14.05.

Both of the above methods indicated a maximum retention storage of approximately 450 points which was therefore adopted for the given watershed.

The minimum retentivity of Lidsdale No. 9 was estimated from the records of the June 1964 storm in which the watershed was evidently "saturated" for much of the wetting phase. During these periods approximately 98% of the rainfall became runoff inflow and the retention

TABLE 3.

DERIVATION OF RETENTION ACCRETION FUNCTION

Column No.	2	3	4	5	6
Date	TIME	EST. Δr p.p.d	EST. ΔR points	n days	Trans or Base
7.12.63	1230				
		1080	103	.10	T
8.12.63	1130				
		412	63	.15	?
9.12.63	0450				
		45	12	.38	B
10.12.63	0600				
22. 4.64	0300				
		605	62	.10	T
	0620				
		500	47	.09	T
	1510				
9. 6.64	0130				
		221	83	.38	B
10. 6.64	1000				
12. 7.64	1420				
		200	24	.12	T
13. 7.64	2400				
16. 7.64	0450				
		16	6.5	.41	B
	2310				
24. 8.64	0010				
		355	44	.12	T
	1420				
		240	18	.08	T
25. 8.64	0230				

TABLE 3 (Contd.)

DERIVATION OF RETENTION ACCRETION FUNCTION

Column No.	2	3	4	5	6
Date	TIME	EST. Δ r p.p.d.	EST. Δ R points	n days	Trans or Base
11. 9.64 12. 9.64	0620 0820	84	29	.34	B
28. 9.64 29. 6.64	0210 1800 0930	182 110	16 20	.09 .18	T ?
30.10.64 31.10.64	2330 0720	76	30	.39	B
27.10.65 28.10.65	1800 0100	164	15	.09	B

Sum of Transient Group = .10 + .10 + .09 + .12 + .08 + .09 + .09

= .79

Mean n' $= \frac{.79}{8} = .098$

Sum of base group = .38 + .38 + .41 + .34 + .39

= 1.90

Mean n' $= \frac{1.90}{5} = .380$

rate was relatively steady at an average value of about 12 points per day, accounting for leakage and evaporation. A relatively small seasonal variation may be expected but there is insufficient data to enable the estimation of this effect, and so the adopted minimum value was 12 points per day.

The retention accretion function is used for the accounting of retention storage during wetting periods in the same manner as the depletion function during drying periods. The difference between retention increments and rainfall is the required runoff inflow. (Table 5 gives the complete accounting for Lidsdale No. 9, reproducing runoff estimates for the rainfall data and watershed functions.)

During limited wetting periods the time lines are ignored but changes in retentivity for various increments of retention are assumed to be the same as for free wetting periods. This assumption may result in significant discrepancies for a prolonged or discontinuous period, in which case an allowance for recovery of retentivity may be desirable. (see 14.07).

In limited wetting periods the actual retention rate is usually equal to the rainfall rate when the latter is much lower than the retentivity. As the rainfall rate approaches the retentivity rate, however, a small amount of runoff may occur and this is either ignored or allowed for by one of the methods suggested in 14.08.

14.07 The Recovery Functions

At the end of a wetting phase retentivity is usually low and it recovers progressively during the following drying phase. The studies of section 11 suggest that the recovery is rapid in the early part of the drying phase and then becomes slower with time.

These conditions may be satisfied if the changes in retentivity are assumed to be proportional to the evaporation losses during the drying period, i.e.:

$$\Delta r = - \frac{\Delta R_v}{j}$$

where j = retentivity recovery time

ΔR_v = change in retention storage due to evaporation

The above "retentivity recovery" function may be derived from the rainfall-runoff data with the following procedure:

(a) Select periods of good records for which the retentivity and retention storage may be estimated at start and finish, (e.g. periods of three months or more between very wet or very dry spells). Changes of retentivity within these periods are due to intermediate rainfall and evaporation.

(b) Calculate the total change in retentivity ($\Sigma \Delta r$) total change in retention storage ($\Sigma \Delta R$), total rainfall (P) and total evaporation ($W = P - Q - \Delta R$, where Q = runoff due to P).

(c) For each period compute the sum of the retentivity decrements due to rainfall ($\Sigma \Delta r_p$) using the retention accretion function. If retention accretion conditions are entirely within the transient region of the function $\Sigma \Delta r_p$ is given by P/n'.

(d) Calculate the retentivity recovery time from:

$$j = \frac{\Sigma \Delta R_v}{\Sigma \Delta r_v} = \frac{W}{\Sigma \Delta r_v} = \frac{W}{\Sigma \Delta r - \Sigma \Delta r_p}$$

These calculations for Lidsdale No. 9 are shown in table 4.

It should be noted that the retentivity recovery time cannot be less than the retention accretion time and is usually somewhat higher as indicated by the slopes of the accretion and recovery lines of fig. 39. The "drift" towards the left hand side of the function during recovery allows for the abstraction of moisture from the lower layers of the soil profile concurrently with depletion from the upper layers.

The "evaporativity recovery" function, when treated similarly, is expressed by:

$$\Delta w = \frac{-\Delta R_p}{h}$$

where h = the evaporativity recovery time

ΔR_p = change in retention storage due to rainfall

It was found for Lidsdale No. 9 that a maximum value of evaporativity was desirable and this was assumed to be 30 points per day which is approximately equal to the maximum potential evaporation rate at Lidsdale. It was also found that the evaporativity recovery time could be regarded as equal to the transient depletion delay (m') so that the transient depletion lines of fig. 37 may also be used in the reverse direction for the recovery of retentivity.

The justification for both of these assumptions is essentially empirical although various interpretations could be devised in terms of effective retention layers and vegetation adaptations. Detailed examination of table 5 suggests that moderate departures from the assumptions would have only small effects on the calculations of retention storage and retentivity.

14.08 The Basic Retention Function

When data is limited, e.g. when only runoff and rainfall records are available, it is usually necessary to assume idealized watershed functions

TABLE 4.

DERIVATION OF RETENTIVITY RECOVERY TIME

Column No.	2	3	4
DATE	RAINFALL	RUNOFF	RETENTION
	Points	Points	Points
12.12.63	24	2	22
16.12.63	200	158	42
26.12.63	42	5	37
29.12.63	60	2	58
12. 1.64	105	4	101
24. 1.64	30	0	30
7. 2.64	20	0	20
15. 2.64	10	0	10
29. 2.64	18	0	18
8. 3.64	51	0	51
8. 4.64	54	0	54
20. 4.64	28	0	28
21. 4.64	312	8	304
26. 4.64	12	0	12
3. 5.64	72	3	69
11. 5.64	19	0	19
26. 5.64	20	0	20
30. 5.64	65	1	64
7. 6.64	13	0	13
Totals:	1.155	183	972

Est. R and r on Dec. 10, 1963: $R = 440$ $r = 40$

" " " " " June 8, 1964: $R = 270$ $r = 1,250$

$\therefore$ $r = 1210$ $R = 972 + 270 - 440 = 802$

R_p (i.e. due to rain) $= 972$

r_p (i.e. due to rain) $= -(972 \times 10.2) = -9920$

R_v (i.e. due to evap.) $= 972 + 440 - 270 = 1142$

r_p (i.e. due to evap.) $= 9920 + 1210 = 11,130$

Retentivity recovery time (j) $= \dfrac{1142}{11130} = .1025$ days

$$1/j = \frac{1}{.1025} = 9.75$$

and convenient forms have been suggested for the retention accretion, retention depletion and recovery functions. The most convenient idealized form of basic retention function is the universal one implied by infiltration theory as shown in fig. 22.

It should be clearly recognized that the actual retention rates and the retentivity are only theoretically equivalent for very high rainfalls when 100% of the watershed is contributing to runoff. The idealized function of fig. 22 is a practical approximation that may apply to homogeneous watersheds but the degree of departure from this form depends on the degree of heterogeneity within the watershed. The greater the variability of retention characteristics, the higher the rainfall intensity must be before 100% of the watershed is contributing to runoff.

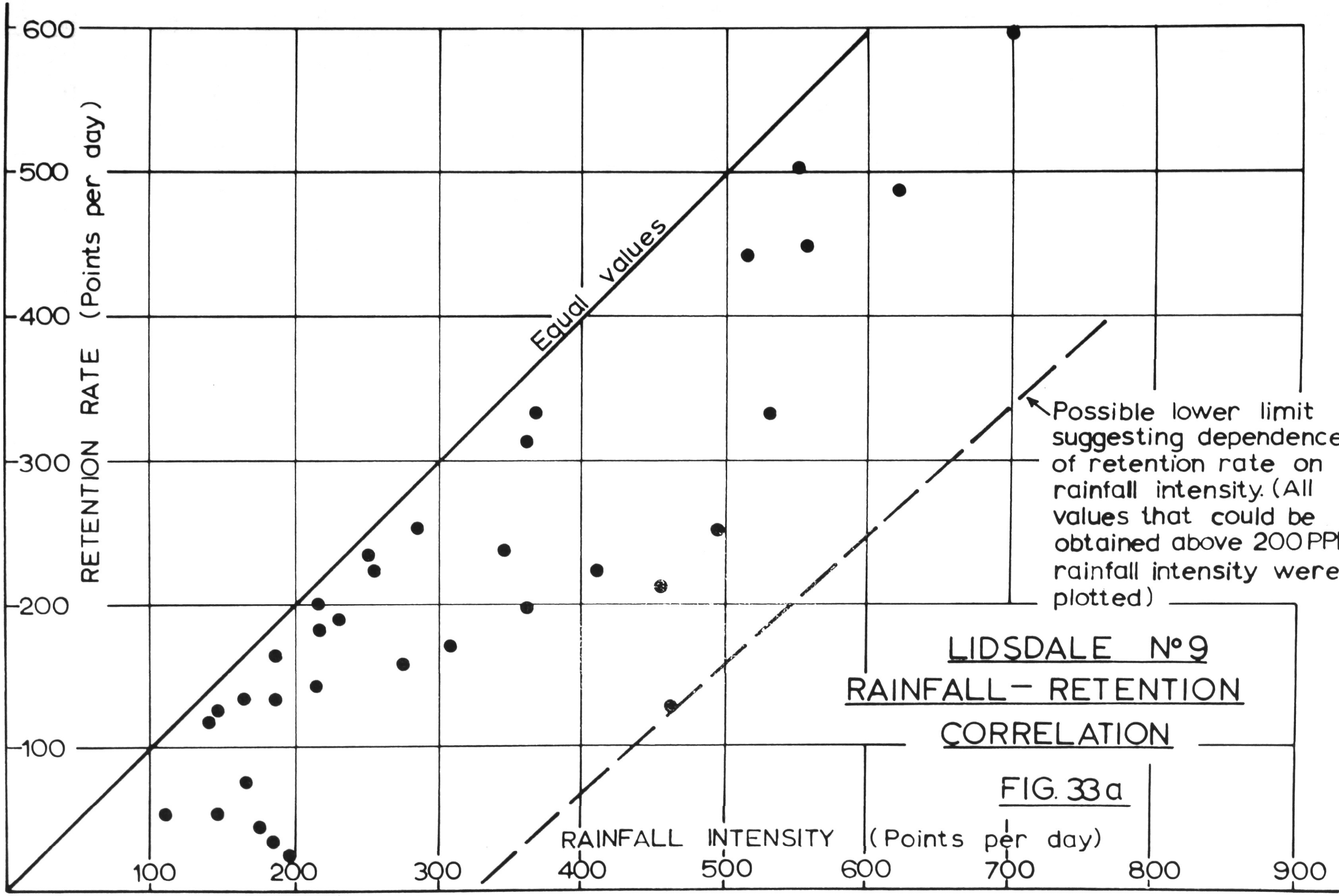
The above is directly allowed for by the infiltration equations of the Stanford University Model (section 8) and indirectly allowed for by the "runoff producing area" of Betson's model (section 10.02). The subdivision of a watershed into relatively homogeneous units (as for the U.S.W.B. Model) also gives indirect allowances and this will be discussed in 14.09.

Some attempts were made to assess the departure of the basic retention function of Lidsdale No. 9 from the idealized form by the following methods:

(a) The correlation between retention rates and rainfall intensities.

(b) An analysis of the bias in reproducing the runoff data from the rainfall records by using the idealized function.

In the correlation of (a), (fig. 33a) there was a notable absence of low retention rates with high rainfall intensities which appeared, superficially, to support the hypothesis of a significant departure. Unfortunately, this could also be due to the seasonal incidence of high intensity storms (see Pilgrim, ref. 184) and no definite conclusions should therefore be drawn from the correlation.



LIDSDALE Nº 9
RAINFALL — RETENTION CORRELATION

FIG. 33 a

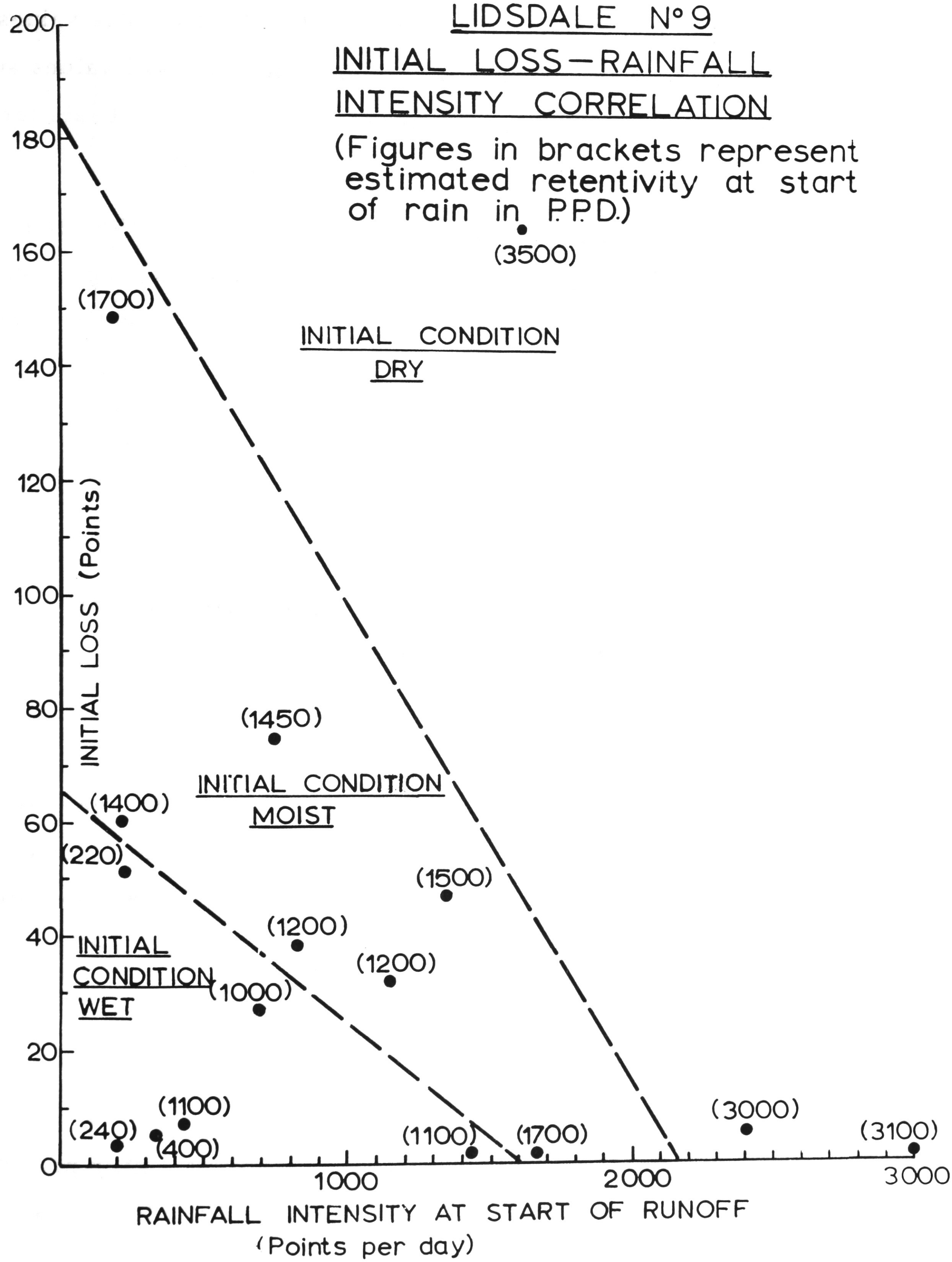

LIDSDALE N° 9
INITIAL LOSS—RAINFALL INTENSITY CORRELATION
(Figures in brackets represent estimated retentivity at start of rain in P.P.D.)
INITIAL CONDITION DRY
INITIAL CONDITION MOIST
INITIAL CONDITION WET
INITIAL LOSS (Points)
RAINFALL INTENSITY AT START OF RUNOFF (Points per day)
(3500)
(1700)
(1450)
(1400)
(220)
(1500)
(1200)
(1200)
(1000)
(1100)
(240)
(400)
(1100)
(1700)
(3000)
(3100)
0
20
40
60
80
100
120
140
160
180
200
1000
2000
3000

FIG. 33 b.

The runoff estimated by the complete operation of this model using the idealized retention function is described in 14.10. The values obtained by this procedure will be compared with the observed values and the bias in the early parts of the wetting phases will provide a basis for modifying the idealized function as shown in figures 34 and 35.

In some cases it may be possible to simplify the expression of the departure from the idealized function by introducing a single constant to represent an "effective impervious factor." The basic retention function would then be of the following form :

$$\frac{dR}{dt} = r \text{ when } P \geq r + b$$

$$= \frac{Pr}{r+b} \quad \text{when } P < r + b$$

where b = the effective impervious factor

P = rainfall rate

R = retention storage

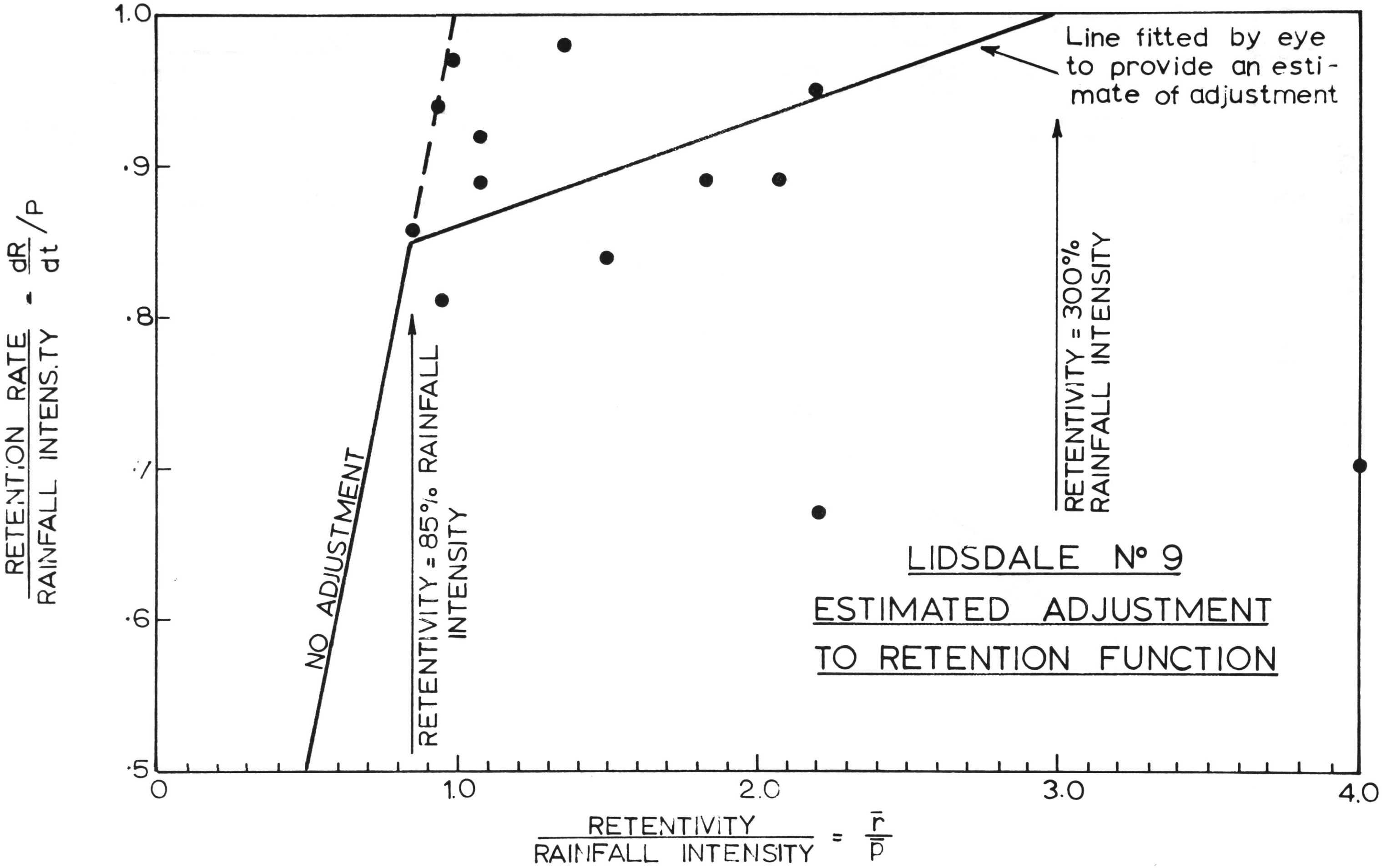
r = retentivity

The work involved in detailed analyses such as the above would probably be unjustified in many circumstances because the departures from the idealized retention function are relatively unimportant when the orders of accuracy of the basic data and total estimates are considered. Nevertheless a theoretical framework has been developed to enable further refinement of this type when required.

14.09 Limited Wetting Periods and Initial Loss

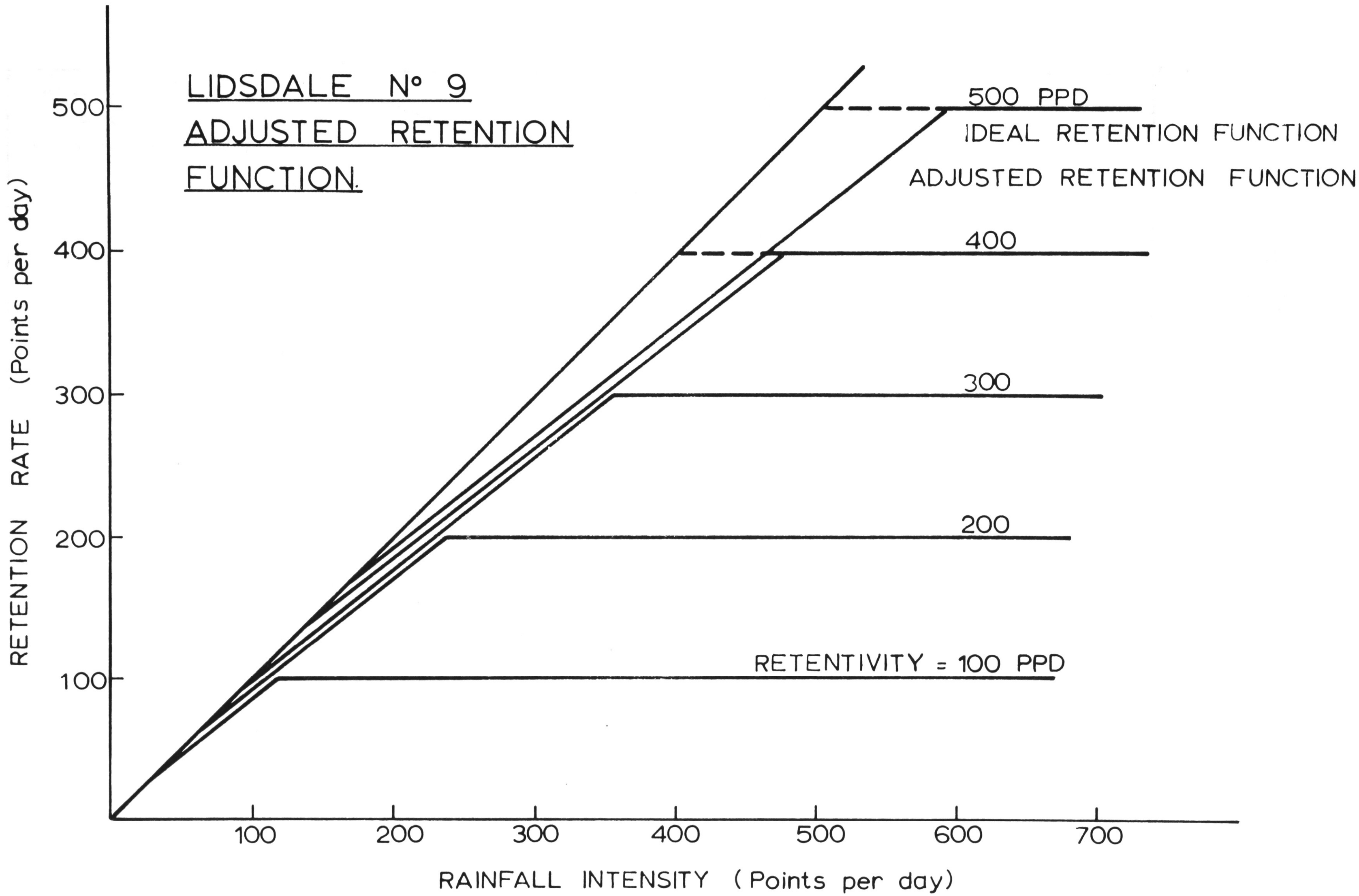
With the idealized retention function, limited wetting periods at the commencement of the wetting phase correspond with initial loss.

Initial loss has been a very useful concept in a number of recent studies (refs.107, 164 and 194) and it is an important component of some of



1.0
·9
·8
·7
·6
·5
0
1.0
2.0
3.0
4.0
RETENTION RATE / RAINFALL INTENSITY - dR/dt /P
RETENTIVITY / RAINFALL INTENSITY = r̄/p̄
Line fitted by eye to provide an estimate of adjustment
NO ADJUSTMENT
RETENTIVITY = 85% RAINFALL INTENSITY
RETENTIVITY = 300% RAINFALL INTENSITY
LIDSDALE Nº 9
ESTIMATED ADJUSTMENT TO RETENTION FUNCTION

FIG. 34.



LIDSDALE N° 9
ADJUSTED RETENTION FUNCTION.
500 PPD
IDEAL RETENTION FUNCTION
ADJUSTED RETENTION FUNCTION
400
300
200
RETENTIVITY = 100 PPD
RETENTION RATE (Points per day)
100
200
300
400
500
600
700
RAINFALL INTENSITY (Points per day)

FIG. 35

the previously described models (sections 9 and 10.04). The term was apparently first used by Snyder in 1939 (ref.195), but it is not mentioned in most of the current text books. Some authorities seem to regard the idea of initial loss as an unnecessary distortion of infiltration theory (ref.196) while others regard it as a practical application of the theory, (Creager, Justin and Hinds, ref.197).

The models of the C.M.B. (section 9) and Boughton (section 10.04) imply that initial loss is dependent only on the initial moisture status of the watershed and is not significantly affected by rainfall intensity. In most of the other models, however, the time and total loss when runoff commences are determined by both the initial moisture status and rainfall intensity.

The retention accretion function sheds some light on the above anomaly as it provides a visual representation of the physical processes. Significant runoff does not commence until the retentivity falls below the rainfall intensity and this undoubtedly occurs earlier and with less total rain when the intensity is higher. Nevertheless large differences in rainfall intensity and retentivity correspond with small differences in retention storage or loss if the slope of the transient lines is steep, i.e. if a low accretion delay is characteristic of the upper layers of the watershed, (e.g. with porous soil and dense vegetation). Therefore, the assumption that initial loss is independent of rainfall intensity may be approximately correct in such cases.

The transient accretion lines for Lidsdale No.9 are not particularly steep however, and under these conditions rainfall intensity would be expected to have a definite influence on initial loss, (i.e. the initial limited wetting periods).

The correlation of fig.33b tends to support this, although it would be desirable to have a greater range of data for a more rigorous test. The rainfall intensities of the wetting periods immediately after commencement of runoff were adopted in the correlation of fig. 33b but from some points of view it may have been preferable to use the intensities of the period immediately before the commencement of runoff. Closely associated with this is the problem of deciding what bursts of rainfall are significant, i.e. how long should wetting periods be?

In 14.04 a wetting period was defined as a period during which the rainfall could be regarded as essentially uniform in intensity. The obvious approximations and subjectivity involved in selecting such periods made this approach unappealing at first but its compensating advantages (in the analysis of data from small watersheds) finally led to its acceptance as explained in 14.10. Experience with the technique and the general consideration of 13.03 and 13.01 suggest the following rules:

(a) Intensities should not be calculated over periods less than the minimum "response time", i.e. the minimum value of $\frac{ds}{d\sigma}$ as given by the runoff depletion function. This is probably related to the time of overland flow and is .4 hours for Lidsdale No. 9.

(b) While the intensities are consistently lower than the retentivity there is no need to consider small or moderate fluctuations at all.

(c) When the intensities and retentivity tend towards the same order of magnitude changes in intensity of 10% or more calculated over the minimum response time should indicate separate wetting periods.

(d) When the rainfall intensities are consistently higher than the retentivity, changes in intensity of 5% or more calculated over the minimum response time should indicate separate wetting periods.

Although (c) above recommends the use of the minimum response time for calculating rainfall intensities, the studies to date suggest that the streamflow hydrograph is not usually very sensitive to fluctuations of runoff inflow over periods less than the response time plus streamflow delay time. This is evidently because the streamflow storage tends to smooth out such fluctuations and the shape of the resulting hydrograph is determined largely by the total inflow during the period, the time distribution of the inflow having only a minor effect.

Boughton (ref.131) reported larger values of initial loss for larger watersheds and suggested that this was caused by channel transmission losses. There are several other possible contributing factors, viz:

(a) Average rainfall intensities are lower over larger watersheds so that the time when retentivites are exceeded tends to be later.

(b) Early runoff generated in remote parts of the watershed does not become significant flow at the outlet because of attenuation over the long period of travel.

(c) The calculated initial loss depends considerably on the estimated time of commencement of runoff. The latter is usually estimated too late from the streamflow hydrograph because of the failure to allow for longer streamflow delays at low flows. The resulting errors are probably insignificant for small watersheds but may approach several hours for large watersheds with corresponding effects on the calculated initial loss.

Under many Australian conditions retention storage is largely recharged during periods of initial loss and this factor is therefore a very important link between rainfall and the corresponding runoff. The derivation and estimation of initial loss present a number of practical and theoretical difficulties, some of which could probably be clarified by further analysis along the lines indicated.

14.10 Rainfall-Runoff Estimation by the Complete Model

After the various functions have been derived for a particular watershed it is possible to synthesize runoff records from suitable rainfall and evaporation data, using relatively simple (but perhaps tedious) accounting methods.

The procedure used for Lidsdale No. 9 is detailed in tables 5 to 7 and is outlined below:

(a) Divide each wetting phase into periods of approximately constant rainfall intensity as discussed previously (page 140) using pluviograph, synoptic or other rainfall data.

(b) Set out the above wetting periods, together with the intervening drying periods as in columns 1 and 2 of table 7, leaving sufficient space between consecutive periods to allow for later subdivisions into free and limited wetting periods, as necessary.

(c) For each wetting period set out the total rainfall (ΔP) in column 13 and the average rainfall intensity (p) in column 5.

(d) For each drying period obtain an estimate of the average potential evaporation and place in column 5. The calculations for Lidsdale No. 9 are not very sensitive to these estimates and the U.S. pan reading of fig.38 appear to be adequate for all years. In cooler and wetter climates it may be necessary to give more attention to this factor.

(e) Start the accounting from a time when retention storage and retentivity are known, e.g. at the end of a prolonged wet period as on December 10, 1963 when the estimated Retention storage was 440 points (column 10) and the estimated retentivity was 40 points per day (column 7). Also, because of the wet state of the watershed the evaporativity was at its maximum value of 30.0 points per day, as adopted in section 14.07.

(f) The first period is a drying period and its initial conditions (evaporativity and retention storage) are represented by point A in figure 37. Because the potential evaporation of 28 p.p.d. (column 5) is less than the evaporativity it is a limited drying period, as designated by "L D" in column 3.

(g) Evaporation in the first period therefore occurs at the potential rate of 28 p.p.d. until the evaporativity falls to 28 p.p.d. which is represented by point B in fig. 37. The total evaporation ($-\Delta R_V$) during this period is $440 - 436 = 4$ which is placed in column 11, and the change in retentivity is $-\frac{1}{f}\Delta R_V = -9.75 \times 4 = 39$ which is placed in column 9. The duration of the period (ΔT) is given by $\frac{4}{28.0} = 0.14$ days and the commencement of the next period is therefore 1830. It is free drying (FD in column 3) because the evaporation rate is now sufficient to satisfy the evaporativity.

(h) As free drying continues until the next wetting period w_o (commencing 1530 December 12) ΔT is 1.88 days and the corresponding change is represented by line BC in figure 37 C being determined from the orthogonal time lines. R and w_o for 1530 Dec. 12 are therefore 404 and 9.8 respectively (columns 10 and 6 $\Delta R . \Delta r$ and r_o are calculated as before.

(i) The period commencing 1530 Dec.12 is a wetting period and its initial conditions (retentivity and retention storage) are represented by point C in fig.39. Because the rainfall intensity of 340 p.p.d. (column 5) is less than the retentivity of 390 p.p.d. (column 7) it is a limited wetting period, (LW in column 3) and continues as such until point D in figure 39. The ΔR_p represented by CD is 409 - 404 = 5 points and Δr is 390 - 340 = 50, or, alternatively, $\Delta r = \frac{-1}{n}$, $\Delta R_p = 5 \times 10.2 = 51 = 50$ approximately. ΔT is obtained from $\frac{\Delta R}{P} = \frac{5 \times 24}{340} = 0.35$ days and the end of the period is therefore 1550 Dec.12.

(j) The next period is free wetting because the rainfall intensity is now sufficient to satisfy the retentivity and any surplus becomes runoff. The duration of the period is .82 hours (ΔT) which is represented by DE in fig.39, E being determined from the time lines. ΔR_p is given by 418 - 409 = 9 and ΔI (column 14) = $\Delta P - \Delta R = 12 - 9 = 3$.

(k) The recovery of evaporativity during the above period is represented in fig.37 by the lines CD and DE, the points D and E corresponding to the appropriate values of R.

(l) The procedure is repeated as before, a continuous account being made of R_o, r_o and w_o which enable the estimation of ΔI during each free wetting period.

The above assumes an idealized basic retention function and tends to underestimate small increments of runoff. If the adjusted retention function is used (fig.35), true free wetting periods only occur when the retentivity is less than 85% of the rainfall intensity, as shown in fig.34. With this function, however, some runoff also occurs in limited wetting periods when the retentivity is between

85% and 300% of the rainfall intensity. The corresponding estimates for Lidsdale No. 9 are compared with those of the idealized function in table 6. A distinct improvement is apparent for small storms but the refinement may not be warranted for flood studies.

Increments of runoff inflow (Δ I) are converted to discharge hydrographs by the runoff accretion, depletion and streamflow functions. These calculations are demonstrated in table 7, the procedure being as follows:

(1) Complete columns 1, 2, 3 and 4 from the previous calculations of table 9.

(2) Obtain Δ S (column 5) and $\Delta\sigma$ (column 6) from the runoff accretion function (fig. 30).

(3) Obtain S' (column 7) and σ' (column 8) from the runoff depletion function (fig. 25) by commencing with previous values of S_t and σ_t (columns 9 and 10) and proceeding along the time lines the distance representing

(4) Calculate S_t (column 9) by adding S' and Δ S.

(5) Calculate σ_t (column 10) by adding σ' and $\Delta\sigma$

The above procedure breaks down when the calculated values of S_t and σ_t plot to the right of, or below the limiting base curve of the retention depletion function. In these cases the flow may be called "saturated" and is regarded as a single function of the storage represented by the limiting curve. Values of S_t and σ_t are then estimated by trial and error as follows:

(1) Select a point on the saturated flow curve about half way between S_o (=S_t for previous period) and a guessed value of S_t. Multiply the corresponding value of σ by ΔT to give a trial value of $\Delta\Omega$. ($\Delta\Omega$ is the volume of outflow for the period).

(2) Calculate a trial value of Δ S from

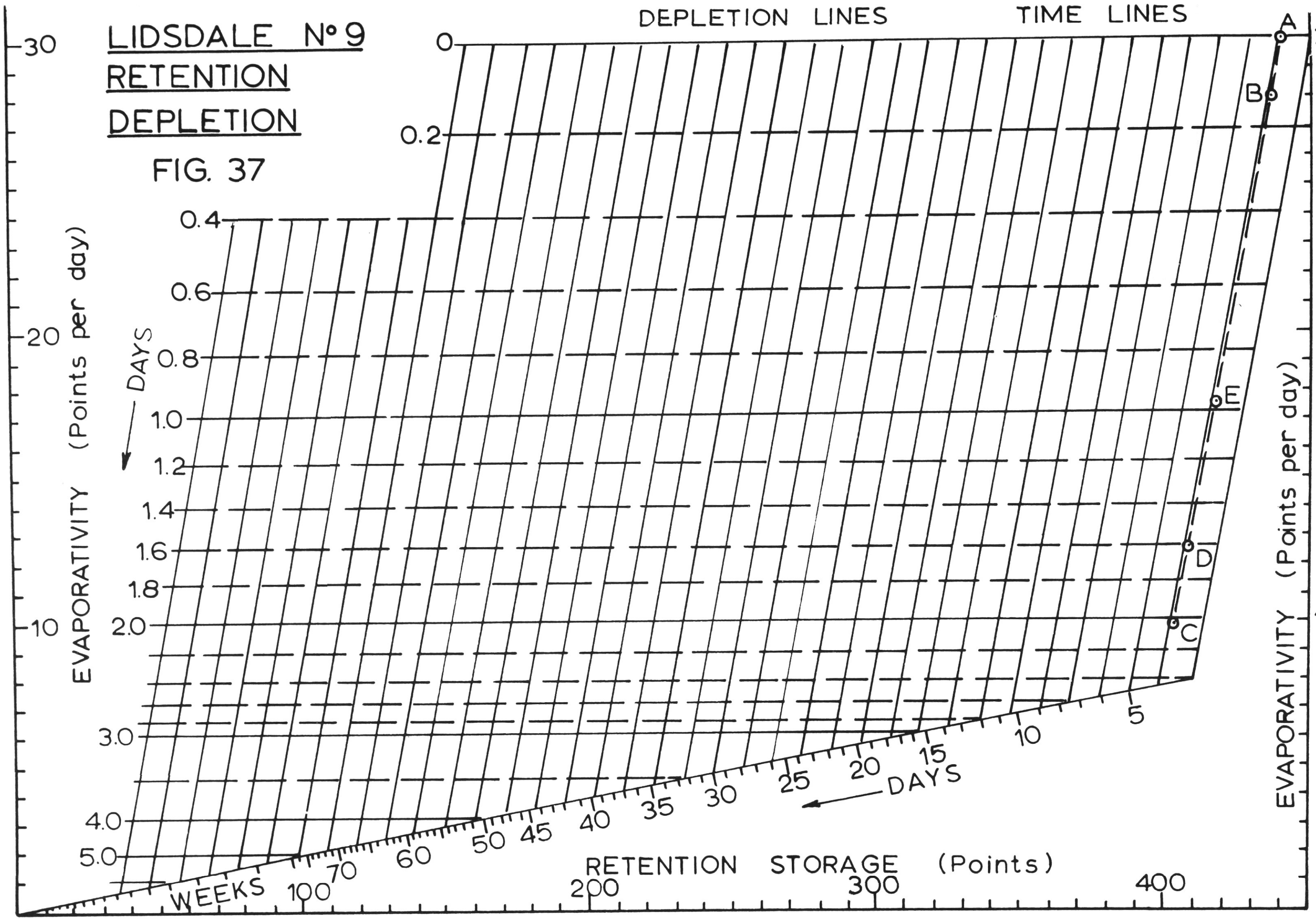
$$\Delta S = \Delta I - \Delta O$$

where ΔO is the trial value of (1) above. A better estimate of S_t is now given by $S_o + \Delta S$.

(3) Select a point on the saturated flow curve half way between S_o and the estimate of S_t from (2). Multiply the corresponding value of σ by ΔT to give the next trial value of ΔO

(4) Repeat the above procedure until stable values of S_t and ΔO are obtained. σ_t is read off the saturated flow curve.

FIG. 36 is not in the original print copy.



LIDSDALE N° 9
RETENTION
DEPLETION
FIG. 37
DEPLETION LINES
TIME LINES
A
B
C
D
E
0
0.2
0.4
0.6
0.8
1.0
1.2
1.4
1.6
1.8
2.0
3.0
4.0
5.0
DAYS
EVAPORATIVITY (Points per day)
30
20
10
5
10
15
20
25
30
35
40
45
50
60
70
100
WEEKS
DAYS
RETENTION STORAGE (Points)
200
300
400

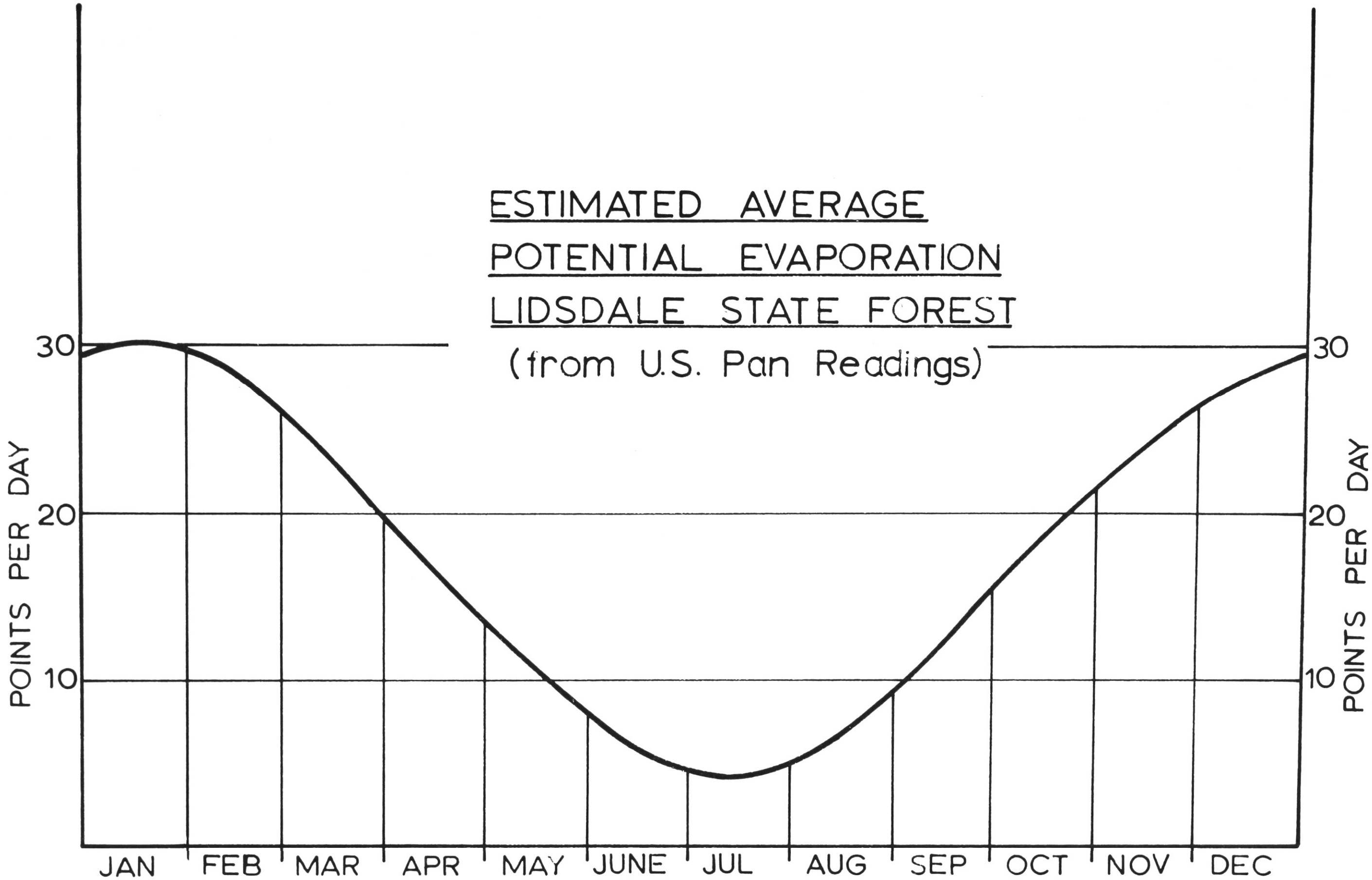

ESTIMATED AVERAGE
POTENTIAL EVAPORATION
LIDSDALE STATE FOREST
(from U.S. Pan Readings)
POINTS PER DAY
30
20
10
JAN
FEB
MAR
APR
MAY
JUNE
JUL
AUG
SEP
OCT
NOV
DEC

FIG. 38

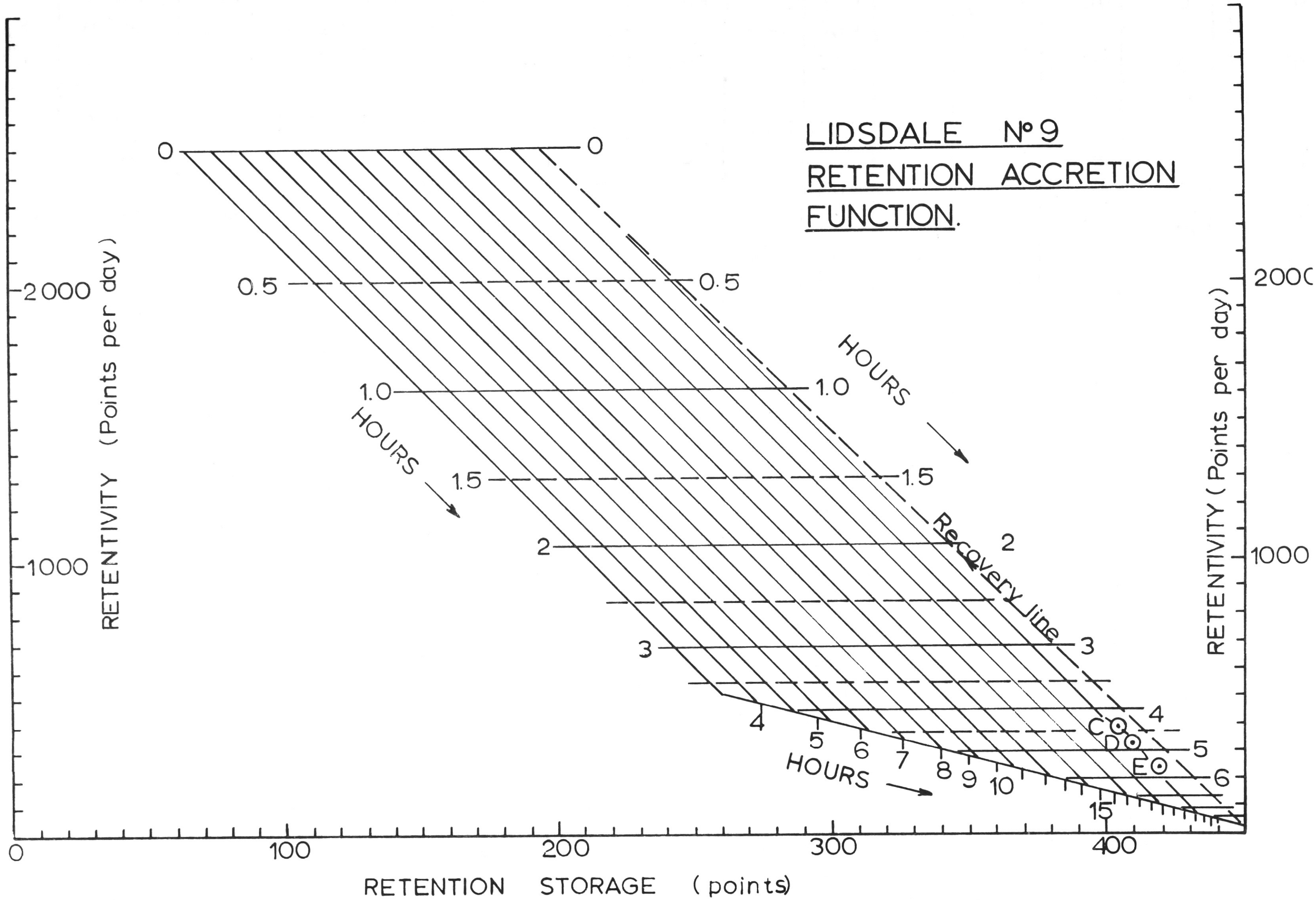

LIDSDALE Nº 9
RETENTION ACCRETION
FUNCTION.
RETENTIVITY (Points per day)
RETENTION STORAGE (points)
HOURS
Recovery line
C
D
E

FIG. 39

TABLE 5 LIDSDALE No. 9 RAINFALL-RUNOFF ACCOUNTING

Column No.	2	3	4	5	6	7	8	9	10	11	12	13	14	15
	Start of Period	L or F	ΔT hrs/ days	P/V ppd.	W_a ppd.	r_o ppd.	Δr / ~~R~~ ppd. -	+	R_o ~~Ro~~ pts.	ΔR pts. -	+	ΔP / ~~P~~ pts.	Calc. ΔI pts.	Obs. ΔI pts.
1963														
Dec. 10	1500	LD	0.14	28	30.0	40		39	440	4				
	1830	FD	1.88	28	28.0	79		311	436	32				
12	1530	LW	0.35	340	9.8	390	50		404		5	5	0)	2
		FW	0.82	340		340	92		409		9	12	3)	
	1640	FD	0.93	28	17.5	248		127	418	13				
13	1500	LW	0.83	202	10.5	375	71		405		7	7	0	0
	1550	FD	3.02	28	14.5	304		266	412	32				
16	1620	FW	1.33	2410	7.3	570	255		380		25	134	109	99
	1740	FD	0.12	28	21.2	315		29	405	3				
	2030	FW	0.40	3690	20.0	344	61		402		6	66	60	59
	2050	FD	9.20	28	23.0	283		859	408	88				
26	0330	LW	0.40	600	6.1	1142	102		320		10	10	0	0
	0350	FD	1.40	29	11.5	1040		117	330	12				
27	1350	FW	0.4	1440	6.1	1157	214		318		21	24	3	*5
	1410	LW	1.0	240	17.5	943	82		339		8	8		
	1510	FD	1.6	29	22.0	861		224	347	23				
29	0540	LW	1.8	622	9.0	1085	463		324		46	46	0)	2
		FW	0.2	622		622	51		370		5	6	1)	
	0740	LW	1.3	148		571	78		375		8	8	0	0
	0900	FD	14.3	30	30.0	493		1015	383	104				
1964														
Jan. 12	1800	LW	8.0	48	5.1	1508	163		279		16	16	0	0
13	0200	FD	0.5	30	13.5	1345		58	295	6				
	1500	LW	3.3	131	10.3	1403	204		289		20	20	0	0
	1820	FW	0.8	1350		1199	356		309		35	43	8)	4
	1910	LW	6.9	90		843	265		344		26	26	0)	
14	0200	FD	10.8		30.0	578			370					
						538	2507	3045	70	317	247	431	184	171

TABLE 5 LIDSDALE No. 9 - RAINFALL-RUNOFF ACCOUNTING (CONTD.)

Column No.	2	3	4	5	6	7	8	9	10	11	12	13	14	15
				See	Page 147									
1964														
Jan. 14	0200	FD	10.8		30.0	578		838	370	86				
24	2120	LW	1.2	600	5.5	1416	305		284		30	30	0	0
	2230	FD	13.9	28	21.4	1111		847	314	87				
Feb. 7	1950	LW	0.7	690	4.4	1958	204		227		20	20	0	0
	2030	FD	7.8	29	15.0	1754		410	247	42				
15	1700	LW	1.0	240	4.0	2164	102		205		10	10	0	0
	1800	FD	13.3	28	9.0	2062		517	215	53				
29	0100	LW	2.5		3.1	2579	183		162		18	18	0	0
	0300	FD	8.0	27	13.0	2396		341	180	35				
Mar. 8	0430	LW	5.0	235	2.7	2737	520		145		51	51	0	0
	0930	LD	0.5	23	30.0	2217		117	196	12				
	2130	FD	30.7	23	23.0	2334		995	184	102				
Apr. 8	1430	LW	4.5	288	1.6	3329	550		82		54	54	0	0
	1900	LD	1.0	19	30.0	2779		195	136	20				
9	1900	FD	10.9		19.0	2974		400	116	41				
20	0500	LW	0.4	1680		3374	285		75		28	28	0	0
	0520	FD	1.2			3089		165	103	17				
21	0830	LW	15.5	188		3254	1220		86		120	120	0	0
	2400	LW	3.0	200		2034	255		206		25	25	0	0
22	0300	LW	0.2	1620		1779	159		231		16	16	0)	2
	0310	FW	0.6			1620	367		247		36	37	1)	
	0350	LD	0.1	17		1253		19	283	2				
	0640	LW	0.5	1030		1272	242		281		24	24	0)	3
	0710	FW	0.2	1030		1030	71		305		7	8	1)	
	0720	LW	1.0	528		959	224		312		22	22		
	0820	LD	0.3	17	30.0	735		58	334	6				
	1510	LW	5.3			793			328					
						215	4687	4902	42	503	461	463	2	5
							215			42		461		

(Contd.)

TABLE 5 LIDSDALE No. 9 RAINFALL-RUNOFF ACCOUNTING (Contd.)

Column No.	2	3	4	5	6	7	8	9	10	11	12	13	14	15
					See Page 147									
1964														
Apr. 22	1510	LW	5.3	236		793	530		328		52	52	0	3
	2030	LD	1.5	17	30.0	263		244	380	25				
Apr. 24	0800	FD	1.8	17	17.0	507		175	355	18				
26	0200	LW	1.0	288		682	122		337		12	12	0	0
	0300	FD	7.0	16	13.1	560		410	349	42				
May 3	0410	LW	1.5			970	102		307		10	10		
	0540	LW	0.6	690		868	175		317		17	17	0)	
	0610	FW	0.3	690		693	72		334		7	9	2)	3
	0630	LW	4.1	187		621	372		341		36	36	0)	
	1040	LD	2.3	13	30.0	249		295	377	30				
5	2000	FD	6.3	13	13.0	544		395	347	40				
11	2310	LW	1.0	456	5.9	939	195		307		19	19	0	0
12	0010	LD	0.7	12	16.0	744		78	326	8				
		FD	13.8	12	12.0	822		720	318	74				
26	1300	LW	5.0	96	4.7	1542	195		244		20	20		
	1800	LD	1.1	10	15.7	1347		107	264	11				
30	0300	LW	7.0	223	10.0	1454	662		253		65	65	0	1
	1000	LD	4.3	9	30.0	792		380	318	39				
		FD	3.3	9	9.0	1172		185	279	19				
Jun. 7	0000	LW			5.0	1357	133		260		13	13	0	0
	1500	LD	0.8	9	12.0	1224		68	273	7				
8	1200	LW	20.8	62		1292	1050		266		103	103		
9	0840	LW	1.7	70		242	42		369		5	5	0	T
	1020	FW	9.7	200		200	125		374		50	81	31)	
	2000	FW	5.1	625		75	35		424		15	133	118)	156
10	0110	FW	5.8			40	20		439		8	23	15)	
	0600	LD	0.1	7		20	0		447	1		0	0	0
	0940	FW				20			446					
						773	3830	3057	118	316	434	598	166	163
							773			118		434		

TABLE 5 LIDSDALE No. 9 RAINFALL-RUNOFF ACCOUNTING (Contd.)

Column No.	2	3	4	5	6	7	8	9	10	11	12	13	14	15
					See Page 147									
1964														
Jun. 10	0940	FW	3.8			20	5		446		2	77	75)	90
	1330	FW	3.5			15	3		448		2	29	26)	
	1700	LD	1.2	7		12			450	1				
	1810	FW	7.0						449	3	4	43	39)	
11	0110	FW	2.3			12			450	1	1	64	63)	208
	0330	FW	2.2			12			450	1	1	47	46)	
	0540	FW	2.2			12			450	1	1	18	17)	
	0800	FW	1.7			12			450	1	1	55	54)	
	0940	LD	0.8	7		12			450	1				
	1030	FW	1.5			12			449	0	1	40	39)	51
	1200	FW	1.3			12			450	1	1	14	13)	
	1320	FW	1.9			12			450	1	1	5	4)	93
	1510	FW	0.5			12			450		0	16	16)	
	1540	FW	2.0			12			450	1	1	10	9)	
	1740	FW	2.2			12			450	1	1	59	58)	
	1950	LD	4.7	7		12			450	1		0	0	0
12	0030	FW	2.9			12			449	1	2	89	87)	93
	0320	FW	5.9			12			450	3	3	12	9)	
	0910	FW	1.1			12			450	1	1	18	17	18
	1020	LD	6.8	7	30.0	12		466	450	48				
19	0430	FW	1.1	480	7.7	478	183		402		18	22	4	6
	0540	LD	1.7	6	17.8	295		107	420	11				
20	2240	FW	0.8	480	11.5	402	112		409		11	16	5	8
	2330	LD	2.1	6	17.8	290		117	420	12				
23	0100	LW	2.5	78	11.2	407	82		408		8	8	0	1
	0330	LD	7.2	5	15.5	325		351	416	36				
30	0900					676			380					
						656	385	1041	66	127	61	642	581	568
							656			66		61		

TABLE 6.

SUMMARY OF ACCOUNTING LIDSDALE No. 9

Column No.	2	3	4	5	6
Date	OBS. ΔI	Calc. ΔI		Acc. Q	
		Ideal Rf	Adj. Rf.	Obs.	Calc.
	Points	Points	Points	Points	Points
12.12.63	2	3	3	2	3
16.12.63	99	109	109	101	112
16.12.63	59	60	60	160	172
27.12.63	5	3	3	165	175
29.12.63	2	1	2	167	177
13. 1.64	4	8	7	171	184
22. 4.64	2	1	2	173	186
22. 4.64	3	1	2	176	188
22. 4.64	3	0	2	179	190
3. 5.64	3	2	3	182	193
9. 6.64	T	0	0	182	193
9. 6.64	156	164	164	338	357
10. 6.64	90	101	101	428	458
11. 6.64	208	219	219	636	677
11. 6.64	51	52	52	687	729
11. 6.64	93	87	87	780	816
12. 6.64	93	96	96	873	912
12. 6.64	18	17	17	891	929
19. 6.64	6	4	4	897	933
20. 6.64	8	5	5	905	938
23. 6.64	1	0	1	906	939
2. 7.64	33	31	31	939	970
3. 7.64	3	0	0	942	970

(Contd).

TABLE 6 (Contd.)
SUMMARY OF ACCOUNTING LIDSDALE No. 9

Column No.	2	3	4	5	6
Date	OBS. ΔI	Calc. ΔI Ideal Rf.	Adj. Rf.	Acc. Q Obs.	Calc.
	Points	Points	Points	Points	Points
12. 7.64	34	41	41	976	1011
13. 7.64	64	56	56	1040	1067
16. 7.64	7	7	7	1047	1074
16. 7.64	36	36	36	1083	1110
19. 7.64	3	4	4	1086	1114
20. 7.64	18	25	25	1104	1139
23. 7.64	1	0	1	1105	1140
24. 8.64	26	19	19	1131	1159
24. 8.64	12	9	9	1143	1168
11. 9.64	6	4	4	1149	1172
13. 9.64	1	0	0	1150	1172
16. 9.64	2	0	0	1152	1172
27. 9.64	1	2	2	1153	1174
28. 9.64	2	0	2	1155	1176
28. 9.64	5	6	5	1160	1181
11.10.64	8	4	4	1168	1185
16.10.64	3	9	9	1171	1194
23.10.64	2	0	2	1173	1196
30.10.64	10	7	6	1183	1202
30.10.64	3	0	3	1186	1205
31.10.64	1	0	1	1187	1206
3.11.64	140	120	120	1327	1326
12.11.64	2	4	4	1329	1330

OBS. ΔI = Observed runoff inflow increment
CALC. ΔI = Calculated runoff inflow increment

IDEAL Rf. = Inflow increment calculated by idealized retention function (fig. 22)

ADJ. Rf. = Inflow increment calculated by adjusted retention function (fig. 35)

ACC. Q. = Accumulated runoff from 12.12.63

T = Trace (less than .005" rain.)

TABLE 7.

STORAGE OUTFLOW COMPUTATIONS LIDSDALE No. 9

1	2	3	4	5	6	7	8	9	10	11	12	13
Date 1964	Time	ΔT hrs.	Δ I points	Δ S points	$\Delta\sigma$ ppd.	S' points	σ' ppd.	S_t points	σ_t ppd	Estimated ΔO (points) Trial 1	Trial 2	Adopted
June 9	0840	1.7	0	0	0	0	0	0	0			
	1020	5.0	16	12	35	0	0	12	35			
	1520	4.7	15	12	32	8	3	20	35			
	2000	5.1	118	60	Saturated			80	600	54	75	58
10	0110	5.8	15	30		"		50	100	38	48	45
	0600	3.7	0			"		41	44			
	0940	3.8	75	34		"		75	490	57	41	41
	1330	3.5	20	-17		"		58	180	30	44	43
	1700	1.2	0			"		50	95			
	1810	5.0	28	5		"		55	140	21	25	23
	2310	2.0	11	0		"		55	140	12	11	11
11	0110	2.3	63	31		"		86	700	38	35	32
	0330	2.2	46	-10		"		76	510	55	57	56
	0540	2.2	17	-16		"		60	210	26	33	33
	0800	1.7	54	24		"		84	680	36	30	30
	0940	0.8	0	0		"		73	450			
	1030	1.5	39	6		"		79	580	28	32	33
	1200	1.3	13	-13		"		66	310	22	25	26
	1320	1.9	4	-13		"		53	120	17		17
	1510	0.5	16	12		"		65	290	4		4
	1540	2.0	9	-10		"		55	140	21	19	19
	1740	2.2	58	27		"		82	650	36	32	31
	1950	4.7	0	0		"		45	60			
12	0030	2.9	87	44		"		89	750	38	43	43
	0320	5.9	9	-46		"		43	50	50	53	55
	0910	1.1	17	12		"		55	145	6	5	5
	1020											

ΔT, ΔI, ΔS, $\Delta\sigma$, ΔO = increments of time, runoff inflow, runoff storage, storage outflow and total outflow volume respectively

S' and σ' = storage and outflow at end of period without inflow increment

S_t and σ_t = " " " " with " "

When saturated runoff applies the flow is a single function of S_t and estimated by trial.

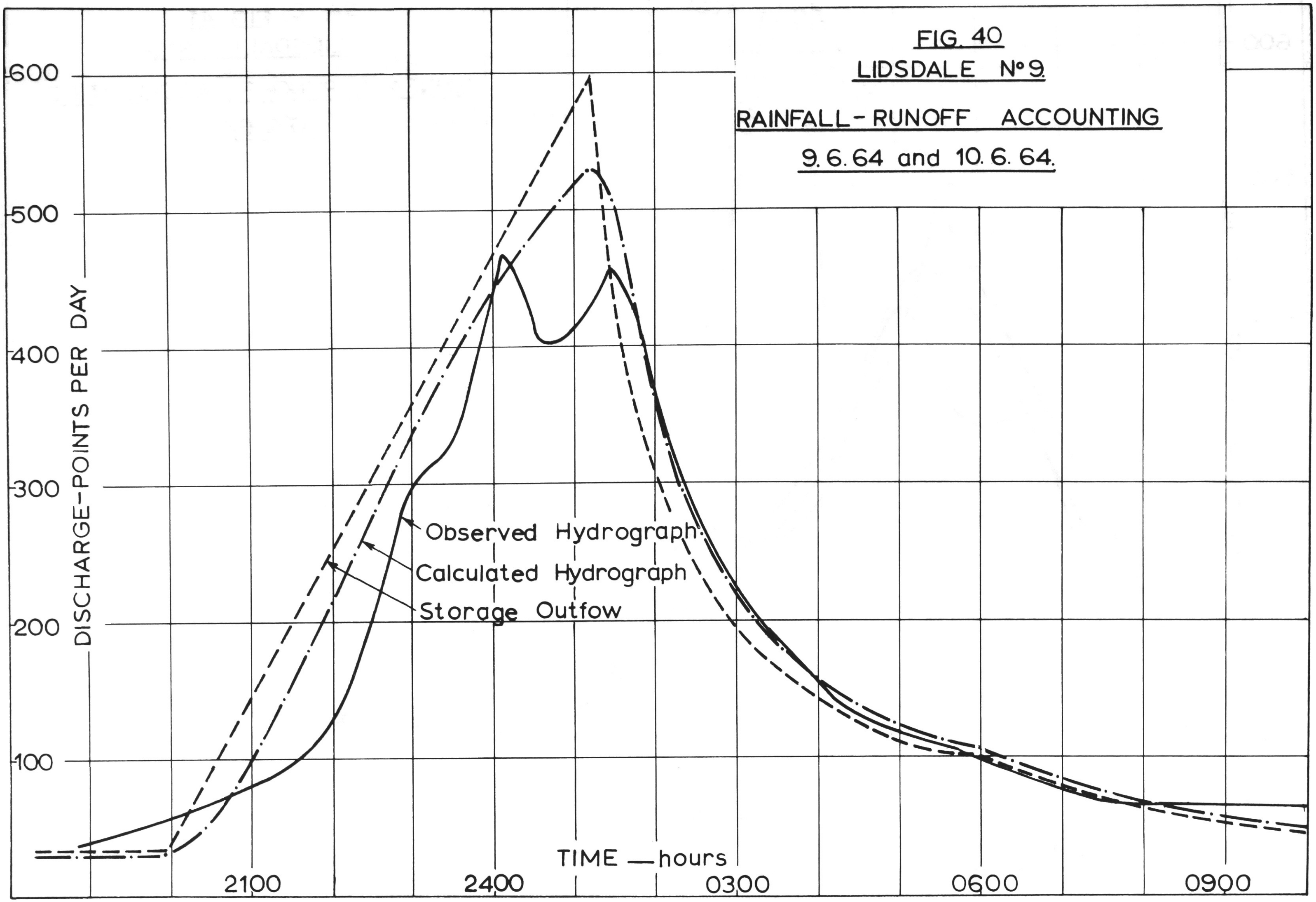


FIG. 40
LIDSDALE No 9.
RAINFALL - RUNOFF ACCOUNTING
9.6.64 and 10.6.64.
DISCHARGE-POINTS PER DAY
600
500
400
300
200
100
Observed Hydrograph
Calculated Hydrograph
Storage Outfow
TIME — hours
2100
2400
0300
0600
0900

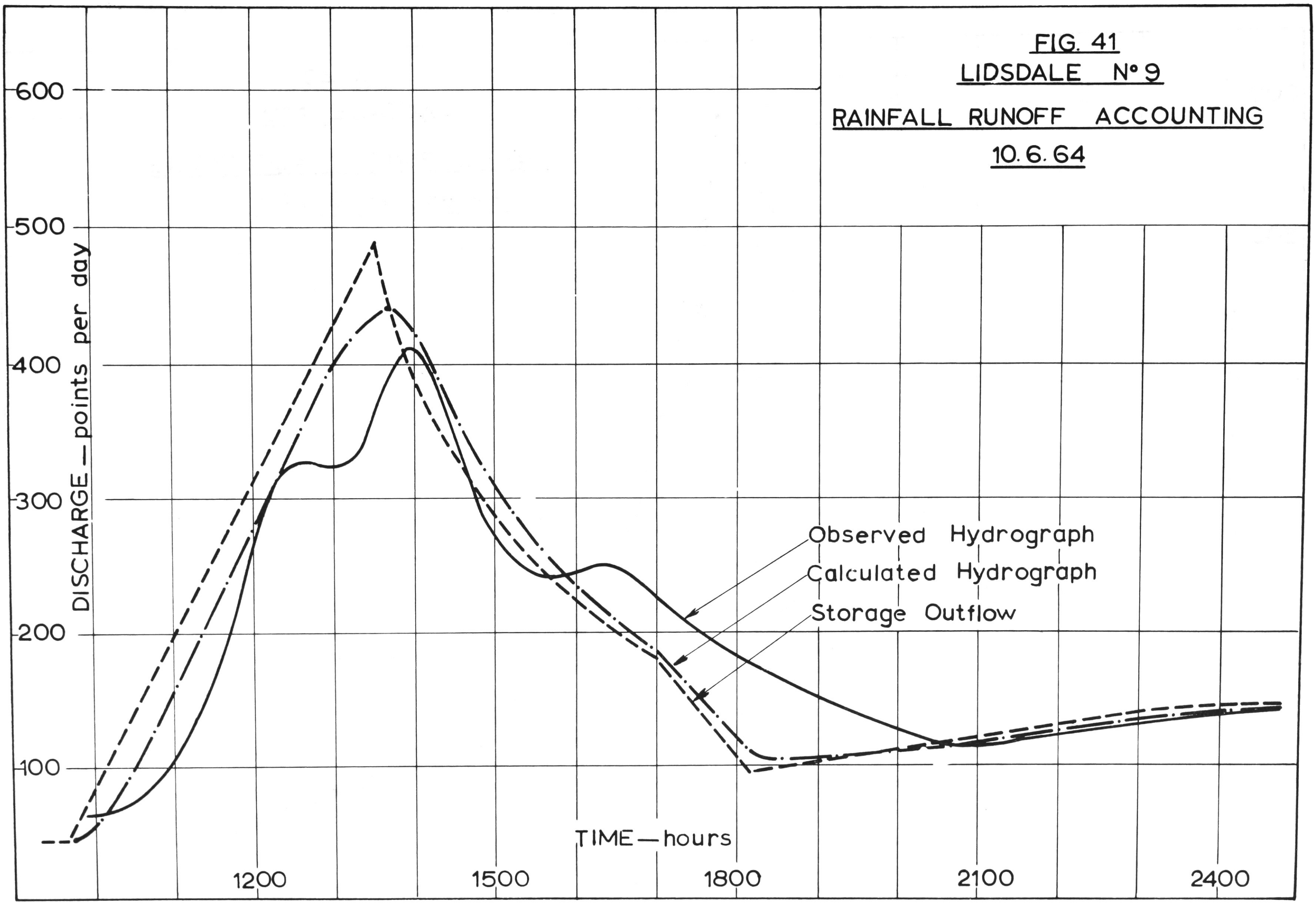


FIG. 41
LIDSDALE N° 9
RAINFALL RUNOFF ACCOUNTING
10. 6. 64
600
500
400
300
200
100
DISCHARGE — points per day
TIME — hours
1200
1500
1800
2100
2400
Observed Hydrograph
Calculated Hydrograph
Storage Outflow

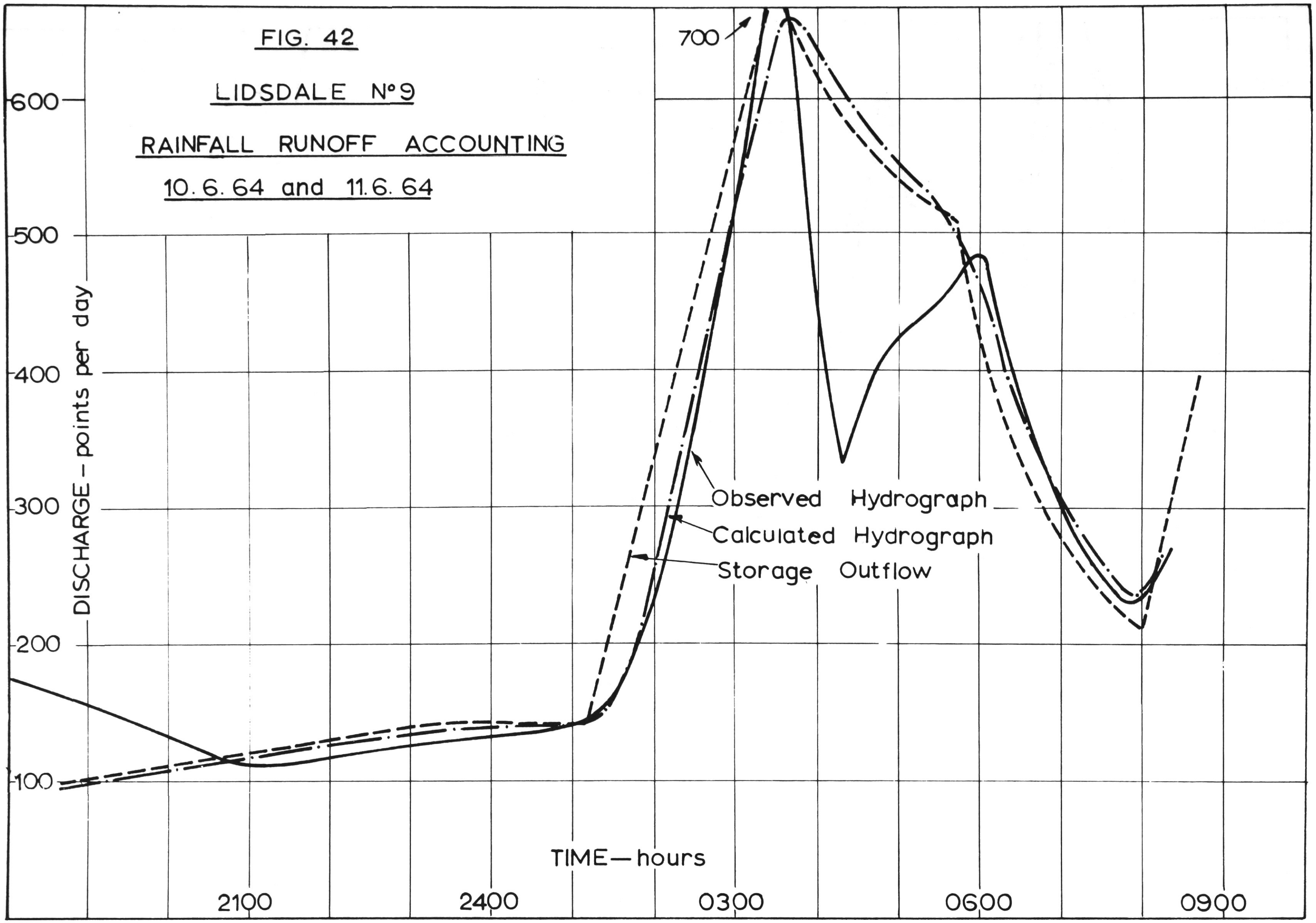


FIG. 42
LIDSDALE No 9
RAINFALL RUNOFF ACCOUNTING
10.6.64 and 11.6.64
700
600
500
400
300
200
100
DISCHARGE – points per day
Observed Hydrograph
Calculated Hydrograph
Storage Outflow
TIME – hours
2100
2400
0300
0600
0900

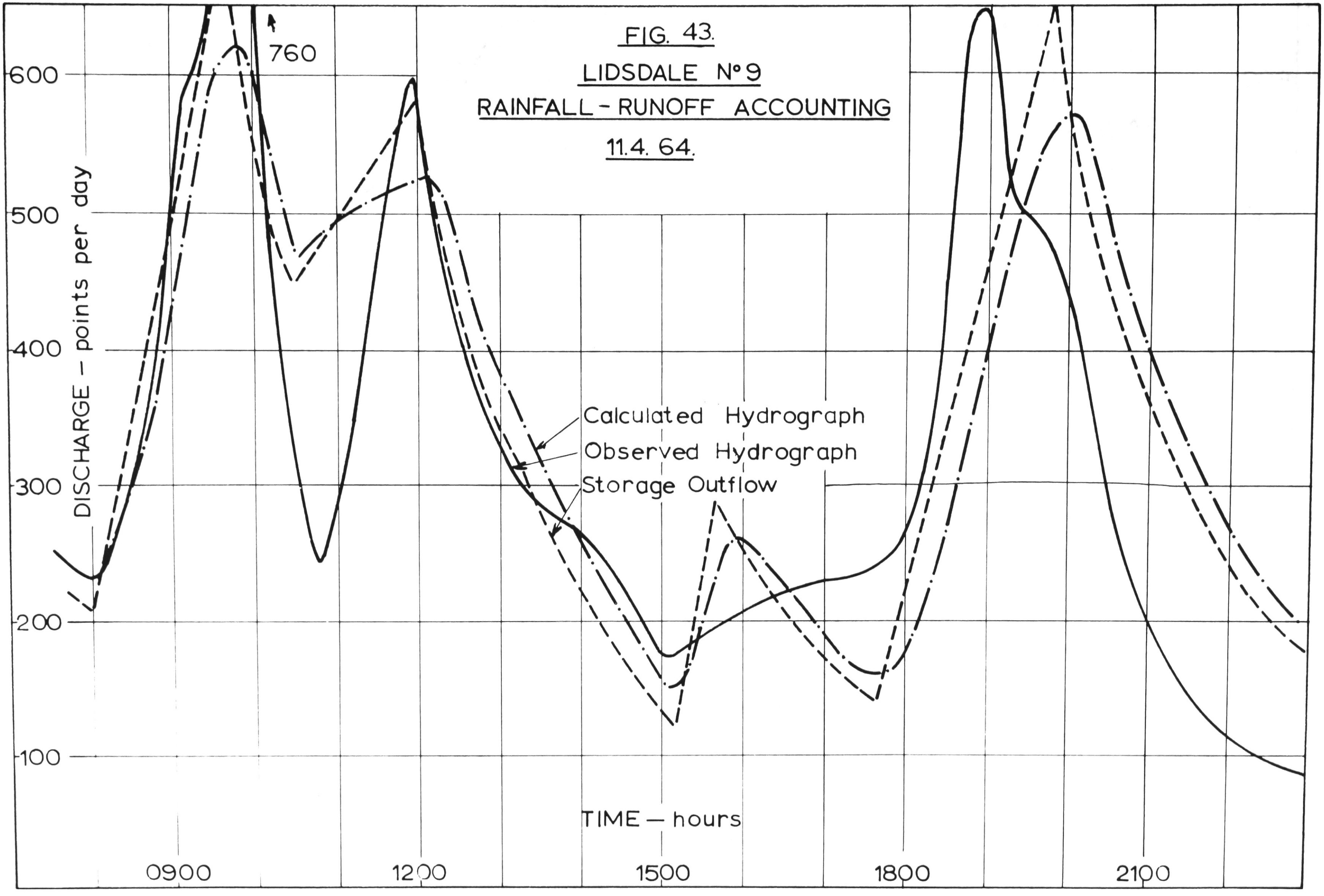


FIG. 43.
LIDSDALE N°9
RAINFALL - RUNOFF ACCOUNTING
11.4. 64.
760
Calculated Hydrograph
Observed Hydrograph
Storage Outflow
DISCHARGE - points per day
600
500
400
300
200
100
TIME - hours
0900
1200
1500
1800
2100

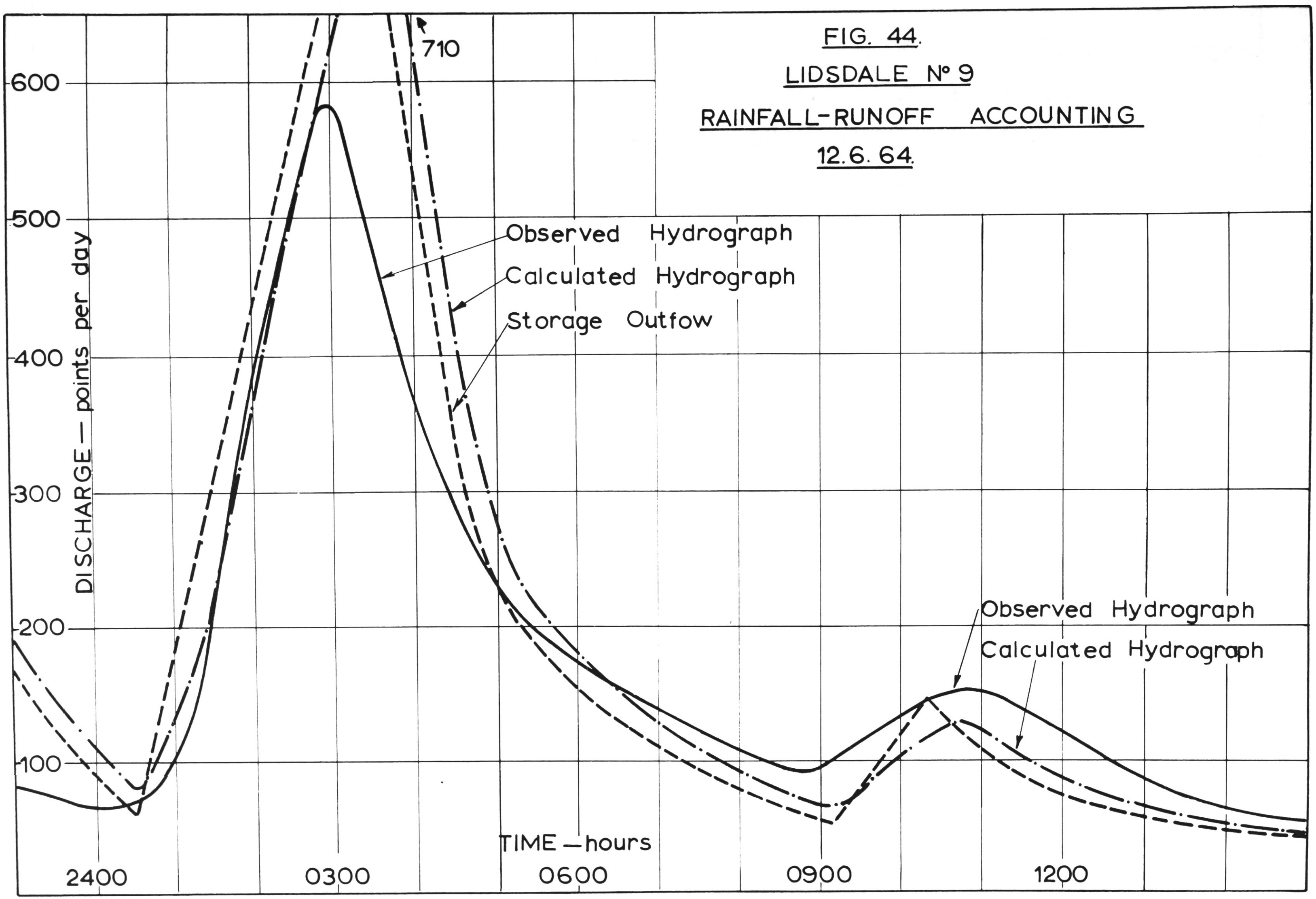
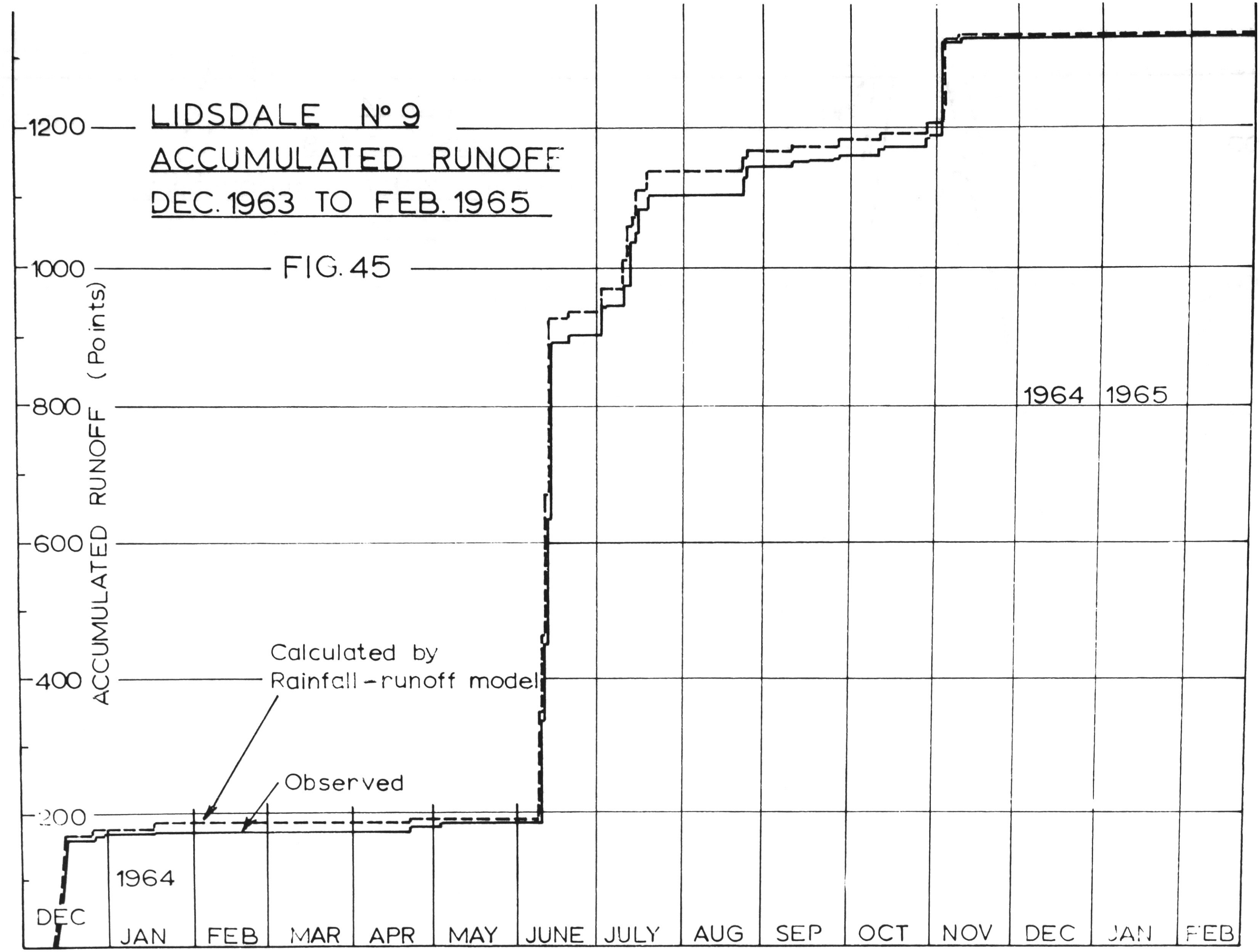


FIG. 44.
LIDSDALE Nº 9
RAINFALL-RUNOFF ACCOUNTING
12.6.64.
710
600
500
400
300
200
100
DISCHARGE — points per day
Observed Hydrograph
Calculated Hydrograph
Storage Outfow
Observed Hydrograph
Calculated Hydrograph
TIME — hours
2400
0300
0600
0900
1200



LIDSDALE Nº 9
ACCUMULATED RUNOFF
DEC. 1963 TO FEB. 1965
FIG. 45
ACCUMULATED RUNOFF (Points)
1200
1000
800
600
400
200
Calculated by
Rainfall-runoff model
Observed
1964
1964 1965
DEC
JAN
FEB
MAR
APR
MAY
JUNE
JULY
AUG
SEP
OCT
NOV
DEC
JAN
FEB

The above gives salient points on the storage outflow hydrograph at the start and end of each period. When Δ I is small, the section of hydrograph between the points may be estimated from the runoff depletion function as it is essentially a depletion curve.

In the case of Lidsdale No. 9 when Δ I was large, the relevant section of the storage outflow hydrograph was assumed to be a straight line between the given points, as shown in figs. 40-44. If required this aspect could be considerably refined by using the runoff accretion function.

Storage outflow hydrographs may be accurately converted to streamflow hydrographs by routing through the streamflow storage. For Lidsdale No. 9, however, this was not considered necessary and a simple lagging procedure was used to reproduce the hydrographs shown in figs. 40-44.

The estimates of runoff given by the complete model appear to be superior to those obtained by more complex models such as Stanford University's, although this may be largely fortuitous. It is realized that further testing is necessary with longer periods of data and different types of watersheds but this is beyond the scope of the present thesis. There are many possible variations of the procedure suggested and some of these would undoubtedly improve both the derivation of the functions and the ultimate accounting. It is desired to focus attention on the versatility and generality of the approach rather than on the details of application.

Some of the success of the example given is due to the small size of the watershed (60 acres) which results in a number of separable hydrographs in each storm and enables variations in retention rates to be analysed during these periods. In larger watersheds only one or two hydrographs may be expected in each storm and it is not possible to derive the retention accretion function in the same way as has been done for Lidsdale No. 9. In such cases the time distribution of runoff inflow must be estimated by reverse routing of the translated hydrograph through the translation and runoff storages using hourly or other fixed time increments. The subsequently derived runoff and retention accretion functions are affected by the approximations inherent in the model to a greater degree than the corresponding functions of Lidsdale No. 9, but in this regard the technique is no worse than any of the others examined earlier.

The complete rainfall-runoff model is readily adapted to fixed time increments which simplify some of the functions and are more suitable for computer simulation. For the small Lidsdale watersheds, however, increments as small as ten minutes are necessary to accurately define the runoff hydrographs and the amount of work in dissecting the data to this extent was found to be prohibitive. Hourly and half hourly increments were tried with Lidsdale No. 9 but the procedure was more tedious and the results less satisfactory than the adopted method with variable increments.

It is claimed that each of the watershed functions may be approximated in terms of one or two constants all of which have recognizable physical significances. It should be possible to relate these constants either empirically or directly to measurable watershed characteristics so that reasonable quantitative assessments are possible even on ungauged streams.

The need for this type of technique in flood studies has led to the development of synthetic unitgraph methods and similar needs now frequently arise in more comprehensive rainfall-runoff studies. The effects of watershed characteristics on hydrologic behaviour, their quantitative expression and their quantitative prediction all form an extensive specialized field of study that may be called "watershed analysis."

Similar sentiments probably prompted the recent advocation of "parametric hydrology" as a specialized sub-branch of Hydrology. This term appears to cover a broader and more diversified range of studies which would include "watershed analysis."

14.11 Introduction to Watershed Analysis

A "simple watershed" is one that may be treated as a single unit and is therefore adequately represented by one retention storage, one runoff storage and one streamflow storage as shown in fig. 23. The technique developed so far has been concerned with simple watersheds, ignoring, in particular, the effects of areal variations in rainfall. There are two possible methods of dealing with the latter complication:

(a) By using another parameter in the streamflow function to account for distribution of outflow from runoff storage, (e.g. by using shorter streamflow delays for storms near outlet).

(b) By representing the watershed with an appropriate system of storages such as in fig. 45A, and accounting each storage separately.

A watershed that cannot be adequately analysed as a single unit may be referred to as a "complex watershed." A different set of functions is not necessarily used for each storage in a complex watershed, indeed, if only rainfall-runoff data is available there would generally be little option but to assume the same function parameters for all storages, with the exception of the streamflow functions. The latter could probably be estimated from channel slopes in a similar way to the storage delay times of Laurenson's model (ref.154).

The possibilities of estimating other function parameters from watershed characteristics seem reasonable but will only be briefly mentioned in this report.

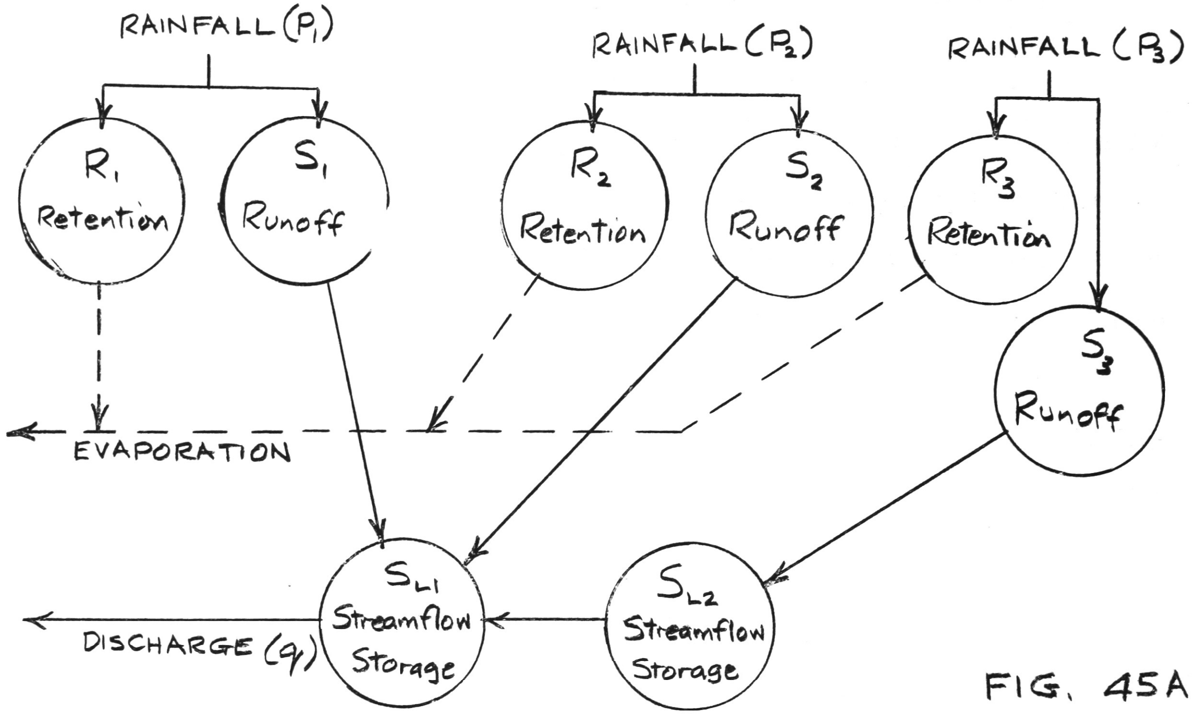


FIG. 45A

The runoff depletion function is approximately specified by three constants representing:

(a) Transient depletion delay

(b) Base depletion delay

(c) Maximum base storage

It should be possible to empirically relate (a) to surface slopes and vegetation if a wide range of suitable measurements and observations were made. Overland flow formulae might also provide satisfactory estimates if suitable adjustments can be derived (ref.1).

Some recent hydrological studies have suggested that (b) and (c) above may be empirically related to geological and channel characteristics, (ref. 5).

The other function parameters and their associated physical characteristics are as follows:

Runoff accretion delay	....	soils and geology
Base retention delays	....	lower soil and vegetation
Transient retention delays	....	upper soil and vegetation
Maximum retention storage	....	soil depth and type channel characteristics foliage area
Maximum evaporativity	....	maximum potential evaporation upper soil and vegetation
Retentivity recovery time	....	vegetation - soil complex
Evaporativity recovery time	....	vegetation - soil complex

Watersheds have a resemblance to the individual specimens of other natural sciences. Each is highly complex and essentially unique; yet each has important features that enable its classification with other specimens and make its behaviour predictable to a certain extent.

Watersheds are hydrologically characterized by their rainfall-runoff functions which also provide a quantitative basis for their classification and prediction.

The field of hydrogeography (see 3.01) includes the classification of watersheds and, in this regard, it seems reasonable to assume that a single set of function parameters applies to a complete hydrologic unit. Similar units would also be expected to have similar function parameters.

Watershed analysis is thus viewed as the detailed study of individual watersheds in order to isolate their particular hydrologic characteristics and to concisely express the effects of these characteristics on rainfall-runoff behaviour. It is therefore complementary to hydrogeography which attends to the general classification of hydrologic characteristics so that analyses for one watershed may be extrapolated to others.

15. CONCLUSIONS

15.01 Estimates of Runoff Statistics

Single, brief records give inaccurate estimates of means and quite unreliable estimates of extremes. Improved estimates are possible however, if the single record can be appropriately supplemented with other information, e.g. the regional period bias and regional frequency ratios. This involves the areal extrapolation of data which appears to be largely neglected as a systematic field of study. The designation "Hydrogeography" is suggested to facilitate the recognition of its scope and necessary outlook.

15.02 Complete Rainfall-Runoff Models

A critical examination of the recognized current rainfall-runoff models suggested that each has some of the following deficiencies:

(a) Artificial components not consistent with the physical processes they are intended to simulate.

(b) Relationships expressed by specific mathematical forms that obscure the degree of approximation involved and restrict the possible range of natural conditions to be represented.

(c) Parameters that are difficult to relate to measurable physical characteristics of the watershed.

(d) Considerable complexity resulting in the dependence on high-speed computers for the general operation and evaluation of parameters.

One factor causing unnecessary complexities in the models is that they endeavour to combine fragmentary concepts used for relatively narrow studies without reference to the rainfall-runoff cycle as a whole. The greatest ultimate improvements in these methods therefore depend on the evolution of an integrated theoretical structure and one possible approach to this has been suggested through a "retention theory."

15.03 An Improved Rainfall-Runoff Model

An improved rainfall-runoff model has been developed from retention theory, with the following advantages:

(a) Each component fits naturally into the total complex but each may also be analysed separately in as much detail as the available data allows.

(b) Relatively simple mathematical or graphical expressions that do not give a false sense of precision.

(c) Analagous (but logical) treatments of several components which simplifies the computations to such an extent that a high speed computer is not essential in many cases.

(c) Parameters that may be derived from a minimum amount of selected data and are therefore not dependent on a long, continuous record.

(e) A recognizable physical significance in all parameters suggesting possible future applications to ungauged watersheds.

The model has been tested with brief data from a 60 acre watershed and the reproductions of runoff agree very closely with the recorded values. The general approach could be adapted to most conditions but further testing is necessary with longer data and a variety of watersheds.

REFERENCES

A. General References

1. LINSLEY, R.K., M.A. Kohler, J.L.H. Paulhus 1949 Applied Hydrology McGraw-Hill, New York.
2. LINSLEY, R.K., M.A. Kohler and J.L.H. Paulhus, 1958 Hydrology for Engineers. McGraw-Hill, New York.
3. WISLER, C.O. and Brater, E.F. 1959 Hydrology. Wiley, New York.
4. JOHNSTONE, D and W.P. Cross. Elements of Applied Hydrology. The Ronald Press Co. New York, 1949.
5. AYERS, H.D. A Survey of Watershed Yield U.N.S.W. Water Research Laboratory Report No. 63. March 1962.
6. FOSTER, E.E. Rainfall and Runoff. McMillan. New York, 1949.
7. AUSTRALIAN ACADEMY OF SCIENCE. Proceedings of Symposium on Water Resources, Use and Management, Canberra 1963 Melbourne University Press.
8. GRADUATE HYDROLOGY NOTES. University of N.S.W. 1955-56.

B. REFERENCES FOR STATISTICAL ESTIMATES

9. CHOW, V.T. University of Illinois Engineering Experimental Station Bulletin 414.

10. CHOW, V.T. The Log Probability Law and its Engineering Applications. Proc.Amer.Soc. Civ.Eng. Vol.80 No.536.1954.

11. ALEXANDER, G.N. Flood Flow Estimation, Probability and the Return Period. Jour. Instn.Eng. Aust. Vol.29 No.10-11, 1957.

12. ALEXANDER, G.N. Inst. Eng., Aust. Civil Eng. Trans. Vol. CE 4 No.1 March, 1962.

13. LANGBEIN, W.B. and HARDISON, C.H. Proc.Amer.Soc.Civ. Eng.81, paper No.826 1955.

14. LANGBEIN, W.B. Discussion of paper by T.G. Chapman in reference 7 page 346.

14a. HALL, L. STANDISH. Trans. A.S.C.E. Vol.84, 1921.

15. WAITT, F.W.F. Jnl. Instn. Eng. Aust., Oct.-Nov.1962.

16. WAITT, F.W.F. Jnl. Instn. Eng. Aust., April-May 1945.

17. CHRISTINOPOULOS, C and LIAKOPOULUS, A. I.A.S.H. Bulletin No.4 Dec.1963

18. BLOKHINOV, E.G. I.A.S.H. Publication No.63 Berkeley 1963.

19. FEDEROV, L.T. I.A.S.H. Bulletin No.20 Dec.1960.

20. FIERING, M.B. A Markov model for low flow Analysis I.A.S.H. Bulletin No. 2 June 1964.

21. GERNEY, J.S. Jnl. Instn. Eng. Aust. March 1962.

22. BEARD, L.R. I.A.S.H. Pubn. No.63. Berkeley 1963.

23. GOULD, B.W. Jnl. Instn. Eng. Aust. Dec.1961.

24. STALL, J.B. and NEILL, J.C. Jnl. Geophys. Res.1961 p.4219.

25. LANGBEIN, W.B. Trans. Amer.Geophys. Union V.30 1949.

26. FISHER, R.A. and TIPPETT, L.H.C. Proc. Cambridge Philosopl. Soc. V.24/

27. GUMBEL, E.J. Statistics of Extremes. Columbia Univ. Press. New York, 1960.

28. GUMBEL, E.J. Statistical Theory of Floods and Droughts. Jnl.Inst. Water Eng. Vol.12. 1958.

29. GUMBEL, E.J. Bulletin. I.A.S.H. April, 1963.

30. GANNON, J.J. I.A.S.H. Pubn. No.63 Berkeley, 1963.

31. BALEK, J. and HOLECEK, J. I.A.S.H. Pubn. No.63 Berkeley, 1963.

32. BERNIER, J. La Prevision Statistique des Bas Debits. I.A.S.H. Pubn. No.63. Berkeley, 1963.

33. KACZMARCK, Z.A. I.A.S.H. Pubn. No.45. 1957 Toronto.

34. CHAPMAN, T.G. Paper in Ref. 7. p.290, 1964.

35. LANGBEIN, W.B. I.A.S.H. Pubn. No.51 Helsinki, 1960.

36. DAVIDSON, D.E. Inst. Eng.Aust. Civil Eng. Trans Sept.1961.

37. DISCUSSION ON REF.36. by Alexander and reply by Davidson. Civil Eng. Trans. March 1962.

38. HERSHFIELD, D.M. Jnl. Geophys. Res. vol.67. 1962.

39. DISCUSSION ON REF. 38 by Alexander Jnl. Geophys. Res. vol.68, 1963.

40. ALEXANDER, G.N. The Use and Abuse of Probability Papers. Hydrology Symposium I.E. Aust. Sydney 1963.

41. KUNKLE, M.K. Jnl. Geophys. Res. v67. No. 4 1962.

42. REINHART, K.G. and ESCHNER, A.R. Trans. Amer. Geophys. Union vol.67 No.6 June 1962.

43. AUSTRALIAN RAINFALL AND RUNOFF: Instn. of Eng. Australia 1958.

44. AMERICAN SOCIETY OF CIVIL ENG. "Hydrology Handbook", 1949.

45. BUREAU OF METEOROLOGY. Bulletin No. 1 Climate and Meteorology of Australia. Revised 1964.

46. DEPT. OF NATIONAL DEVELOPMENT. Review of Australia's Water Resources 1963.

47. GIBBS, W.J. Some Notes on Rainfall in Australia. Commonwealth Bureau of Meteorology. Working Paper 62/163. 1963.

48. U.S.A. WEATHER BUREAU. Technical Paper No.40. Rainfall Frequenc Atlas of the United States 1961.

49. BELL, F.C. Water Research Foundation of Australia. Bulletin No. 8 1964.

50. MC ILLWRAITH, J.F. Jnl. Instn. Eng., Aust. 25: 7-8 July-Aug.1953.

51. HERSHFIELD, D.M. AND KOHLER, M.J. Jnl. Geophys. Res.65 1960 p.1737.

52. REICH, B.M. Jnl. Hydrology 1:1 March 1963.

53. KAROLY, A. Statistical Analysis of Point Rainfall in Victoria and N.S.W. W.R.F.A. Report 1962.

54. Mc CUTCHAN, A.J. Short Cuts in Hydrology Institution of Eng. Aust. Hydrology Symposium, Cooma, N.S.W. 1962.

55. McARTHUR, A.G. and CHENEY, N.P. Water Res. Foundation of Australia, Report No.17 1964.

56. TURC, L. I.A.S.H. Pubn. No.38 Rome 1954.

57. THORNTHWAITE, C.W. Geog. Rev.38: 1948.

58. U.S. GEOLOGICAL WATER SURVEY W.S.P. 846.

59. CORDERY, I. Undergrad. Thesis. University of N.S.W. 1963.

60. SAMRA, S. Estimation of Monthly Runoff from Rainfall and Temperature Records, W.C. & I.C. Report Oct 1960.

61. McDONALD, N.S.W. The Water Balance of Three Catchment Areas in New South Wales. Monograph No. 1 University of New England, 1963.

62. LEWIS, D.C. and BURGY, R.H. Jnl. Hydrology II No. 3 March, 1963.

63. COULTER, G.C. Report No.60 Water Research Laboratory University of New South Wales.

64. REICH, B.M. Jnl. Hydrology. Vol. III Nos. 3/4. November, 1965.

65. WOOD, I.R. Civil Eng. Trans. March, 1959.

66. BINNIE, A. R. Proc. Inst. Civil Engs. (London) Vol. 109, 1892.

67. TURNER, A.K. Jnl. Instn. Eng., Aust. Jan-Feb.1960

68. COOK, H. U.S. Soil Conservation Service Agr. Handbook No.57 1953.

69. Mc DONALD, J.C. and GREEN, C.R. Jnl. Geophys. Res. vol. 65 No.8 1960.

70. HIDORE, J.J. Jnl. Hydrology, Vol.1, No. 4 1963.

71. HAYES, W.J. Rainfall Depths in N.S.W. Undergrad. Thesis University of N.S.W. 1965.

72. MATALAS and LANGBEIN, Jnl. Geophys. Res. Aug. 1962.

73. Deleted.

74. ALEXANDER, G.N. Inst. Eng. Aust. Hydrologic Studies and Investigations, 1955.

75. URBAN, J. I.A.S.H. Pubn. No.63 Berkeley, 1963.

76. WHITMORE, J.S. Jnl. Hydrology, vol.1 No. 2, 1963.

77. O'MAHONY, G. Time Series Analysis of Some Australian Rainfall Data. Commonwealth Bureau of Meteorology Study No.14 May, 1961.

78. MARVIN, C.F. Supplement 15. Geog. Rev. p.666, 1923.

79. LASZLOFFY, W. I.A.S.H. Pubn. No.63 Berkeley, 1963.

80. LEEPER, G.W. Jnl. Aust. Inst. Agric. Sci. March, 1953.

81. RODIER, J. Irregularite temporelle dans cours d'eau et les facteurs qui la conditionment. I.A.S.H. Pubn. No.63 Berkeley 1963.

82. YEVDJEVICH, V.M. I.A.S.H. Pubn. No.63 Berkeley 1963.

83. DIXEY, F. Jnl. Hydrology, Vol.II No.1 1964.

84. STEWART, J. Jnl. Soil Conservation Service of N.S.W. Oct.1960.

85. BENSON, M.A. I.A.S.H. Bulletin No.19 1960.

86. MOORE, W.G. A Dictionary of Geography. Penguin.

87. HARTSHORNE, R. Perspective on the Nature of Geography. John Murray, London 1960.

88. GENTILLI, J. A Geography of Climate. Univ. of Western Australia Press. 1958.

89. MILLER, A.A. Climatology. Methuen London 1957.

90. CHRISTIAN, C.S., STEWART, G.A. and PERRY, R.A. Australian Geographer Vol. VII No. 6 1960.

91. CHRISTIAN, C.S. Regional Land Surveys. Aust. Inst. Agric.Sci. Jnl. vol.18. Sept.1952.

92. CHRISTIAN, C.S. Discussion in Proc. of 9th Pacific Science Congress, Bangkok. 1960.

93. McCALL, J.E. Streamgauging Network in the U.S.A. Proc. A.S.C.E. Vol. 87 HY 2 Mar. 1961.

94. SOMOV, N.V. I.A.S.H.Pubn. No.63 Berkeley 1963.

95. COMMONWEALTH BUREAU OF METEOROLOGY. Results of Rainfall Observations for N.S.W. (to 1944).

96. N.S.W. PREMIER'S DEPT. A Preliminary Survey of Resources. The Mitchell Region 1953.

C. <u>RAINFALL-RUNOFF MODELS AND PHYSICAL PROCESSES</u>

97. AMOROCHO, J. and HART, W.E. Trans. Amer. Geophys. Union, Vo.45, No. 2 June 1964 pp.307-21.

98. THORNTHWAITE, C.W. and MATHEW, J.R. Publications in Climatology, vol. 8, 1955.

99. LAMOREAUX, W.W. Monthly Weather Review, Vol.90, 1962.

100. PENMAN, H.L. Commonwealth Bureau of Soils, Harpenden, England, Tech. Communication No.53, 1963.

101. KOHLER, M.A., NORDENSON, T.J. and FOX, W.E., U.S.Weather Bureau, Research Paper No.38, 1955.

102. COOK, H.L. Trans. Amer. Geophys. Union, Vo.27 No.5 Oct.1946.

103. HORTON, R.E. Proc. Soil Sc. Soc. Amer., Vol.5, 1940.

104. HORTON, R.E. Trans. Amer. Geophys. Union Vo.18, Pt.II 1937

105. CRAWFORD, N.H. and LINSLEY, R.K. Dept. Div. Eng. Stanford Univ. Tech. Report No.12 July, 1962.

106. CRAWFORD, N.H. and LINSLEY, R.K., I.A.S.H. Pubn.No.63 Berkeley 1963.

107. BODY, D.N. and WATSON, B. Aust. Bureau of Meteorology, Working Paper 61/782 April, 1962.

108. BODY, D.N. Jour. I.E. Aust., Vol.34, No.1-2 Jan-Feb.1962.

109. SHARP, A.L. and Others. Jour Geophys. Res. Vol.65 April 1960.

110. HARRIS, B and Others. Jour. Geophys Res. Vol.66 Oct. 1961.

111. JAMIESON, L.N. and CANTWELL, B.E. Jour. I.E. Aust. Vol.30, No. 6 June 1958.

112. SNYDER, W.M. Jour. Geophys. Res. Vol.67 No.2 1962.

113. CHAPMAN, T.G. Aust. C.S.I.R.O. Div. Land Research and Regional Survey, Canberra, A.C.T. Tech. Memor.61/4, 1961.

114. CHAPMAN, T.G. Civ. Eng. Trans. I.E. Aust., Vol.CE5, Mar.1963.

115. BETSON, R.P. Jour. Geophys. Res. Vol.69 No.8 April 15, 1964.

116. TENNESSEE VALLEY AUTHORITY, OFFICE OF TRIBUTARY AREA DEVELOPMENT. Res. Paper No. 1 Dec.1961.

117. TENNESSEE VALLEY AUTHORITY, OFFICE OF TRIBUTARY AREA DEVELOPMENT. Res. Paper No. 2 Feb.1963.

118. TENNESSEE VALLEY AUTHORITY, OFFICE OF TRIBUTARY AREA DEVELOPMENT. Res.Paper No. 3 June 1963.

119. VEIHMEYER, F.J. and HENDRICKSON, A.H. Trans. Amer. Geophys Union Vol.36. No. 3 June 1955.

120. THORNTHWAITE, C.W. and MATHER, J.R., Publications in Climatology Vol.7 1954.

121. PENMAN, H.L. Jour. Soil Science, Vol.1 1949.

122. VAN BAVEL, C.H.M. Agronomy Jour. Vol.45 No. 4 April 1953.

123. CROSS, R.L. Proc. UNESCO Symposium on Climatology and Microclimatology, Canberra 1958.

124. GARDNER, W.R. Soil Sciences, Vol.89 Feb.1960.

125. PHILIP, J.R. Trans. Third Congress on Irrigation and Drainage, San Francisco, 1-4 May 1957 Vol.3 Question 8.

126. SLATYER, R.L. Netherlands Jour. Agric. Sc. Vol.4 1956.

127. DENMEAD, O.T. and SHAW, R.H. Agronomy Jour. Vol.54, Sept-Oct 1962.

128. PALMER, J.H., TRICKETT, E.S. and LINACRE, E.T. Agric. Meteorology, Vol. 1 No. 4 Oct. 1964.

129. SLAYTER, R.O. and McILROY, I.C. Practical Microclimatology Melbourne, C.S.I.R.O. 1961.

130. VISSER, W.C. Int. Assoc. of Scientific Hydrology, Surface Waters Symposium, General Assembly, Berkeley, 1963 Publn.62.

131. BOUGHTON, W.C. Civ. Eng. Trans. I.E. Aust. Vol. CE8, 1 April 1966.

132. BELL, F.C. Paper presented at 1963 Institution of Engineers, Aust. Hydrology Symposium Sydney June 13, 1963.

133. HORTON, R.E. Monthly Weather Review, Vol.51 1923.

134. STOUT, B.B. and McMAHON, R.J. Jour. Geophy. Res. Vol.66, No.6 June 1961.

135. JOHNSTON, R.D. Australian Jour. Botany Vol.12 No.2 Dec.1964.

136. PENMAN, H.L. and SCHOFIELD, R.K. Jour. Agric.Sc. Vol.31 1941

137. PENMAN, H.L. Proc. Roy.Soc.London, Series A, Vol.193, 1948.

138. PHILIP, J.R. and DE VRIES, D.A. Trans.Amer.Geophys.Union Vol.38 No. 2 April 1957.

139. GARDNER, W.R. and HILLEL, D.I. Jour. Geophys. Res. Vol.67, No.11 Oct.1962.

140. MAKKINK, G.F. Jour. Inst. Water Engineers (London) Vol.11, 1957.

141. WEST, E.S. and PERKMAN, O. Australian Jour. Agric. Res. Vol.4 No. 3, July, 1953.

142. WATSON, K.K. Civ. Eng. Trans. I.E. Aust. Vol.CE8 No April 1966.

143. WATSON, K.K. Jour. of Hydrology, Vol.3, No.1 May, 1965.

144. PHILIP, J.R. Soil Science, Vol.77, Feb.1954.

145. YOUNGS, E.G. Soil Science, Vol.86 No. 4, Oct.1958.

146. HEWLETT, J.D. and HIBBERT, A.R. Jour. Geophys. Res. Vol.68, No.4 Feb.15, 1963.

147. HEWLETT, J.D. and HIBBERT, A.R. Bulletin Int. Assoc.Scientific Hydrology, No. 6 Sept.1961.

148. O'DONNELL, T. Int.Assoc. of Scientific Hydrology, General Assembly, Helsinki 1960 Publicn. No.51.

149. NASH, J.E. Proc. I.C.E. Vol.10, May-Aug.1958 (Paper No.6282)

150. BARNES, B.S. Proc. A.S.C.E. Jour. Hydraulics Div. Vol.85 No HYS Aug.1959 (Paper No. 2128).

151. SILVESTER, R. ed. - Hydraulics and Fluid Mechanics. Proc. First Australasian Conf., University of Western Australia, 6-13 Dec.1962 Oxford, Perganon, 1964; pp 173-190. Paper entitled Hydraulic Mechanism of Runoff by Y. Ishihara.

152. HENDERSON, F.J. and WOODING, R.A. Jour. Geophys. Res., Vol.69 No.8 April 1964.

153. SUGAWARA, M. Japanese Jour. Geophys., Vol.2 No. 4 March 1961.

154. LAURENSON, E.M. Water Research Laboratory, University of New South Wales, Report No. 66 Dec.1962.

155. COLMAN, E.A. and BODMAN, G.B. Proc. Soil Sc. Soc.Amer.Vol.9, 1945.

156. VASQUEZ, R and TAYLOR, S.A. Proc. Soil Sc.Soc.Amer.Vol.22 1958.

157. HOLMES, R.M. and ROBERTSON, G.W. Monthly Weather Review, Vo
87, 1959.

158. LAURENSON, E.M. and PILGRIM, D.H. Jour. I.E. Aust. Vol.35 No.1-2, Jan-Feb.1963.

159. COSTIN, A.B., WIMBUSH, D.J. and KERR, D. Aust. C.S.I.R.O. Div
Plant Industry, Tech Paper No.14, 1960.

160. TURNER, A.K. Jour. of Hydrology, Vol.1 No.2 Sept.1963.

161. DITTMER, H.J. Amer. Jour. Botany Vol.24, July 1937.

162. GARDNER, W.R. and EHLIG, C.F. Agronomy Jour, Vol.54 No. 5, Sept.-Oct.1962.

163. KOHLER, M.A. Simulation of Daily Catchment Water Balance. Paper in ref.7, 1964.

164. BODY, D.N. and RAINBIRD, A.F. Paper in ref.7 1964.

165. SLATYER, R.O. and DENMEAD, O.T. Paper in ref.7.1964.

166. PHILIP, J.R. Paper in ref. 7 1964.

167. KOHLER, M.A. Rainfall-Runoff Models, I.A.S.H. Pubn. No.63 Berkeley, 1963.

168. BELL, F.C. Water Research Foundation of Australia Report No.15 1965.

169. BOUGHTON, W.C. Water Research Lab. Report No.78 University of New South Wales, 1965.

170. LAURENSON, E.M. Jnl. Hydrology, Vol.II No. 2 1964.

171. RANGANA, G. M. Tech. Thesis. University of N.S.W. 1961.

172. COULTER, C.G. Notes on Baseflow Separation. Cyclostyled University of N.S.W. 1960.

173. HERTZLER, R.A. Civil Eng. Vol. 9, P.487, 1939.

174. ZOCH, R.T. Monthly Weather Rev.62, (1934), 64 (1936) and 65 (1937).

175. LEVI, E and VALDES, R. Jnl. Hydrology, Vol.II No. 2 1964.

176. HOLTEN, H.N. and OVERTON, D.E. Jnl. Hydrology, Vol.I No.3, 1963.

177. HOLTEN, H.N. and OVERTON, D.E. Jnl. Hydrology, Vol.II No. 4, 1964.

178. COURT, C.J, Jnl. Geophys. Res, Vol.67, No.11, Oct.1962.

179. VELIKANOV, M.V. I.A.S.H. Pubn. No.63, Page 77, Berkeley 1963.

180. WOODING, R.A. Jnl. Hydrology, Vol.III, Nos.3/4, 1965.

181. DOOGE, J.C. I.A.S.H. Pubn. No. 52 Helsinki, 1960.

182. KOELZER, V.A. Discussion on ref.102. Trans. Amer. Geophys Union Vol.27, No. V. 1946.

183. LASSON, L, HULL, H.W. and FRANK, B. U.S.D.A. Circular No.910 1952.

184. PILGRIM, D.H. I.E. Aust. Civil Eng. Trans. April 1966.

185. MUNRO, C.H. U.N.S.W. Water Res. Lab. Report, No.8 1959.

186. SNYDER, F.F. Trans. Amer. Geophys. Union 1939 Part IV.

187. LINSLEY, R.K. Discussion on ref. 164 See ref.7, 1964.

188. CREAGER, W.P., JUSTIN, J.D. and HINDS, J. Engineering for Dams Vol.I Wiley. New York 1945.

189. BELL, F.C. and GATENBY, M.G. Are Water Yields decreased when Catchment Vegetation is converted to pinus radiata? Water Research Foundation of Australia (in preparation)

APPENDIX

GLOSSARY OF SPECIAL TERMINOLOGY

ACCRETION DELAY: Characteristic time for a storage to increase by the usual processes of accretion. It is given by the ratio of the storage volume to the rate of increase.

BASE EVAPORATION: Evaporation from the more remote parts of the retention storage and therefore characterized by high depletion delays. (q.v.) It comprises the minimum values of evaporativity corresponding with the various values of retention storage.

BASE RETENTION: Retention by the more remote parts of the retention storage and therefore characterized by high accretion delays (q.v). It comprises the minimum values of retentivity corresponding with the various values of retention storage.

BASE RUNOFF: Discharge from the more remote parts of the storage and therefore characterized by high depletion delays (q.v.). It comprises the minimum rates of flow corresponding with the various values of runoff storage.

CLIMATIC REGION: An area of the earth with comparatively uniform climatic characteristics that distinguish it from adjoining regions. In the systems widely accepted by geographers, climatic regions generally exceed 50,000 square miles in area.

COMPLEX WATERSHED: A watershed that cannot be adequately analysed as a single unit because of heterogeneity in rainfall or other characteristics. It may be considered as an appropriate system of storages which are analysed in separate but inter-related groups.

DEPLETION DELAY: Characteristic time for a storage to diminish by the usual processes of depletion. It is given by the ratio of the storage volume to the rate of decrease.

DRYING PERIOD: An increment of time during which the retention storage is continually depleted by evaporation. It is analysed as a single unit but is not necessarily of a fixed duration.

DRYING PHASE: A group of drying periods and minor wetting periods during which the dominant process is evaporation. It is preceded and followed by wetting phases (q.v.).

EVAPORATIVITY: The evaporation that would occur with a high potential rate i.e. when the available enery is not restricted. It is intended to express the limiting rate of evaporation as controlled by the availability of the water and is therefore generally assumed to be uniquely related to the retention storage.

EVAPORATIVITY RECOVERY TIME: The ratio of the retention storage decrement to the resulting increment of evaporativity.

FREE DRYING PERIOD: A drying period throughout which the evaporation rate is equal to the evaporativity (q.v.).

FREE WETTING PERIOD: A wetting period throughout which the retention rate is equal to the retentivity (q.v.)

HYDROGEOGRAPHY: The study of the distribution of hydrological phenomena over the earth's surface.

HYDROLOGIC REGION: A group of hydrologic units (q.v.) corresponding closely with a rainfall region (q.v.).

HYDROLOGIC UNIT: An area of land generally less than 200 square miles, having comparatively uniform hydrologic characteristics that distinguish it from adjoining regions.

LIMITED DRYING PERIOD: A drying period in which the available energy is not sufficient to satisfy the evaporativity and evaporation therefore proceeds at a rate approximately equal to the potential.

PERIOD BIAS: Ratio of a statistic from a particular period of data to the same statistic from a specified standard period of data.

LIMITED WETTING PERIOD: A wetting period in which the rainfall is not sufficient to satisfy the retentivity and retention therefore proceeds at a rate approximately equal to the rainfall intensity.

RAINFALL REGION: A subdivision of a climatic region, generally between 500 and 50,000 square miles in area, in which rainfall characteristics have a comparatively high degree of uniformity.

RESPONSE TIME: Ratio of flow to rate of change of flow.

RETENTION ACCRETION FUNCTION: An expression giving the changes in retentivity caused by increases in retention storage.

RETENTION DEPLETION FUNCTION: An expression giving the changes in evaporativity caused by decreases in retention storage.

RETENTION FUNCTION: An expression giving the rates of increases in retention storage caused by particular rainfall rates.

RETENTION RATE: Actual rate of increase of retention storage during a wetting period.

RETENTION STORAGE: The volume of water in a watershed that is unlikely to become runoff.

RETENTIVITY: The retention rate that would occur with rainfall of high intensity, i.e. when the supply of rainfall is not restricted. It is intended to express the limiting rate of retention as controlled by the watershed.

RETENTIVITY RECOVERY TIME: Ratio of retention storage decrement to corresponding increment of retentivity.

RUNOFF DEPLETION FUNCTION: An expression giving the rates of decrease in outflow caused by depletion of the runoff storage.

RUNOFF STORAGE: The volume of water in a watershed (excluding streamflow storage) that is likely to become runoff at some future time.

SIMPLE WATERSHED: A watershed that can be adequately analysed as a single group of storages, e.g. with one retention storage, one runoff storage and one translation storage. In general simple watersheds should not exceed about 100 square miles.

TRANSIENT EVAPORATION: Evaporation from less remote parts of the retention storage (from the watershed surface) and therefore characterized by low depletion delays.

TRANSIENT RETENTION: Retention by the less remote parts of the retention storage and therefore characterized by low accretion delays.

STREAMFLOW STORAGE: A hypothetical non-linear, reservoir-type storage in series with the runoff storage and intended to simulate the effects of the channel system in rapidly conveying water from all parts of the watershed to the outlet. In specialized analyses this storage may be replaced by more comprehensive hydraulic mechanisms.

WATERSHED ANALYSIS: The detailed study of individual watersheds in order to isolate their particular hydrologic characteristics and to concisely express the effects of these characteristics on rainfall-runoff phenomena.

WETTING PERIOD: An increment of time in which there is continuous rainfall of approximately uniform intensity.

WETTING PHASE: A group of wetting periods and minor drying periods during which the dominant process is rainfall. It is preceded and followed by drying phases. (q. v.).