Aspects of the hydrology of the Browney basin North East England

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ASPECTS OF THE HYDROLOGY OF THE BROWNEY BASIN, NORTH-EAST ENGLAND

by

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ABSTRACT

A hydrological study was undertaken to investigate precipitation, evapotranspiration and runoff in the Browney basin, north-east England The areal and seasonal distribution of precipitation during the period 1968-1972 were analysed and frequency analysis were carried out on the amount of rainfall and the number of raindays per month during the period 1939-1973 at Durham Observatory. There was a high correlation between yearly precipitation and altitude October and September were the driest months of the year while February was the wettest month. The relatively high intensity rainfall in the late summer months gave evidence for the occurrence of convectional rain during this time of the year A wide range in the amount of rainfall and in the number of raindays for any given month at Durham Observatory was observed

To measure evapotranspiration two sets of evapotranspirometers were installed at two locations just outside the extreme eastern and western margins of the catchment Other methods for measurement or estimation of evapotranspiration in the catchment were also used These were the Penman and Thornthwaite formulae, the catchment water balance method and simple hydraulic lysimeters The results of measured evapotranspiration at the two locations indicated greater evapotranspiration at the higher elevation The use of simple hydraulic lysimeters for the measurement of actual evapotranspiration was discarded because of significant differences in the results of two replicates A comparison of the catchment water balance evapotranspiration with the results from the other methods revealed that there was some moisture deficit in the catchment especially during the late summer months

Studies of runoff data from the catchment showed significant variations in the yearly, seasonal and short term patterns These differences were explained by the differences in the amount and distribution of precipitation, evapotranspiration and antecedent soil moisture conditions

The overall hydrology of the catchment was studied by the simulation of the flow records during the period 1969-1973, using the Stanford Watershed Model IV The results obtained revealed high correlations between the monthly recorded and simulated flows The mean monthly recorded flow during the five year period exceeded the simulated flow by 2 per cent The actual evapotranspiration and groundwater components of the hydrologic cycle were also studied using the results of the simulation method

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CONTENTS

LIST OF TABLES

LIST OF FIGURES

No. Page

INTRODUCTION

This thesis attempts to study a number of aspects of the hydrology of the Browney basin, north-east England. The main emphasis, however, is on the study of evapotranspiration and the digital simulation of stream discharge using the Stanford Watershed Model IV.

In the first chapter, the catchment characteristics are briefly explained. The characteristics considered are geology, soil, land use, topography, shape and drainage net. The hydrological importance of each of these factors is also mentioned.

In the second chapter, the spatial and temporal variations of precipitation within the catchment are discussed. For this purpose, the available precipitation data from the seven raingauges inside or close to the basin are used. The effect of altitude on precipitation is then investigated and the results are presented.

In view of the availability of long term precipitation data at Durham Observatory, a 35 year study of daily rainfall amounts greater than or equal to 0.2 mm, 1 mm, 2 mm, 5 mm and 10 mm is made. From the results obtained, frequency curves of the number of days per month with different rainfall amounts over the period 1939-1973 are plotted. Frequency curves of yearly rainfall, annual 24 hr maximum rainfall and monthly rainfall are also presented. Later in chapter two, the intensity of hourly rainfall in the catchment is studied. For this purpose the available recording charts from the Casella recording raingauge at Durham Observatory over the period 1969-1974 are used. The hourly rainfall data are also grouped into four classes, greater than or equal to 0.2 mm, 1 mm, 2 mm, 5 mm and 10 mm. From these data frequency curves of the monthly percentage of hours with precipitation of different amounts are drawn. Evidence for the occurrence of convective precipitation during the late summer months is given by a study of hourly precipitation data.

Evapotranspiration, the methods of its measurement in the Browney basin and the results obtained are discussed in chapters three to five

Thus, the third chapter starts with a definition of the evapotranspiration process and then the effects of different climatic, soil and plant factors upon it are briefly explained Later in this chapter, the theoretical, empirical and water balance methods for measurement of evapotranspiration are explained and the mathematical derivation of the formulae are presented

In chapter four, the methods applied to measure evapotranspiration in the Browney basin are outlined and the procedures followed are discussed in detail These methods are divided into two groups. Those employed to measure or estimate evapotranspiration under unlimited moisture supply conditions (Potential Evapotranspiration), and those to measure or estimate evapotranspiration under prevailing soil moisture conditions in the field (Actual Evapotranspiration)

In chapter five, the results of the evapotranspiration study are shown The presentation of the results commences with those of potential evapotranspiration at the two different locations, Durham and Honey Hill Variations of potential evapotranspiration over a ten year period are then studied Finally the results of the actual evapotranspiration study are discussed

In chapter six, variations of yearly, monthly and mean daily runoff are studied and the distribution of runoff resulting from two similar storms during the winter and summer seasons in the catchment are discussed The duration curves of yearly and monthly runoff during the period October 1956 to September 1973 and the mean daily flows during selected periods are also presented The available data are then used to estimate the 100-year drought and the 100-year flood for the catchment

The overall hydrology of the basin is studied by simulation modelling, using the Stanford Watershed Model IV. This model had been progressively developed at the Department of Civil Engineering, University of Stanford, U S A. The final report of this pioneering work was reported in 1966 Since then the model has been applied to many watersheds in the U S A and elsewhere In the United Kingdom, the application of the Stanford Watershed Model is limited to studies on the River Clyde in Scotland (Fleming, 1970, Bunny, 1973). One of the reasons for the limited use of the model may be the large amount of data which the model requires for proper simulation of a river Another reason for the lack of widespread use of the model is explained by the difficulty which is often experienced in optimizing a relatively large number of parameters by manual or automatic optimization methods The Stanford Watershed Midel IV was originally developed for the prediction of runoff from ungauged basins and the extension of short term streamflow records in gauged basins However, it has been successfully used in catchment water balance studies and in studies of the effects of artificial changes of land use on the hydrology of a catchment.

In this study there are two main objectives in the application of the Stanford Watershed Model IV to the Browney basin

1. - To test the applicability of this model in simulating hydrographs of British rivers, using the Browney river as a case study

2 - To study the water balance equation of the Browney catchment, in particular the groundwater component of the total runoff and actual evapotranspiration

Thus, the section on simulation starts in chapter seven with a review of some of the methods for studying and predicting runoff prior to the development of the Stanford Watershed Model IV

In the following chapter, the application of the Stanford Watershed Model IV is discussed, its different parameters are defined and the derivation of these parameters (measured and fitted ones) are explained in detail.

In chapter nine, the sensitivity of a parameter is defined and the process of optimization is discussed. The results of sensitivity tests of fitted parameters as applied to the Browney basin are then presented and the optimized values ultimately used in the simulation of the basin are shown.

The results of simulation modelling for the five year period commencing January 1969 and ending September 1973 are presented in chapter ten Later in this chapter the monthly and yearly variations of groundwater flow and actual evapotranspiration, as determined by this model, are studied. Subsequently the importance of accurate input potential evapotranspiration data in the model are investigated. For this purpose, the variations in the yearly, monthly and daily runoff resulting from two different sets of potential evapotranspiration data (Penman evaporation and evapotranspiration) are studied

In the last chapter of this thesis, the major conclusions reached as the result of this study are presented.

It should be mentioned that throughout this study the term 'water year' is used to denote the year starting October of the preceding year and ending September of the current year. Thus water year 1972 refers to the year starting October 1971 and ending September 1972

It has been attempted to express the results of this study in terms of metric system of units. However in the section on simulation, since all the input and output data of the Stanford Watershed Model IV are in English units, the diagrams in this section are expressed in English units also. In the text, on the other hand, the equivalent metric values are included in parenthesis Similar procedure is followed in the first chapter where some map information is in terms of English system of units.

CHAPTER ONE

CATCHMENT CHARACTERISTICS

The Browney River is a tributary of the River Wear in County Durham, north-east England (Figs 1 and 2) It rises in the Pennine Uplands (Alston Block) and joins the River Wear in the Wear Lowlands. The Alston Block is composed of a rigid block of sediment in the western part of the County. It falls gradually in height in an easterly direction, until at around 120 m, it merges into the Wear Lowlands (Beaumont, 1970).

The Browney River with its tributary, the Deerness, drains an area of 69.7 mi^2 (178 km^2) in an easterly direction The area was determined by planimetering on a 1 25000 O.S map, using the topographic boundary as the basis of separation of the catchment from its adjacent ones It has been assumed that the watershed and the groundwater boundaries coincide.

The Browney River rises on the moors near Burn Hill (082 444, sheet 84), while the Deerness rises at Tow Law (122 392, sheet 84), (Calvert, 1884 as reported by Smith,pm1972). The catchment is 9 8 mi (15 8 km) long and its average width is about 7 1 mi (11 4 km)

Geology. The geology of Durham County, of which the Browney catchment is a part, is the result of tectonic activity causing fracturing and folding rocks which has brought them to their present position. These rocks range from Carboniferous to Jurassic in age The oldest rocks outcrop in the western hill margin and dip eastward under a progressively younger rock cover Much of the solid rock, particularly in the lowland zone is covered by deposits of glacial drift (Cairney et al, 1970). The thickness of the drift cover differs even within relatively short distances and the thickness is appreciable below a height of 150 m. This fact was confirmed by Francis (1970) who stated that drift cover on the ridges is generally thin and impersistent He thus concluded

Fig.1 LOCATION MAP OF THE BROWNEY BASIN

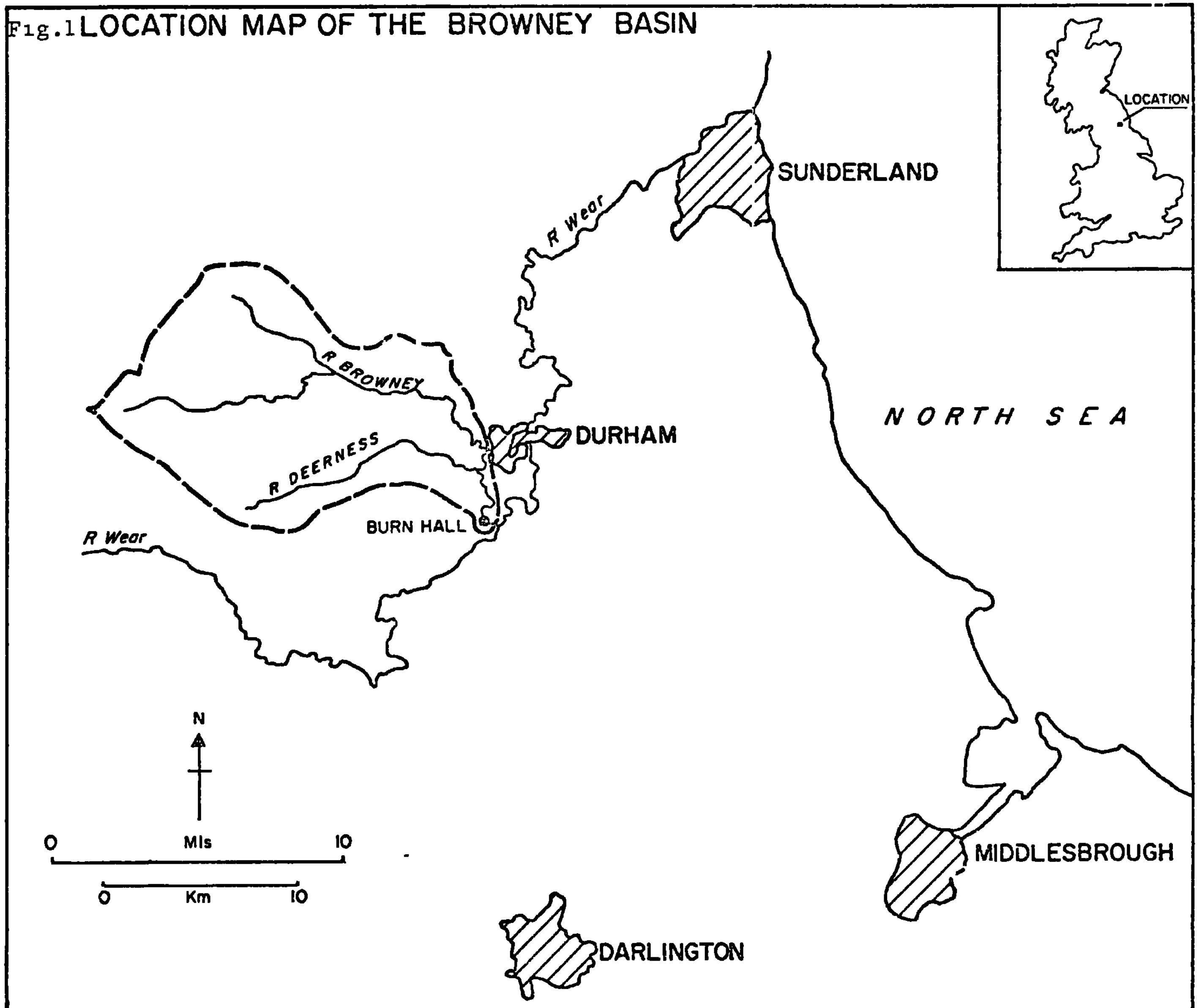
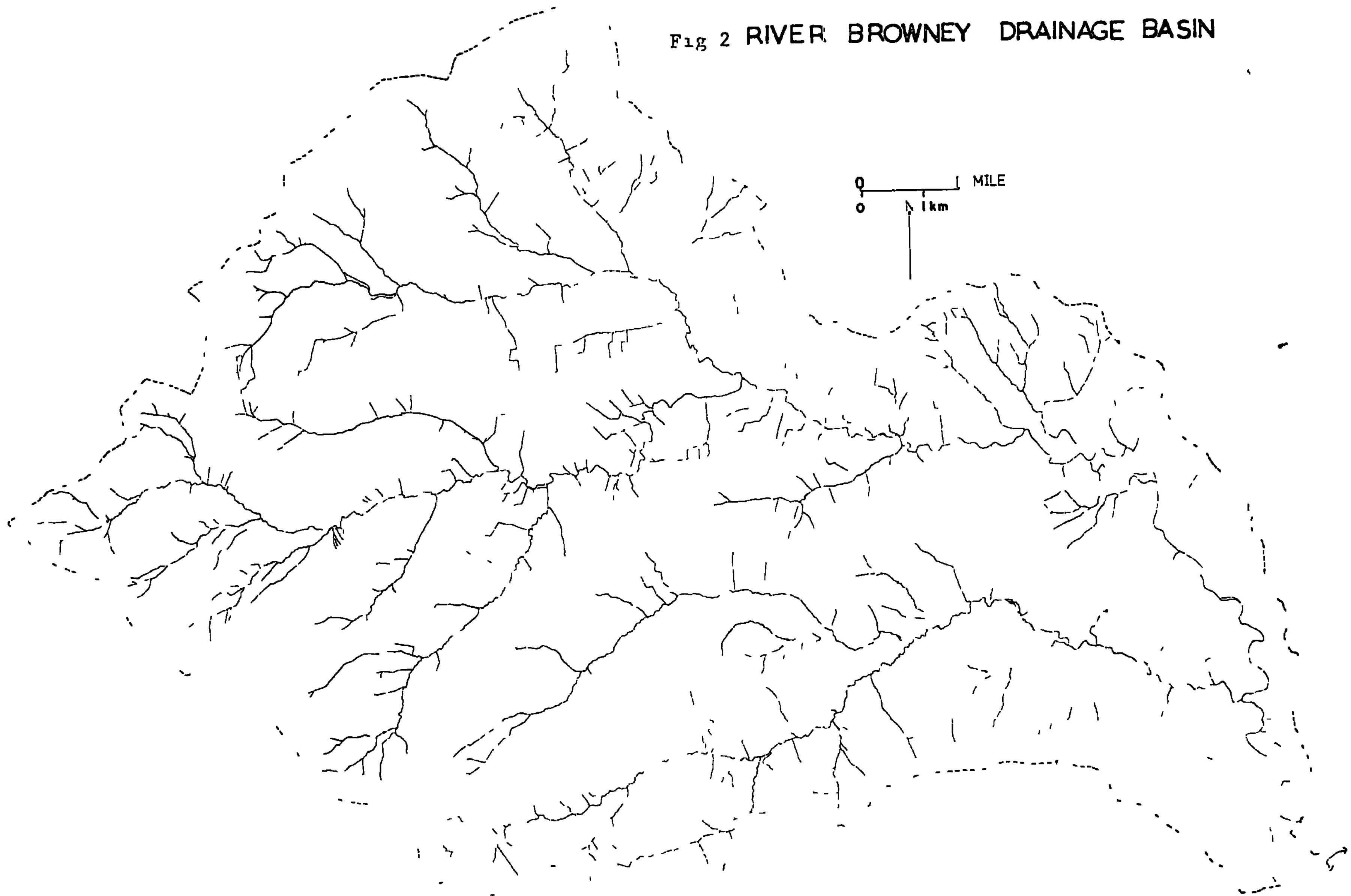


Fig 2 RIVER BROWNEY DRAINAGE BASIN



that the principal drift deposits are situated along the lines of the major valleys.

The glacial deposits have in fact buried the valleys associated with the present rivers and according to Beaumont (1970), the most widespread of these drift deposits is the lower boulder clay which nearly always rests upon solid rock. This till is a stiff dark grey or grey-brown sandy clay and in the Wear lowlands contains much Carboniferous material In many places, it is compact and tough with large boulders

The Carboniferous rocks forming the solid geology of the Browney catchment belong to the Coal Measures Group. The term Coal Measures is given to a series of sedimentary cycles each composed of the general sequence - shale - siltstone - sandstone - seatearth - coal - in ascending order (Johnson, 1970). The Coal Measures are divided into Lower, Middle and Upper divisions. The Lower Coal Measures are a sandy series with mainly thin coal seams in the west of Durham The Middle Coal Measures contain thick coal seams. The Upper Coal Measures are mostly shaly and are not represented in the Browney area (Smith, 1972) The Lower and Middle Coal Measures which are represented in the study area are shown diagrammatically in Appendix 1.

The geology of a catchment affects the hydrologic cycle by controlling the baseflow component of runoff. If the rocks are of high storage capacity, then the baseflow contribution is high. If the rocks are not porous, then there will not be any substantial baseflow and, therefore, the yearly hydrograph would be more peaked and the river would be flashy

For the Browney basin the rocks are not the best for the storage of groundwater. However due to coal mining large cavities have been created in some of the aquifers which can store large volumes of water. Some of the water thus stored in the mining cavities is being pumped out into the river.

<u>Soil</u> The soils, and in particular their physical properties are a very important factor affecting the hydrology of a catchment Soil texture and soil structure determine the rate of infiltration, and thus affect runoff volume and runoff distribution with respect to time They also determine the capacity of the soil to store moisture. Soil moisture storage affects actual evapotranspiration during dry periods. It also is considered as an index of flood potential.

In the area under study the major soil type is formed by the heavy grey till of the Carboniferous rocks, though there are soil types which are developed on alluvium, fluvio-glacial deposits and sandstones. The principle characteristics of the soil of the County of which the catchment is a part, consist of a grey-brown sandy loam or loam surface horizon, a mottled yellow-brown sandy clay loam B horizon, which overlies a grey sandy clay loam or clay loam till (Stevens and Atkinson, 1970)

The reconnaissance map of Durham by the Soil Survey of England classifies the soil of the district under three main series (Jarvis and Stevens, 1969, quoted by Smith, 1972)

1. Newburn Series

Type sandy loam, loamy sand

Topography high ridges

Site drainage seasonal drought

Profile drainage excessive to free

Parent material Coal Measures, sandstone

Profile 0-17.5 cm, yellowish brown, few stones, nutty structure, porous 17.5 cm, bright yellowish brown micacious sandy brash Weathering rock.

2. Improved Croxdale Series

Type loamy, sandy loam

Topography rolling slopes

Site drainage seasonally wet

Profile drainage imperfect

Parent material Coal Measure till

Profile 0-25 4 cm, dark brown to brown, stony, crumb structure, porous and friable.

25.4-53.3 cm, bright yellowish-brown loam to sandy loam, very strong, fine cloddy structure, slightly compact, mottles, occasional MnO_2

53 3 cm, yellow-brown true sandy clay loam with greyish tinge, numerous boulders, cloddly structure, compact MnO_2, mottling and grey coating on faces of structural elements.

3. Croxdale Series

Type sandy loam, loam

Topography undulating

Site drainage seasonally wet

Profile drainage impeded

Parent material Coal Measure till

Profile 0-20 3 cm, dark grey-brown, stoneless, fine cloddy structure, compact, much rusty mottle

20.3-30.5 cm, dark grey-brown loamy sand similar to above layer.

30.5-61 cm, yellowish-brown coarse sandy clay loam with grey marbling, many stones, prismatic structure, fissures and tenacious rust and MnO_2 mottling

61 cm, dark brown clay loam with grey marbling, many stones and occasional boulders.

The soils of about twenty square miles (51 km^2) of the Browney-Deerness valley were studied in detail by Smith DM (1972). According to the results of this study, the soils of the region are divided into seven series, namely Esh, Broom, Opencast, Ushaw, Witton and Gilbert, Browney and Deerness. Three of these series e.g Esh, Broom and Opencast are derived from Carboniferous rocks or clay parent material with a heavy to moderately fine textured soil These soils have impeded or imperfect drainage and they occur mostly on slopes of varying steepness and on the interfluves

The soils of the Ushaw series are developed on sandstone parent material and occur on the ridge tops They are of limited extent and have free drainage

The soils of Witton and Gilbert series are developed on fluvio-glacial gravels. Generally the amount of clay present in the profile is low resulting in free drainage. These soils occur along valley bottoms.

The soils of the Browney Deeiness series are developed on alluvium and they provide the deepest rooting zones in the area. They are of moderately coarse texture with sandy loam predominating.

The dominant soils in the area of Smith's detailed survey are the Esh and Broom series which cover more than 60% of the soils studies. These two soil series are very similar to those of the improved Croxdale and Croxdale Series, mentioned earlier

p^f Curve One of the most important physical properties of a soil is its capacity for holding moisture and in particular the moisture within the range between field capacity and permanent wilting point The amount of moisture in this range is called available water and is used by crops and vegetation during a drought period

In view of this importance of soil moisture content, an attempt was made to prepare moisture retention curves of the soils at Durham Observatory (267 415) and Honey Hill (052 468) (Fig.3). These are the locations at which evapotranspiration data were collected.

The soil moisture retention curve (p^f curve as suggested by Schofield in 1935 for ease in graphical representation and tabulation) shows the relation between moisture tension and water content in the moisture range associated with plant growth To construct the curve, a pressure membrane apparatus was employed (Richard,1952) Nine moisture content values corresponding to nine tensions applied were determined.

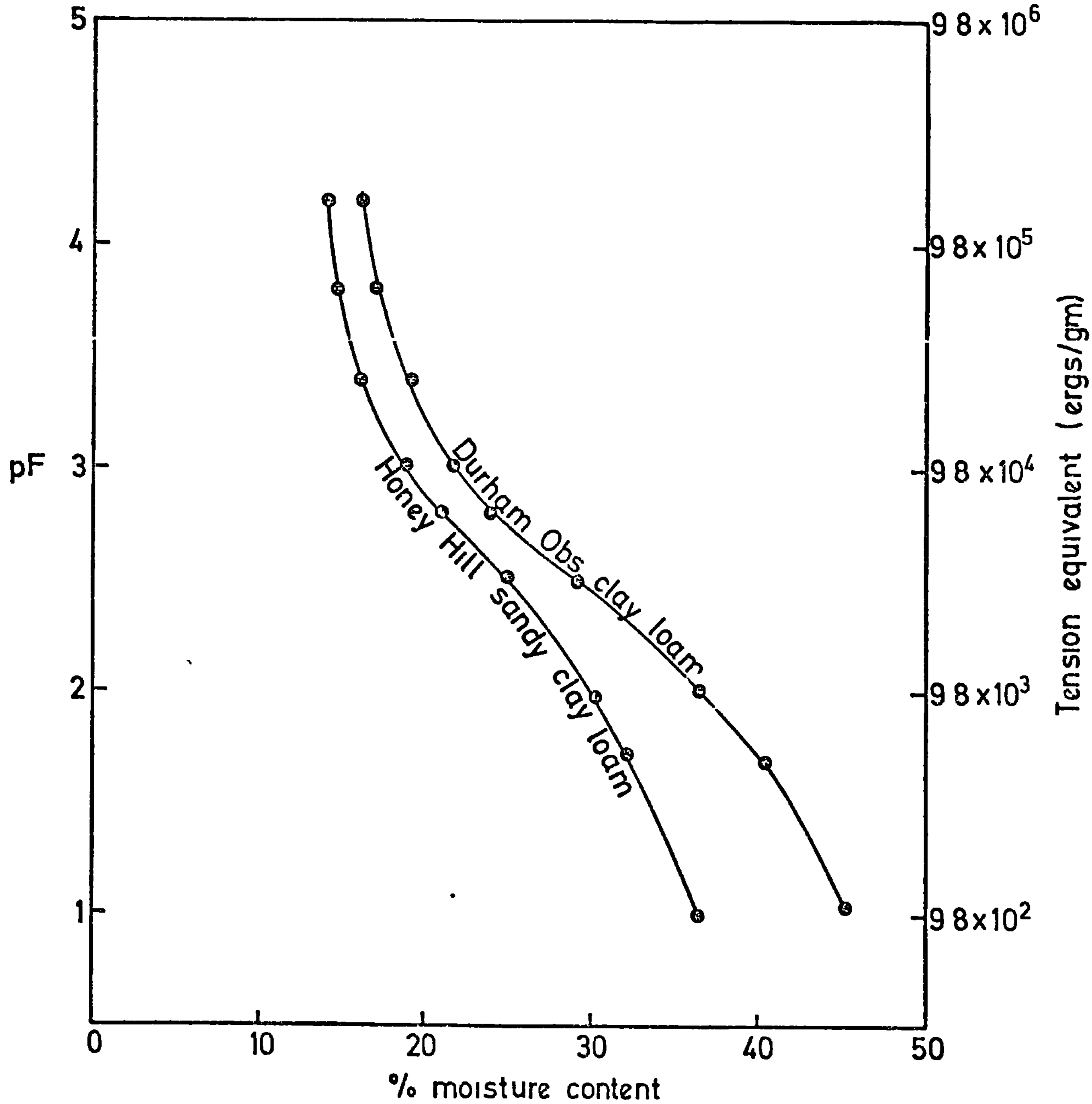


Fig 3 Water retention (pF) curves of soils at Durham Observatory and Honey Hill

For each determination two samples from each location were used.

Tables 1 and 2 show the results of moisture retention curve and textural analysis (Hydrometer method) of the soils from Durham Observatory and Honey Hill. These results show that the total available water for the Durham Observatory soil is 13 2 per cent, while that for the Honey Hill soil is 11.3 per cent These figures were obtained by deducting the values of the permanent wilting point, (corresponding to 15 atmospheres)

Table 1 p^f values and the corresponding moisture content of soils at Durham and Honey Hill

p^f values Atmosc Eqv	1 0 1/100	1.7 5/100	2 0 1/10	2.5 1/3	2 8 2/3	3 0 1	3.4 2 5	3.8 6.4	4.2 15.0
% moisture content of soils									
Durham Obs.	45 5	40.6	36 5	29 3	24.2	22 0	19 0	17.1	16.1
Honey Hill	36 3	32 4	30.1	24 9	21 2	18 6	16 0	14 7	13 6

from those of field capacity (corresponding to 1/3 atmosphere). Thus the maximum available water at Durham Observatory would be 132 mm per metre of soil, while at Honey Hill, the corresponding figure is 113 mm per metre of soil The difference in the amount of available water at Durham and Honey Hill is explained by the textural differences of the soils. The soil at Honey Hill is lighter in texture, sandy clay loam, compared with clay loam at Durham It has a lower capacity for holding moisture and, therefore, during a drought period. the Honey Hill soil can support the water need of the crops for shorter periods than the soil from Durham Observatory.

Table 2 Results of textural analysis of soils at Durham and Honey Hill

Soils	% sand	% silt	% clay	texture
Durham	33.0	31 5	35 5	clay loam
Honey Hill	60 5	15.0	24 5	sandy clay loam

<u>Land Use</u> Land use or land management is another factor significantly affecting runoff from a catchment. A vegetation cover lowers the velocity of rain drops striking the soil surface It also improves the structure of the soil and increases the infiltration capacity. Therefore runoff will be lower from a soil under vegetation than from a soil with no plant cover. Other effects of land use on the hydrology of a catchment are due to processes of interception and evapotranspiration, e.g. as the percentage of vegetation increases, the volume of water stored by interception and used by evapotranspiration increases.

In the Browney catchment, the main land use is agriculture, though there are some portions of the area taken by woodland, settlement, mine working or derelict land. The percentage of these different factors of land use for about twenty square miles (51 km^2) of the Browrey Deerness valley is given by Smith (1972) as follows

Agriculture	73%
Forestry plantation	7.25%
Settlement including roads	15%
Recreation areas	1.4%
Mining and its legacy	2.6%
Industrial land	0.75%

For the catchment as a whole the area of settlement is 7.0 percent as determined by planimetering on a 1 25000 O S map, while the total area of forestry plantation and agriculture is higher than in the sample study.

<u>Shape</u> The shape of a catchment governs the form of the hydrograph and the peak flow rate. Long narrow catchments would be expected to have attenuated flood discharge periods, whereas round basins would be expected to have sharply peaked hydrographs. In spite of the difficulty of expressing shape numerically, several indices have been suggested and used. Wisler and Brater (1959) refer to two indices given by Cravelius (1914) The first, termed the compactness coefficient, is

the ratio of the perimeter of the watershed to the circumference of a circle whose area is equal to that of the drainage basin. This coefficient is a measure of its flood potential The closer the index is to unity, the greater is the circularity of the basin and the more susceptibility to flooding The value of this index for the Browney catchment is calculated as $k_p = {}^{P}/_{2}\sqrt{\pi A} = 1\ 45$ (P is the perimeter with the value of 43 0 miles (69 2 km) and A is area which is 69 7 m^2 (178 km^2)) The second index also proposed to Gravelius in 1914, is the form factor derived by dividing the average width of the catchment by its axial length The axial length is the length measured from the outlet to the most remote point in the catchment The value of the form factor for a compact catchment is close to unity, while a long narrow catchment has a low form factor value. Comparing two catchments of similar size, the one with a lower form factor, generally yields a lower peak runoff, since a heavy rainfall is less likely to fall simultaneously over the entire area For the Browney, the form factor is ${}^{7.1}/_{9\ 8} = 0.72$

A third index of shape is Miller's circularity ratio (Rodda, 1971) represented by $Rc = \frac{A_b}{A_c}$ in which A_b is the area of the basin, and A_c is the area of the circle with the same length as the perimeter of the drainage basin Miller's index is unity for a circular basin, and for two basins of equal size the flood potential would be greatest for the one with the smallest A_c value. Miller found that the circularity ratio remained remarkably uniform in the range 0 6 to 0 7 for first and second order basins in homogeneous shales and dolomites, indicating the tendency of small drainage basins in homogeneous geologic material to preserve geometrical similarity (Strahler, 1964) Miller's index was employed in analysing the flow from a number of Appalachian basins, but it was found to have a low correlation with peak discharge (Rodda, 1971). The value of Miller's index for the Browney basin is $\frac{69\ 7}{147\ 1} = 0\ 47$

Elevation. For an analysis of the elevation characteristics of the basin, the area-elevation distribution was determined by analysis of contour information from a 1 63360 O S map (Fig.4) The area lying between successive 100 ft (30.5 m) contour intervals was measured. The percentage of the total that each of these areas form was then computed and the percentage of the total area lying above and below each different contour was obtained by summation (Table 3 and Fig.5)

Table 3 The percentage of area above or below the indicated elevations

Contour elevation		Area between contours in km^2	% of total area	% total over lower limit	% total lower than upper limit
ft	(m)				
150-200	(45.7 - 60 9)	1.8	1 0	100.0	1.0
200 300	(60 9 - 91 4)	9 4	5.2	99 0	6 2
300-400	(91 4 - 121.9)	19 8	11.0	93.8	17 2
400-500	(121 9 - 152.4)	21 2	12.0	82 8	29 2
500-600	(152 4 - 182 9)	24.3	13 8	70 8	43.0
600-700	(182 9 - 213.4)	26.6	15 1	57 0	58.1
700-800	(213.4 - 243 9)	28 1	15.9	41.9	74 0
800-900	(243 9 - 274 4)	21.6	12.1	26 0	86 1
900-1000	(274.4 - 304.9)	12 9	7.2	13.9	93.3
1000-1100	(304.9 - 335 4)	7.7	4.3	6.7	97 6
1100-1200	(335.4 - 365 9)	2 3	1.2	2.4	98.8
1200-1250	(365 9 - 381 1)	2 3	1.2	1 2	100 0

In the Browney catchment elevation ranges from 150 ft (45.7 m) at Low Burn Hall (252 384) to about 1,250 ft (381.1 m) near Moss Hole Corner (056 447), Honey Hill The mean elevation of the catchment was obtained using the formula $E = \sum ae/A$ in which

a - is the fraction of area between successive contours,

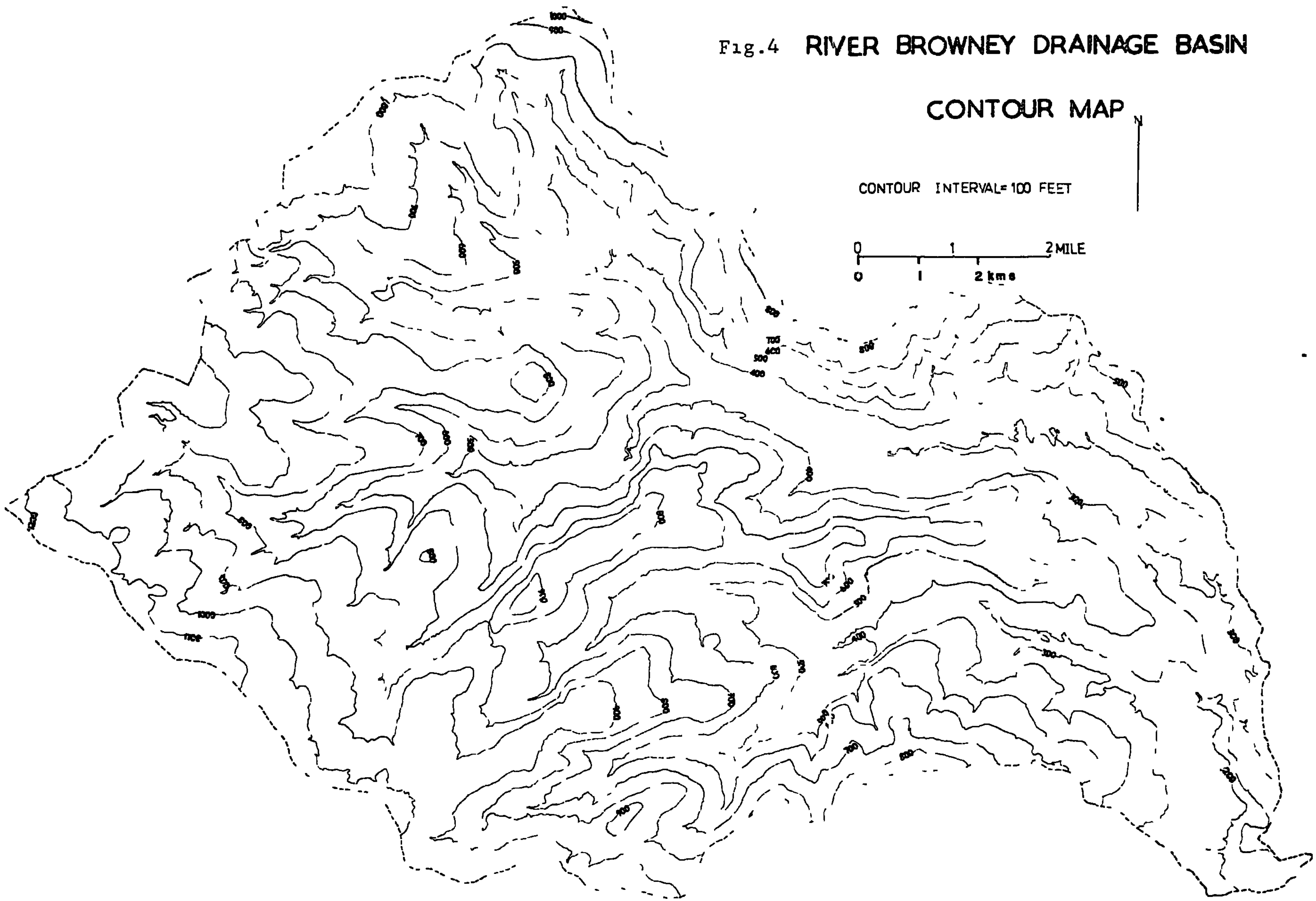
e - is the mean elevation of contours and

A - is the total area of the basin

The mean elevation of the catchment from the above formula is 642 ft (195 7 m).

Fig.4 RIVER BROWNEY DRAINAGE BASIN

CONTOUR MAP



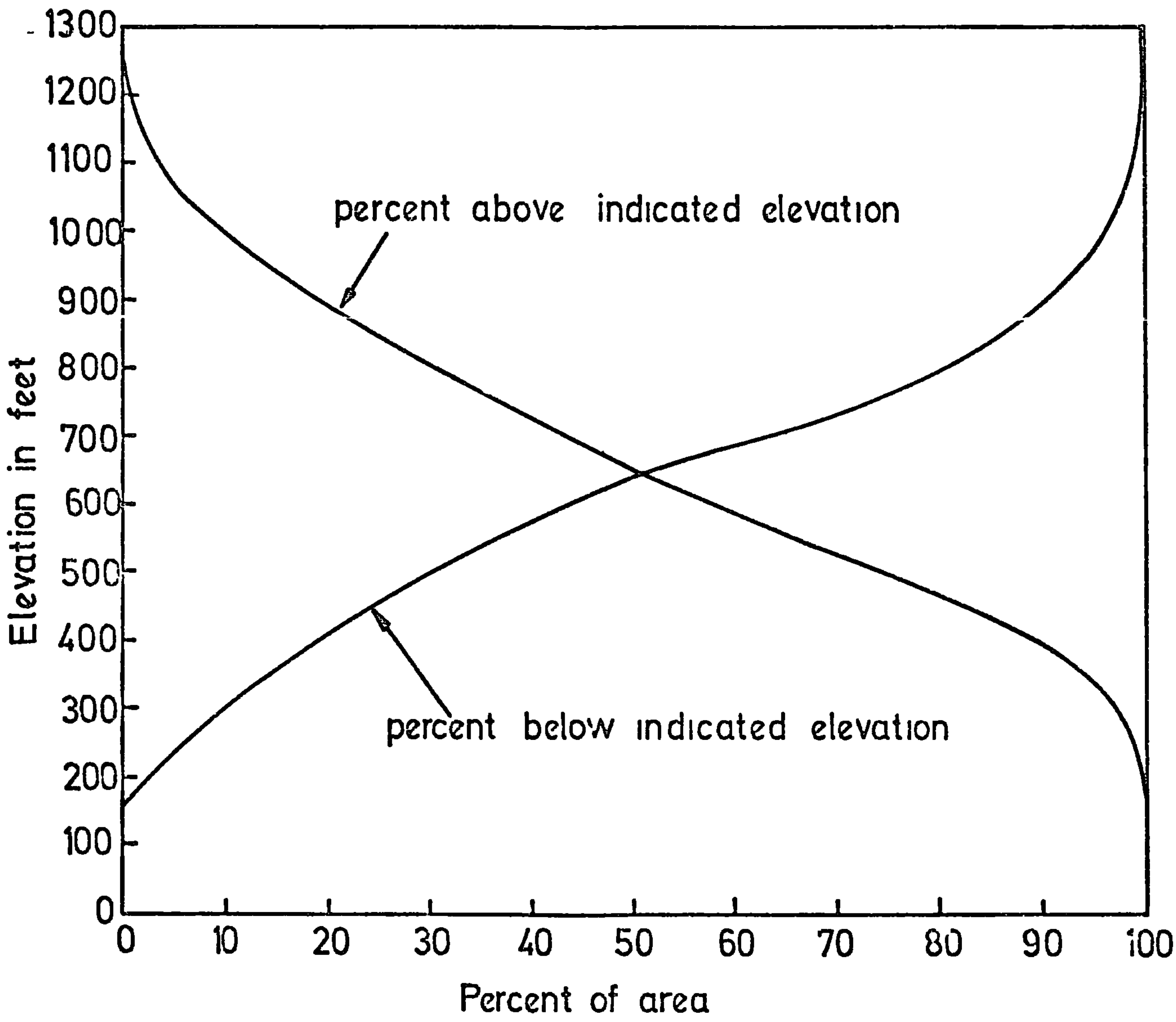


Fig 5 Hypsometric curve of Browney Basin

As for the effect of elevation upon the hydrology of a catchment, it is clear that an increase in elevation results in a decrease in temperature and an increase in precipitation depth The effect of elevation upon precipitation depth is very large within the Browney basin and their relationship will be studied in the following chapter. Elevation also affects the precipitation form. However its effect upon the evapotranspiration process is not well understood This has also been investigated in this study and the results are presented in the chapter on evapotranspiration

<u>Slope</u> The following procedure was adopted to determine the slope of the catchment (Nash,1966).

At the intersection points of the grid lines on a 1 25000 O S. map of the catchment, the distances between adjacent 25 ft (7 6 m) contours were measured. These distances were then used to take the slope at each point. An analysis of the frequency distribution of these slope values was made and the results are tabulated in Table 4 and shown in Fig.6. The number of intersection points was 152.

<u>Table 4</u> Percentage of area with slopes above the indicated values

limit of % slope	Area in sq mls (km^2)	% of total	% of area over lower limit
0-5	18.5 (47 2)	26.50	100.0
5-10	35.7 (91 2)	51.20	73.5
10-15	10.5 (26.8)	15.10	22.3
15-20	4.1 (10 5)	5.90	7 2
20-	0.9 (2 3)	1 30	1.3

The mean slope of the catchment is 5 9%, while the median slope is 6.4% The importance of the slope of a catchment is on its control over the time of overland flow and its indirect effect on infiltration and runoff magnitude. In other words with an increase in slope, the

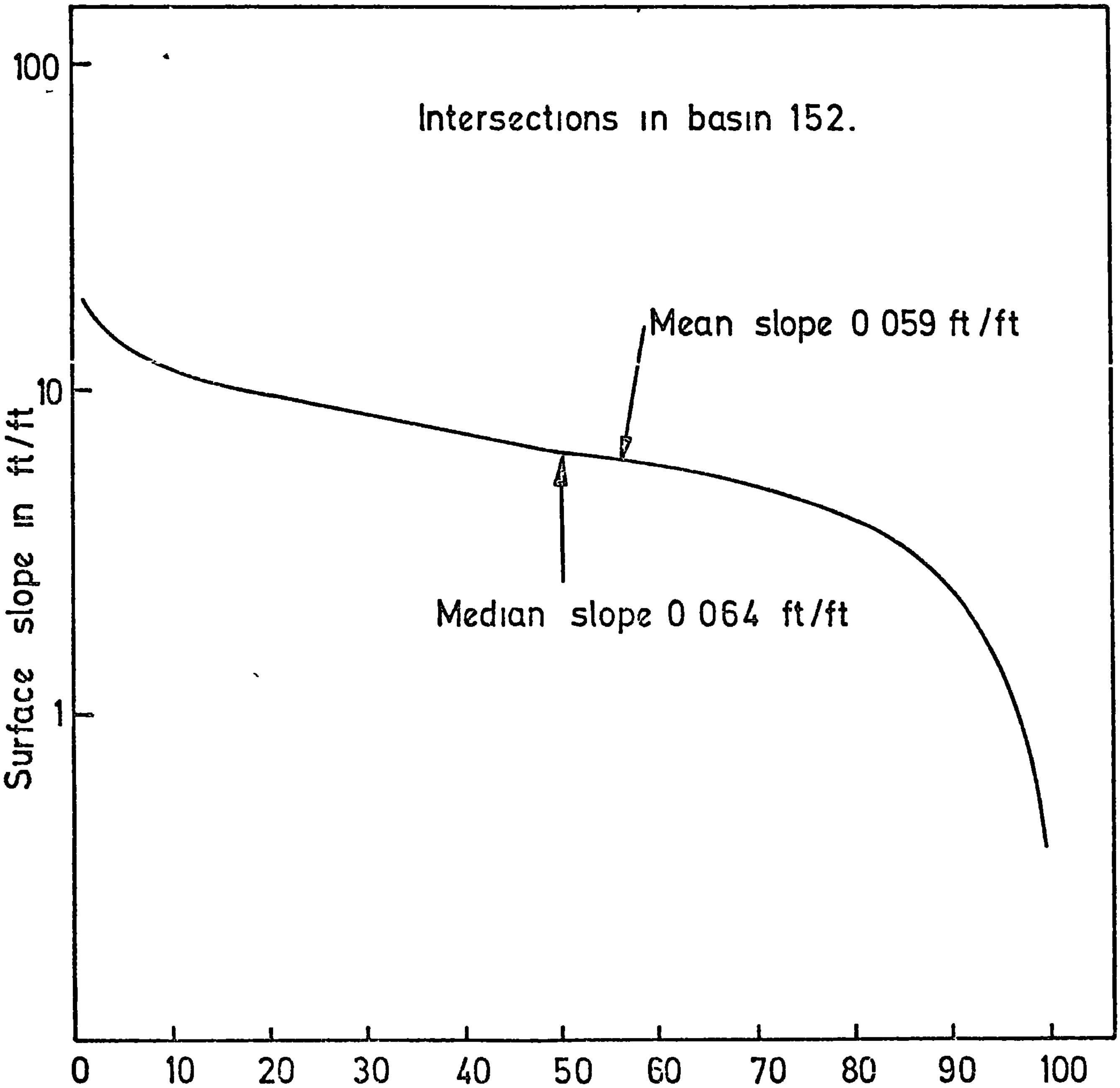


Fig 6 Percent of area with slope equal to or greater than indicated value

velocity of overland flow increases. Thus, the time during which moisture can infiltrate into the soil decreases, and as a result, runoff magnitude increases.

Drainage network The drainage network is described by the order of the streams, length of tributaries, stream density, drainage density and length of overland flow The drainage network governs the shape of the hydrograph and the efficiency of the drainage system For example, the larger the number of branches and segments of each branch, the more attenuated are the flood discharge periods and vice versa.

Stream order It is this index which reflects the branching or bifurcation in the basin. Horton (1945) (Wisler and Brater,1959) has classified stream order by assigning 1 to the small unbranched, finger tip tributaries, 2 to those streams with two branches of first order, 3 to those with two branches of second order, etc.- Strahler (1964) adopting this system considers the unbranched tributaries as order one, where two branches of first order meet, that forms a second order and where two second order branches, third order is formed.

Shreve (1966) suggested another ordering method in which the value of any order reflects the number of first order streams feeding into it.

In this study the Strahler system of classification has been applied to the Browney River using a 1 25000 O.S. map which is the basis of any large scale investigation The results of this analysis are presented in Table 5.

It is shown that with an increase in stream order, the number of segments decrease. The plot of the number of segments versus order on a semi-logarithmic paper gives an approximate straight line relationship (Fig.7) The antilog of the regression coefficient, known as the bifurcation ratio is the index for the degree of branching (3 20) and is important in controlling the peakedness of the runoff hydrograph.

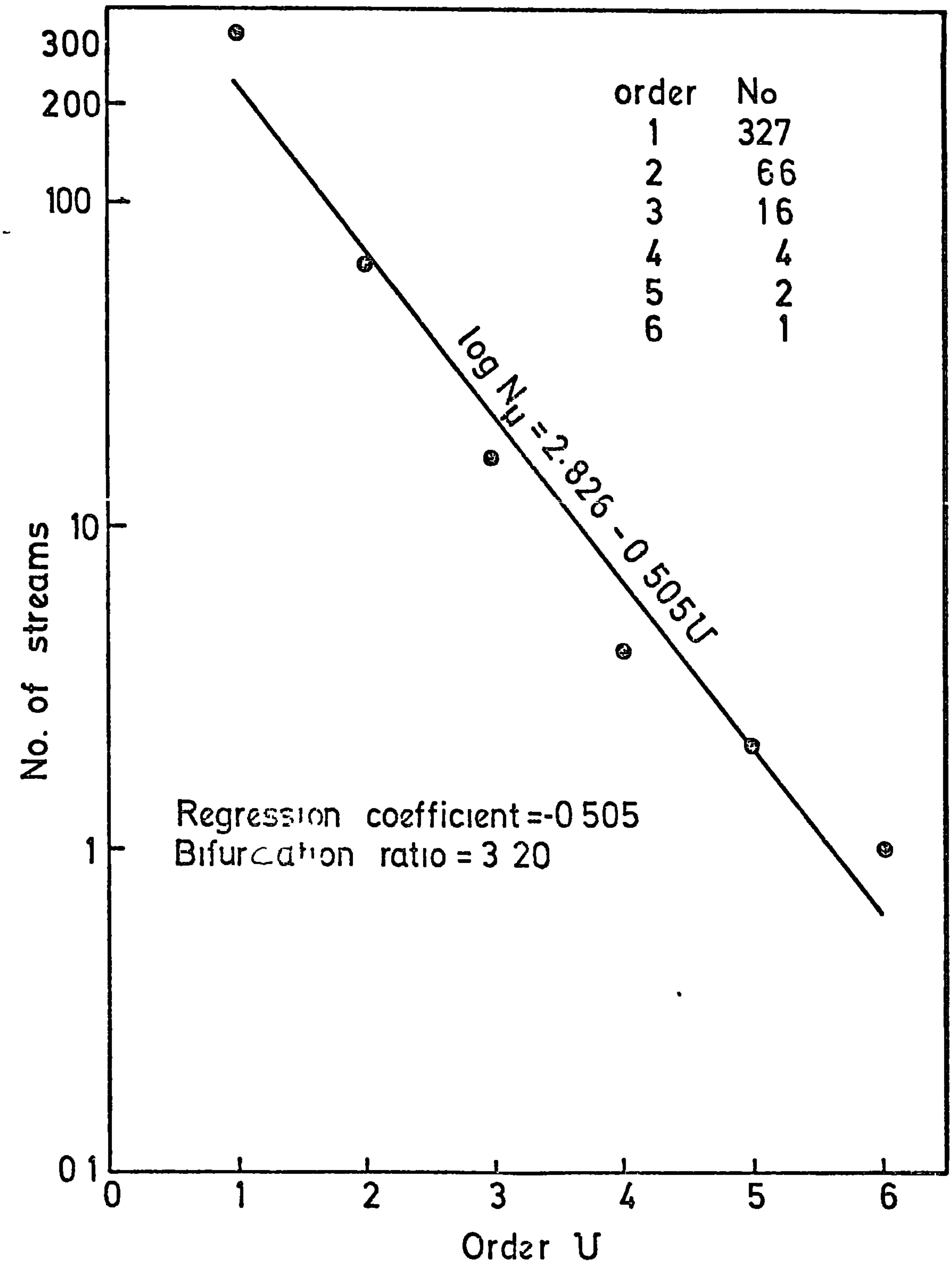


Fig 7 Plot of number of stream segments versus order with a fitted regression

Table 5 Stream order, number of segments in each order, total length and mean length of segments of each order for the Browney River

Stream order	Number of segments in each order	Total length miles (km)	Mean length of segments miles (km)
1	327	83 44 (134 2)	25 (0 40)
2	66	43 76 (69 9)	.66 (1.1)
3	16	29 56 (47 6)	1 84 (2.9)
4	4	20.60 (33 1)	5 15 (8.3)
5	2	15.56 (25 0)	7 78 (12 5)
6	1	3.96 (6 4)	3 96 (6 4)

Chorley (1971) states that "The bifurcation ratio for a given density of drainage lines is very much controlled by basin shape and shows very little variation (ranging between 3 and 5) in homogeneous bed rock from one area to another Where structural effects cause basin elongation, however, this value may increase appreciably" For the Browney catchment the bifurcation ratio is 3.20 which means that on the average there are 3.20 times as many channel segments of any given order as of the next higher order.

Length of streams The length of tributary streams is a function of the order of tributaries. When the mean lengths of tributaries are plotted versus their orders on semi-logarithmic paper, an approximate straight line results and the antilog of the regression coefficient is the length ratio. The value of length ratio for the Browney catchment is 1.89,(Fig 8)

It should be stated that the above mentioned law has not been fully applicable to the Browney River, and that the mean length of the two fifth order streams has been higher than that of order 6 (Table 5).

The importance of the length of tributaries is ascribed to their indication of the degree of drainage and the steepness of drainage basin Steep well drained areas usually have numerous small tributaries, whereas in plain regions where the soils are deep and permeable, only relatively

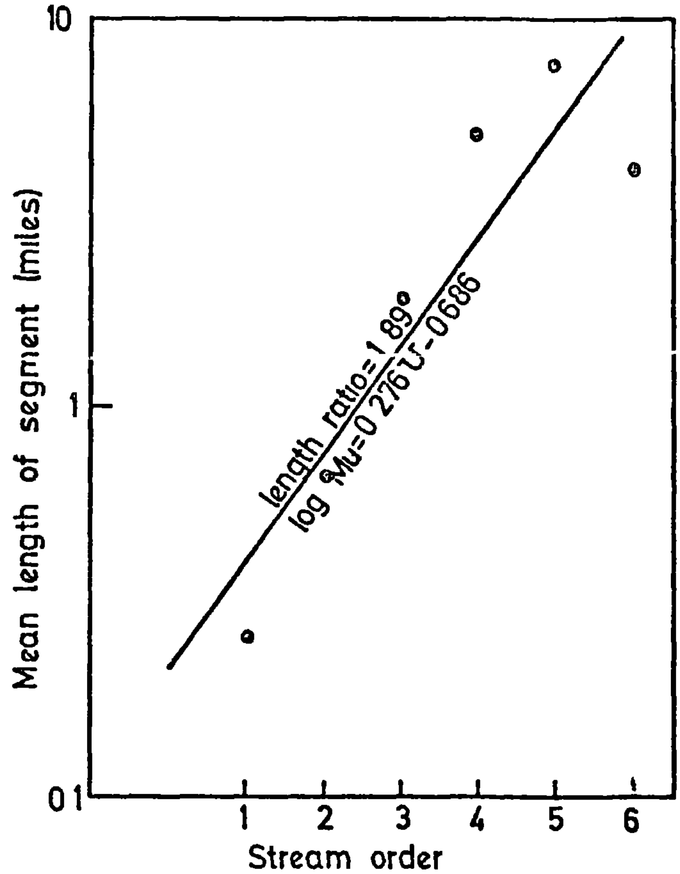


Fig 8 Plot of mean length of segment versus order with a fitted regression for River Browney

long tributaries are maintained as perennial streams (Wisler and Brater, 1959).

Drainage Density. This is the average length of streams per unit area within the basin i e. $D = \frac{L}{A}$ where L is the total length of streams in the basin and A is the drainage area Drainage densities vary from well below 1 mile/sq. mile (0 63 km/km^2) in poorly drained basins to more than 5 mile/sq mile (3 2 km/km^2) in exceptionally well drained basins (Linsley et al, 1949). In general low drainage densities are found in regions of highly permeable subsoil material under dense vegetation cover and where relief is low High drainage densities occur in regions of impermeable subsurface materials, sparse vegetation and mountainous relief.

The value of drainage density is 2 82 mi/mi^2 (1 73 km/km^2) for the Browney catchment The drainage map of the basin (Fig.2) shows that the drainage density is highest in the headwater section of the catchment.

Length of overland flow This parameter is a measure of stream spacing or the degree of dissection and is determined from the formula

$L_g = {}^1/_{2D}\sqrt{1-({}^{S_c}/{S_y})^2}$ in which

L_g - is length of overland flow

D - is drainage density

S_c - is channel slope

S_y - is the average ground slope (Horton, 1945, Strahler, 1964)

The value of this parameter for the Browney River is 0.177 miles (0.285 km).

Longitudinal profile of the stream channel The stream profile is a plot of horizontal distance along the stream versus elevation The slope of a stream between any two points is taken as the fall between the points, divided by the stream length If the segments of a given order are combined into a single average segment, then the average slope of the channel segments of that order is the average vertical fall over the mean horizontal length This procedure has been followed for finding the slopes of different orders of the Browney River and the composite profile is plotted in Fig 9. It is shown that the average slope for a given order is more than that of a higher order and less than a lower one This relationship is known as the law of stream slopes and is similar to that of stream slopes (Horton,1945, Strahler,1964) The plot of mean stream slopes in ft/mile versus stream order is given in Fig.10.

As observed from Fig 9 the profile of the Browney River has got an exponential shape with a marked concavity.

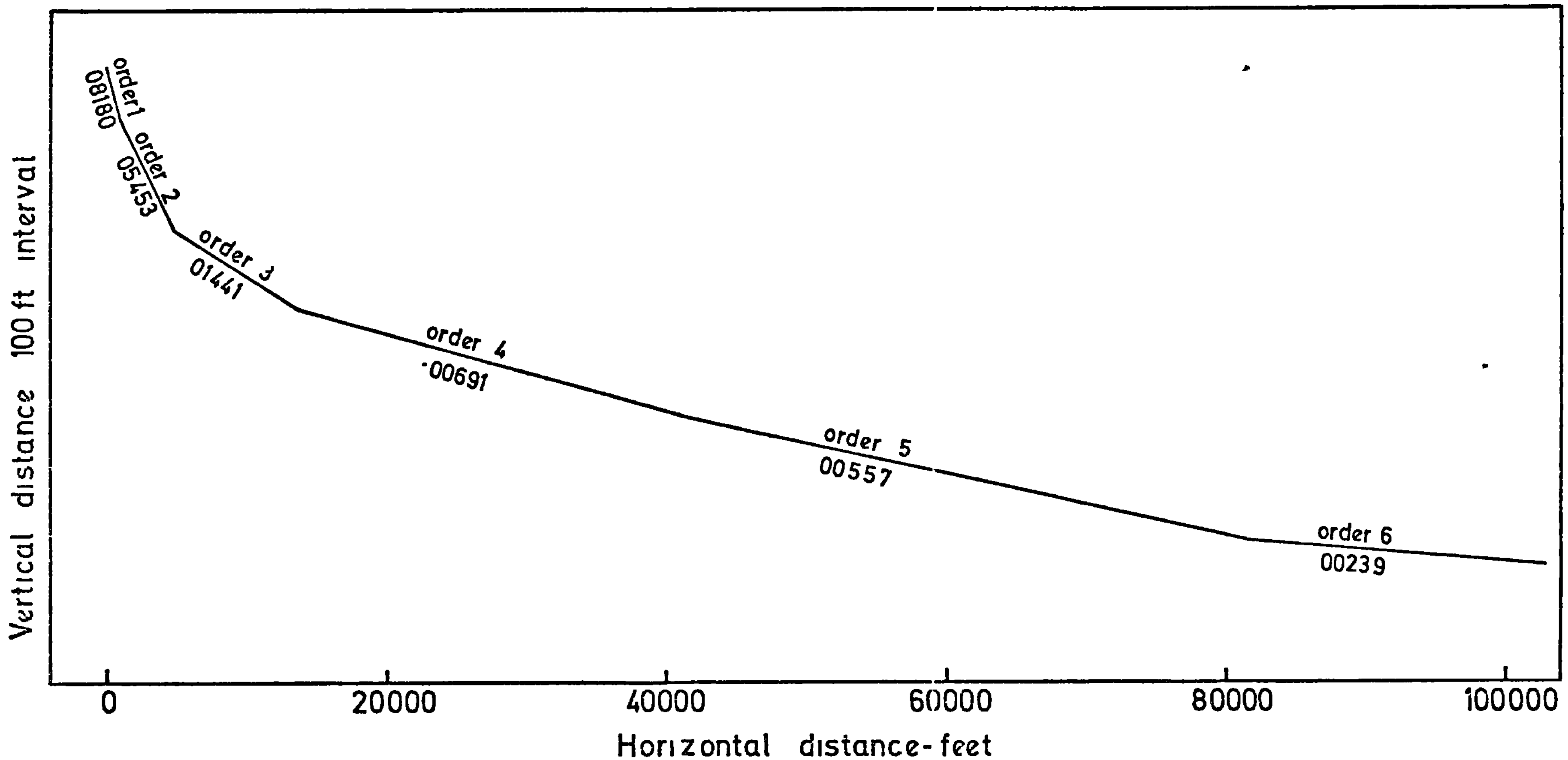


Fig 9 Plot of composite profile of Browney River showing the difference in mean slope of each of the six segments of differing order

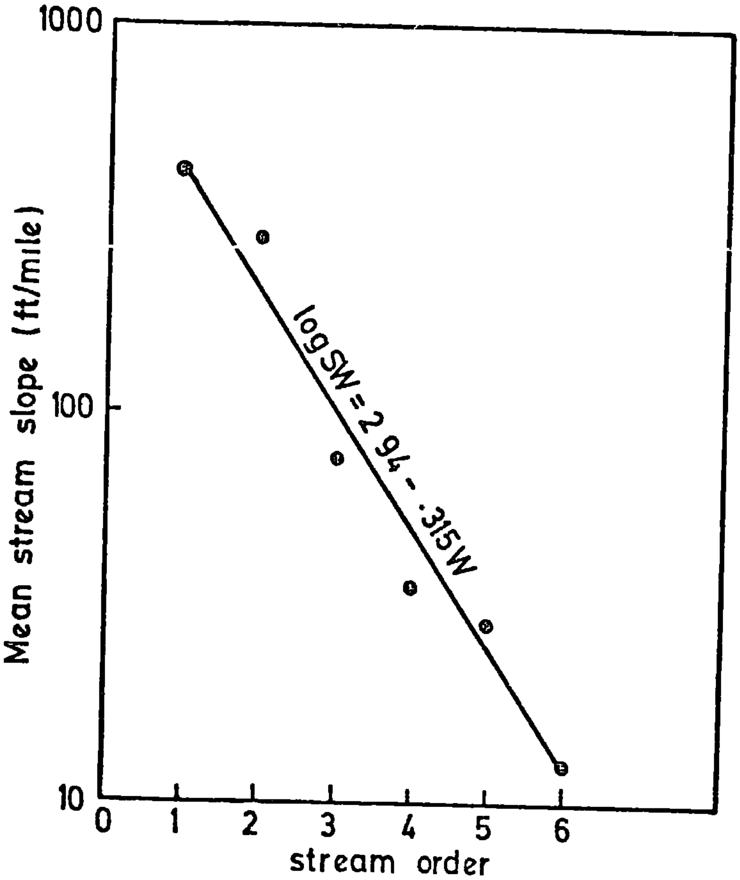


Fig 10 Regression of mean channel slope versus order for River Browney

CHAPTER TWO

PRECIPITATION

Precipitation refers to all forms of moisture falling to the ground. The necessary conditions for precipitation to occur are

1. A sufficient supply of water vapour in the atmosphere
2. Lifting of the water vapour to higher altitudes

This second factor results in cooling with subsequent condensation of the vapour Condensation occurs at the dew point of the temperature at which saturation occurs if air is cooled at constant pressure.

Precipitation is divided into three types, depending on the means of uplift of the water vapour These types are

1. cyclonic
2. convectional
3 orographic

Cyclonic precipitation is classified into non-frontal and frontal precipitation. In the non-frontal type, lifting of air is caused by convergence of air streams towards a low pressure region. Frontal precipitation occurs as a result of lifting of warm air over cold air. This occurs in warm or cold fronts In the warm front, warm air moves upwards over a wedge of cold air In the cold front, however, warm air is replaced by moving cold air and therefore it rises Warm front precipitation is of lower intensity than cold front precipitation

Convectional rains are caused by the differential heating of the air at the surface As a result the warmer air rises and cools adiabatically It, therefore, causes condensation and precipitation This type of precipitation is spotty in terms of areal distribution and usually of high intensity

Orographic precipitation is that which results from cooling of moisture laden air masses by the mechanical lifting of mountain barriers

This type of precipitation is usually more pronounced on the windward side of mountains than on the leeward.

Within the Browney basin, precipitation is due to all the three types mentioned above Smith (1970) mentions that most of the precipitation over the County Durham is ultimately cyclonic. However, as will be shown, the effect of orographic lifting is quite apparent when the relationship between precipitation and altitude is considered. Evidence for the existence of convectional precipitation during the summer season is shown by the analysis of rainfall intensities later in this chapter

Measurement of precipitation within the Browney Basin

One of the problems associated with the accurate estimation of mean precipitation over a catchment is the density and distribution of the raingauges A number of workers have discussed the accuracy required for the measurement of precipitation within a catchment and Ward (1971) refers to some of them

Clearly the density of raingauges required for measuring rainfall with a certain accuracy varies with catchment characteristics The number of gauges in an area with steep slopes should probably be different from an area of flat terrain Under conditions with frequent occurrence of convectional precipitation, more raingauges are needed than when precipitation is mostly of cyclonic type This is explained by the fact that cyclonic precipitation might in some cases be uniform over a radius of 25 km^2, whereas convectional precipitation can vary widely over a short distance.

Thus Hershfield (1967) noted that in major summer storms in relatively flat areas, it was not unusual for the isohyetal pattern to show gradients of 2 inches or more per mile (32 mm per km) and that it was not unreasonable to expect storm precipitation differences of about 1 inch (25.4 mm) between raingauges separated by distances of only 0.1 mile (0 16 km) (Ward, 1971b)

For the Browney basin, precipitation is measured by an autographic raingauge and eight daily gauges The locations of these gauges are shown in Fig 11 and their National Grid References, altitudes and the dates since the records were available are given in Table 6. The records from Ushaw raingauge which were available for this study date back to 1971. This station, therefore, has not been included in the estimation of mean catchment rainfall because of its limited available data.

Thus the density of raingauges whose records have been employed in this study is about one per 10 mi^2 (25 5 km^2) Six of the raingauges are distributed within the altitude range of 336-741 feet (102-226 m) The other raingauge is at an elevation of 1,179 feet (359 m) However, it is located outside the catchment In view of the high degree of association between altitude and rainfall, as will be shown in a later section, it can be stated that additional raingauges in the catchment within the elevation range of 800 to 1,250 feet (244 to 381 m) would probably have given a more representative picture of precipitation at the higher elevation. Therefore the accuracy of estimation of mean areal precipitation could have been increased

Description of daily and autographic raingauges

The daily raingauges which are used to measure rainfall are the Meteorological Office Mark II gauges These have a collecting funnel aperture 5 inches (127 mm) in diameter and more than 4 inches (101 mm) deep The funnel leads down to a glass bottle placed either in an inner can or in the base itself. The gauge is set with its rim 12 inches (30.5 cm) above the ground level.

The autographic raingauge, which is situated at the Durham Observatory was installed in 1962. The autographic raingauge is an electrical one designed to record total rainfall at a distance from the rain receiver and so obviate the necessity of frequent outdoor visits for changing the charts The rain is received in an aperture defined by an 8 inch

Table 6 National Grid Reference, altitude of the raingauge stations and the date since when the records were available for the Browney Basin in this study

Location of raingauge	National Grid Reference	Elevation ft	(m)	Record available since
Durham	NZ 267415	336	(102)	1886
Consett	NZ 125504	710	(216)	1967
Edmondsley	NZ 232488	500	(152)	1968
Lanchester	NZ 175482	680	(207)	1968
Satley	NZ 117437	741	(226)	1961
Waskerley	NZ 022444	1179	(359)	1895
Waterhouses	NZ 189412	438	(133)	1960
Ushaw	NZ 219436	597	(182)	1971

(203 mm) diameter brass rim It is then fed down by a funnel into one half of a bucket, which when it is full, tilts and empties itself, so allowing the other half of the bucket to fill until the process is repeated. Each measurement of the bucket is transmitted electrically to a moving chart so that the rainfall record consists of a number of steps of 0.02 inches or 0 5 mm of rainfall When two inches or 51 mm of rainfall has fallen, the recording pen reaches the top of the chart and then automatically returns to the baseline.

There are several objections to the use of such records. The main objection is that rainfall less than the capacity of the bucket is not recorded. Another objection is that any rain falling during the tilting of the bucket is also not measured. The instrument, however, has the advantage of not being damaged by the frost. It also records rainfall at a distance from the rain receiver which is an advantage in adverse weather conditions

Estimation of mean areal rainfall

Mean precipitation over the Browney basin was determined by three methods

1. Arithmetic method

2. Thiessen polygon method

3. Isohyetal method

1. Arithmetic method In this method, the values of precipitation from all the seven gauges were added together and the total thus obtained was divided by the number of raingauges e g. $P = \frac{P_1 + P_2 + P_3 + \ldots . P_7}{7}$, in which P is the mean areal precipitation and P_1, P_2 . .. and P_7 are the precipitation depths at each station. The mean areal precipitation over the catchment for each of the five years during the period 1968-1972, and the five year average for the same period is shown in Table 9.

2. Thiessen polygon method This method was devised by Thiessen (1911) to determine the mean depth of precipitation over a given area. The fundamental principle followed to accomplish this purpose consists in weighting the precipitation value of each station by a suitable proportion of the area for which that station value is considered to be representative

To determine the weighting factor for each station in the Browney basin, the geographical location of each station was plotted on a map of the basin The adjacent stations were joined by straight lines. The perpendicular bisectors were then drawn from the mid-points of each line These perpendicular bisectors formed the boundaries of a number of polygons, except where the area was limited by the boundary of the catchment (Fig.11). Each of the polygons represented the area surrounding one rainfall station. It was assumed that the rainfall at each station was the same as any other location within the polygon surrounding it The area of each polygon was then measured and was expressed as a percentage of the total area of the drainage basin. These dimensionless ratios formed the weights given to the rainfall stations and they are shown in Table 7.

The mean areal precipitation for each period (year and month) was

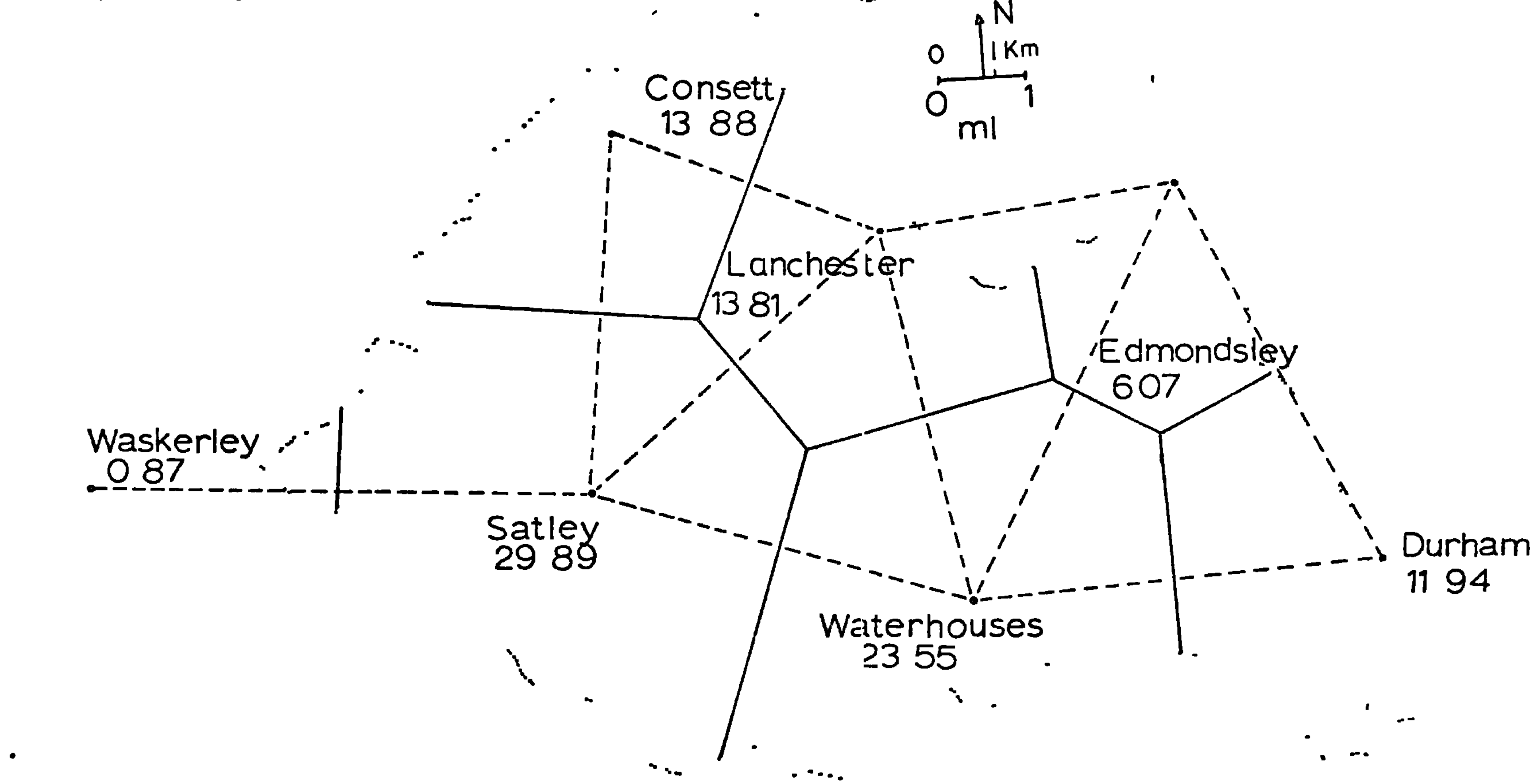


Fig 11. The Thiessen Polygons for average precipitation in the Browney Basin

obtained by multiplying the depth of precipitation by the dimensionless ratios for each gauge and summing these values e g

$$P \text{ (mean precipitation)} = \frac{A1}{A} P_1 + \frac{A2}{A} P_2 + \frac{A3}{A} P_3 + \frac{A4}{A} P_4 + \ldots \frac{A7}{A} P_7$$

Table 7

Raingauge stations and their Thiessen indices	
Durham Observatory	11 94
Waterhouses	23 55
Edmondsley	6 07
Lanchester	13 81
Consett	13 87
Satley	29 89
Waskerley	0.87

3. Isohyetal method Mean precipitation over the basin was also determined by the isohyetal method Isohyets are lines joining points of equal precipitation To draw the isohyetal map for any basin, the precipitation depth over any period of time at each station is plotted at its corresponding geographical location Points with elevations equal to those of selected isohyets are then fixed, and lines drawn through points of equal elevation represent the isohyets.

To determine the mean areal precipitation by this method, the area between any two successive isohyets is measured and expressed as a percentage of the total area. The value thus obtained is multiplied by the mean values of the two isohyets The products are then summed and the result is a weighted mean depth of precipitation

This procedure was followed to determine the mean depth of precipitation for each year of the period 1968 to 1972 and for the mean monthly values over the same period The calculation used for the estimation of mean precipitation for the year 1972 is shown in Table 8.

Table 8 Calculation of mean areal precipitation by the isohyetal method for the Browney basin during the year 1972

Isohyetal value, upper limit mm.	Isohyetal value, lower limit mm.	Columns $\frac{1+2}{2}$ mm	Ratio of enclosed area to total	Columns 3x4 mm
550	541	545.5	0.00129	0.70
575	550	562.5	0.01488	8.37
600	575	587.5	0.04398	25.84
625	600	612.5	0.06145	37.64
650	625	637.5	0.12095	77.11
675	650	662 5	0.18110	119.99
700	675	687.5	0.25220	173.43
725	700	712.5	0 27040	192 64
750	725	735.5	0.02520	18 60
775	750	762 5	0 02134	16.27
800	775	787.5	0.00711	5.60
			TOTAL	676.19

Yearly mean rainfall of the catchment

The values of mean areal precipitation of the catchment for each of the five years, 1968 to 1972 and the five year average is shown in Table 9. From this table it can be observed that arithmetic mean precipitation exceeds the values obtained by both the isohyetal and the Thiessen methods The Thiessen method also reveals a lower estimate than the isohyetal method The difference in the five year average values between the isohyetal and Thiessen is less than one per cent, whereas that of the arithmetic method exceeds the isohyetal and Thiessen methods by 2 2 and 3.1 per cent respectively. The relatively high value of the arithmetic mean precipitation is due to the inclusion of the Waskerley raingauge in the estimation of mean areal rainfall This gauge is located outside the Browney basin at an elevation of 1,179 ft (359 m) and the rainfall at this station is the highest of all the gauges. This therefore, results in an overestimation of mean rainfall by the arithmetic method In the Thiessen method, the weight given to this station is only 0 87 per cent (e g., the precipitation at this station is representative of only 0 87 per cent of the area) whereas in the arithmetic method, its weight is equal to that of any other gauge, or is about 14 3 per cent.

Table 9 Mean areal precipitation of the catchment, 1968-1972 in mm

Year	Arithmetic	Thiessen	Isohyetal
1968	856.1	829 0	835.5
1969	901 3	872 7	872.8
1970	685 3	653.6	669.1
1971	645 6	624 0	628 8
1972	676.3	667 5	676.2
Average	752.9	729.4	736.5

Considering the application of these different methods for the estimation of mean areal rainfall, the arithmetic method is the simplest. It could possibly be used for basins in which the density of raingauges is rather high and where the distribution of the gauges closely follows the physiography of the catchments For the Browney basin, however, since the density of the gauges is about 1 per 10 mi^2 (25.5 km^2) and the distribution of gauges is rather less at higher elevations, the results obtained by the arithmetic method overestimate those of the Thiessen and isohyetal methods.

The Thiessen method is an objective method because it gives weighting factors to each of the gauges. The weights once obtained can be applied for any number of computations. It is, therefore, easy to use and because of this reason when a large number of means are to be determined for the same area and when extreme accuracy is not required, the Thiessen method would be suitable

The isohyetal method is considered to be the most accurate method of the three if drawn accurately. However, it is the most tedious of them all. One of the advantages of the isohyetal method is that it shows the pattern of the rainfall. However, since the rainfall pattern varies from storm to storm, a new isohyetal map should be drawn for each storm.

Mean monthly precipitation over the catchment

The average areal value of mean monthly precipitation using the isohyetal and Thiessen methods are shown in Table 10.

Table 10 Average areal values of mean monthly rainfall by the isohyetal and Thiessen methods in mm

Month / Method	J	F	M	A	M	J	J	A	S	O	N	D	Total
Isohyetal	76.5	60.8	61.8	50.0	63.4	61.8	61.0	74.5	53.6	33.0	78.3	61.8	736.5
Thiessen	76.8	60.1	59.6	49.6	63.3	62.2	62.5	73.3	52.1	31.3	77.6	61.0	729.4

It can be seen that the isohyetal values exceed those of the Thiessen method during nine months. In January, June and July, the Thiessen values are more than those of the isohyetal method. November, January and August are respectively the months with the highest precipitation amounts, while October, is the month with lowest precipitation.

Precipitation pattern over the catchment

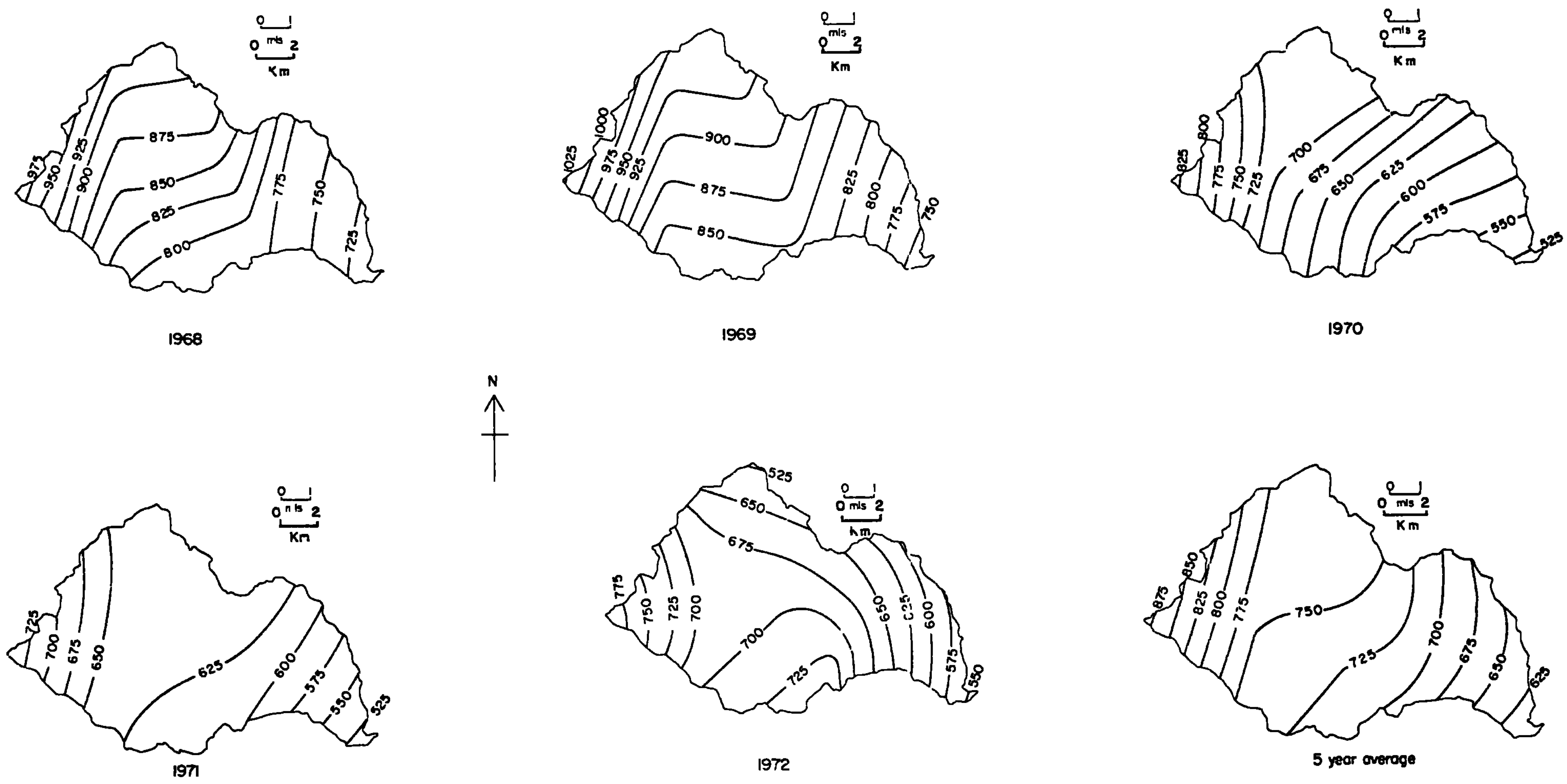
(a) Yearly pattern. A study of the yearly isohyetal maps (Fig 12) shows that in general there are three different rainfall zones in the Browney basin.

1. A westerly and north westerly zone of rapidly increasing rainfall.
2. An easterly and south-easterly zone of declining rainfall.
3. A central area with relatively uniform precipitation.

The first zone is characterized by high elevations and steep slopes. The elevation in this zone is almost everywhere above 700 feet (213 m).

Fig 12

ISOHYETAL MAPS OF THE YEARLY RAINFALL (mm) IN THE BROWNEY BASIN DURING 1968 - 1972



In view of the great influence of topography on precipitation in this area, therefore, precipitation amounts are highest in this part of the catchment The precipitation gradient in this zone averaged about 35 mm per mile (22 mm per km) as judged by the five year isohyetal map

The second rainfall zone is characterized by low elevations and low slopes The altitude in this zone is mostly below 500 feet (152 m) The precipitation gradient is less in this zone than in the first zone

The third zone is the zone with relatively uniform precipitation as judged by the five year average isohyetal map However, when the yearly isohyetal maps are considered, it is observed that precipitation for some years e g 1968, decreases from north to south with a relatively modest gradient of about 15 mm per mile (9 4 mm per km) The trend for the year 1972, however, changes During this year precipitation increases from north to south. This lack of consistency might be due to the changing directions of the depression or random fluctuations

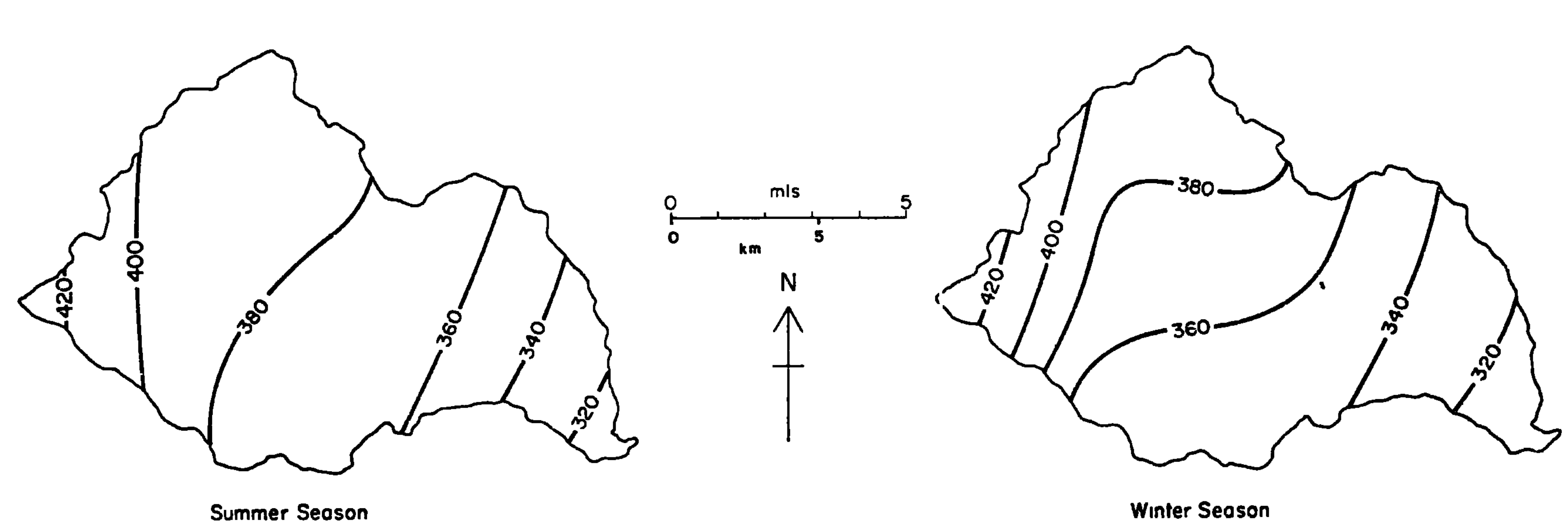
(b) Seasonal pattern of precipitation Isohyetal maps of the winter and summer seasons are shown in Fig 13 It can be observed that the precipitation patterns of the two seasons are identical in having increasing westerly and decreasing easterly components. However, during the winter season, there is a decreasing precipitation from north to south, though such a decrease is not obvious during the summer season

(c) Monthly pattern of precipitation The pattern of precipitation for the three months of September, November and December is more or less similar to that of the winter season For the particular month of November, there is no easterly gradient of rainfall, but the decline of rainfall from north to south is quite sharp

The rainfall patterns for other months show a decreasing rainfall depth from west to east, with the gradient being steeper for the months of January, March and August These gradients range from a maximum of

Fig 13

ISOHYETAL MAPS OF AVERAGE SEASONAL RAINFALL (mm) IN THE BROWNEY BASIN DURING 1968-1972



5 mm per mile (3 mm per km) in the western part of the catchment during March to about 1 mm per mile (0 6 mm per km) in the central part of the catchment in August. For the months of February, April, May, June, July and October, the gradient of precipitation from east to west is low, and in some places is less than 0 1 mm/mile (0 06 mm/km) (Figs.14 and 15).

Precipitation versus altitude

Throughout the previous sections it has been emphasized how the elevation influences the distribution of precipitation within the catchment. In order to study this relationship more clearly, regression equations of precipitation versus altitude were derived for each year in the period 1968-1972 as well as the mean of the five year period The equations derived are of the form

$P = a + bh$ in which

P - is the precipitation depth,

h - is the height in feet,

a - is the ordinate's intercept and,

b - is the coefficient of regression

The equations thus obtained are shown in Table 11.

Table 11 Yearly regression equations of precipitation versus elevation

Year	Regression Equation
1968	$P = 588.7 + 0\ 4084\ h$
1969	$P = 653.3 + 0.3784\ h$
1970	$P = 418.8 + 0\ 4069\ h$
1971	$P = 483\ 9 + 0\ 2622\ h$
1972	$P = 481.9 + 0.2968\ h$
Average	$P = 524\ 3 + 0\ 3490\ h$

Using these equations, it has been possible to derive the mean precipitation value for each year by substituting the value of mean elevation of the catchment The mean elevation of the catchment is 642 feet (196 m).

Fig 14

ISOHYETAL MAPS OF AVERAGE MONTHLY RAINFALL (mm) IN THE BROWNEY BASIN DURING 1968-1972

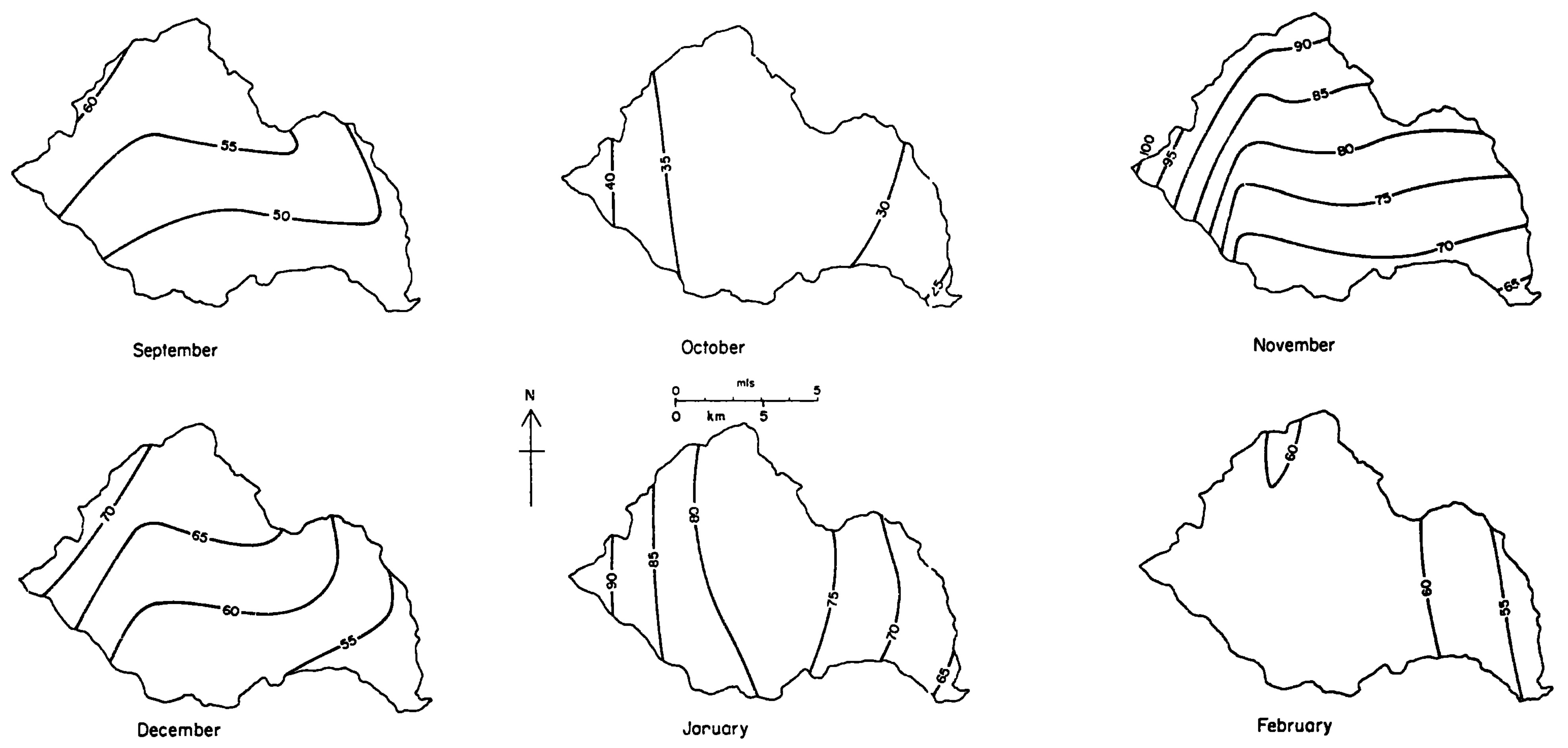
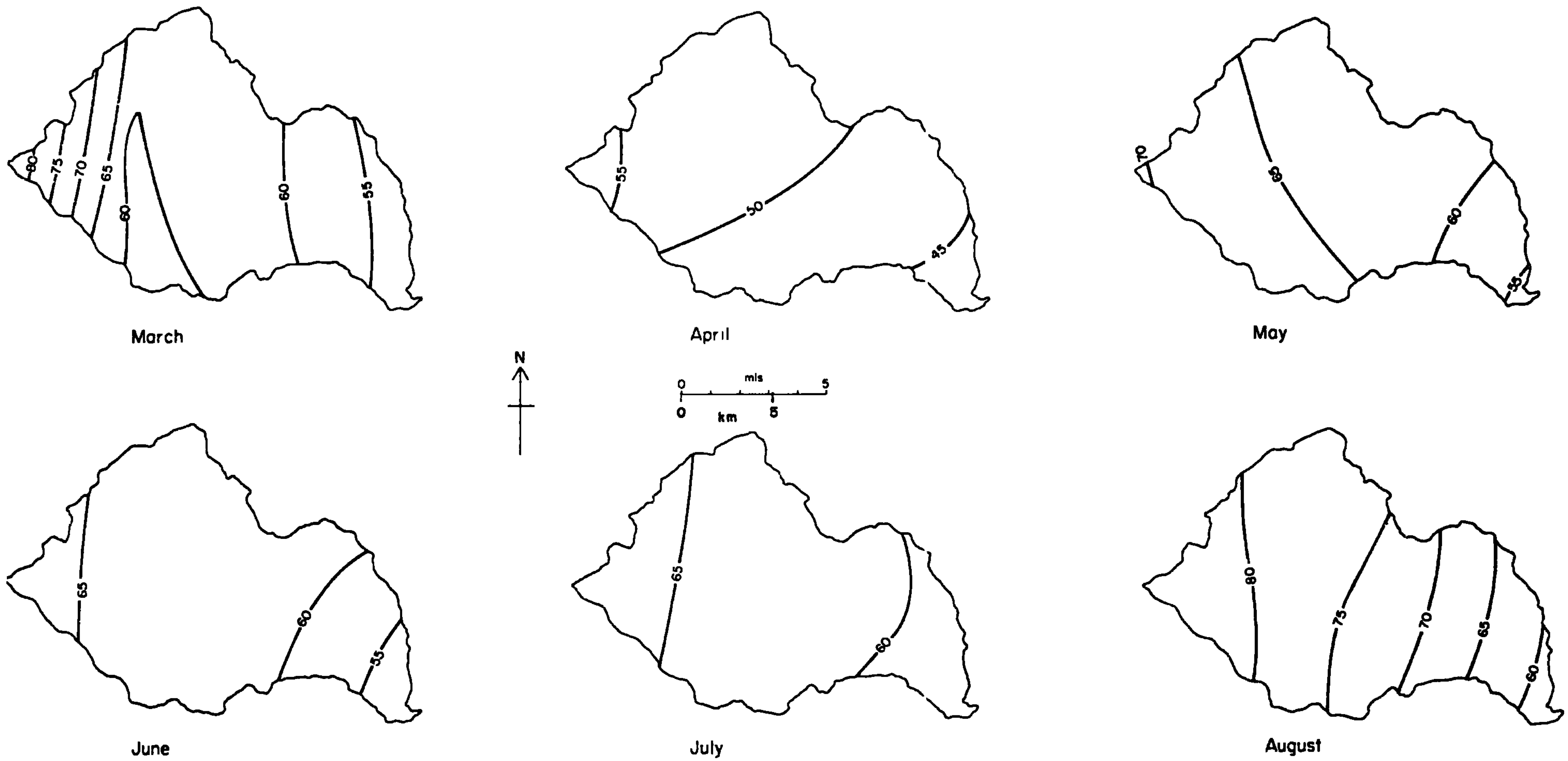


Fig 15

ISOHYETAL MAPS OF AVERAGE MONTHLY RAINFALL (mm) IN THE BROWNEY BASIN DURING 1968-1972



This procedure was followed and the mean precipitation of the catchment was determined by this method These values are given in Table 12 with those from isohyetal and Thiessen method This table shows that the mean precipitation from the rainfall-altitude relationship over the five year period exceeds both those of isohyetal and Thiessen method In fact each of the yearly values from the precipitation-altitude relationship (except that of year 1972) exceeds those of the isohyetal and Thiessen methods. However the difference in the five year average values of isohyetal and altitudinal method is about 1 6 per cent and that of Thiessen and altitudinal methods is 2.5 per cent.

Table 12 Yearly mean areal precipitation in the Browney basin during the period 1968-1972

Year	Isohyetal mm	Thiessen mm	Altitudinal mm
1968	835.5	829 0	850.8
1969	872.8	872 7	896.2
1970	669.1	653 6	680 0
1971	628 8	624.0	642.2
1972	676.2	667 5	672.5
Average	736.5	729.4	748.4

Correlation coefficients for the relationship between precipitation and altitude range from 0 85 in 1972 to 0 99 in 1970 (Table 13). The correlation coefficient of the average yearly rainfall and altitude is 0 97. The coefficients of determination range from 0 71 to 0.98 and that of average yearly is 0.95. Based on the average yearly value of the coefficient of determination, it can be stated that 95 per cent of the variance is explained by the regression equation

Table 13 Correlation coefficients between precipitation and altitude and coefficients of determination

Year	Correlation coefficients	Coefficients of determination
1968	0.97	0 94
1969	0.96	0.91
1970	0.99	0.98
1971	0 89	0 79
1972	0.85	0.72
Average	0 97	0.95

Long term means of annual and monthly precipitation

To study the long term means of annual and monthly precipitation for the Browney catchment, the Durham Observatory records were used The mean yearly and mean monthly precipitation values over the 35 year period 1939-1973 were calculated. The choice of a 35 year period was first made by Edward Bruckner (Tannehill,1955) and is now also used by the Meteorological Office. The period 1939-1973 was selected in order to include the latest available record in the study

The mean annual precipitation over the 35 year period was 639.2 mm. This value is lower than the 83 year mean of 645 mm at the Durham Observatory (Smith,1970) by less than one per cent. The highest annual value during the 83 year period was 886 mm in 1930 and the lowest value was 440 mm in 1959. For the 35 year cycle considered the minimum was the same, but the maximum value of 813 mm was 5.6 per cent lower than the highest of the 83 year period.

The distribution of mean monthly precipitation in millimetres and as a percentage of the annual total during the 35 years period is given in Table 14. October is the month with the least amount of precipitation, while February has the highest amount of precipitation. The mean monthly precipitation for these two months are 36.2 mm and 65.7 mm respectively. These results vary from those reported by Smith (1970), who from the records of 1881 to 1950, stated that February or March was the driest month

Table 14 Mean monthly values of precipitation during the period 1939-1973 at Durham Observatory

Month / Depth	J	F	M	A	M	J	J	A	S	O	N	D	Total
mm	58.6	65.7	51.9	56.0	61.9	48.7	59.1	54.1	43.2	36.2	56.5	47.3	639.2
%	9.2	10.3	8.1	8.8	9.7	7.6	9.2	8.5	6.8	5.7	8.8	7.3	100.0

On a seasonal basis, winter (December, January and February) with a total of 171.6 mm is the wettest, while autumn (September, October and November) with a total of 135.9 mm is the driest. The spring (March, April and May) total is 169.8 and the summer (June, July and August) total is 161.9 mm.

Precipitation frequency studies

The frequency of precipitation refers to the occurrence of a given depth of rainfall during a defined period. Frequency studies are important in various fields such as the design of sewage and drainage works, the design of the flood control structures, storage reservoirs, farm terraces, highway and railway culverts (Foster, 1949). Due to this importance a frequency study of rainfall records during the 35 year period of 1939 to 1973 was made.

Annual precipitation frequency. For the study of the frequency of annual precipitation, the 35 yearly values were arranged in descending order and they were plotted in Fig 16. The return period defined as the average time interval within which a given rainfall might be equalled or exceeded once was calculated after Chow (1964), using the formula

$Tr = \frac{n}{m}$ in which

n - is the number of years of record and

m - is the rank of the annual rainfall according to its magnitude

From this graph it is observed that annual values of 630 mm, 752 mm,

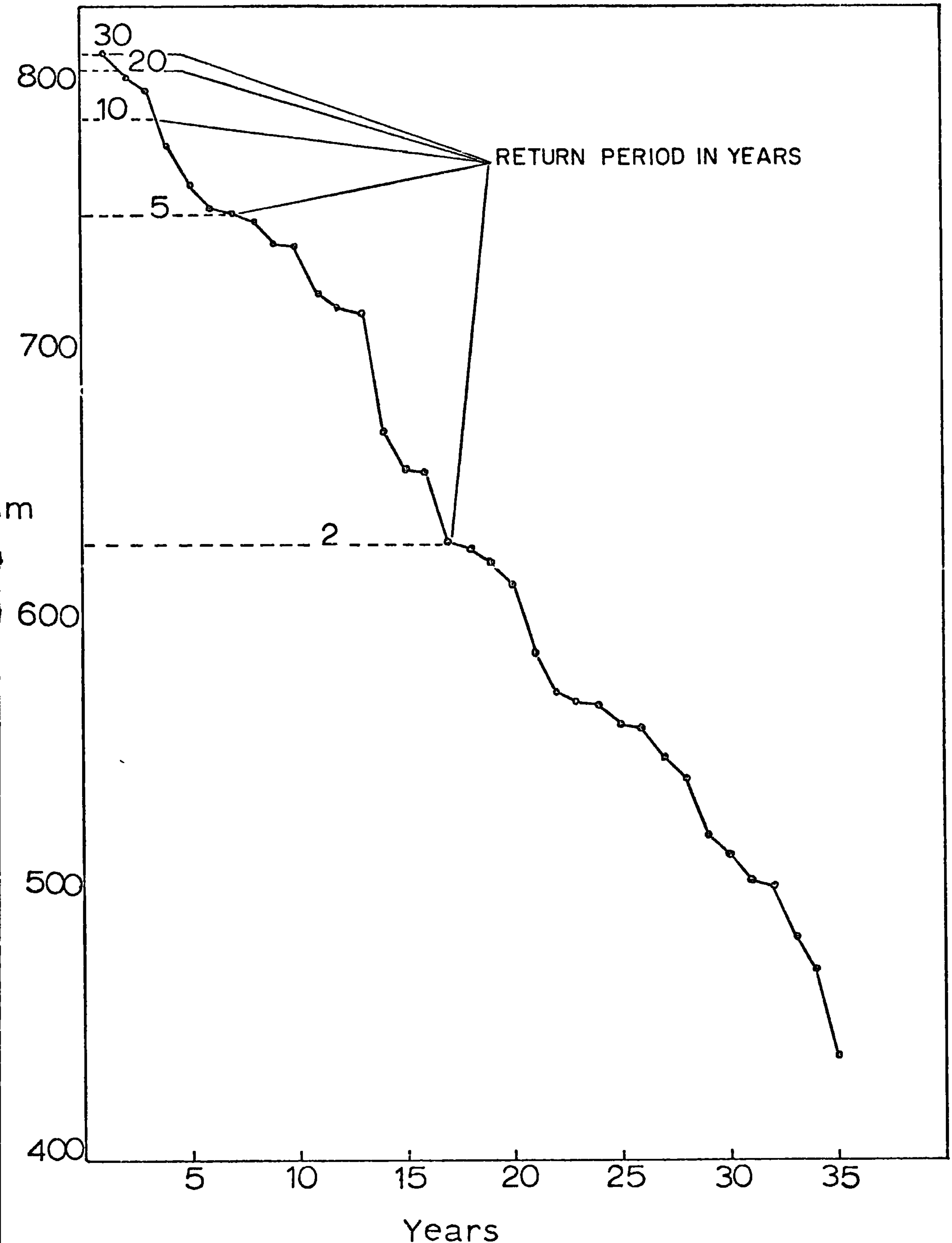


Fig 16 Frequency curve of annual precipitation at Durham Observatory (1939 - 73)

788 mm, 806 mm and 812 mm have got return periods of 2, 5, 10, 20 and 30 years respectively In other words every two years, there is a chance of getting 630 mm or more rainfall and every 30 years, there is a chance of getting 812 mm or above

<u>Frequency curves of monthly values</u> The plot of frequency of occurrence or exceedance of monthly precipitation is shown in Fig 17 The curves of the months of each season are shown in one figure.

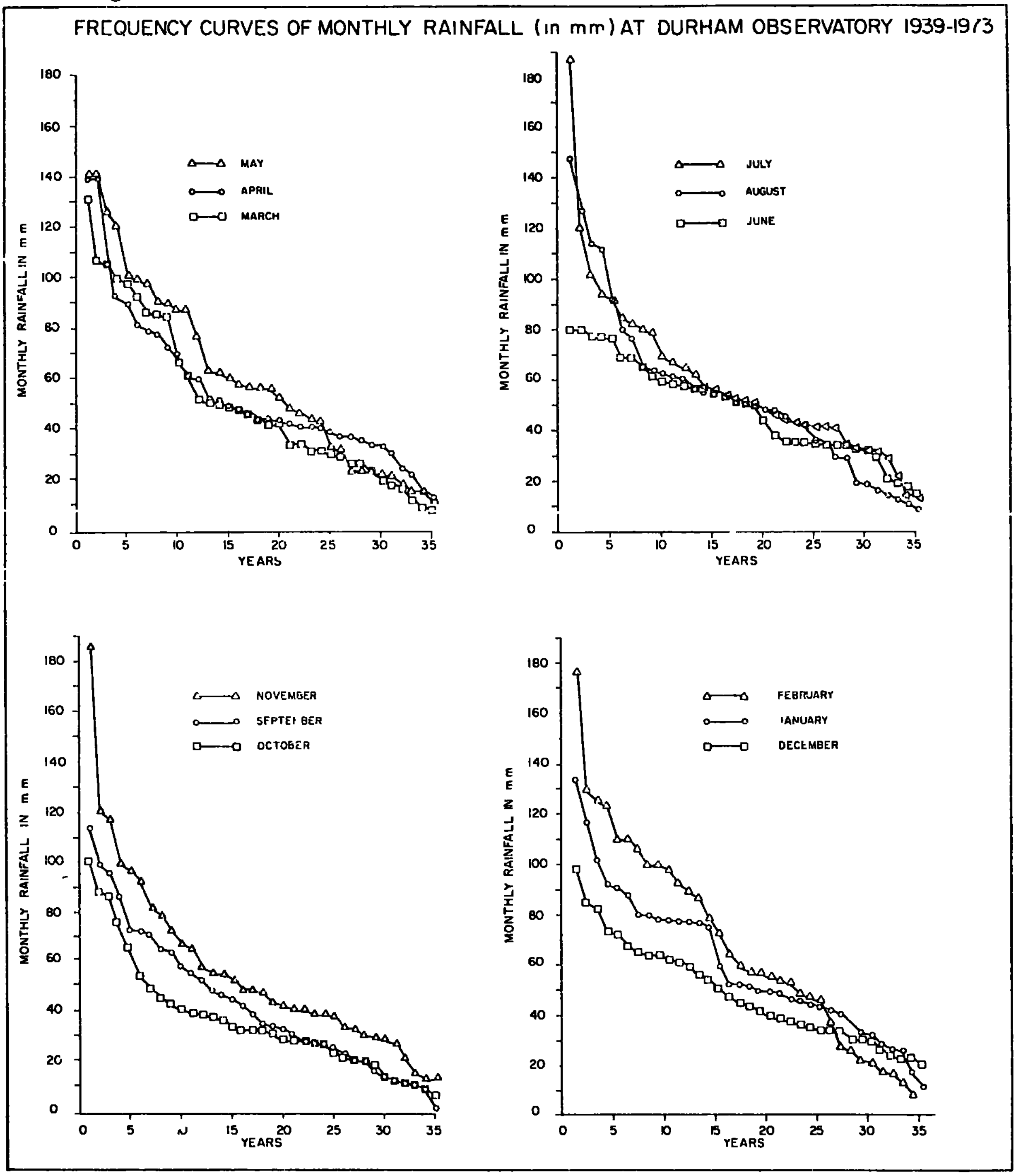
During the summer season, within the 35 year period, July has got the maximum monthly value of 187 mm, while, the maximum value of August is 148 mm and that of June is only 80 mm July values are in general higher than August and August values are more than June June values, however exceed those of July during the two lowest years (Fig.17) They also exceed the August values during the nine lowest years The minimum monthly values during the period have been 15 mm, 13.5 mm and 8 5 mm for June, July and August respectively

During autumn, November is the month with values which are consistently higher than October and September in all the 35 years The highest monthly value in November is 186 mm as compared with 113 5 mm for September and 100 mm for October September is also a wetter month than October and only in two years have the October values exceeded those of September. The lowest monthly values during these months were 14 mm, 7 mm and 2 mm respectively

Of the winter months, February has got values exceeding January and December in 25 and 26 years out of the 35 respectively The values of January are consistently higher than December except for the lowest two The maximum monthly value for February is 173 5 mm, while that of January is 133 5 mm and that of December is 97 mm. The minimum values are 19 5 mm for December, 10.5 mm for January and 6 0 mm for February

During the spring months of March, April and May the three curves are more or less similar, though the values of May generally tend to

Fig 17


FREQUENCY CURVES OF MONTHLY RAINFALL (in mm) AT DURHAM OBSERVATORY 1939-1973
MAY
APRIL
MARCH
JULY
AUGUST
JUNE
NOVEMBER
SEPTEMBER
OCTOBER
FEBRUARY
JANUARY
DECEMBER
MONTHLY RAINFALL IN mm
YEARS

exceed April, and those of April are in most cases higher than March. The highest values are 131 5 mm, 140 mm and 142 mm respectively for the three months of March, April and May, and the lowest are 9 mm, 12 mm and 14 mm respectively for the same months

For a comparison of frequencies among all the twelve months, a frequency table (Table 15) was prepared showing the number of times the value of each month equals or exceeds a given limit From studying this table, the following conclusions were drawn

Table 15 Frequency of monthly precipitation amounts equalled or exceeded during the period 1939-1973 at Durham Observatory

Month / Amount	J	F	M	A	M	J	J	A	S	O	N	D
180 mm	-	-	-	-	-	-	1	-	-	-	1	-
160 mm	-	1	-	-	-	-	1	-	-	-	1	-
140 mm	-	1	-	2	2	-	1	1	-	-	1	-
120 mm	1	4	1	2	4	-	2	2	-	-	2	-
100 mm	3	7	4	3	6	-	4	3	1	1	3	-
80 mm	6	13	9	7	11	2	8	6	4	3	7	3
60 mm	14	15	11	12	15	9	13	12	9	5	11	12
40 mm	25	25	20	24	24	20	27	24	16	10	22	18
20 mm	33	30	30	33	32	32	33	28	22	22	32	34
0 mm	35	35	35	35	35	35	35	35	35	35	35	35

1. July and November are the two months with the highest values. These are 187 mm and 186 mm respectively.

2. October, September and June are three months with the frequency of their monthly values generally being lower than the other months for a given limit.

3 The three spring months of March, April and May have a higher frequency of monthly values exceeding 100 mm In fact the total number of monthly values exceeding 100 mm during spring is 13 times compared with 10 for winter, seven for summer and five for the autumn months

The results of the frequency analysis of monthly precipitation given in Table 15 can be used to estimate the frequency of different soil moisture contents and to predict the possibilities of physiological drought or soil moisture deficit. This is because the monthly values of evapotranspiration do not change appreciably from year to year, and thus by using the monthly rainfall data and evapotranspiration, some indication on the frequency of the varying amounts of soil moisture can be obtained. Such information, thus, can be used in feasibility studies for the construction of structures such as irrigation systems or storage reservoirs

Similarly the data in Table 15 can give some estimate of the frequency of runoff during each month by using evapotranspiration data.

Annual 24-hour maximum precipitation and daily amounts In order to study the annual 24-hr maximum precipitation two methods were considered These were (1) annual exceedance series and (2) annual maximum series. In the annual exceedance series, a number of 24-hr values of precipitation with magnitudes greater than a certain base value are selected The base value is selected such that the number of daily rainfall amounts equals the number of years of record

In the annual maximum series, however, the largest value from each year is selected. There are advantages and limitations associated with each of these methods. These are discussed in detail by Chow (1964). For practical purposes, however, Chow (1964) mentions that the two series do not differ much except in the values of low magnitude.

The relationship between the recurrence intervals of the annual maximum series and the annual exceedance series has been derived by Chow (1950) and is given by the formula

$$Te = \frac{1}{\ln Tm - \ln (Tm - 1)}$$ in which

Tm and Te are the recurrence intervals of the annual maximum and the annual exceedance series respectively

Table 16 The annual 24-hr maximum rainfall at Durham Observatory during the period 1939-1973

Rank	Annual 24-hr maximum mm	Return period	Month and year of occurrence	Rank	Annual 24-hr maximum mm	Return period	Month and year of occurrence
1	55 6	36 0	Sept. 1948	19	28 2	1 9	Dec. 1973
2	50 3	18.0	Aug. 1971	20	27 7	1.8	Oct 1966
3	50 0	12 0	Feb 1941	21	27.7	1 7	March 1964
4	46 4	9.0	Aug. 1953	22	27.2	1 64	Oct. 1949
5	43.9	7 2	July 1961	23	26 7	1.56	April 1947
6	40.2	6.0	Nov 1951	24	26 3	1.50	Aug. 1956
7	39 4	5.1	Nov. 1950	25	25.2	1.44	Aug. 1943
8	39.1	4.5	Oct. 1967	26	24.3	1 38	May 1958
9	37.4	4 0	Oct. 1945	27	24.1	1 33	Sept. 1957
10	37.3	3.6	July 1965	28	23.9	1 29	Jan. 1972
11	36.0	3 3	Oct. 1939	29	23 9	1.24	June 1969
12	35 6	3 0	Aug. 1960	30	23.3	1.20	July 1940
13	35.2	2.8	May 1954	31	23 2	1 16	June 1955
14	33 8	2 6	July 1962	32	22.4	1.12	April 1944
15	33.7	2 4	July 1973	33	21.3	1 09	Sept. 1952
16	33 3	2.25	Aug. 1970	34	18.3	1 06	July 1959
17	32.0	2 12	March 1964	35	16 0	1 03	Sept. 1942
18	29.8	2 00	July 1946				

In this study the annual maximum series was used. For this purpose, the highest value for each year was found from the records. These values were then arranged in decreasing order and the return period for each value was calculated The calculation of the return period was by the formula

$Tr = \frac{n + 1}{m}$, n being the number of years of record and m being the rank of the rainfall according to its magnitude. The frequency curve and the probability plot of the annual 24-hr maximum rainfall are shown in Figs. 18 and 19 and the numerical values of these maxima with their corresponding return period is shown in Table 16 The probability was calculated by the formula $P = {}^{1}/_{Tr}$ in which P is probability and Tr is the return period

It is observed that the highest value of annual 24-hr maximum rainfall during the 35 year period is 55.6 mm and the lowest is 15 00 mm Values of 29 8 mm, 39 2 mm, 47 8 mm and 51 4 mm have got return periods of 2, 5, 10 and 20 years respectively

Considering the months of occurrences of these annual daily maxima, it is noticed that in more than one third of the 35 year period, the annual daily maximum has occurred in July and August October and September have been the next two months with highest frequencies of maximum 24-hr precipitation (Table 17).

Table 17 Frequency of occurrence of annual 24-hr maximum rainfall in each month at Durham Observatory, 1939-1973

Month	J	F	M	A	M	J	J	A	S	O	N	D
Frequency	1	1	2	2	2	2	7	6	4	5	2	1

The fact that annual 24-hr maximum rainfall values have a higher frequency during the summer can also be proved by referring to the results of a study of daily rainfall amounts recorded at Durham Observatory

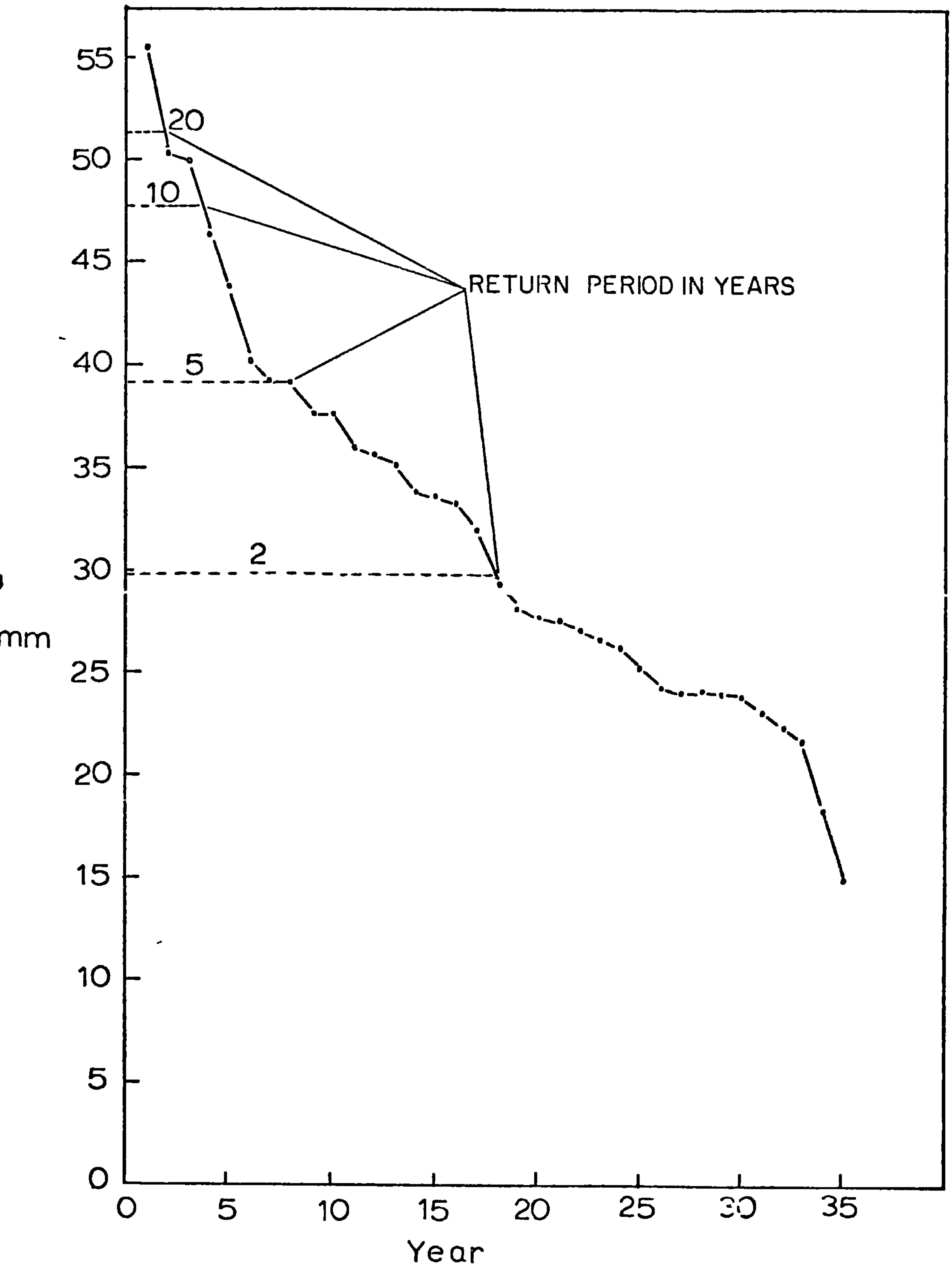


Fig 18 Frequency of occurrence of annual 24-hour maximum precipitation at Durham observatory (1939 - 73)

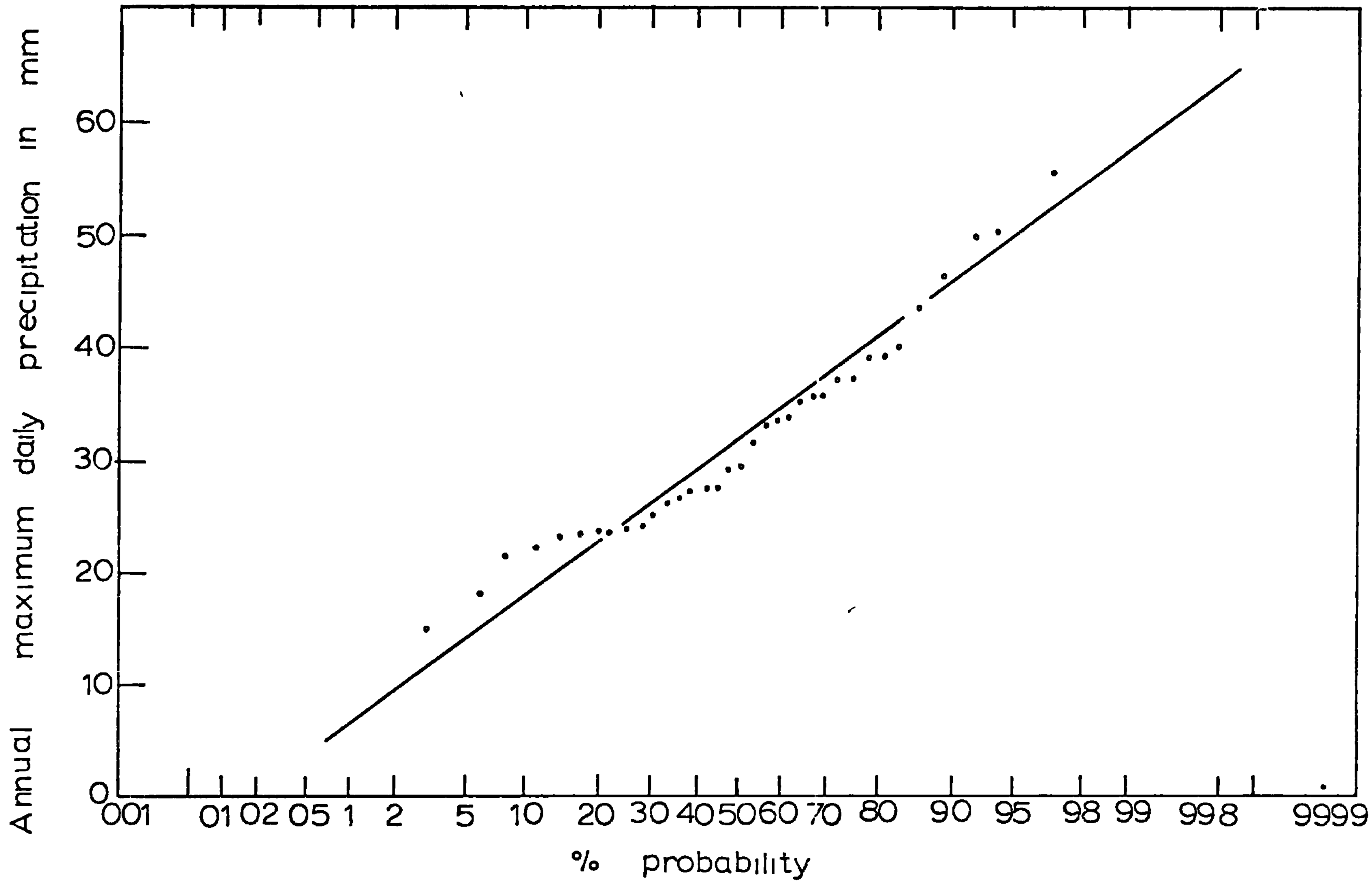


Fig 19 Annual maximum daily precipitation at Durham observatory (1939 – 1973)

during the period 1916 to 1950 (Smith,1970) From this study, Smith concluded that there was the tendency for the heavy falls to be concentrated in the summer season He further confirmed his results by referring to the occurrence of 17 noteworthy storms lasting two hours or less between 1865 and 1960. Ten out of 17 of these rainstorms which were classified by British rainfall as rare, were during July and August and none occurred later than October or earlier than May. Thus all these results give evidence for the occurrence of convective precipitation during the summer

Daily rainfall amounts during the period 1939-1973 were also studied by grouping them into four classes e.g greater than or equal to 0 2 mm, 1 mm, 2 mm, 5 mm and 10 mm A frequency distribution table for each month of the year was prepared The average frequency of days with precipitation greater than or equal to the indicated values during each month over the 35 year period was expressed as the percentage of days per month and from these results Fig. 20 was drawn

Table 18 Frequency of daily rainfall amounts equalled or exceeded during the period 1916-1950 at Durham Observatory, after Smith (1970)

Month / Amount	J	F	M	A	M	J	J	A	S	O	N	D	Year
63 mm	0	0	0	0	1	0	0	0	0	0	0	0	1
51 mm	0	1	0	0	6	0	0	0	6	0	0	0	13
38 mm	0	6	0	0	11	6	5	5	16	1	1	0	50
25 mm	6	12	0	2	18	20	32	25	32	15	11	1	174
13 mm	75	55	52	42	86	81	162	187	168	99	96	66	1179

This figure shows that November is the month with the highest percentage frequencies of daily rainfall equal to or greater than 0 2 mm, 1 mm, 2 mm and 5 mm These percentage frequency values are 57 0, 40 0 , 29 5, and 15 5, for the respective daily amounts The lowest percentage frequency values are 43 0 (April), 27 0 (March), 18 0 (April) and 7 3 (March) for the above mentioned daily amounts

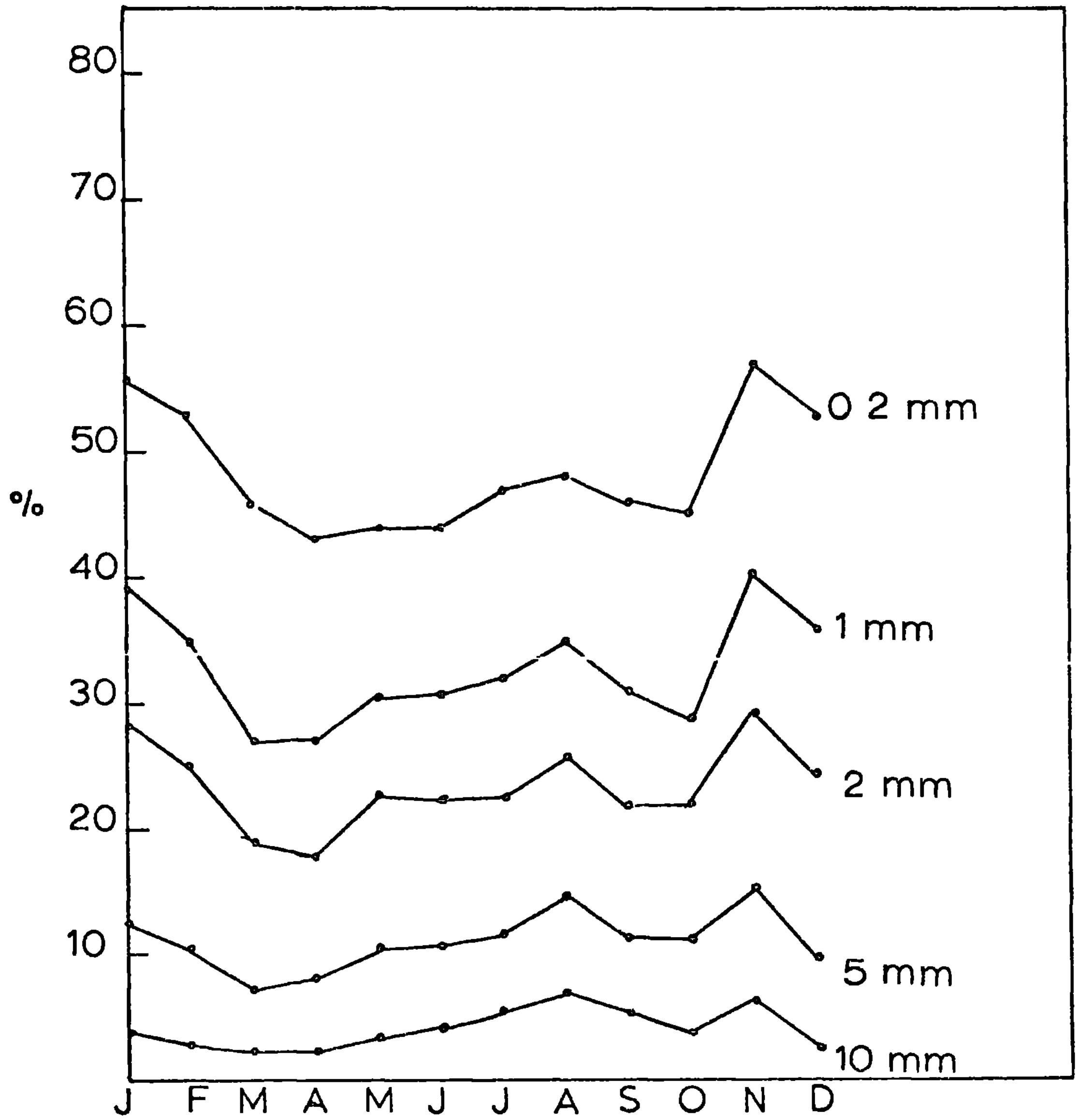


Fig 20 Average frequency (% per month) of daily rainfall equal to or greater than the indicated amount at Durham observatory (1939 - 73)

For the percentage of daily rainfall greater than or equal to 10 mm, August has got the highest value of 7 0 per cent and April has got the lowest value of 2 4 per cent.

The points that can be mentioned about this figure are

1 The per cent frequencies of low daily amounts (0 2 mm, 1 mm and 2 mm) during the winter season are higher than those during the summer months

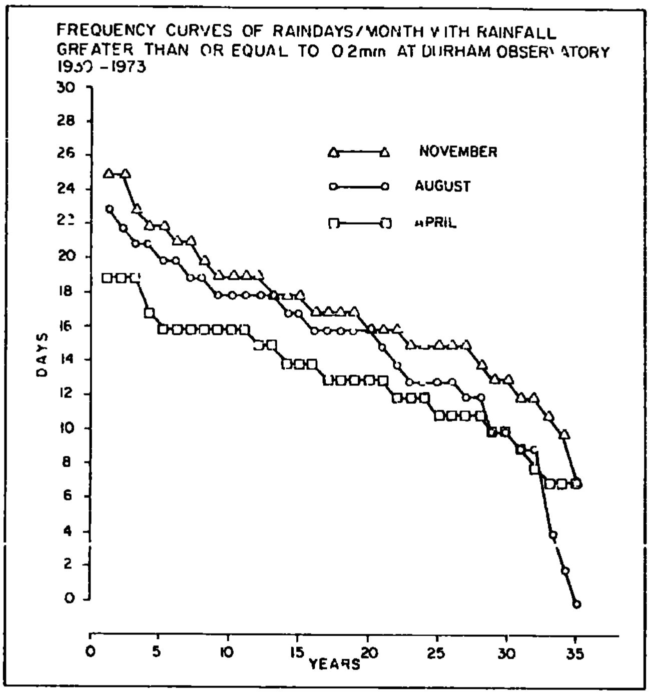
2 With an increase in the daily rainfall amount, the gap between the percentage frequencies of the summer and winter months, which is observed for low daily amounts decreases In fact for high daily amounts of 10 mm or more, August has got the highest per cent frequency, and the frequencies of other summer months are in general higher than during the winter months.

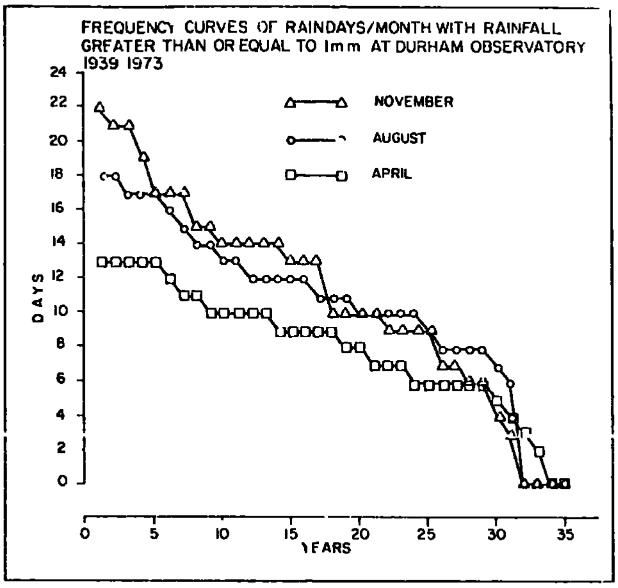
For a closer study of the occurrence or exceedance of days with different rainfall amounts, the frequency curves of the months of August, November and April were plotted (Figs. 21 and 22) The choice of these months was based on the average percentage frequency curves discussed in the preceding paragraphs. August was, thus, considered because it had the highest percentage frequencies of daily amounts among the summer months. November was considered because it had the highest percentage frequencies of daily amounts among the winter months and April was chosen because it had the lowest percentage frequencies of all months for all the daily amounts except one (5 mm)

The figures show that the frequency curves of April are below those of Novembe and August for all the daily amounts. Comparing the curves of August and November, however, it can be observed that for rainfall amounts up to 5 mm, the frequency curves of November are in most cases above those of August For a daily amount of 5 mm or more, the frequency curve of August tends to exceed that of November in a majority of years.

The range of values of occurrence or exceedance of days with pre-

Fig 21


FREQUENCY CURVES OF RAINDAYS/MONTH WITH RAINFALL GREATER THAN OR EQUAL TO 0 2mm AT DURHAM OBSERVATORY 1939 -1973
NOVEMBER
AUGUST
APRIL
DAYS
YEARS


FREQUENCY CURVES OF RAINDAYS/MONTH WITH RAINFALL GREATER THAN OR EQUAL TO 1mm AT DURHAM OBSERVATORY 1939 1973
NOVEMBER
AUGUST
APRIL
DAYS
YEARS

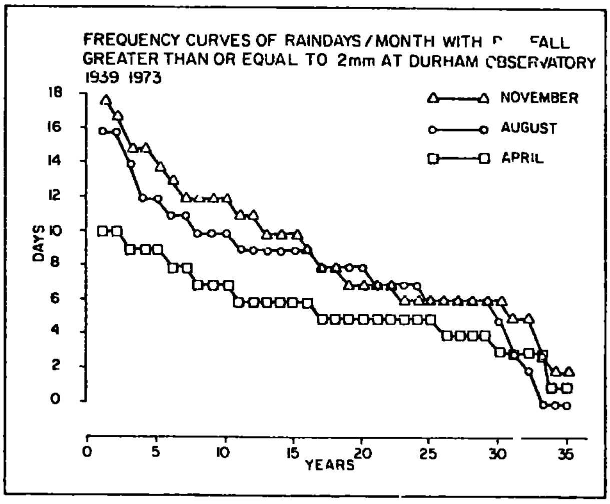
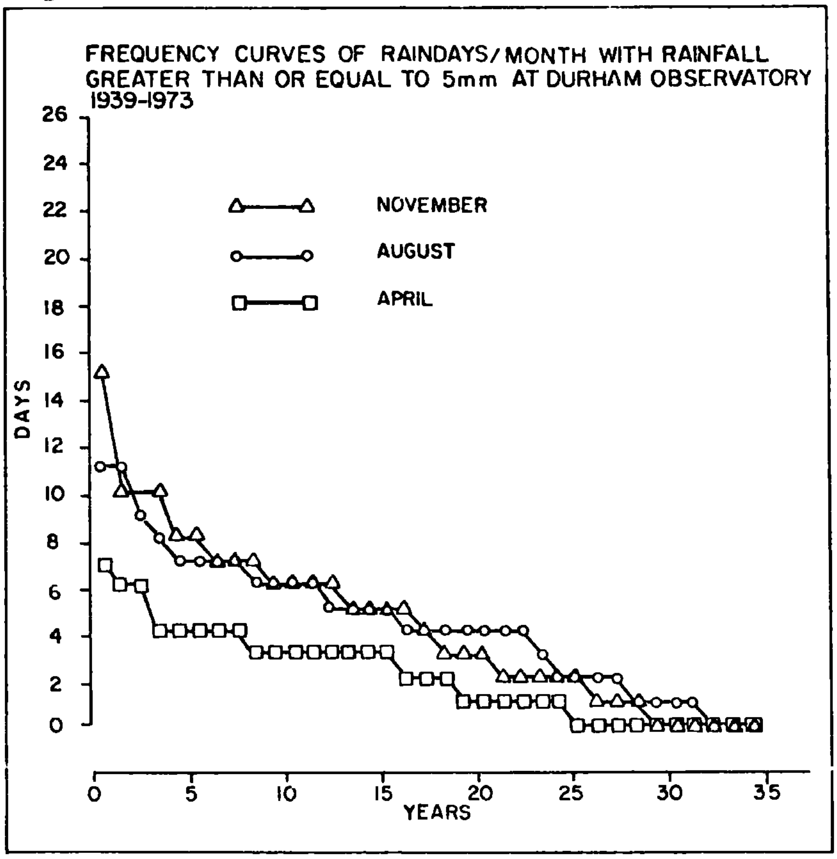
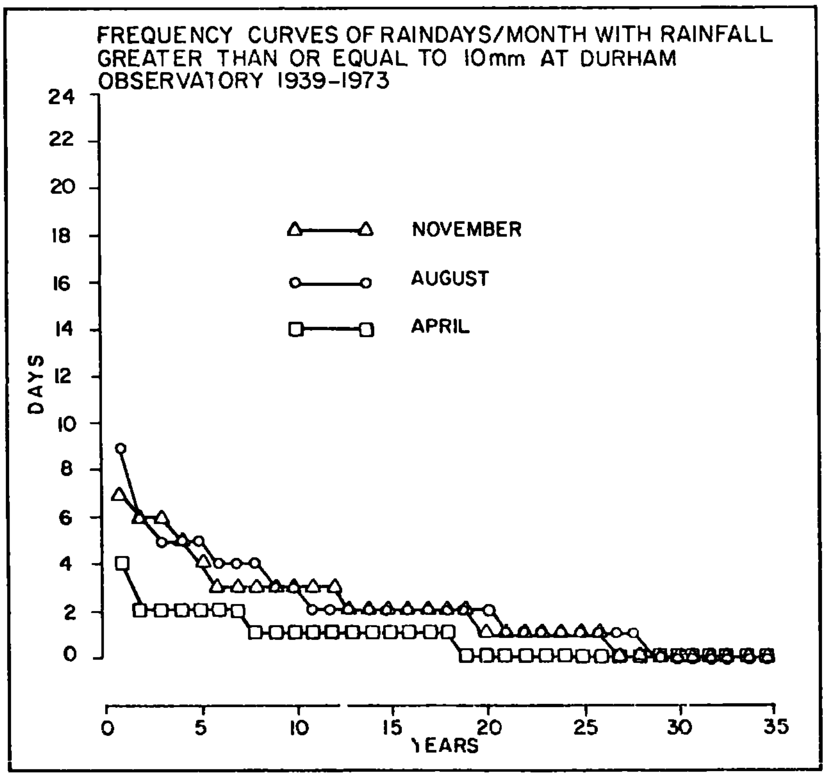

FREQUENCY CURVES OF RAINDAYS/MONTH WITH RAINFALL GREATER THAN OR EQUAL TO 2mm AT DURHAM OBSERVATORY 1939 1973
NOVEMBER
AUGUST
APRIL
DAYS
YEARS

Fig 22


FREQUENCY CURVES OF RAINDAYS/MONTH WITH RAINFALL GREATER THAN OR EQUAL TO 5mm AT DURHAM OBSERVATORY 1939-1973
NOVEMBER
AUGUST
APRIL
DAYS
YEARS
0
2
4
6
8
10
12
14
16
18
20
22
24
26
0
5
10
15
20
25
30
35


FREQUENCY CURVES OF RAINDAYS/MONTH WITH RAINFALL GREATER THAN OR EQUAL TO 10mm AT DURHAM OBSERVATORY 1939-1973
NOVEMBER
AUGUST
APRIL
DAYS
YEARS
0
2
4
6
8
10
12
14
16
18
20
22
24
0
5
10
15
20
25
30
35

cipitation greater than or equal to 0 2 mm during the 35 year period has been 7 to 25 for November, 0 to 23 for August and 7 to 19 for April The number of days with daily precipitation equal to or greater than 10 mm during the 35 year period ranges from 0 to 7 in November, 0 to 9 in August and 0 to 4 in April

An application of the study of frequencies of daily precipitation amounts is in the estimation of soil moisture content and runoff This is because the daily falls of less than 2 mm are probably not important in increasing soil moisture content, because much of the water will be lost as evaporation before reaching the soil surface. On the other hand, daily falls of more than 10 mm might exceed the average infiltration rate of the soil and thus result in immediate surface runoff. The minimum infiltration rate of the soils of the area is given by Edmonds et al (1970) as about 2 8 mm/hr Therefore it is only the rainfall with moderate intensity that can increase the soil moisture content

Hourly rainfall In the preceding section it was stated that daily rainfall over 10 mm might result in immediate surface runoff This in fact depends to a large extent on hourly or shorter period distribution of the rainfall If, for example, a daily fall of 30 mm falls during a period of an hour, most of that will result in direct runoff. However, if this amount of rainfall is uniformly distributed over the 24-hr period, and assuming a depleted soil moisture reservoir, almost all of it might enter the soil and thus increase the soil moisture content.

Intensity of rainfall expressed in mm or inches per hour is thus a very important aspect of the hydrology of the catchment For this reason an attempt was made to study the rainfall intensities within the Browney basin The available records from the Durham Observatory recording gauge were studied These records date back to 1963. However some of the charts for the period 1963-1969 were missing or were incorrect due to the malfunctioning of the pen mechanism For this reason, the rainfall

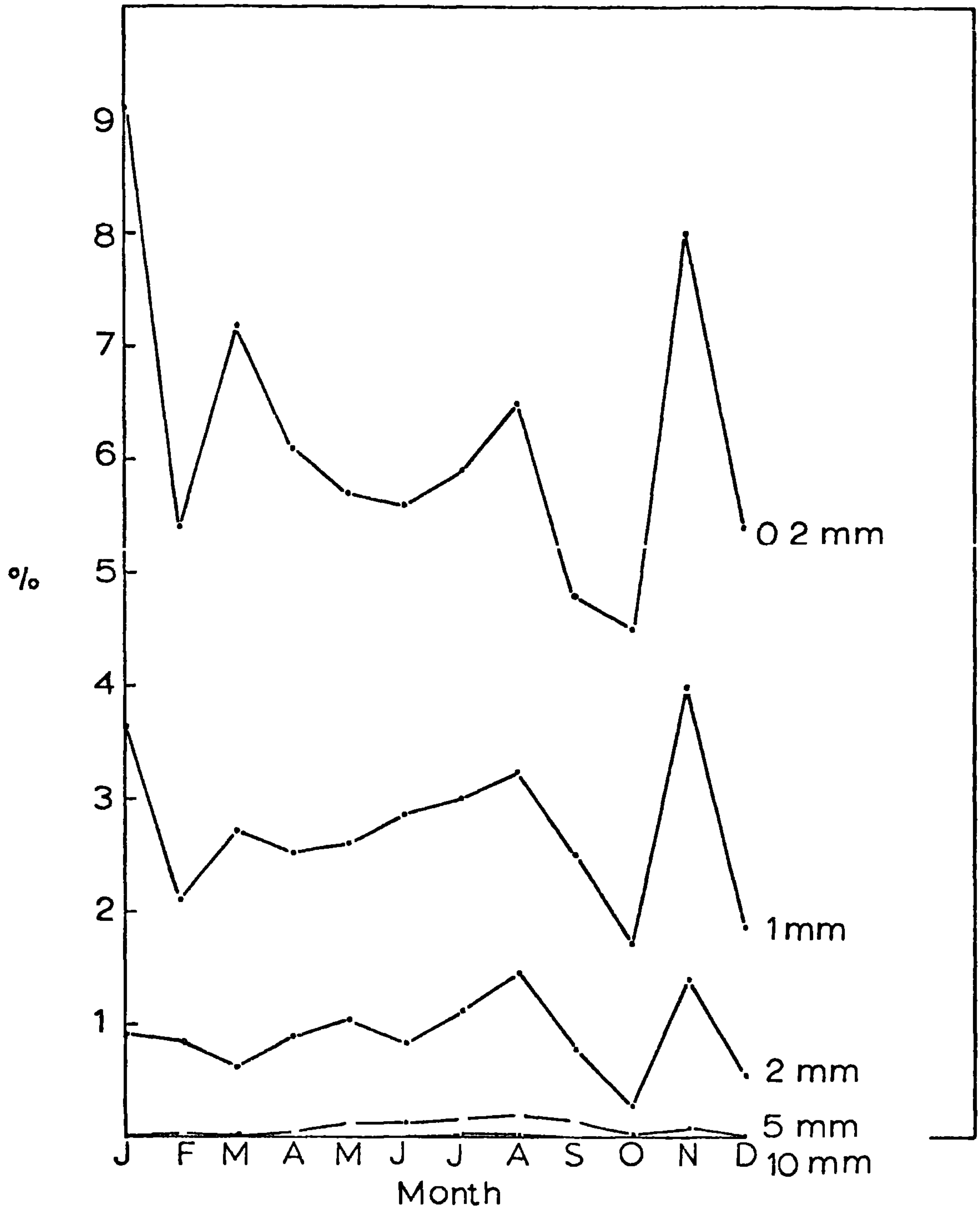


Fig 23 Average frequency (% per month) of hours with precipitation equal to or greater than indicated amount at Durharn Observatory from 1969 to 1974

intensity study was limited to the period October 1969 to September 1974, a period of six years The hourly intensities were read from the charts A frequency distribution table of hourly intensities equal to or greater than 0.2 mm, 1 mm, 2 mm, 5 mm and 10 mm was prepared. The frequency of hours with precipitation greater than or equal to the indicated value during each month was expressed as a percentage of the total hours during the month and from these results Fig 23 was drawn From this figure, several points can be made

1 In general the winter months have got a higher percentage of low intensity hourly rainfall January with a percentage frequency value of 9.1 for hourly rainfall of 0 2 mm or more is the highest of all months. October on the other hand has got the lowest value of 4.5 per cent.

2 For the hourly rainfall intensities equal to or greater than 2 mm, 5 mm and 10 mm, August exceeds all the other months. During August the frequencies of hourly rainfall greater than or equal to 2 mm, 5 mm and 10 mm are 1 45, 0 40 and 0 03 per cent

Thus from this intensity study it can be observed that high intensity rainfall and in particular intensities equal to or greater than 10 mm/hr are very low in frequency The maximum intensity value for each month during the six year period of study is given in Table 19

Table 19 Maximum monthly intensities during the six year period, 1969-1974 at Durham Observatory

Month	J	F	M	A	M	J	J	A	S	O	N	D
Maximum Intensity in mm/hr	4 8	5 1	3.8	6.1	8 6	6 6	20 3	17.3	7.1	11 2	6 1	4 1

Thus the highest hourly intensity occurred in July with a value of 20 3 mm The next highest value is in August with a value of 17.3 mm. Considering these intensity values and using the definition of excessive precipitation referred to by Gilman (1964), it can be stated that there has been only one incidence of excessive hourly rainfall e.g. 20.3 mm in July. This is because excessive precipitation is referred to as any precipitation that falls at a rate equalling or exceeding that indicated by the formula $P = t + 20$ where P is precipitation in hundredth of an inch and t is the time in minutes By this definition,

$P/_{100} = 60\ (\text{min}) + 20 = {}^{80}/_{100} = 0\ 8$, therefore any hourly precipitation equal to or greater than 0 8 inch (20 3 mm) would be excessive

Longest run of dry days To investigate the longest run of dry days, a procedure similar to the other frequency studies was adopted The longest run of days for each month during the 35 year period was considered. These were arranged in decreasing order Considering the distribution of longest run of dry days in each month, it was found that the frequency curve of October was more or less similar to September, November and May, that of March was similar to June, July and August and that of January was similar to December, February and April, consequently only the frequency curves of March, October and January were drawn (Fig 24) From the study of frequency curves of the months of March, October, and January it was found that the longest run of dry days during the 35 year period was 30, and that this occurred in March 1953 In fact this length of run of dry days has been the longest in all the months throughout the 35 years. It has been followed by 27 days in August 1947 and 25 days in July 1955 (Table 20)

Considering the definition of absolute drought given by Meteorological Office (1963), which is "a period of at least 15 days to none of which is credited 0.01 inch (0 2 mm) of rain or more", it can be concluded that, during the 35 year period, there has been three incidences of absolute droughts in March, two absolute droughts in October, and none in January.

It should be mentioned that the mere tabulation of the number of consecutive dry days does not give much information about the occurrence

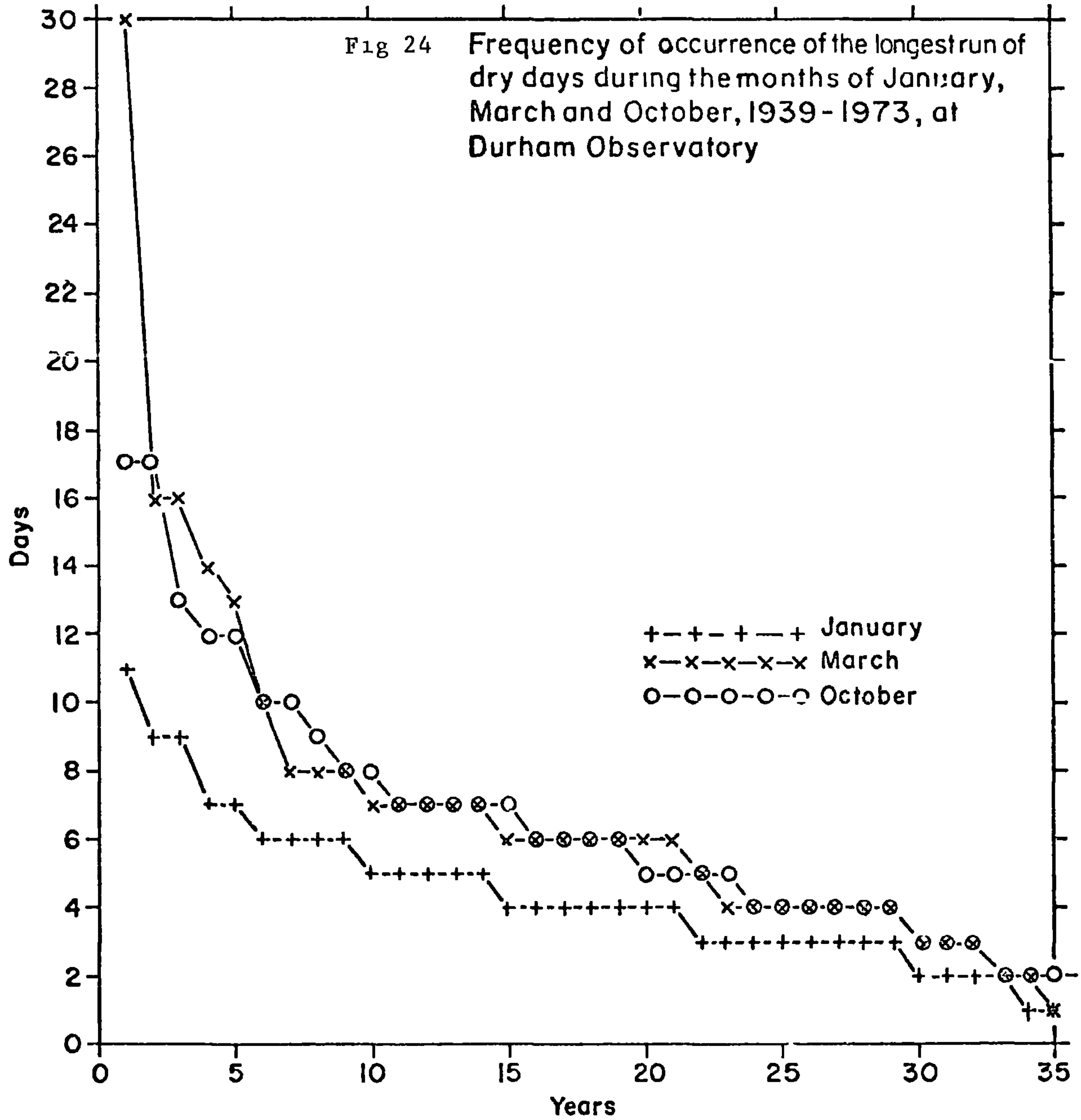


Fig 24 Frequency of occurrence of the longest run of dry days during the months of January, March and October, 1939-1973, at Durham Observatory
Days
Years
January
March
October
0
5
10
15
20
25
30
35
2
4
6
8
10
12
14
16
18
20
22
24
26
28
30

of physiological drought This is because there might be a lot of moisture available in the soil at the beginning of a dry period. Thus assuming the soil moisture to be at field capacity, it could possibly supply some 75 mm of rainfall and, therefore, prevent the occurrence of any physiological drought during the rainless days.

Table 20 Annual series of longest run of dry days/month, 1939-1973 at Durham Observatory.

Length of Period	Month and Year of Occurrence	
30	March	1953
27	August	1947
25	July	1955
21	August	1972
21	June	1970
19	May	1952
18	July	1971
17	October	1969
17	October	1951
16	March	1973
16	March	1966
15	September	1959
15	November	1957
15	June	1949
14	March	1943
14	July	1961
14	June	1960
14	May	1963
13	March	1948
13	October	1962
13	September	1941
12	July	1946
12	June	1950
12	June	1967
11	June	1964
11	May	1956
11	May	1939
11	April	1954
10	March	1945
10	October	1965
10	April	1944
9	July	1968
9	April	1942
8	October	1958
8	May	1940

CHAPTER THREE

EVAPOTRANSPIRATION - A LITERATURE REVIEW

Evapotranspiration is the name given to the combination of two physically similar processes i e. evaporation and transpiration. Evaporation is the conversion of water in liquid form to vapour from water and soil surfaces. This vapour is then transferred to the atmosphere. Transpiration is defined as evaporation from plant surfaces.

Evaporation from water, soil or plant surfaces occurs because some water molecules acquire sufficient energy to overcome the cohesive forces that bind them together Some of these molecules when transferred into the air collide with other molecules of the air, and hence they lose some energy and fall back into the liquid form. The evaporation process will thus, be continued until an equilibrium state is reached, when the number of outgoing molecules is balanced by incoming ones This state of equilibrium is called saturation and the vapour pressure (pressure created owing to the motion of water molecules) of the evaporating surface and that of overlying air would then be equal

Assuming no limitation of water, two requirements must be satisfied so that either or both of these processes could occur

1. Some source of energy must be available to be used as latent heat of vaporization, or to be more specific, 580 calories are needed to evaporate one gram of water at 25^{o}C
2. There must be some mean of transport of the vapour to the atmosphere. This is done by the turbulence of the atmosphere.

Studies of evapotranspiration are of prime importance in many desciplines of science e.g. hydrology, agriculture, meteorology and geography Design of water storage reservoirs, irrigation systems in semi-arid and arid regions, prediction of frequency of agricultural droughts in humid regions and flood studies are all dependent upon knowledge of evapotranspiration This explains the reasons for the active involvement of scientists in this field, the results of whose work have filled the literature, especially during the past four decades. However, owing to the complexity of the soil-water-plant system, there are still some shortcomings to the solutions and approximations so far reported and consequently more work needs to be undertaken.

Problems still existing in evapotranspiration studies were also recognised by our predecessors and Penman (1956) in reviewing some of their views quotes Symons (1867) saying "evaporation is the most desperate branch of desperate science of meteorology", and Cleveland Abbé (1905) mentioning, "of course we need to know the loss of water by evaporation, but in nature it is so much mixed up with seepage, leakage and consumption by plants that our meteorological data are of comparatively little importance".

Factors affecting evapotranspiration

There are three groups of factors which affect evapotranspiration These factors are

1. Climatic factors or conditions of the atmosphere i e. radiation, temperature, humidity, atmospheric pressure and wind
2. Soil factors i e. soil water content and capillary conductivity
3. Plant factors i.e degree of plant cover, rooting depth, plant height, number and arrangement of stomata and their opening and closing.

Climatic factors affecting evapotranspiration

(a) Radiation For the evaporation of 1 cm^3 of water at 25^oC, 580 calories are needed. The original source of all the energy involved in the transformation from liquid to water vapour is the sun. The amount of radiation which would reach the earth from the sun if there was no atmosphere is about 1 94 $calories/min/cm^2$ of surface normal to the direction of radiation This amount of radiation, known as the solar constant, at any point on the earth's surface, is reduced by several factors. These factors are the angle of incidence of the rays, which determines the intensity of radiation at any location, scattering of solar radiation by the constituents of atmosphere (air molecules, water vapour and dust), absorption of radiation by atmospheric clouds and gases (ozone, molecular oxygen, carbon dioxide) and reflection The difference between the total radiation arriving at a surface (global and long wave from the sky) and the total radiation leaving the surface (albedo and long wave emitted by the surface) is called net radiation Net radiation can be further divided into that used for heating the soil, a part for heating the air and the rest for evaporating water. The small amount of net radiation used in photosynthesis and stored in plants can be neglected because the error in measurement of the other terms is more than that used for metabolic activities of plants

Both global and net radiation are highly correlated with evapotranspiration, though correlation of evapotranspiration with net radiation is much higher than with global radiation (Tanner,1967)

(b) Temperature Temperature is a climatic factor which is highly correlated with evaporation However radiation and evapotranspiration, which are nearly in phase on a daily basis, may vary considerably from day to day In contrast air temperature changes slowly because of high thermal storage of the earth and the atmosphere and because of the large amounts of net radiation used for evapotranspiration (Pelton et al,1960). Thus the correlation between mean daily temperature and daily net radiation is poor. This thermal lag effect is presented in Figs 25a and 25b which show that two temperatures are associated with each solar radiation or net radiation value The effect of thermal lag is least during the period when both temperature and net radiation are maximum. This lag effect may be observed from the Fig.25c which shows that mean monthly evaporation for months having the same temperature is higher during spring and summer months than during late summer and autumn (Veihmeyer,1964).

(c) Humidity. The effect of humidity upon evapotranspiration has been

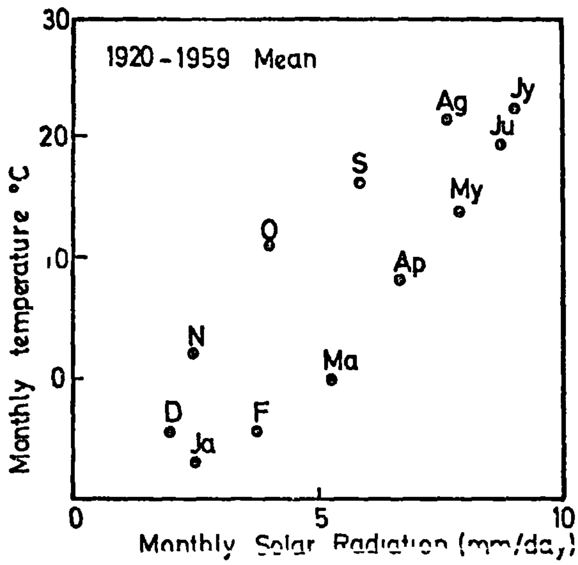


Fig 25a Mean monthly solar radiation (equivalent evapotranspiration) and mean monthly Thornthwaite PE at Madison, Wis
(From an original diagram by Pelton et al 1960)

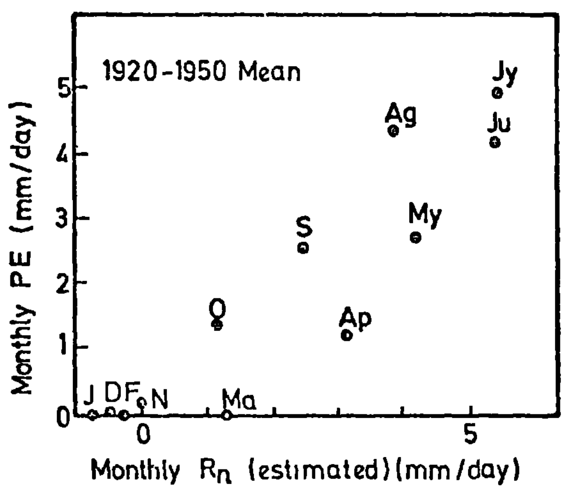


Fig25b Mean monthly net radiation (equivalent evapotranspiration) and mean monthly Thornthwaite PE at Madison, Wis
(From an original diagram by Pelton et al 1960)

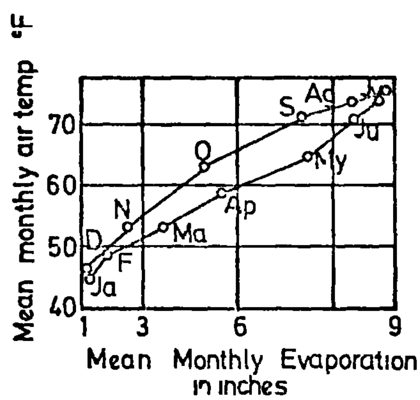


Fig 25c Evaporation from U S Weather Bureau Class A Pan at Davis, Calif averaged Pan for 1926-1959
(From an original diagram by Veihmeyer 1964)

known since 1802 when Dalton suggested his law This states that with a given condition, evaporation is proportional to the deficit in vapour pressure at the evaporating surface and the overlying air From this law, it is clear that the evaporation rate will be greater under dry air conditions than under moist ones, also that any increase in humidity decreases the amount of evaporation

(d) Wind Wind removes moist air from evaporating bodies and replaces it with dry air which is capable of holding more water Clearly the higher the velocity of the wind, the more the evaporation rate will be up to a point when other factors (e g energy) become limiting Under such conditions any increase in wind velocity does not increase evaporation.

A phenomenon which should be considered in any evapotranspiration study is advection It occurs when a pre-heated or pre-cooled air passes over a well watered field Thus, such a pre heated or pre-cooled air becomes a source or a sink for energy which results in increase or decrease of the air temperature (Fig 26) Advection might result in sizable errors when it is caused by warm air from deserts or oceans or cold air from cool ocean currents Van Wijk and De Vries (1954) show examples from Ireland, Norway, and the Netherlands in which advection errors due to warm air currents occur.

(e) Atmospheric pressure Any increase in barometric pressure is expected to decrease the evaporation rate and vice versa. This is explained by the fact that with a decrease in atmospheric pressure, the density of overlying air decreases Variation of atmospheric pressure, however, is accompanied by changes in other meteorological factors affecting evapotranspiration, so that the resultant effects are not well understood

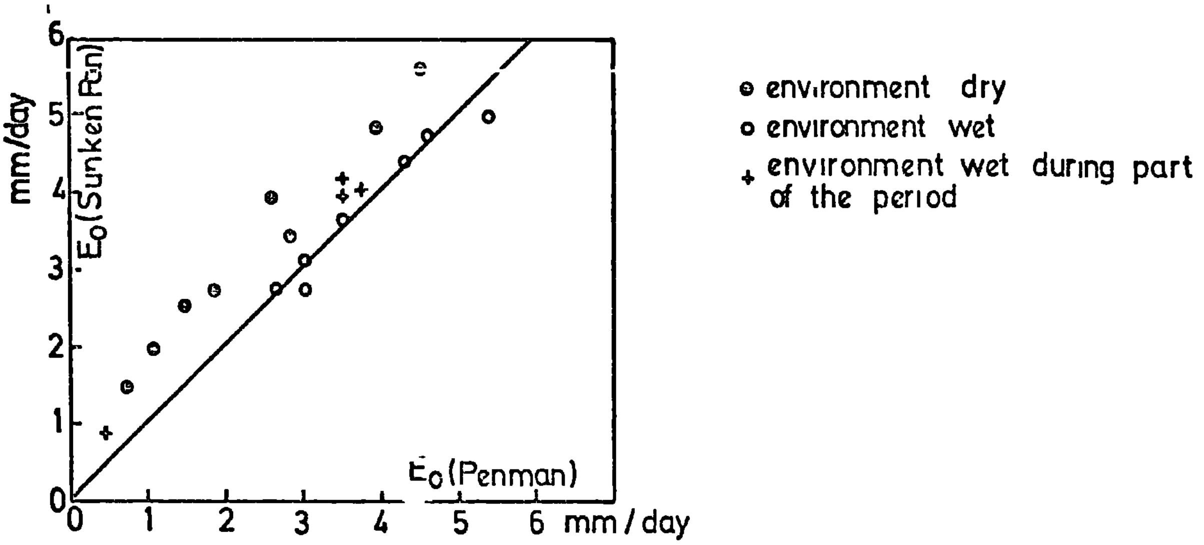


Fig 26 Relation between evaporation calculated with the modified Penman equation and data measured with sunken pans Mean values for 10 days in 1959
(From an original diagram by Rijtema 1965)

<u>Soil factors affecting evapotranspiration</u>

The soil is the source of water for evapotranspiration and so it has a definite effect on the rate at which a plant is able to transpire and a soil surface to lose water as evaporation. Two factors combine together to determine the rate at which the bare soil supplies water to the evaporative site These factors are the <u>water content</u> and <u>capillary conductivity of the soil</u> Both of these factors are texturally controlled, i.e. a coarse-textured soil like sand retains less water at any suction level than a fine-textured soil like clay. This results in a small supply of water near the surface readily available for delivery to the evaporative site for coarse textured soil than that which occurs in fine textured soils Capillary conductivity for fine textured soils is also more than that of coarse textured ones. The capillary rise for a fine textured soil may reach several feet whereas that of coarse textured soil may be only a few inches

When the water content of the soil is at field capacity or over, evapotranspiration will occur at the potential rate. At the permanent wilting point (lower range of water availability) and in the absence of any contribution of moisture due to capillary rise, there will not be any water to be transpired by the plant or used for evaporation by the soil surface. The exact nature of the relation between evapotranspiration and soil moisture variation over the range from field capacity to permanent wilting point is still a matter of controversy and apparently conflicting evidence Some of the views on this relationship will be discussed in the section on the estimation of actual evapotranspiration

<u>Plant factors affecting evapotranspiration</u>

These factors are

(a) percent cover

(b) rooting depth

(c) height

(d) stomata closure and pattern

(a) Per cent cover As the per cent cover of the soil increases, the magnitude of Et increases. This could be ascribed by reflection. Since the colour of the crop affects radiation, and since colour varies with the degree of cover, it is implied that an increase in the percentage cover increases evapotranspiration Another explanation that may be given for this direct effect of percentage cover on evapotranspiration is that with drying of the soil after rain, or irrigation, the rate of evaporation is checked and the supply of water from below cannot satisfy the evaporation rate However, under similar conditions, a dense cover of plants can supply water for transpiration for at least two weeks before there is any decrease in Et (Gates et al ,1967)

(b) Rooting depth Under conditions in which soil water is limited in the top root zone, a well extended deep rooted crop can extract moisture from the lower zone, thus satisfying the water need for evapotranspiration

(c) Plant height An increase in plant height appears to increase evapotranspiration by greater interception of advected heat (Gates et al, 1967). In a study by Rijtema (1968), it was found that the coefficient of turbulent exchange increased by a factor of 2 with a change in vegetation height from 10 cm to 90 cm and by a factor of more than 5 with a change from a short cut green surface at about 2 or 3 cm to a vegetation height of 90 cm (Ward, 1975)

(d) Stomata closure pattern Evapotranspiration from plants occurs by the movement of vapour from the intercellular spaces within the leaf to the free atmosphere by diffusion through stomata, hence any reduction or closure of the stomata results in a decrease in Et Penman (1956) considering the effects of stomata opening on Et, included a factor f in his formula which is based on stomata geometry, and population and day light. Reduction of the aperture of the stomata is caused by internal water deficit in the plant and climatic factors like light, temperature and wind

Potential versus actual evapotranspiration

Potential evapotranspiration is defined by Penman (1956) as, " evaporation from an extended surface of short green crop, actively growing, completely shading the ground, of uniform height and not short of water" By this definition the effects of soil and plant factors on evapotranspiration are removed and climatic factors are considered to be the only variables upon which evapotranspiration depends Actual evapotranspiration is however, evaporation from soil, water and plant surfaces as dictated by combination of all the three groups of factors Measurement of potential evapotranspiration under conditions defined by Penman, is relatively easy However, owing to complex soil-water-plant relationships, the measurement of actual evapotranspiration is difficult

The methods of measuring evapotranspiration may be divided into three main groups

1. Theoretical methods
2. Empirical methods
3. Water balance methods

1 Theoretical methods of estimating evapotranspiration

These methods are divided into two sub-groups, aerodynamic methods and energy balance methods The first of these two is self-contained, while the second depends on some of the principles of the first

Eddy correlation (aerodynamic) A parcel of air (eddy) with an upward movement causes an upward flux of water vapour. If the velocity of the eddy and its vapour content is measured over a period of time the summation of the products of the two terms give the upward flux of water vapour in the atmosphere The instrument for measuring eddies and their characteristics i e temperature, humidity and vertical wind velocity is called an evapotron This instrument measures fluctuations in the temperature, humidity and wind velocity of minute eddies This information may be fed directly into a computer and an output of net upward movement of water vapour for the evaporating surface obtained (Ward,1971)

Profile method (aerodynamic) In this method the vertical flux densities of water vapour, sensible heat and momentum (shearing stress) are represented by a one dimensional form of a general steady state equation i e.

$E_x = K_v \rho \delta q / \delta Z$ (1), $H = K_h C_p \rho \delta T / \delta Z$ (2)

$\tau = K_m \rho \delta U / \delta Z$ (3)

in which E_x is the evaporation rate in $gm/cm^2/sec$

K_v is the eddy transfer coefficient of water vapour in cm^2/sec

ρ is the density of air in gm/cm^3

q is the specific humidity of the air in gm/gm of moist air

Z is the elevation above the ground surface

H is the vertical flux density of sensible heat in $cal/cm^2/sec$

K_m is the eddy transfer coefficient of sensible heat in cm^2/sec

C_p is the specific heat capacity of dry air at constant pressure (0 242 $cal/gm/C^o$)

T is the temperature in C^o

τ is the vertical flux density of momentum in $dyne/cm^2$ ($gm/cm\ sec^2$)

K_m is the eddy transfer coefficient of momentum in cm^2/sec

U is the wind velocity in cm/sec

Vapour pressure and specific humidity are related by the formula

$q = \varepsilon e/P$ (ε is the ratio of the densities of water vapour and dry air, e is the vapour pressure and P is the atmospheric pressure) Since it is the vapour pressure which is usually measured in evapotranspiration studies, equation (1) results in

$E_x = (K_v \rho \varepsilon / P)(\delta e / \delta Z)$ (4)

To determine the evaporation, sensible heat and momentum, transfer co-coefficients should be evaluated Applying the equation for an increase of wind speed with height i e

$\delta U/_{\delta} Z = \sqrt{\tau/_{\rho}}/KZ$ (5) in which K is the Von Karman constant (0 41) and equation (3) we get

$$K_m = K^2 (U_2 - U_1) (Z_2 - Z_1) / (\ln Z_2/Z_1)^2 \qquad (6)$$

If we now assume that each eddy transports water vapour and sensible heat and momentum proportionally, then $K_m = K_v = K_h$. Consequently evaporation, sensible heat and momentum can be determined from equations (4), (2) and (3) respectively For evaporation the equation is then in this form

$$E_* = (K^2 \rho \varepsilon /P) (U_2 - U_1) / (\ln (Z_2/Z_1)^2 \qquad (7)$$

For the validity of equation (6) and similar ones for sensible heat and momentum, several assumptions should be made and Tanner (1967) discusses them in detail

(a) Steady state The assumption of a steady state appears to exclude transient conditions at the surface (e g vapour pressure or temperature changes brought about by radiation variation) This assumption might not cause serious error in agronomic work, though in research with sensitive instruments it will result in some error

(b) Adiabatic conditions. Surface heating and cooling produces wind profile curvature and causes changes in profile slope, consequently equation (5) would be invalid

(c) Homogeneity of surface Surface homogeneity affects the ratio of Kv/Km and Kh/Km Over heterogeneous surfaces some spots are primarily sources of heat and others may be sources of water vapour and sinks for momentum If the scale of wind eddy is the same size or smaller than the heterogeneities, heat and water vapour are separated in eddies and the individual eddies act selectively in transporting heat and water vapour as a result of the greater upward acceleration of hotter and drier eddies, thus Kv ≠ Kh ≠ Km

(d) Vertical transport restriction To restrict transport to a vertical direction, there should be no horizontal gradient of vapour, heat and momentum This is explained by the fact that as air moves from a surface of a given wetness, temperature, and roughness, to a different surface, the velocity, air temperature and vapour profile change from those representing properties of the first surface to those resulting from the properties of the second

Energy balance methods Principles of conservation of energy when applied to the evaporative surface, can help one to understand fully the energy aspects of this process The simple energy balance of a surface is expressed by

$$Rn = LE_{x} + H + S \qquad (8)$$

in which Rn is the net radiation (difference between total incoming radiation, and outgoing long wave and reflected radiation), E is the energy used for evaporation ($E = L\ E_x$, L being the latent heat of evaporization in cal/gm and E_x is the evaporation rate in gm/cm^2/time) having the unit of cal/cm^2/time, H is the amount of energy used for heating the air and S is that portion of net energy used for heating the soil (all in cal/cm^2/time) This formula neglects small energy terms such as that used in metabolic activities (photosynthesis and respiration) and storage of heat in plant tissues Rn may be measured by net radiometers or alternatively determined by using weather parameters The value of S can either be measured by soil heat flux plates, or disregarded (Penman et al, 1967) Thus net energy should be partitioned into the amount used for evaporation and that used for heating the air.

Bowen (1926) determined that for a given surface the ratio of energy partitioned into sensible heat (H), to that partitioned into evaporation is relatively constant Thus H/E = β(9) Replacing the values of H + E from equations (2) and (4) into (9) we get

$$K_h\ C_p\ P\ (T_2 - T_1)\ /K_v\ \varepsilon L\ (e_2 - e_1) = \beta \qquad (10)$$

K_h is assumed to be equal to K_v (valid for homogenous surfaces) and if $C_p\ P/\varepsilon L = \gamma$, then $(T_2 - T_1) / (e_2 - e_1) = \beta/\gamma$ (11)

from equation (8) and equation (9), then, we get,

$E = (R_{n/1+\beta})$ (12)

Hence actual evaporation can be obtained from the above equation by knowing R_n and measuring the temperature and vapour pressure at two heights Values of β over homogeneous irrigated areas range between zero and 0 2 (Penman et al,1967).

2 Empirical methods

Various empirical formulae have been suggested for the prediction of evapotranspiration The development of these equations has been based on the correlation of evapotranspiration with one or more climatic factors, and the degree of empricism of the formulae varies depending on the number of factors considered and the simplifying assumptions made The climatic factors which are mostly adopted as variables are radiation (solar and net), mean temperature, humidity and wind.

Radiation methods of evapotranspiration assessment fall into two groups (Tanner,1967) Those which are based on rational energy balance, and those which relate evapotranspiration with solar or net radiation by simple regression techniques Examples of the former method are those of Penman (1948) and McIlory (Slatyer and McIlory,1961) reported by Rijtema (1965), which in principle are not empirical but become so because of assumptions made to utilize existing weather data An example for the latter method is that of Jensen and Haise (1963). In view of the fact that solar radiation is the main source of energy for supplying the latent heat of vaporization of water, even in arid areas, good correlations between evapotranspiration and net or solar radiation are to be expected and Pelton et al,(1960) suggest that they are the best methods, if radiation data are available

Mean temperature methods can be exemplified by those of

Thornthwaite (1948) and Blaney and Criddle (1950). The widespread use of these formulae is due to their simplicity and availability of temperature and latitude data at any location and any period of time

Humidity methods are a variation of the Dalton formula in which the vapour pressure gradient is replaced by saturation deficit data and the empirical wind functions are different An empirical formula which uses humidity as well as temperature to estimate evapotranspiration is given by Halstead (1951) Some of these methods are discussed in the following paragraphs

Dalton formula. This is the oldest aerodynamic formula and is based on an empirical equation of the form

$E = f(U_z)\ (e_s - e_z)$ (13) in which E is the evaporation rate, $f(U_z)$ is an empirical wind speed at height z, e_s is the saturated vapour pressure at the evaporating surface, and e_z is the mean actual vapour pressure at the height z Clearly this formula gives values for potential evaporation or evapotranspiration, since the vapour pressure at the surface is assumed to be saturated The wind speed function is commonly given in the form of $f(U) = (a+bu)$ or $f(u) = bu$, a and b being constants

Penman formula Penman in 1948 suggested a formula which was based on his work at Rothamsted Experimental Station This formula is by far the most complete of all empirical formulae in which the principles of both energy balance and aerodynamic methods are incorporated The derivation of Penman formula is as follows -

using equation (12) developed earlier i e. $E = Rn/(1+\beta)$ and an aerodynamic equation similar to that of Dalton,

$$E = f(U_z)\ (e_s - e_z)$$

similarly $H = f(U_z)\ (T_s - T_z)$ and

then $R_n = E + f(U_z)\ (T_s - T_z)$ (14)

Since the slope of the saturation vapour pressure curve, Δ, is given by

$\Delta = \frac{es - ez^o}{Ts - Tz}$ (e_z^o is the saturation vapour pressure at height z),

then

$$R_n = E + {}^{\gamma}/_{\Delta}\ f(U_z)\ (e_s - e_z^o) \qquad (15)$$

But $e_s - e_z^o = e_s - e_z^o + e_z - e_z = (e_s - e_z) - (e_z^o - e_z)$

therefore, $R_n = E + ({}^{\gamma}/_{\Delta})\ f(U_z)\ (e_s - e_z) - (\gamma/_{\Delta})\ f(U_z)\ (e_z^o - e_z)$

or

$$R_n = E + (\gamma/_{\Delta})\ E - ({}^{\gamma}/_{\Delta})\ E_a \qquad (16)$$

E_a being equal to $f(U_z)\ (e_z^o - e_z)$ (17)

and finally $E = ((\Delta/\gamma)\ R_n + E_a)\ /\ ((\Delta/\gamma) + 1)$

or

$$E = (R_n + (\gamma/\Delta)E_a)\ /\ (1 + \gamma/\Delta) \qquad (18).$$

The empirical wind speed function ($f(U_z)$) that Penman first suggested (1948) is represented in the following form

$$f(U_2) = 0.35\ (1+U_2/100) \qquad (19)$$

However later he suggested a new wind speed function (Penman 1956)

$$f(U_2) = 0\ 35\ (0\ 5+U_2/100) \qquad (20).$$

The value of R_n (net radiation) is measured by net radiometer or calculated from the formula

$$R_n = R_1\ (1-r) - R_B \qquad (21)$$

where R_1 is short wave radiation reaching the surface, r is the reflection factor and R_B is the long wave radiation.

The short wave radiation can be measured with reasonable ease and accuracy using solarimeters Since the number of places at which it is measured is small, it can be calculated by an empirical relation of the form

$$R_1 = R_A\ (a+bn/N) \qquad (22)$$

in which R_A is the theoretical maximum radiation if there was no atmosphere, a and b are latitude dependent constants and n/N is the ratio of actual to maximum possible hours of sunshine

Deacon (1958) compared measured and calculated data of short wave radiation and found a variation of 15 per cent (Rijtema, 1965). r is dependent on the kind of surface. Its approximate value for a water surface is about 0 05 and for green vegetation is 0 25

R_B, the long wave radiation leaving the surface, is expressed quantitatively by the Stefan-Boltzmann Law i e

$R_B = \sigma T^4$ in which σ is a constant and T is absolute temperature However this back radiation is reduced by atmospheric vapour and cloud Penman (1956) assuming a complementary relation between mean cloudiness and the mean sunshine factor n/N, expresses the net back radiation by the following formula

$$R_B = \sigma T^4 \, (0.56 - 0.09\sqrt{e_z}) \, (0\ 10 + 0.90\ n/N) \qquad (23).$$

In his formula Penman removed the necessity of two level measurements of temperature, wind velocity and vapour pressure by assuming that the lower height measurement has been shifted to the evaporating surface, that the evaporating surface is saturated, and that the surface temperature is equal to that at level Z_2

To use this formula, measurement of temperature (maximum, minimum, dry and wet bulb) and wind velocity at a height of two metres should be made In addition measurements should be made of net radiation (by net radiometer) or total radiation (by solarimeter). In the absence of these instruments, data from sunshine recorders should be used in the empirical equation (22) to calculate the total radiation term

The source of empiricism in the Penman formula lies in the use of an empirical aerodynamic equation and neglect of heat storage below the evaporating surface. Furthermore the assumption that the evaporating surface is saturated, neglects the vapour pressure deficit at the non-saturated evaporating surfaces and consequently the Penman formula yields values for potential evapotranspiration In spite of these points the formula is an excellent tool for soil moisture estimation in the range of field capacity conditions and the inaccuracy involved is no worse than that in the estimation of rainfall (Penman et al,1967). Ward (1967) refers to the results of application of this formula by Gilbert et al (1954) who found it successful for five-day period estimation and by Pearl et al (1954) over a seven-day period In general as the length of period increases, the accuracy of the formula increases

Thornthwaite formula Thornthwaite (1948) suggested an empirical formula for estimation of potential evapotranspiration. This formula is based on mean temperature data and has been developed from rainfall and run-off data of drainage basins. The simplicity of the formula and the availability of the temperature data for long periods at many locations have been the main reason for its widespread use Thornthwaite equation is in the form

$P.E^{*} = 1.6\ (10T/I)^{a}$ in which $P.E^{*}$ is unadjusted potential evapotranspiration in cm, T is monthly mean air temperature (C^{o}), $I = \sum_{i=1}^{12} i$, and $i = (T/5)^{1.514}$ (i is defined as monthly heat index).

$a = 6.75\ (10^{-7}I^{3}) - 7.71\ (10^{-5}I^{2}) + 1.792\ (10^{-2}I) + 0.4923$. The formula gives unadjusted rates of potential evapotranspiration based on a 12 hour day and thirty-day month and is corrected by the actual day length in hours (h), and the number of days in month (N) to get adjusted values, i e $P.E = P.E^{*}\ (^{h}/12)\ (^{N}/30)$.

a log-plot of P.E versus T is a straight line passing through ($P.E^{*} = 1.6$ and $T = {}^{I}/10$). The applicability of this formula is based on the correlation between radiation and temperature and between radiation and evapotranspiration The formula is most useful for monthly or longer period estimates It has been found unreliable for daily, three-day and six-day period estimates (Pelton et al,1960) Tanner (1967) recommends the calibration of the formula for any given region to derive local monthly or seasonal coefficients which include temperature lag, vegetation and local climatic corrections

Evaporation pans Evaporation pans are widely used to integrate the overall effects of climatic factors on evapotranspiration. Since the same factors that result in evapotranspiration, cause evaporation from pans, some sort of equation could be developed for relating pan evaporation data to evapotranspiration This seems to be highly desirable because none of the calculated formula take into account all the climatic conditions and further the appraisal of the results obtained should be based on the effect of soil and plant factors which appear as empirical factors It is, therefore, preferable to use open pan evaporation as an estimate of the potential evaporation and then an empirical correction factor can be applied to measured evaporation to relate it to actual evapotranspiration

There are many different types of pans and they differ in size and exposure. Consequently the results obtained are different, though in general all evaporation values thus obtained, exceed that from large water surfaces under similar conditions. To convert pan data to true evaporation, coefficients with values less than one are applied. Among the different evaporation pans used, the two most popular, are the Sunken Pan from the United Kingdom and the Class A evaporation Pan from the United States

3. Water balance methods

These methods of measuring evapotranspiration are based on the application of water balance equation Water balance equation is presented in the following simple form

$$P = R + Et + \Delta S + \Delta G$$

in which P - is precipitation

R - is runoff

Et - is evapotranspiration

ΔS - is change in soil moisture level and

ΔG - is change in groundwater storage

Water balance methods may be divided into the following types -

evapotranspirometer method

lysimeter method

soil profile method

catchment method

Evapotranspirometer method. This method has been used by Thornthwaite (1955) to measure potential evapotranspiration An evapotranspirometer is a water-tight tank filled with soil It works on the principle that the difference between the water added and that drained represents the volume of water used as evapotranspiration The assumption which is made is that the moisture content of the soil is always kept near field capacity For this purpose, the tank is irrigated daily by an amount of water which exceeds the potential evapotranspiration Hence, considering the water balance equation, $Et = P + I - D$ in which

P - is precipitation

I - is irrigation

D - is the excess water drained

By using this formula, it is assumed that change in the soil moisture content is zero The results of studies of potential evapotranspiration by this method are reported by Green, (1957, 1959), Ward (1963) and Pegg et al (1972)

Lysimeter method Lysimeters are isolated blocks of soil used to measure actual evapotranspiration Lysimeters are of many sizes and vary depending on the depth of the soil profile, the type of crop, the accuracy required and expense.

To measure actual evapotranspiration by lysimeters, the soil container is measured periodically to determine the weight changes. Depending upon the method of measurement, lysimeters are divided into weighing and hydraulic types In the weighing type, the soil container is placed inside a tank and is free for weighing Weighing is done by mechanical or electrical balances of different designs One of the best examples of weighing lysimeters is the large Coshocton weighing lysimeter (Harold and Dreibelbis,1951, as reported by Winter,1963). This lysimeter consists of a 65 ton monolith block of soil, 8 1 m^2 in area and 2.44 m in depth and is weighed by a recording balance which is sensitive to 0.25 mm Et.

Another example of this type is the 1½ ton instrument developed by the National Institute of Agricultural Engineering and installed at the Rothamsted Experiment Station (Morris, 1959). These lysimeters are highly desirable and useful for the measurement of actual evapotranspiration, however, they are very expensive and so the amount of experimental replication is limited

A hydraulic lysimeter is one in which the soil container is placed on some sort of hydraulic load cell and changes in weight are shown as changes in the pressure of the load cells. Winter (1963) described two lysimeters based on the above principle. One has got most of the weight of the soil container supported by bouyancy in water, while the remainder is supported by a flexible water filled bag on which it rests The other one described has got all the weight of the soil filled container supported by water filled bags. These lysimeters are simple to build and the cost of construction is low, and hence replication can be made.

Soil profile water balance method In this method, the equation of water balance is applied to the root zone of the plant and the major parameters which are considered are rainfall, runoff and Δs, which is the change in soil moisture content over the short period of consideration Two important factors affect the accuracy of the results obtained by this method. These are deep percolation and capillary movement of water from a high water-table to the surface which occurs mostly during dry seasons.

The instrument recommended for determination of changes in soil moisture content in this method is the neutron probe. This instrument allows rapid measurement of soil moisture with a high degree of accuracy

<u>Catchment water balance method</u> Evapotranspiration may be measured by applying the water balance equation to a catchment However, there are two main problems in adopting this procedure These are -

(i) the catchment should be water tight so that no leakage may occur from it to adjacent ones or vice versa.

(ii) owing to spatial and temporal variations of soil moisture content and groundwater storage, the areal values of these two terms cannot be measured Considering long term evapotranspiration, however, the values of Δs (changes in soil moisture content) and ΔG (changes in groundwater level) may approach zero and, therefore, actual evapotranspiration would be equal to the difference between precipitation and runoff Reports of some water balance studies have been given by Imeson et al (1972) and Pegg et al (1972).

<u>Estimation of actual evapotranspiration</u>

Measurement of actual evapotranspiration by lysimeters is expensive and rather difficult The values of actual evapotranspiration obtained are valid only for the combination of soil and plant for which the measurement is made Therefore, to avoid the difficulty and high expense of measuring actual evapotranspiration by lysimeters, several estimation methods have been suggested

Thornthwaite and Mather (1955) suggested that actual evapotranspiration drops below potential as soon as the water content of the soil drops below field capacity At any time the value of actual evapotranspiration is determined by the product of potential evapotranspiration and the ratio of current moisture content to that at field capacity

Veihmeyer and Hendrickson (1955) reported the results of their work with small peach trees grown in tanks They found that actual evapotranspiration was the same for peaches grown in tanks near field capacity as those near wilting point Their conclusion was that actual evapotranspiration is the same as potential just above permanent wilting point

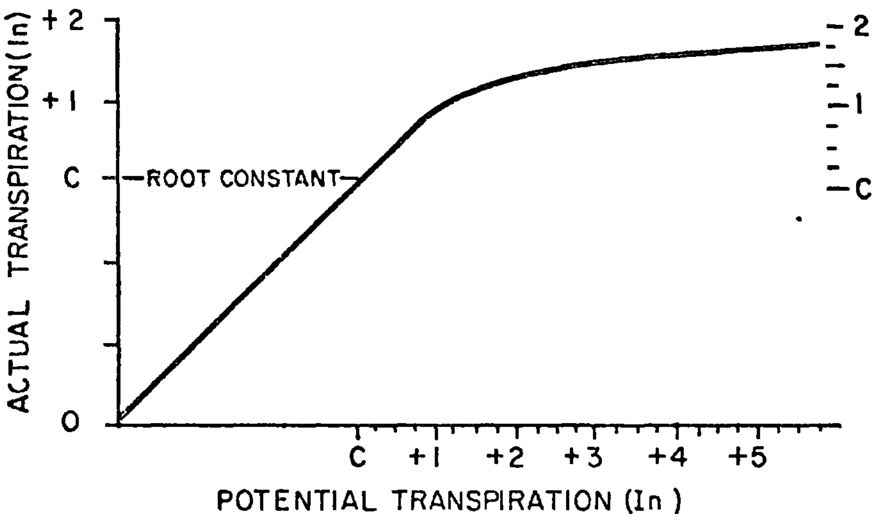
Penman (1949) suggested a method which was based on a composite drying curve for saturated and non-saturated conditions This curve was derived in the laboratory by applying drying potentials to bare soil initially at field capacity. This drying curve compared the evaporation rate from bare soil with that from open water exposed to the same conditions, and since the curve was the same for both sandy and clay soils, he assumed that it could safely be used for a range of dry soils. This curve shows that as the soil moisture is depleted due to evapotranspiration, actual evapotranspiration remains the same as potential up to a point, the abssisa of which is equal to a value called the root constant.

The root constant is defined as the amount of water readily available for a plant to draw on within its root zone without needing to depend on supplies below the rooting depth and without any limitation to the transpiration rate caused by water shortage The specification of the root constant is somewhat of a guess as mentioned by Penman (1949) and depends on soil and crop conditions and management He assumed a value of 75 mm for turf and 200 mm for deep rooted plants

Fig.27 also shows that as soil moisture depletion is continued beyond the root constant for another 25 mm, actual Et remains almost the same as potential Et Penman assumed that this extra amount of moisture would be supplied from depths below the root zone After that point there would be a sharp drop in the ratio Eta/Etp which then gradually becomes constant at a figure of 1/12

This method of Penman has been adopted by the Meteorological Office (Grindley,1967, 1970) for estimation of soil moisture deficit, which is then used as a form of warning to the river authority Soil moisture deficiency data are valuable because they indicate when soil moisture is approaching field capacity Under such conditions appreciable amounts of runoff can be expected to be produced by a given rainfall Table 21 represents the typical values of potential and actual evapotranspiration for a root constant of 75 mm.

Fig 27 COMPOSITE DRYING CURVE FOR SATURATED AND NON SATURATED CONDITIONS (ROOT CONSTANT OF THREE INS)



AFTER PENMAN 1949

Table 21 Typical values of potential and actual evapotranspiration for a root constant of 75 mm used in the Penman model (After Grindley,1970)

Potential Evapotranspiration (mm)	75	100	125	150	175	200	225	250
Actual Evapotranspiration (mm)	75	99	109	113	115	117	119	121

CHAPTER FOUR

EVAPOTRANSPIRATION - METHODS OF MEASUREMENT IN THE BROWNEY BASIN

To measure evapotranspiration in the Browney basin, all the various methods discussed in the preceding section were considered. However, the approaches adopted involved the use of evapotranspirometers, the Penman and Thornthwaite formulae for measuring or estimating potential evapotranspiration, and the catchment water balance, lysimeters, and the Penman drying curve method for measuring or estimating actual evapotranspiration The choice of these methods was based on their relative accuracy, expense and practicality The specific details of the methods and the procedures followed are explained in the following paragraphs

1. Potential Evapotranspiration

(a) Evapotranspirometers This instrument was used to measure potential evapotranspiration The choice of this method relative to evaporation pans was based on the lower cost of the evapotranspirometers The evapotranspirometer's set up and the technique adopted was similar to that described and used by Green (1957) and Ward (1963, 1967) Fig 28 shows the constructional details of the instrument It consists of three water-tight oil drums, two of which are filled with soil containing a small amount of gravel (15 cm) at the base to assist free drainage. The third tank houses the receptacles which collect the excess water flowing through the outlet pipes attached to the bottom of the soil tanks.

A pair of evapotranspirometers was set up at the Durham Observatory The choice of Durham Observatory for the evapotranspiration measurement was because it housed the meteorological instruments whose records were to be used in the Penman and Thornthwaite formulae In addition Durham Observatory was easily accessible for daily readings and management

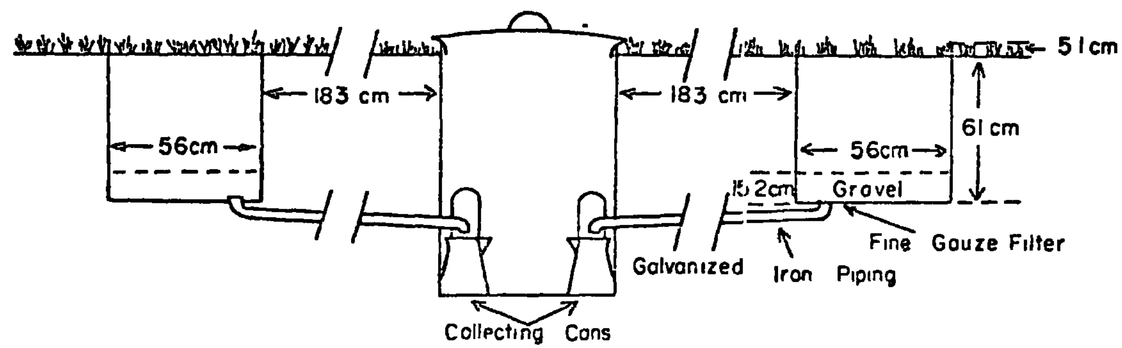


Fig.28. A pair of evapotranspirometers From an original by Ward (1967)

The Observatory is situated on a slight hill, at a height of 102 m above O.D The site is open and well exposed The location of the site is given by the following co-ordinates National Grid Reference NZ 267 415, latitude 54° 46' 06" N and longitude 01° 35' 05"W

The instruments are located on the lawn in front of the south elevation of the Observatory Temperature thermometers are exposed in a standard Stevenson screen. Precipitation is measured in a 5 inch (127 mm) diameter raingauge A Campbell-Stokes Sunshine recorder is sited 10 m above ground level Wind speed and direction are measured by a Dines Pressure Tube Anemograph (height of vane 16 m above ground level, 118 6 m above O D , effective height 10 m)

Preparations for the installation of the tanks started in late December 1972 Soil layers of 10 cm depth were dug out, labelled and covered In January 1973, the tanks were installed so that the rim of each tank protruded 5 cm above ground level in order to prevent water from the surrounding area running into the tanks. A layer of 15 cm of gravel was then put at the bottom of tanks to help in the drainage of excess water and to prevent the clogging of the drainage pipes by the soil particles The soil layers were then put into the tanks in a reverse order to that during digging They were compacted one by one so that when all the layers were replaced, the soil inside the tanks had a profile similar to that of the adjacent soil The surface of the soil in each tank was later covered with turf identical to that of surrounding lawn.

In order to allow for the settlement of the soil, the tanks were frequently watered during the period January to April 1973, and the actual collection of data was delayed until the beginning of May 1973. From 1st May 1500 cc of water were sprinkled onto each of the tanks daily at 9 30 a m and at the same time the volume of the percolate from the preceding day was measured and recorded These data, supplemented by the depth of precipitation from the standard 5 inch (127 mm) raingauge, were used to calculate the daily evapotranspiration, i e Et = Input-Output in which the summation of precipitation and irrigation is input and excess water is the output Two values were thus obtained daily from the two replicates and they were averaged in order to cancel any accidental errors.

It should be mentioned, however, that the daily values thus obtained were unadjusted values and owing to the lack of constancy in soil moisture storage and the time lag in the percolation of excess water, especially following heavy rain, some corrections had to be applied. The correction method is discussed in the following paragraphs

Green (1959) mentioned that about 90 per cent of all the excess water percolates in the first 24 hours and after 72 hours the amount still to drain is negligible Since this time lag effect of excess water could affect the weekly, 10-day and even monthly values, Green (1959) suggested a smooth curve method for adjustment of such errors

This consisted of plotting the cumulative curve of daily evapotranspiration for one period and drawing a smooth line which passes below the peaks in the cumulative graph which are caused by excessive precipitation This method enables assessment of weekly, ten day and monthly totals to be made and, therefore, it has been applied in this study.

This technique has not been applied during long periods of dry weather In these circumstances, the day to day variations of evapotranspiration could be indicated from the daily measurements discussed earlier.

The measurement of evapotranspiration which had started on 1st May, was discontinued in early October. The main reason for the discontinuation of the measurements was the appearance of a tiny hole at the bottom of the middle drum which housed the collecting cans As the groundwater level was being raised by frequent rain, water was finding its way through the hole and causing the containers to float For sometime this excess water, when it was a couple of cms, was emptied daily Several attempts were made to seal the tank to prevent water entry, but all were in vain

The measurement of evapotranspiration was also carried out by a second set of evapotranspirometers, which were installed close to the western end of the catchment at an elevation of 1100 ft (333 4 m) in order to evaluate the variation of evapotranspiration over the catchment area and over an elevation range of about 800 ft (242 7 m)

Several locations within the catchment were considered with a view of finding an open and well exposed site for this second set of evapotranspirometers It was also important to find a volunteer observer who would accept the responsibility for taking the daily measurements Eventually a suitable location was found at the Treatment Works of the Durham County Water Board at Honey Hill (National Grid Reference 052 468)

The management of the Water Board allowed the installation of the tanks and arranged for the daily (excluding the Saturday and Sunday) measurements of evapotranspiration for a period of one year During this period frequent visits were made by the author to check the accuracy of the measurements

The establishment of the evapotranspirometers at Honey Hill was completed in May 1973 using a procedure similar to that followed at Durham Observatory The measurement of precipitation at Honey Hill was carried out by the Water Board as part of its daily duties

The collection of data began on 14th July, 1973 and it continued for one year with measurements ending on 13th July 1974 Continuity in the measurement of potential evapotranspiration during 24th to 30th November and 8th to 14th December,1973 was broken The lack of measurement during these two periods was due to freezing conditions in the field

Therefore, evapotranspiration for these two periods was estimated by extrapolation of the cumulative potential evapotranspiration curve.

During the measurement period, frequent cuttings of the grass in the evapotranspirometers at both locations were made to maintain the height at about 4 cm in length.

(b) Empirical formulae of Penman and Thornthwaite. Potential evaporation was also calculated by both the Penman and the Thornthwaite formulae. For the Penman formula, the calculation was facilitated by the application of a computer program, supplied by the Institute of Hydrology, which uses the daily values of temperature (maximum, minimum, dry and wet bulb), run of wind and radiation data (either of solar radiation, net radiation or hours of sunshine) as input and produces daily values of Penman EO and Et with empirical wind functions [i.e. $f(u) = 0.35\,(1+\frac{u}{100})$ or $f(u) = 0.35\,(0.5+\frac{u}{100})$] as output. The formats of a typical lead card, control card and data card employed for running the program are shown in Fig 29. The program has thus been used to determine daily values of evaporation and evapotranspiration over a period of ten years.

To calculate potential evapotranspiration by the Thornthwaite formula, a simple program was written by the author to calculate the monthly potential evapotranspiration over the same ten year period. The values thus obtained had to be adjusted by a correction factor for day and month length. The appropriate correction factor was based on the formula

$$PE\ (\text{adjusted}) = PE\ (\text{unadjusted})\ \frac{(h\ \text{day length})}{12}\ \frac{(N\ \text{month length})}{30}.$$

In view of the fact that the estimation of evapotranspiration by the use of mean temperature methods, which includes that of Thornthwaite, is not reliable over periods shorter than one month (Pelton et al,1960 and Tanner,1967), no attempt was made to determine the weekly or ten day period values by this formula. The only short period study of eva-

Fig 29

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1      C
2          MEDIAN EVAPORATION AND EVAPOTRANSPIRATION INPUT FORMAT
3                                                                  93          1   174
4          172719  +542  102                     25 2  3  10 1 0 0 3 1 1 1 8   131 174
5          52    36    34    35   292     0                        3 1 1       1[illegible]174
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Description of the Cards in Fig 29

Card 1 (one for each month) Lead card

(a) Columns 61 and 62 always 93

(b) Column 74 is having the catchment number

(c) Columns 77-80 bear the month and year number

Card 2 (one for each month) Control card

(a) Catchment area in hectares to one decimal place in columns 1-10

(b) Latitude in degrees (+=N, -=S) to one decimal place in columns 11-16

(c) Altitude in metres, columns 17-21

(d) Albedo normally 25 over grassland, columns 41-45

(e) Temperature index, 2=degrees c, columns 46-47

(f) Wind index, 3=knots columns 48-50

(g) Height of anemometer in metres, columns 51-54

(h) Sun index, 1=hours, columns 55-56

(i) Total radiation index, 0=no data, columns 57-58

(j) Net radiation index, 0=no data, columns 59-60

(k) Data type, always 3, column 62

(l) Source index, 1=manual station, columns 63-64

(m) Number of sources, 1 for manual station, columns 65-66

(n) Input frequency, 1 for wet and dry at 9 00 hours only, columns 67-68

(o) Output frequency, 8 for manual station, columns 69-70

(p) Catchment number, columns 72-74

(q) Number of days in month, columns 75-76

(r) Month and year columns 77-80

Card 3 Data card (for each day of the month)

(a) Maximum temperature columns 1-6	Could be negative values to one decimal place.
(b) Minimum temperature, columns 7-12	
(c) Morning dry, columns 13-18	
(d) Morning wet, columns 19-24	
(e) Wind run, columns 25-30	No decimal places.
(f) Sunshine, columns 31-36	To one decimal place

(g) Data type, 3, column 62

(h) Source index, 1 columns 53-64

(i) Number of sources 1 columns 65-66

(j) Catchment number, columns 72-74

(k) Day of month, columns 75-76

(l) Month and year, columns 77-80

potranspiration was that of the evapotranspirometers compared with different estimates of Penman

2 <u>Actual evapotranspiration</u>

(a) <u>Catchment water balance method</u> To determine the values of evapotranspiration by the catchment water balance method, the monthly values of average rainfall over the catchment and runoff from it for the period October 1963 to September 1973 were considered The average rainfall over the catchment was based on the Thiessen method, using the data from the available raingauges There were seven raingauges available during the period 1968 to 1973 and four raingauges during the period 1963 to 1968 The available raingauges during 1963 to 1968 were those at Durham (National Grid Reference 267 415), Waterhouses (National Grid Reference 189 412), Satley (National Grid Reference 117 437) and Waskerley (National Grid Reference 022 444).

The runoff data used were obtained by dividing the monthly discharge by the area of the catchment. The groundwater area and topographic area were assumed to be coincident

The effect of mine water pumped into the river was not considered This was because the data of discharge of mine water into the river was only available for the year 1973 The total discharge of mine water for this year was about 13 6 mm or about 4 4 per cent of the mean yearly value.

In applying the water balance equation, it was assumed that the Browney catchment is water tight and that no leakage of moisture occurs from it to the adjacent basins or vice versa

(b) <u>Lysimeters</u> The best method for measurement of actual evapotranspiration is by a weighing lysimeter with a large volume of soil to minimize the restriction of plant roots and to reproduce soil suction Alternatively a neutron probe could be used to measure changes in the soil moisture content over short intervals However both of these methods are costly and attempts to borrow a neutron probe were not successful Therefore, it was decided to use simple hydraulic lysimeters suggested by Winter (1963) for point measurement of actual evapotranspiration. These lysimeters were developed at the National Vegetable Research Station at Wellesbourne and had been recommended by Winter (1963), (1966), Ward (1967), and Pegg (1970)

These lysimeters suffer the disadvantage of not containing large volumes of soil and not reproducing the actual soil-plant-water conditions in the field However it was thought, that if the results of the initial testing of the lysimeters were promising, then lysimeters could be employed to study the effects of different soil moisture contents on actual evapotranspiration

The constructional details of the simple hydraulic lysimeter are shown in Fig.30 A polythene bin 42 cm deep and 38 cm in diameter is used as the container for the soil and plant The "plastic bag" on which the soil container rests consists of an inner tube for a motor scooter type ('350-8' i e 13 inches (32 5 cm) outside diameter and 7 inches (17 5 cm) inside diameter The manometer consists of a length of 0 13 inches (3 3 mm) - glass tubing with a pressure head of 20 inches (51 cm) of water above the container. The manometer deflection is approximately 1.5 times the corresponding addition or loss of water to or from the whole soil surface of the lysimeter, (e.g. water addition or loss of 10 mm on the lysimeter produces a deflection of 15 mm). The calibration is not precisely linear, because distortion of the inner tube cross-section with increasing soil container weight might cause a change in the contact area between the tube and the container bottom

For the initial testing two of these lysimeters were constructed and were installed in a relatively open field in the western part of Durham Observatory in previously prepared pits The soil of the pits was dug and covered in 10 cm layers so that when filling the containers,

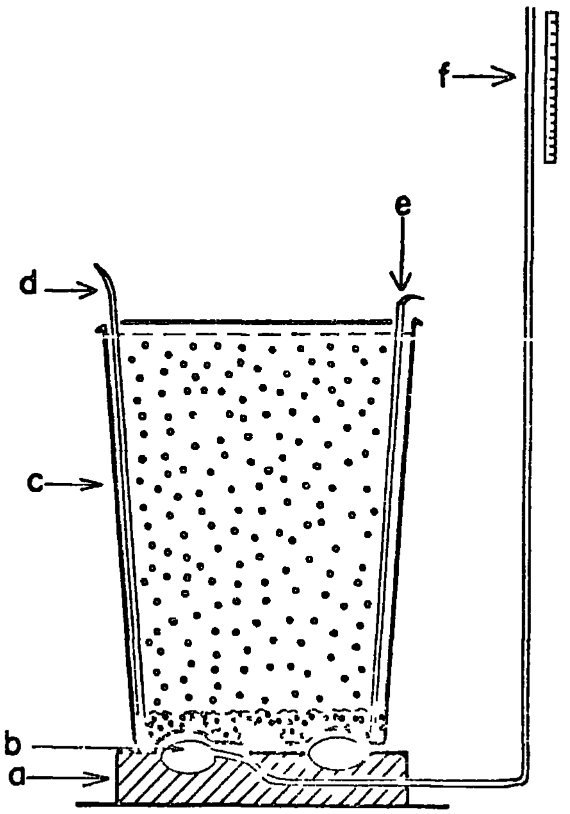


Fig 30 Non-buoyed Lysimeter

a Expanded polystyrene base, b Water-filled motor-scooter inner tube, c Soil container (plastic domestic refuse bin), d Pipe for removal of drainage by suction, e Air-vent pipe, f Vertical manometer

AFTER WINTER 1963

the soil profile would be similar to that of the surroundings The sides of the pits were cemented so that if the groundwater level rose during the period of observation, the lysimeters would remain undisturbed

The height of the pit was such that the soil surface of the containers when set on the expanded polystyrene base, was at the same level as the surrounding soil. Before filling the container with soil, a layer of pebbles was laid at the bottom and this was then separated from the above soil profile by a layer of coarse grit The thickness of the pebble and grit layers was about 5 cm These layers were laid to ease the sucking of excess water by the vacuum pipe The surfaces of the lysimeters were covered with turf identical to that of surrounding area

These lysimeters were installed in late March 1973, but measurements did not begin until 15th May 1973 The measurements were continued until early October 1973. During the period of observation and testing which lasted for four and a half months, one of the lysimeters lost its balance, leaned against the side of the pit and it had to be re-established This happened on 8th June 1973 This may have resulted from distortion of the inner-tube cross-section caused by changes in the weight of soil container bringing about a change in the contact area between the tube and the container bottom

On 17th August 1973 the water level of one of the manometers dropped below the scale This was due to continued loss of water from the lysimeter due to evapotranspiration without adequate replenishment of the soil moisture by rainfall to compensate for the loss Consequently the lysimeter container was removed, the scooter tube was refilled with water and the container was reset

The recording of water level in the lysimeter was done daily, except for a few days in September and a couple of days in other months (Appendix III) During the whole period the length of grass was maintained at a height of about 4 cm

(c) Penman method To determine average values of actual evapotranspiration over the catchment using the drying curve method of Penman, several assumptions had to be made. These were that 85 per cent of the catchment was covered by short rooted vegetation with a root constant of 75 mm, 10 per cent of the catchment was covered by the deep rooted trees with a root constant of 200 mm, and the remaining 5 per cent was taken to be riparian (shallow water table areas where evapotranspiration occurs at potential rate at all times) The percentage value of each zone was estimated using the distribution of land use in the catchment These values were then slightly modified to represent the spatial variations in the soil moisture holding capacity of the three levels

Penman (1950) in his study of the water balance of the Stour catchment and Grindley(1967) in applying the method for the estimation of soil moisture deficit to other catchments adopted values of 50, 30 and 20 to represent the percentage distribution of short rooted crops, deep rooted crops and the riparian zone These values, of course, apply to lowland catchments, where the soils are deep and water table zones constitute larger percentages of area

A typical working example of this method is shown in Table 22 Column one represents the month of the year In column (2), the Thiessen average of monthly rainfall for the raingauges throughout the catchment is given In column (3) the monthly potential evapotranspiration (Penman EO_2) is shown The difference between rainfall and evapotranspiration is given in column (4). From this column it is observed that for this particular year (1971), moisture deficiency starts in May when the potential evapotranspiration exceeds rainfall by 45 9 mm This deficiency is further increased to 110 6 mm (column 5) which is the cumulative moisture deficiency in July In the following month part of this deficiency is compensated for by excess rainfall (76 1 mm)

In September and October, rainfall is exceeded by evapotranspiration,

Table 22 Calculation of the areal value of actual evapotranspiration over the Browney catchment using the Penman method for the year 1971 (mm)

Month	Precipitation (R) (mm)	Evapotranspiration (Penman EO_2) (mm)	$R-EO_2$ (mm)	Cumulative moisture deficiency (mm)	Actual Evapotranspiration in (mm)			
					shallow water table zone (5%)	Deep rooted crop zone (10%)	shallow rooted crop zone (85%)	Areal value
January	53.3	6.6	+46.7	0	6.6	6.6	6.6	6.6
February	18.3	8.9	+ 9.4	0	8.9	8.9	8.9	8.9
March	70.2	31.0	+39.2	0	31.0	31.0	31.0	31.0
April	61.0	53.2	+ 7.8	0	53.2	53.2	53.2	53.2
May	47.9	93.8	-45.9	45.9	93.8	93.8	93.8	93.8
June	71.6	83.6	-12.0	57.9	83.6	83.6	83.6	83.6
July	50.0	102.7	-52.7	110.6	102.7	102.7	96.7	97.6
August	143.3	67.2	+76.1	34.5	67.2	67.2	67.2	67.2
September	12.8	48.2	-35.4	69.9	48.2	48.2	48.2	48.2
October	22.0	37.2	-15.2	85.1	37.2	37.2	37.2	37.2
November	53.3	11.9	+41.4	43.7	11.9	11.9	11.9	11.9
December	20.7	12.1	+ 8.6	35.1	12.1	12.1	12.1	12.1

and so the moisture deficiency is increased by the difference From November, owing to low Et values, however, the deficiency is gradually reduced The actual evapotranspiration from the zone with a shallow water table (column 6) is shown to be the same as potential evapotranspiration. This is because there is unlimited supply of moisture available in this zone

For the deep rooted zone, also, the rates of actual evapotranspiration and potential evapotranspiration are the same (Table 22, column 7) The explanation for this observation is that the cumulative moisture deficiency is always less than 225 mm. This value is the maximum moisture deficiency above which actual Et drops below potential Et.

For the zone with shallow rooted crops the value of actual evapotranspiration drops below potential only in July This is because July is the only month for which the cumulative deficiency is above the limit of readily available water in the soil reservoir For this month the cumulative moisture deficiency is 110 6 mm, while the total readily available water for the shallow rooted zone is 100 mm Therefore, actual evapotranspiration will be 96 7 mm using the Penman drying curve, as compared with 102 7 mm of potential evapotranspiration The areal value of actual evapotranspiration for the catchment is the sum of the product of the area of each zone times the actual evapotranspiration value

CHAPTER FIVE

EVAPOTRANSPIRATION - RESULTS AND DISCUSSION

(a) Durham Observatory results

The adjusted weekly, and monthly values of potential evapotranspiration from the evapotranspirometers have been obtained through the application of the smooth curve technique (Green,1957) Fig 31 shows the plot of cumulative values of daily unadjusted potential evapotranspiration and the smooth line drawn to bypass the peaks caused by excessive precipitation for the period of observation at Durham Observatory

The total amount of Et measured by the evapotranspirometers for the period 1st May to 30th September 1973 is 395 mm, while the Penman EO_1, EO_2 and Et values are 410 mm, 387 mm and 324 mm respectively (Table 23) EO_1 and EO_2 are values of potential evaporation (albedo 0 05), the former having an aerodynamic term with an empirical function of wind speed of $f(U_2) = 0.35\ (1+ \frac{U_2}{100})$, while the empirical wind function of the latter is $f(U_2) = 0.35\ (0\ 5+ \frac{U_2}{100})$ Et refers to potential evapotranspiration (albedo of 0 25) with a wind function similar to EO_1

Table 23 Monthly values (mm) of measured Et, estimated Penman values of Et, EO_1, EO_2 and estimated Thornthwaite Et at Durham Observatory during the year 1973

Month	Measured Et (evapotranspirometers)	Estimated			
		Penman			Thornthwaite
		Et	EO_1	EO_2	
May	82 0	70 5	88 2	83 7	71 8
June	86 0	76 1	96 7	92.6	94 2
July	96 0	68 1	86 3	81.4	111 3
August	83 0	68 5	86 5	81 2	97 3
September	48 0	40 3	51 9	48 3	69 0
	395 0	323 5	409 6	387 2	443 6

These results show that measured Et at Durham Observatory exceeds Penman Et and Penman EO_2 by 18 per cent and 2 per cent respectively, however it is less than EO_1 by 4 per cent In a similar study comparing

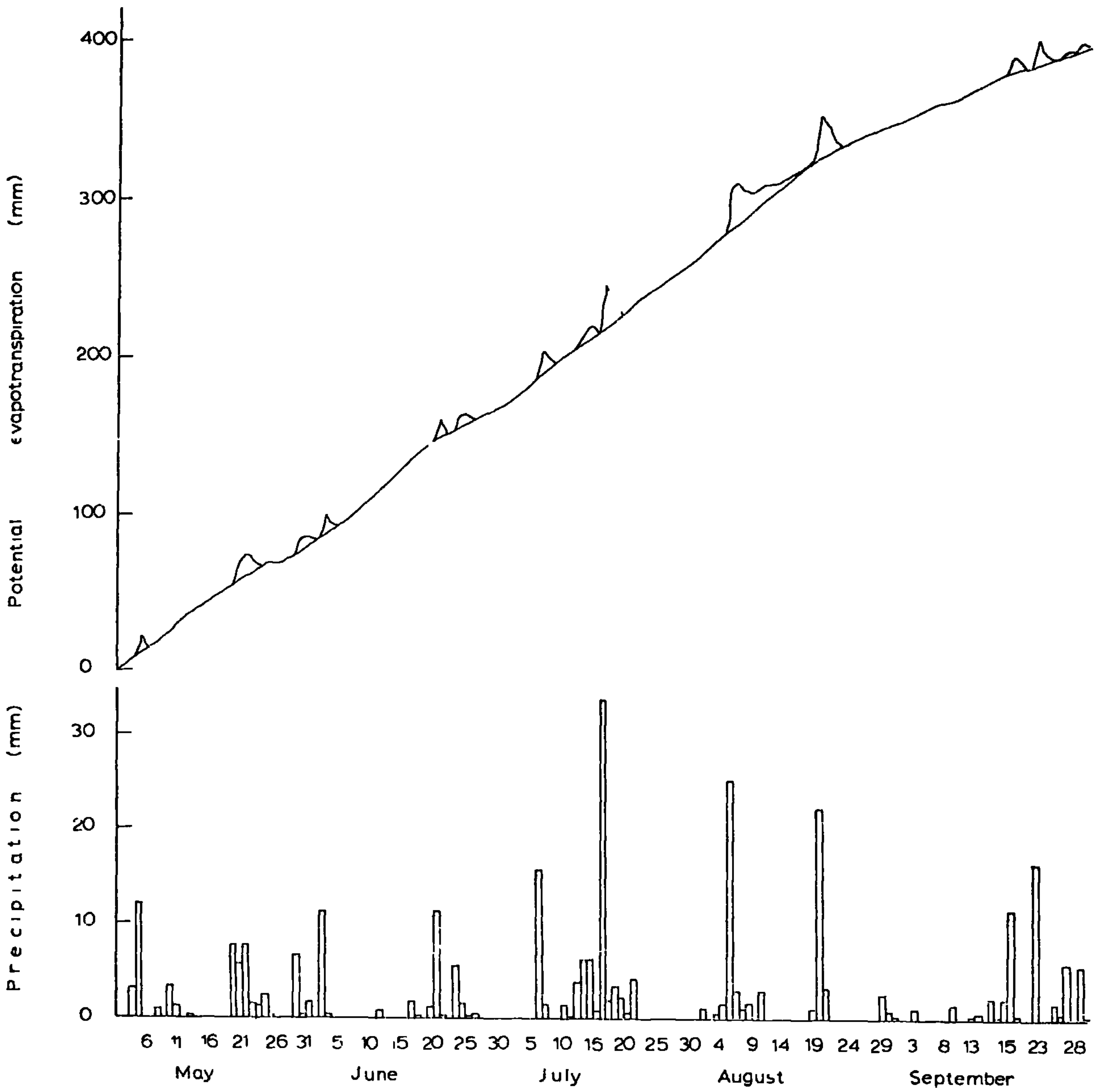


Fig 31 Cumulative potential evapotranspiration its smooth curve and depth of precipitation at Durham observatory from 1st May to 30 September 1973

measured evapotranspiration with Penman Et and EO (the wind function was not mentioned), Pegg et al (1972) found that measured Et exceeded Penman Et by about 13 per cent, while the Penman EO value for the year was more than measured Et by 8 per cent.

The reasons for the low values of Penman Et could possibly be ascribed to the lack of measured data of radiation for use in the formula. Hassan (1973) refers to the results of a comparative study of computed and measured evapotranspiration for a ten year period at one station in south-east England In this study, it was found that estimates made with measured radiation were 10 per cent to 15 per cent higher than those made with theoretical radiation

Referring to Table 23 and Fig 32 for comparison of the individual monthly values, it is observed that the highest monthly measured value does not coincide with the highest Penman estimates Measured Et is highest during July (96 mm), while that of EO_1, the highest value of the Penman estimates is 86 mm in this month On the other hand, the estimated Penman values are highest during June (EO_1 = 97 mm, EO_2 = 93 mm and Et is 76 mm) The corresponding measured Et for this month is 86 mm

Several factors might explain the lack of coincidence of highest monthly values of measured Et and Penman values The most important factor, however, is the neglect of soil heat storage in the Penman formula.

During June, when the monthly Penman estimates have their maximum values, some of the net energy received could have been used for heating the soil However, since the storage of heat is neglected by the Penman formula, the net energy available is used as heat of vaporization and for heating the air Thus evapotranspiration will be over-estimated

The effect of the neglect of soil heat upon evapotranspiration estimates by the Penman formula was studied by Edwards (1970) In comparing soil moisture changes determined by the neutron scattering method and the Penman formula, he found that the Penman Et value during summer

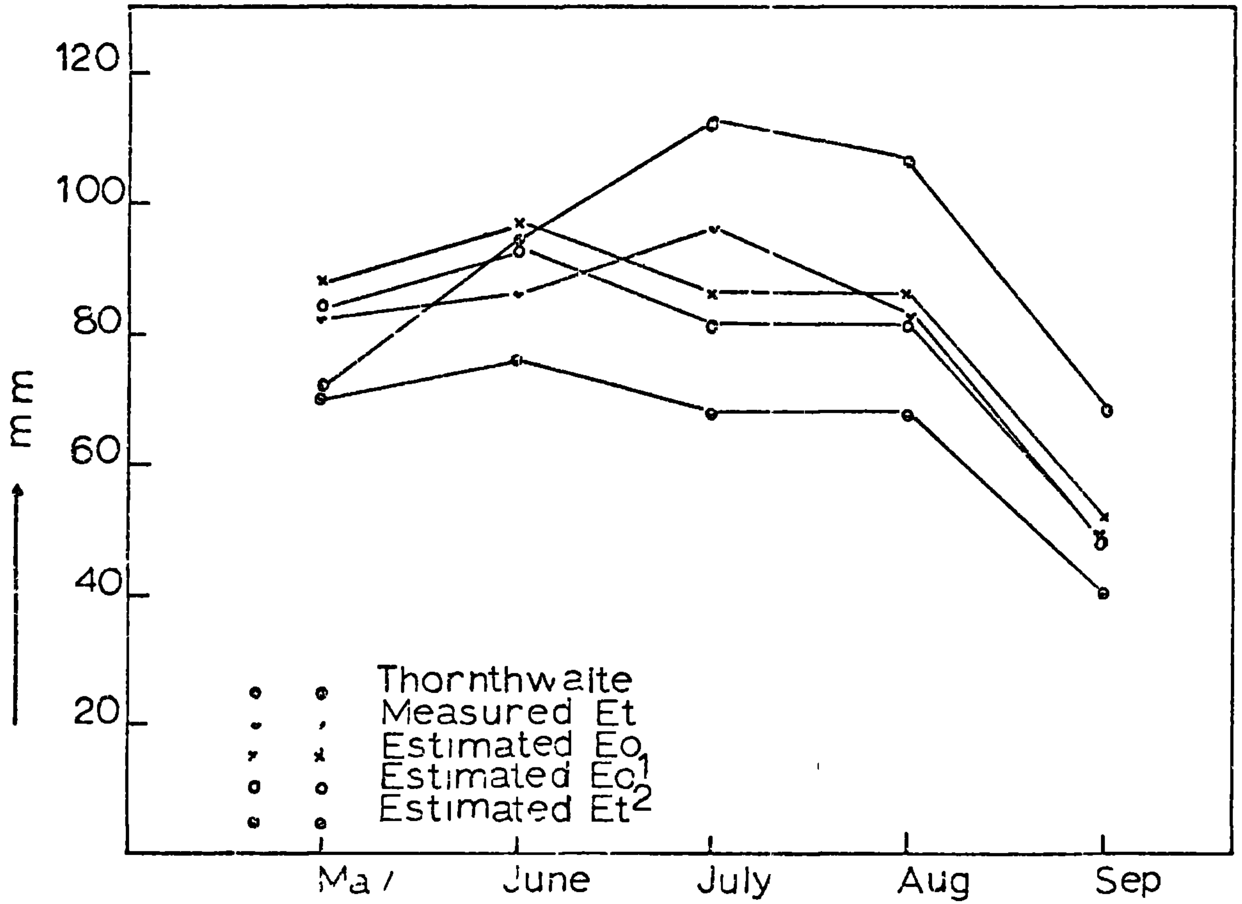


Fig 32 Monthly values of measured Et, Penman Eo_1 and Eo_2 and Et and Thornthwaite Et at Durham Observatory during the period May to Sept 73

was above that measured by moisture meter by 40 mm. This size of error, he mentioned, could not have arisen from other possible sources of errors in the measurement of soil moisture, and thus he thought that the heat budget of the evaporation formula was in error. Using soil and earth thermometers, he obtained an approximation of the heat storage and then he found good agreement between calculated and measured evapotranspiration.

The high value of measured Et during July is explained by the lack of measured data during 17th, 18th and 19th July. As a result of high rainfall (33 7 mm) on 16th July, the water table level was raised and upward movement of water through the tiny hole in the bottom of the middle tank prevented any measurement being made for the three following days of 17th, 18th and 19th July Therefore, the measured evapotranspiration during this period was obtained by extrapolating the cumulative evapotranspiration curve This procedure has apparently resulted in an overestimation of the total for the week ending 23rd July i e. estimation of measured Et is 22 mm as compared with 15 mm, 14 mm and 12 mm for the values of Penman EO_1, EO_2 and Et respectively This, therefore, explains the high value of measured Et during July.

During September, measured Et and the Penman estimates of Et, EO_1 and EO_2 reach their lowest values For each of the five months, measured Et exceeds Penman Et, but is less than Penman EO_1 (except in July, when the measured value is higher than the Penman EO_1)

The total value of Thornthwaite evapotranspiration during this period exceeds the measured Et by some 12 3 per cent It is also higher than the estimated value of Penman EO_1 by 8.3 per cent. Thornthwaite monthly Et exceeds measured Et during June, July, August and September, but is less during May It also has monthly values less than the measured EO_1 during May and June, but it exceeds Penman EO_1 during July, August and September

Weekly totals for measured Et and the Penman estimates are plotted in Fig 33. From a comparison of these data, the following conclusions are drawn

1 Measured Et is higher than Penman Et in 17 out of 22 weeks. Totals of estimated Et for the weeks ending 28th May and 3rd September exceed the measured values by 1 and 2 mm respectively For the remaining 3 weeks the values are equal

2. Measured Et is higher than Penman EO_2 in 11 weeks, lower in 8 weeks and they are almost equal during the remaining 3 weeks

3 Measured Et is higher than Penman EO_1 in 9 weeks and lower in 11 weeks, being approximately equal in the remaining 2 weeks.

4 The curves of weekly values of measured and estimated Et have maximum divergence for the week ending 23rd July The reason, as explained earlier is due to lack of measured data on 17th, 18th and 19th July

5. The week of 12th to 18th June has the maximum average daily Et of 3 7 mm derived from the evapotranspirometers, whereas that for Penman EO_1, EO_2 and Et is 4.3 mm, 4 1 mm and 3 4 mm respectively.

The scatter diagrams of seven day totals of measured Et versus estimated Penman values are shown in Fig 34 The correlation coefficients between measured and estimated Penman values are the same to two significant figures i e 0 80

For the monthly values the correlation coefficients between the measured Et and Penman estimated values are shown to be the same (0 90), while that of Thornthwaite's estimated value for Et is lower e g. 0 77. Thus it is shown that the Penman values are more closely correlated with measured Et than those of the Thornthwaite formula

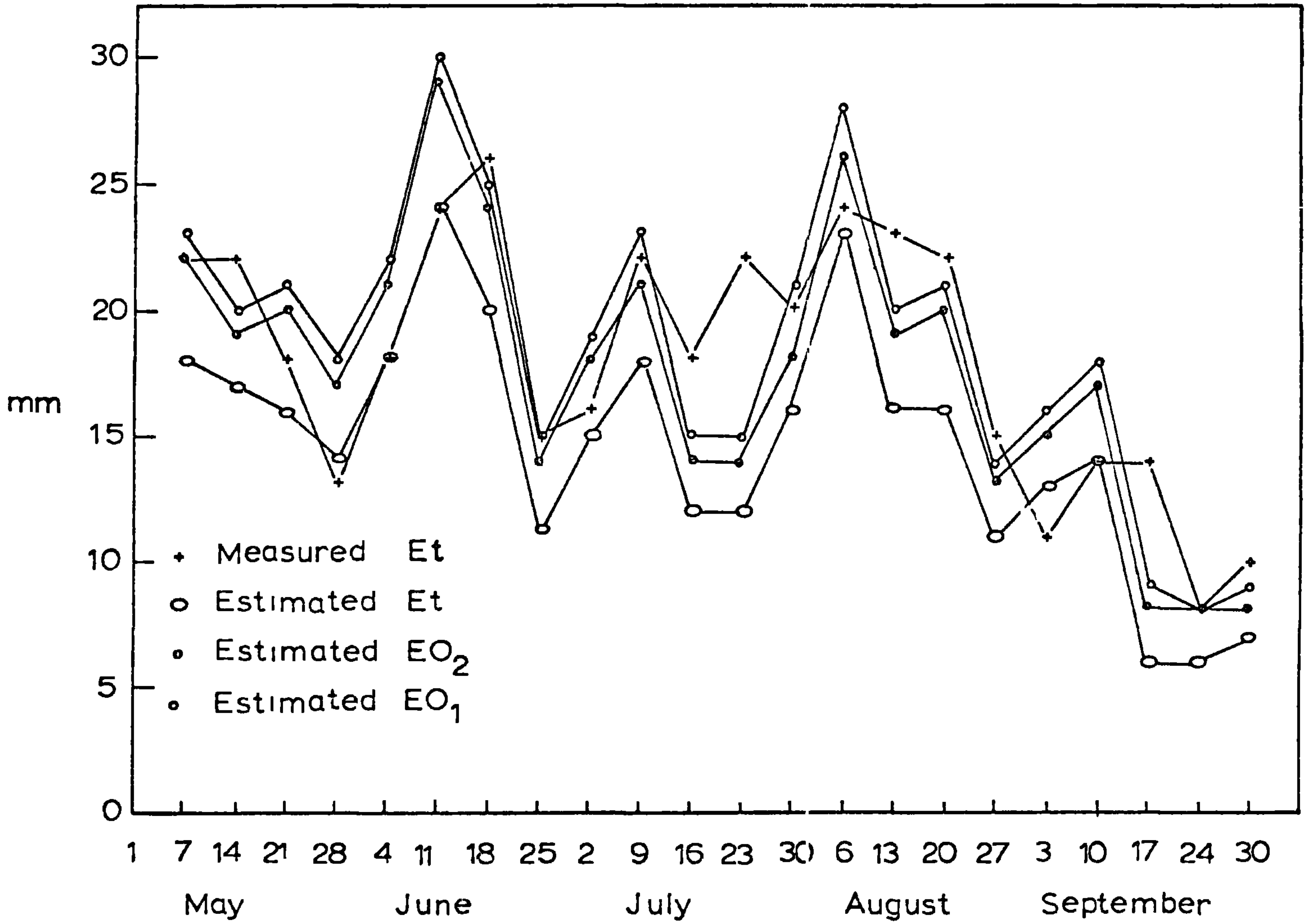


Fig 33 Weekly totals of measured Et and estimated Penman values of Et, EO_1 and EO_2 for the period May 1st to September 30, 1973 at Durham observatory

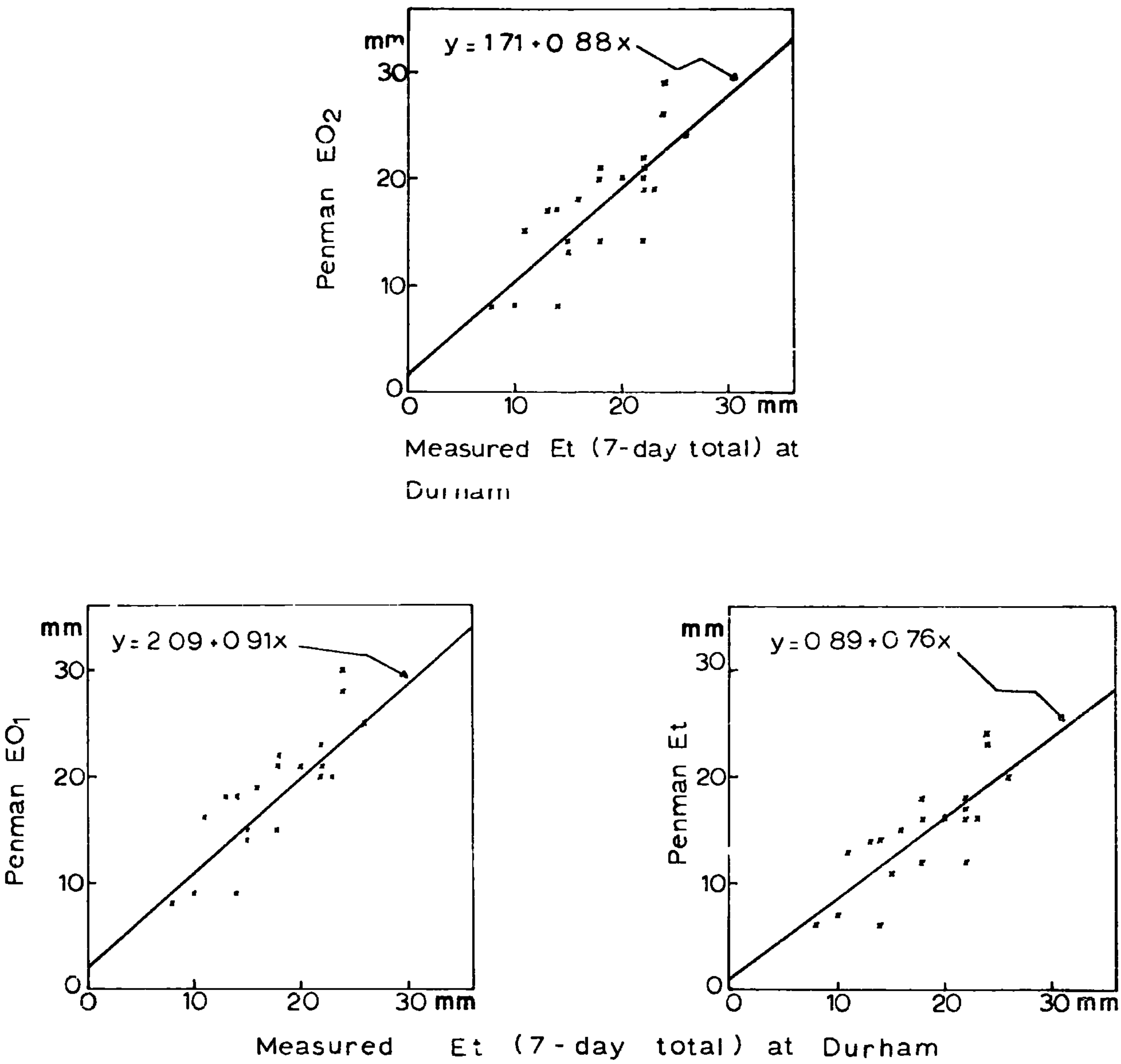


Fig 34 Scatter diagrams of measured Et versus Penman Et, EO_1 and EO_2

Table 24 Correlation coefficients between measured Et and estimated Penman and Thornthwaite values at Durham Observatory

	Weekly totals	Monthly totals
Measured Et versus Penman Et	0 80	0 90
Measured Et versus Penman EO_1	0 80	0 90
Measured Et versus Penman EO_2	0.80	0 90
Measured Et versus Thornthwaite Et	-	0 77

(b) Honey Hill results

The total potential evapotranspiration measured at Honey Hill by the pair of evapotranspirometers from 14th July, 1973 till 13th July, 1974 is 624 mm The corresponding values for the Penman estimates of EO_1, EO_2, Et and Thornthwaite Et are 582 mm, 544 mm, 467 mm and 622 mm respectively These estimates are calculated by using meteorological data at Durham Observatory The results thus show that measured Et at Honey Hill exceeds Penman EO_1 by about 7 per cent, EO_2 by 13 per cent and Et by 25 per cent and is slightly above the Thornthwaite estimate. Considering the seasonal values (Table 25), it is observed that measured evapotranspiration at Honey Hill during winter (1st September to 28th February) is 151 mm or 24 per cent of the yearly measured value, while measured Et during summer (March - August) is 473 mm or 76 per cent of the total Et Penman evaporation and evapotranspiration values for winter are 21 to 22 per cent of the total, and so those of summer are 79 to 78 per cent respectively Thornthwaite values for winter and summer are also 180.4 mm (29 per cent) and 442.1 mm (71 per cent) respectively

Table 25 Seasonal values of measured and estimated evapotranspiration at Honey Hill (mm)

	Measured	Estimated			
		Penman			Thornthwaite
		EO_1	EO_2	Et	
Winter (Sept, Oct, Nov	95	87 8	80 0	68 7	129 6
(Dec, Jan, Feb	56	39 8	35 1	34 4	50 8
	151	127 6	115 1	103 1	180.4
Summer (Mar, April, May	182	174 8	164 4	137 9	141 4
(June, July, Aug	291	281.4	264 5	226 2	300 7
	473	456.2	428 9	364 1	442 1

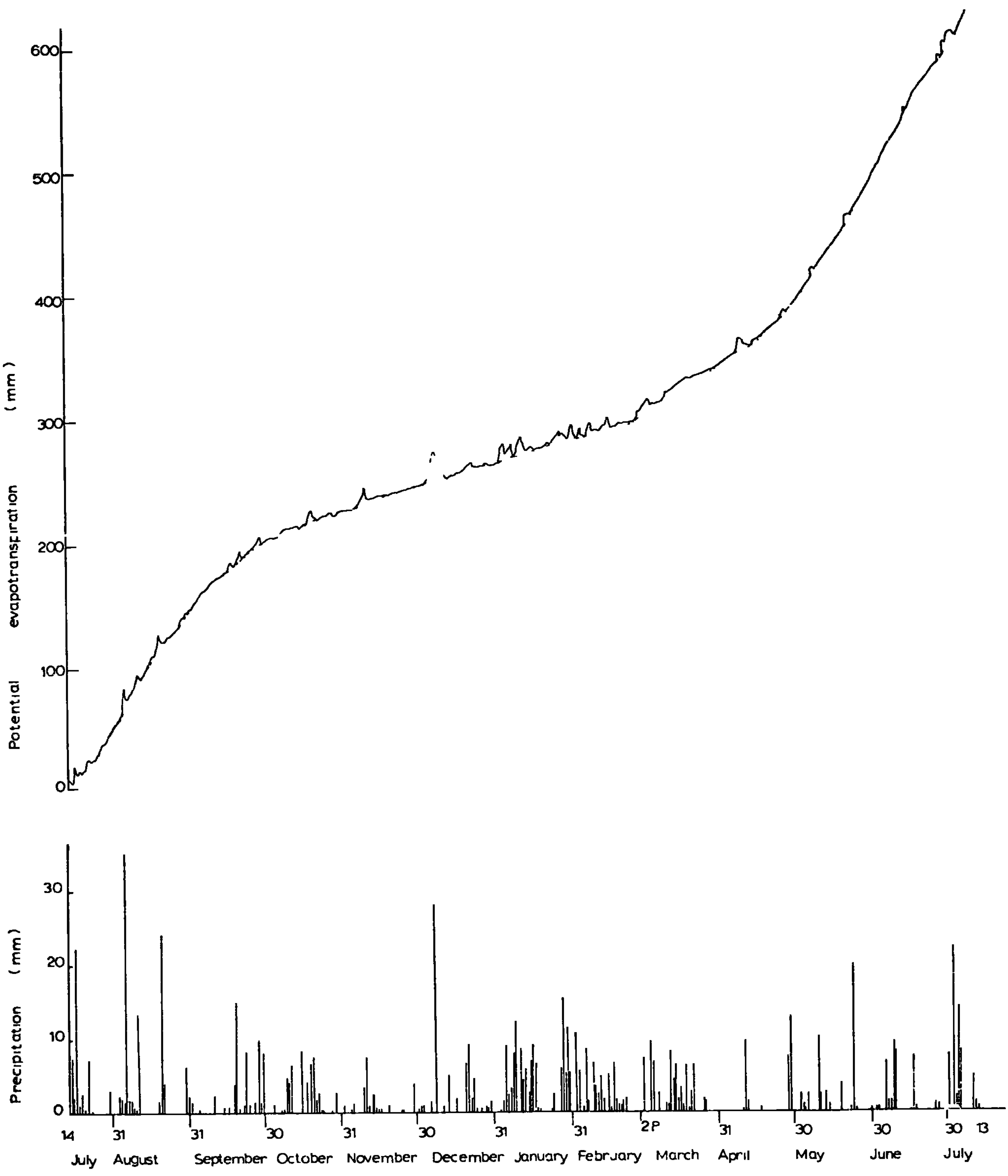


Fig 35 Cumulative measured potential evapotranspiration its smooth curve and depth of precipitation at Honey Hill from 14 July 1973 to 13 July 1974

Comparing the seasonal values of measured Et and Penman EO_1, it is concluded that measured Et during winter exceeds that of Penman EO_1 by 15 5 per cent while during the summer the excess amount is 3.5 per cent.

Dividing the year into four seasons of autumn, winter, spring and summer (Table 25), it is shown that measured evapotranspiration during autumn is 15 per cent of the yearly value, that of winter is 9 per cent, that of spring is 29 per cent and that of summer is 47 per cent of the total for the year

For the Penman EO_1, the total evaporation during autumn, winter, spring and summer is 15, 7, 30 and 48 per cent of the yearly value respectively. These figures show very clearly that evapotranspiration during the three summer months makes up less than half of the yearly total

Comparing the measured Et and estimated EO_1 values, it is concluded that the total measured evapotranspiration during autumn, winter, spring and summer exceeds that of the Penman EO_1 value by 7 5, 29, 4 and 3 per cent respectively. Measured Et during each of the four seasons of autumn, winter, spring and summer exceeds the Penman EO_2 values by 16, 37, 10 and 9 per cent and it exceeds Penman Et by 28, 39, 24 and 22 per cent respectively These results, therefore, show that Penman estimates of evaporation and evapotranspiration are too low during the winter season They also confirm work by Robin (1958), (reported by Smith,1964a), who claimed that the Penman formula is least valid for the winter part of the year when vegetation is dormant

The Thornthwaite evapotranspiration values exceed the measured evapotranspiration values by 36 per cent during autumn and 3 per cent during summer, however they are lower than the measured Et values by 9 per cent during the winter and 22 per cent during the spring. The fact that the Thornthwaite potential evapotranspiration value is above measured Et by 36 per cent in autumn and is lower than measured Et by 22 per cent in spring, shows that the spring and autumn estimates of evapotranspiration by the Thornthwaite formula could be in error. The source of error is attributed to the lag of temperature behind radiation which arises from thermal storage of the soil Fig 36 shows the monthly values of evapotranspiration measured by evapotranspirometers, at Honey Hill and estimated by Penman EO_1, EO_2 and Et and Thornthwaite Et at Durham Observatory. These values are also shown in Table 26.

It is observed that measured monthly evapotranspiration is greater than Penman EO_1 for nine months, and is lower during May by 7 mm (7 per cent), during June by 4 mm (4 per cent) and during September by 1.9 mm (4 per cent). The divergence between these two curves is highest during December, when measured evapotranspiration exceeds EO_1 by 10 mm or 54 per cent Considering Penman EO_2 and Et, the former is exceeded by the measured Et values in eleven months, being lower in May by 1 per cent while the latter is exceeded by measured Et in all twelve months

Considering the 10-day totals (Fig 37), measured Et is higher than Penman Et in all the 10-day periods except the week ending 19th January when measured evapotranspiration is 6 mm, compared with 8.6 mm of estimated Et

The estimated Penman EO_2 values are higher than the measured Et values during three periods, i e. 10th to 19th January (9.1 mm EO_2 and 6.0 mm measured Et),20th to 29th May (5.4 mm EO_2 and 5.0 mm measured Et) and 30th June to 8th July (41 8 mm EO_2 and 38.0 mm measured Et).

The estimated Penman EO_1 value is higher than the measured Et value in ten of these 10-day periods, six during the summer and four during the winter (Fig 37).

From a study of Table 27, it is observed that the measured evapotranspiration values and the Penman estimates are highly correlated on a 10-day basis. The scatter diagrams of the 10-day totals of measured Et and the Penman estimates are shown in Fig 38

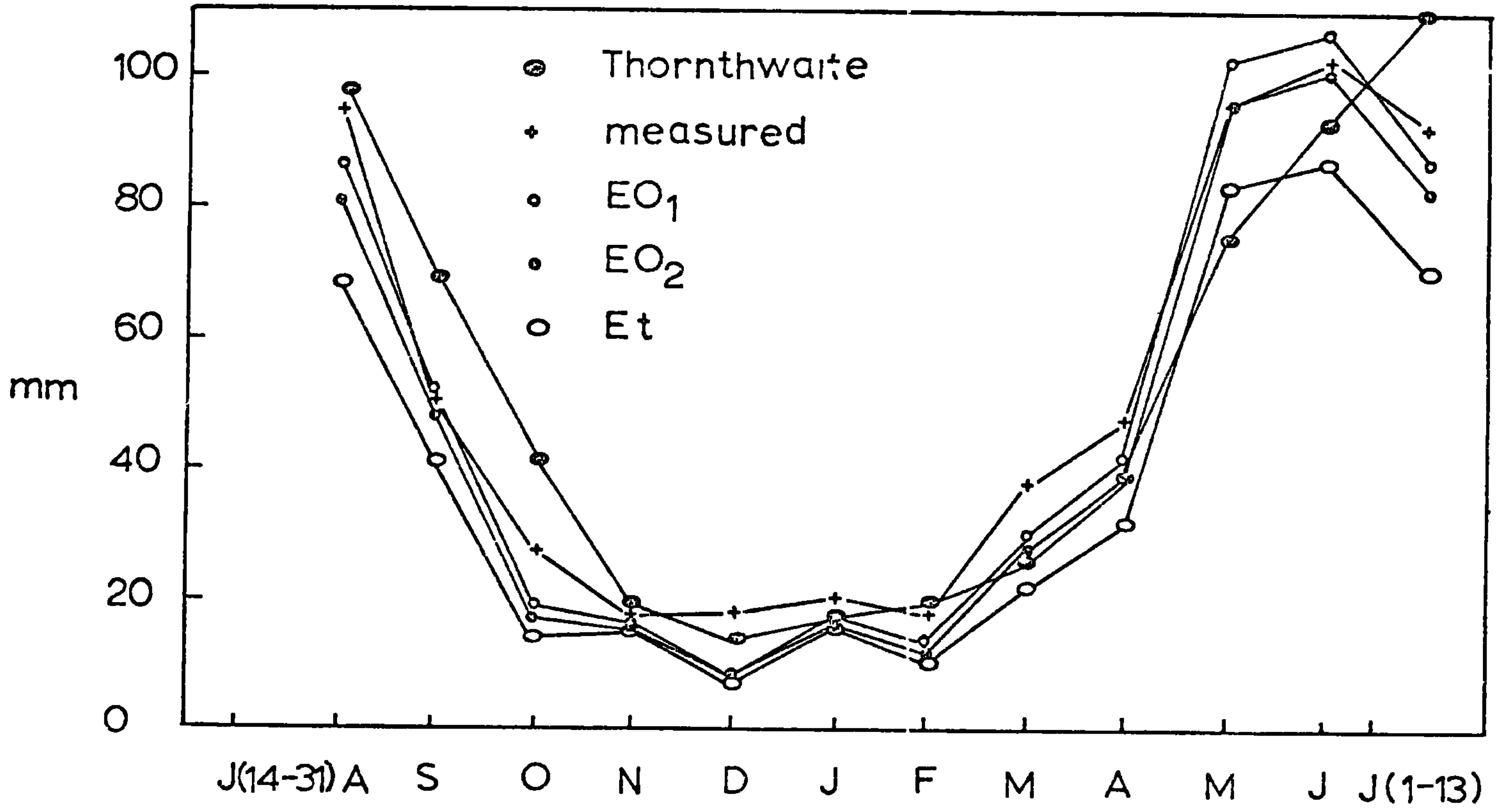
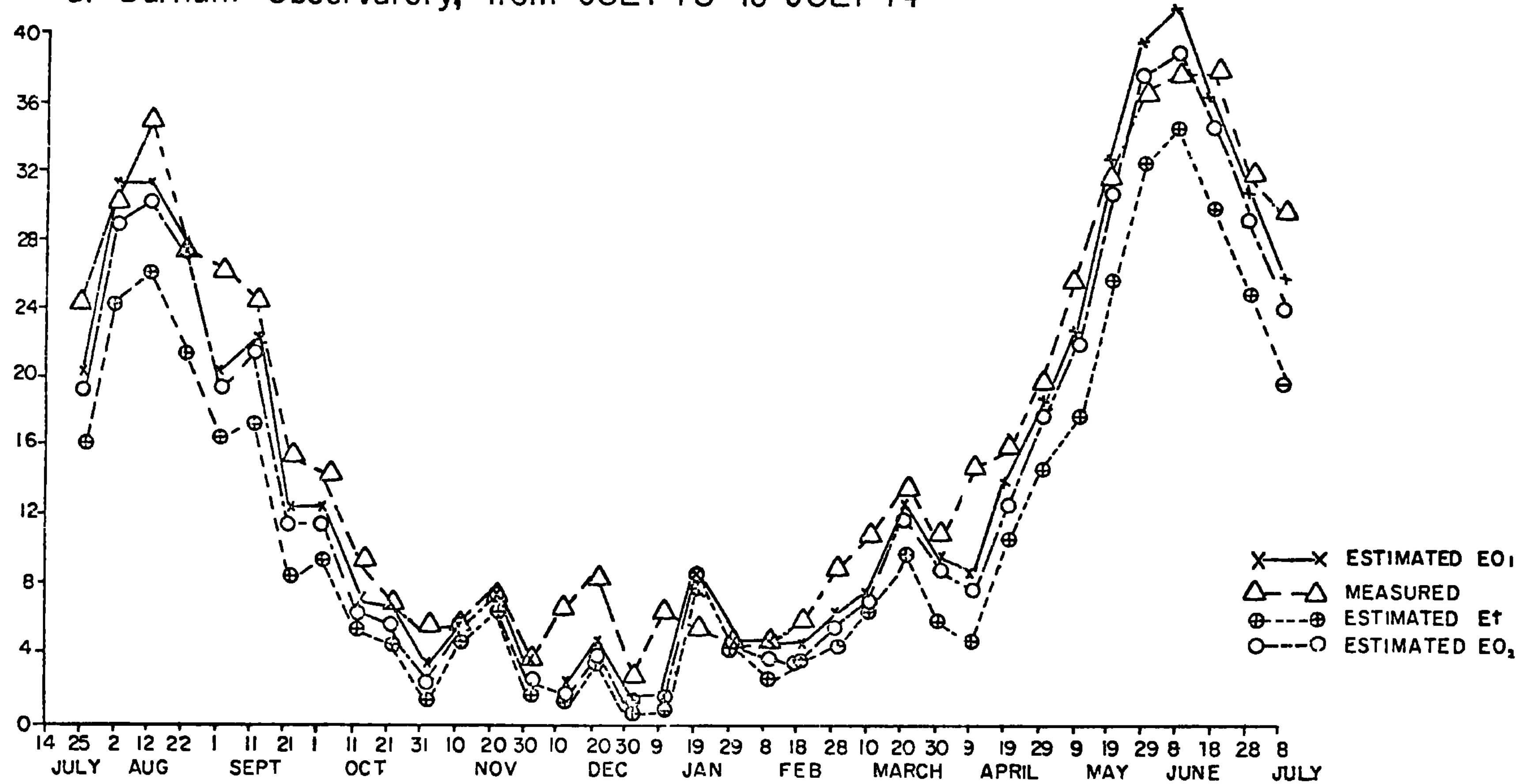


Fig 36 Monthly values of measured Et at Honey Hill and Penman EO_1, EO_2 and Et at Durham from 14 July 73 to 13 July 74

Table 26 Monthly values (mm) of measured Et at Honey Hill and estimated Penman values of Et, EO_1, EO_2 and Thornthwaite Et at Durham

Month		Measured Et	Estimated			
			Penman			Thornthwaite
			Et	EO_1	EO_2	Et
July (14-31)	1973	48.0	36.4	46.7	44 2	64.6
August	1973	95 0	68.5	86.5	81 2	97.3
September	1973	50 0	40 3	51.9	48 3	69 0
October	1973	27 0	13.6	18.9	16 9	41 2
November	1973	18 0	14.8	17 0	14 8	19 4
December	1973	18 0	8 0	8.3	7 3	13.9
January	1974	20.0	15.9	17 6	15.5	17.2
February	1974	18.0	10.5	13.9	12 3	19 7
March	1974	38 0	22.4	29 7	27.9	26.5
April	1974	48 0	32.2	42 0	39 7	39.1
May	1974	96.0	83 3	103 1	96.8	75.8
June	1974	103.0	87 7	107 0	100 4	93.8
July (1-13)	1974	45.0	33.6	41.2	38 7	45 0
TOTAL		624 0	467 2	581 8	544 0	622.5

Fig .37 Ten-day totals of measured Et at Honey Hill and estimated EO1, EO2 and Et at Durham Observatory, from JULY 73 to JULY 74



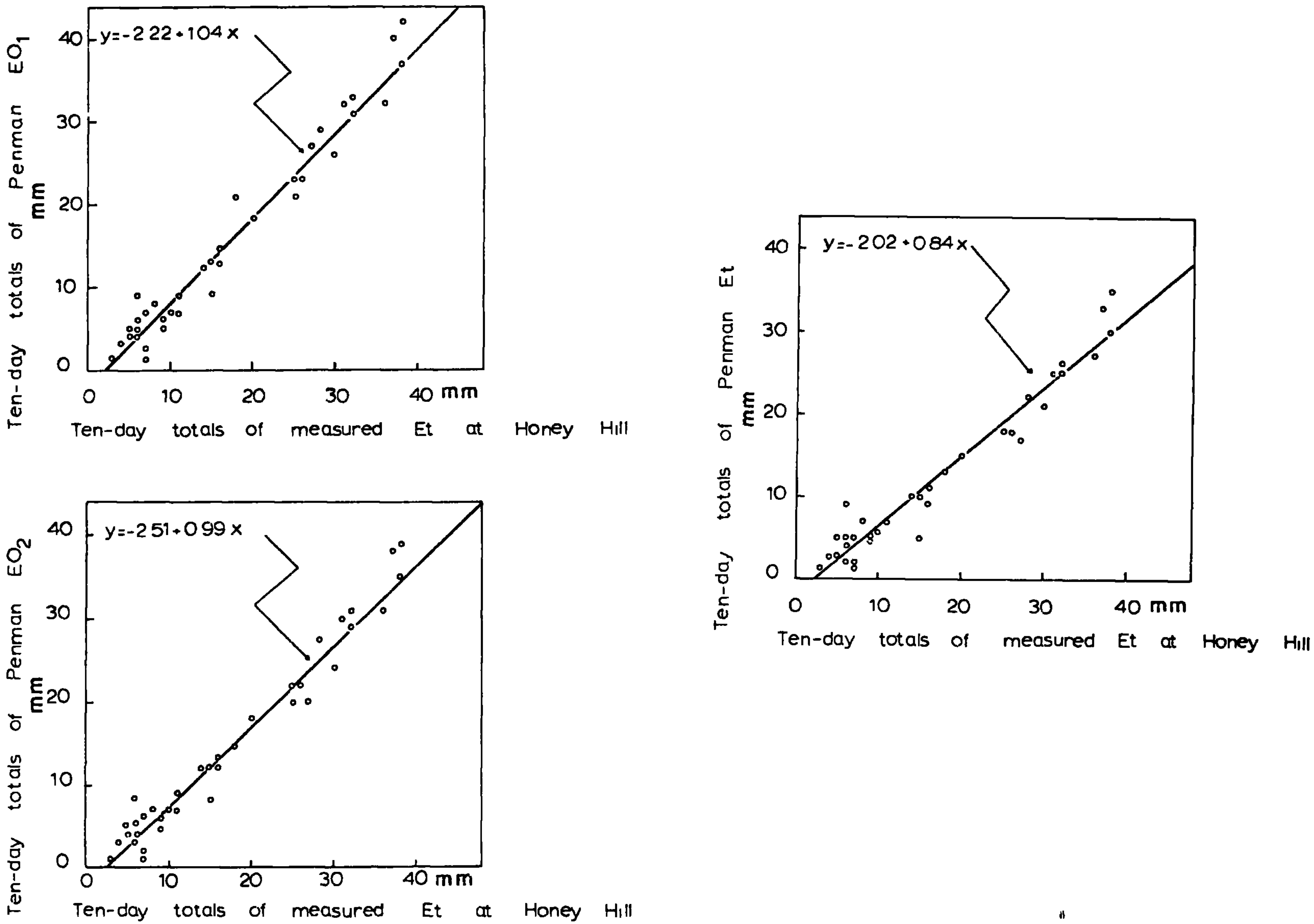


Fig 38 Scatter diagrams of ten-day totals of measured Et versus Penman Et, EO_1 and EO_2

Table 27 The correlation coefficients between measured Et at Honey Hill and estimated Penman and Thornthwaite values at Durham Observatory

	10-day totals	Monthly totals
Measured Et versus Penman Et	0 97	0 98
Measured Et versus Penman EO_1	0 97	0.99
Measured Et versus Penman EO_2	0.98	0.99
Measured Et versus $Thornthwaite_{Et}$		0.81

Comparison of the results of the evapotranspirometers at Durham Observatory and at Honey Hill

The main objective of setting up the two sets of evapotranspirometers at Durham and Honey Hill was to study the variation of evapotranspiration over the catchment However care should be taken in making any conclusion from the results This is explained by the fact that there are errors which might have influenced the results One of the main sources of error is due to exposure variation at the two sites of the measurement. The evapotranspirometer set at Durham has been installed inside the meteorological fence close to the instruments. The fence is bordered on one side by the Observatory building and on the other sides by some trees The evapotranspirometers at Honey Hill, on the other hand, have been set up at a point, which is well exposed and surrounded by an open area without any obstacles Thus these variations at the sites of measurement could possibly have affected the results.

Edwards(1970) mentions an error of some 12 per cent in the measurement of evaporation from two identical sunken evaporation tanks exposed on the meteorological site at Wallingford These two tanks were separated by only 30 metres. The explanation which was found to account for this difference was the effective lowering of the rim of the tank due to a slight surface undulation.

With this introduction, the results obtained from these two sets of evapotranspirometers are compared Referring to Tables 23 and 26 it is observed that the total evapotranspiration during August and September

1973 (the period during which evapotranspiration at Durham and Honey Hill had been measured concurrently) is 131 mm for Durham compared with 145 mm for Honey Hill, an increase of about 10 per cent at Honey Hill. The higher evapotranspiration rate at Honey Hill compared with Durham could also be observed by considering that the Penman EO_1 value has been above the measured Et value at Durham by 4 per cent but lower than the measured Et at Honey Hill by 7 per cent during the respective periods of measurements at these stations.

Thus assuming that there are no significant errors due to variation in the exposure of the evapotranspirometers or in the measurement of input and output moisture at the two locations it would seem that evapotranspiration increases with increasing height.

The results of similar experiments concerned with evapotranspiration-height relationships are reported elsewhere and according to Peck et al (1963), they are inconclusive Fortier (1907) reported that water loss from six evaporation pans decreased with height to about 10,000 feet (3048 m) above which there was little change (Linsley et al,1949) Green (1959) in a study of potential evaporation during (1955 to 1957) obtained results indicating a general increase in evaporation rate with elevation. Meyer (1942) found that evaporation from water surfaces in the mountainous areas of the coast ranges of California increased with elevation and that evaporation pan data from Los Angeles County indicated a similar relationship from sea-level to 4,000 feet (1219 m) (Linsley et al,1949) Grindley (1970) in an attempt to construct a map showing the distribution of average annual evaporation over the British Isles used an inverse evapotranspiration-height relationship According to this relationship, there is a decrease of 0 35 inches (8 9 mm) in evapotranspiration for an increase of 100 feet (30 5 m) in elevation.

These contradictory results of evaporation-height studies can partly be explained by the variation of meteorological elements with height Thus as elevation increases, the temperature decreases (dry adiabatic lapse rate of 0.54 F /100 feet or about $1^{o}C$ per 100 m), and consequently it is estimated that there will be a decrease in evaporation due to lowering of temperature with height An increase in elevation results in an increase of relative humidity and therefore evapotranspiration decreases Atmospheric pressure also decreases with increasing height, therefore, the density of water molecules in the air decreases. Since the rate of evapotranspiration depends on the vapour pressure gradient between the evaporating surface and that of overlying air, evapotranspiration will be expected to increase with decreasing pressure.

Another important factor that could affect the interpretation of the results is that of advection. Advection occurs when pre-heated air from the surrounding area passes over the evapotranspirometers, thus, due to additional energy, the evapotranspiration rate increases. The effect of advection under British climatic conditions and a short crop cover is usually ignored (Edwards et al, 1970), however there are instances when the tanks might become over-exposed An extreme case is that given by Green (1957) for a location in Scotland when potential evapotranspiration during August 1955 was about 180 mm. This value, however, was adjusted to 81 mm to remove the effect of advection

In this study the effect of advection has been ignored However, it is quite possible that advection might have affected the results, especially those at Honey Hill This is explained by the better exposure of the evapotranspirometers and the higher wind velocities at this location

<u>Water balance evapotranspiration versus Penman and Thornthwaite estimates and measured Et</u>

The yearly values of measured and estimated evapotranspiration and evaporation, and the average yearly value of rainfall minus runoff over the period October 1963 to September 1973 are shown in Table 28.

Table 28 Yearly values of measured and estimated evapotranspiration, evaporation and the average yearly value of rainfall minus runoff over the period October 1963 to September 1973 within the Browney basin

Rainfall-runoff (mm)	Measured Et at Honey Hill (mm)	Penman method (mm)			Thornthwaite meth (mm)
		EO_1	EO_2	Et	Et
439.0	624.0	581 8	544 0	467.2	622.5

These results show that the measured Et and Thornthwaite Et value exceed the water balance Et by 185 mm and 183 5 mm respectively or by about 42 per cent The Penman EO_1, EO_2, and Et values also exceed the water balance Et value by 142 8 mm (33 per cent), 105 mm (24 per cent) and 28 mm or 6 4 per cent respectively

Thus, assuming the Browney Catchment to be water-tight and that the topographic divide and groundwater divide are coincident, then it can be concluded that there is some deficiency of moisture within the catchment This is because the value of evapotranspiration from the water balance method which represents actual evapotranspiration within the catchment is lower than potential evaporation and evapotranspiration values from other methods.

Summary To summarize the results of evapotranspiration measurements at the two locations, the following statements can be made

1. Evapotranspiration within the Browney catchment, as measured by two pairs of evapotranspirometers, is greater at the higher elevation The higher value of evapotranspiration at the higher elevation, however, could have been due to differences in the exposure of the tanks, observational errors or the effect of advection.

2 Based upon the data available, the Penman EO_2 values are assumed to represent the mean evapotranspiration over the catchment. This assumption is made because the Penman EO_2 values are closest to the measured Et at

Durham Observatory during the period May to September 1973, closer to the average yearly water balance Et from the catchment (compared with Penman EO_1 values) and are most highly correlated with measured Et at Honey Hill. Whether this assumption is valid and, therefore, that the Penman EO_2 values are closest to true areal evapotranspiration value, is subject to question Pegg et al (1972) mention that, "as with discussion of variation of rain catch with different types of raingauges, there is no absolute standard against which to verify the selected values It is seldom possible to say more than that two of the methods agree closely and are, therefore, more likely to be correct than the other widely discrepant results, an argument whose illogicality needs no emphasis"

Long period study of values of Penman Et and EO_2, Thornthwaite Et and water balance Et

The mean yearly and mean seasonal data of the Penman Et, Penman EO_2, Thornthwaite Et and water balance Et over a 10-year period are presented in Table 29

Table 29 Mean yearly, maximum, minimum and coefficient of variation of the yearly values of Penman Et and EO_2, Thornthwaite Et and water balance Et over a 10-year period October 1963 to September 1973

Method	Mean yearly (mm)	Maximum (mm)	Minimum (mm)	C of variation
Penman Et	470 5	568 6	344 0	14.52
Penman EO_2	548 9	633 5	442 1	11 02
Thornthwaite	616.1	632 5	598 5	1 75
Water-balance	439 0	500.2	369.2	11.43

Referring to Table 29 it is observed that mean yearly values of Penman Et, Penman EO_2, Thornthwaite Et and water balance Et have the same order of magnitude as those for the period 14th July 1973 to 13th July 1974, which were discussed in the previous section. Thus the Thornthwaite estimate of evapotranspiration is above that of Penman EO_2 by 10.9 per cent Penman EO_2 is above that of Penman Et by 14 3 per cent and

Table 30 Mean seasonal values of Penman EO_2, Et, Thornthwaite Et and water balance Et in mm and as a percentage of their respective yearly values (Oct 1963 - Sept 1973)

Method	Mean Seasonal (Oct to Mar)	% Mean Seasonal	Mean Seasonal (Apr. to Sept)	% Mean Seasonal
Penman Et	86 3	18.3	384.2	81.7
Penman EO_2	92 7	16.9	456.2	83.1
Thornthwaite	128.3	20.8	487 8	79.2
Water balance	153 3	34 9	285.7	65.1

Penman Et is above that of the water balance Et by 6.7 per cent.

Considering the maximum yearly values of these estimates during the 10-year period under study, it can be observed that the maximum Penman EO_2 is slightly above that of the Thornthwaite Et value e.g 633.5 mm for Penman compared with 632 5 mm for Thornthwaite Both of these maximum values, belong to the water year 1964 (Tables31 and 32). The minimum yearly value of Thornthwaite Et, on the other hand, is greater than that of the Penman EO_2 value by 26 per cent The minimum yearly values from these two methods occur in two different years. Thus these results show that during the 10-year period of study the Thornthwaite potential evapotranspiration values have a smaller range than those of the Penman EO_2 values The range of values for Thornthwaite Et, Penman EO_2, Penman Et and the water balance Et are 34 0 mm, 191.4 mm, 224 6 mm and 131 0 mm respectively.

The coefficient of variation of the Thornthwaite yearly values is 1.75 per cent, while those of the Penman EO_2, the water balance method and the Penman Et are 11 02, 11 43 and 14 52 per cent respectively The reason for the smaller range and the low dispersion of the yearly Thornthwaite values is explained by the low variation of mean yearly temperature.

The mean seasonal values of the Thornthwaite Et, the Penman EO_2 and Et and the water balance Et in mm and as a percentage of the yearly values are presented in Table 30. According to this table the value of the Penman Et during the summer expressed as a percentage of the yearly

Table 31 Monthly values of potential evaporation, EO_2 (mm) by the Penman formula (Oct.63 - Sept.73)

Month / water year	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL
1964	18.3	8 9	6 2	8 1	18 0	24 8	69 4	113 6	104 4	112.3	85 6	63.9	633 5
1965	22.7	15 2	13 5	14 0	20.6	29 7	67.1	95 4	110 1	79.4	84 4	44.6	596 7
1966	22.2	10 4	5 1	8 1	14 8	57 2	50 2	91 4	88 7	110.2	77 0	43.4	578 7
1967	14.4	7.8	3.4	2.3	10.3	28 2	48 1	60 4	84 3	81.2	64.7	37.0	442.1
1968	14.2	3.2	4.5	10 2	6 6	40 0	56.1	66 8	97 4	67 0	67 3	41 0	474.3
1969	21 8	6 7	7 9	5.5	8 3	23 2	63.0	65.3	94 2	104 0	82 3	46 8	529 0
1970	31.6	13.9	5 7	3 9	15.8	42 1	68.1	96.3	116.3	106 9	73.0	44.2	617.8
1971	18 5	9.6	4 5	6 6	8 9	31 0	53.7	93.8	83 6	102 7	67.2	48 2	528 3
1972	37 2	11 9	12.1	7 5	9.0	31 8	60 0	83.3	80.5	87.9	81 8	43.7	546.7
1973	20.0	11 2	5.7	9 7	16.1	37 3	55.1	83.7	92 6	81.4	81.2	48.3	541.8
Total	220.9	98 8	68.6	75 4	128.4	345.3	590.8	850.0	952 1	933.0	764.5	461.1	5488 9
Mean	21 1	9 9	6.9	7 5	12.8	34 5	59.1	85 0	95.2	93 3	76 4	46.1	548 9
S D	7.2	3.5	3.4	3 3	4 8	10.1	7.6	16 7	11 9	15.8	7.8	7.1	60.5
C.V.	32.58	35.35	49.28	44 00	37 50	29.28	12 86	19 65	12 50	16 93	10 21	15.40	11.02
Max.	37.2	15.2	13 5	14 0	20 6	57.2	69 4	113 6	113 3	112.3	85.6	63 9	633.5
Min	14.2	3 2	3 4	2.3	6 6	23 2	48 1	60.4	80 5	67 0	64.7	37 0	442 1

Table 32· Monthly values of potential Et (mm) by Thornthwaite formula (Oct.63 - Sept.73)

Month / water year	O	N	D	J	F	M	A	M	J	J	A	S	Total
1964	47.0	20.5	12.5	11.2	14.5	16.3	50.6	86.4	93.9	109.9	96.5	73.1	632.5
1965	40.4	24.8	7.4	9.7	15.0	22.3	47.3	76.8	102.6	97.7	93.3	69.5	606.8
1966	49.3	10.8	10.4	6.1	13.7	34.5	32.8	75.6	106.5	107.3	92.9	67.5	607.4
1967	43.9	21.6	14.0	11.5	2.0	35.3	45.3	63.8	96.1	106.1	99.1	70.4	609.1
1968	44.4	18.6	13.2	12.1	5.3	32.5	44.5	59.3	99.5	98.7	96.4	74.0	598.5
1969	54.4	27.4	9.3	14.1	1.0	12.2	23.2	72.8	97.3	119.1	106.2	75.1	612.1
1970	56.7	15.5	7.5	8.0	7.9	16.6	35.1	82.0	108.5	107.7	103.8	75.3	624.6
1971	45.8	22.5	14.4	12.6	17.6	36.3	42.8	60.0	82.2	117.1	97.1	76.3	624.7
1972	48.0	21.0	21.7	11.2	14.1	30.2	50.5	71.4	85.0	108.7	98.1	64.2	624.1
1973	46.2	21.9	16.1	12.4	14.3	25.2	41.3	71.8	94.2	111.3	97.3	69.0	621.0
Total	476.1	204.6	126.6	108.9	105.4	261.4	413.4	719.9	965.8	1083.6	980.7	714.4	6160.8
Mean	47.6	20.5	12.7	10.9	10.5	26.1	41.3	72.0	96.6	108.4	98.1	71.4	616.1
S D	4.9	4.7	4.3	2.4	6.0	8.9	8.6	8.9	8.4	6.8	4.17	3.9	10.78
C.V	10.29	22.93	33.86	22.02	57.14	34.10	20.82	12.36	8.70	6.27	4.25	5.46	1.75
Max	56.7	27.4	21.7	14.1	17.6	36.3	50.6	86.4	108.5	119.1	106.2	76.3	632.5
Min	40.4	10.8	7.4	6.1	1.0	12.2	23.2	59.3	82.2	97.7	92.9	64.2	598.2

value does not differ much from those of the Penman EO_2 and the Thornthwaite Et, but it is above that of the water balance Et by 17 per cent. During the winter season, however, the value of the water balance Et as a percentage of the yearly value is above those of the Penman Et by 17 per cent, and it exceeds those of the Penman EO_2 and the Thornthwaite Et by 18 per cent and 14 per cent respectively

The high winter value of evapotranspiration obtained from the water balance method is due to the fact that evapotranspiration is assumed to be the difference in the values of precipitation and runoff during this season. Since there is a time lag in the release of excess rainfall from groundwater storage, the winter runoff is, therefore, underestimated. Thus evapotranspiration, representing the difference between precipitation and runoff, is over-estimated.

During the summer, on the other hand, some of the moisture stored in groundwater during the preceding winter, will be released as baseflow. The value of total runoff, therefore, will increase by this amount and evapotranspiration during the summer will be lower.

Referring to the curves of mean monthly data (Fig.39), it is observed that evapotranspiration from the water balance approach exceeds Thornthwaite Et during the period November to February, and is greater than Penman Et and EO_2 during October to February.

The Thornthwaite Et is higher than the Penman EO_2 value during the months of June to January inclusive. In October and November, the Thornthwaite formula gives values which are more than double those calculated by Penman EO_2 formula For example during October the Thornthwaite Et value is 47.6 mm while that of the Penman EO_2 is 22.1 mm. For November the respective values of the Thornthwaite Et and the Penman EO_2 are 20 5 mm and 9 5 mm.

During the spring months, however, the Thornthwaite Et value is below that of the Penman EO_2 (Tables 31 and 32) The explanation for the

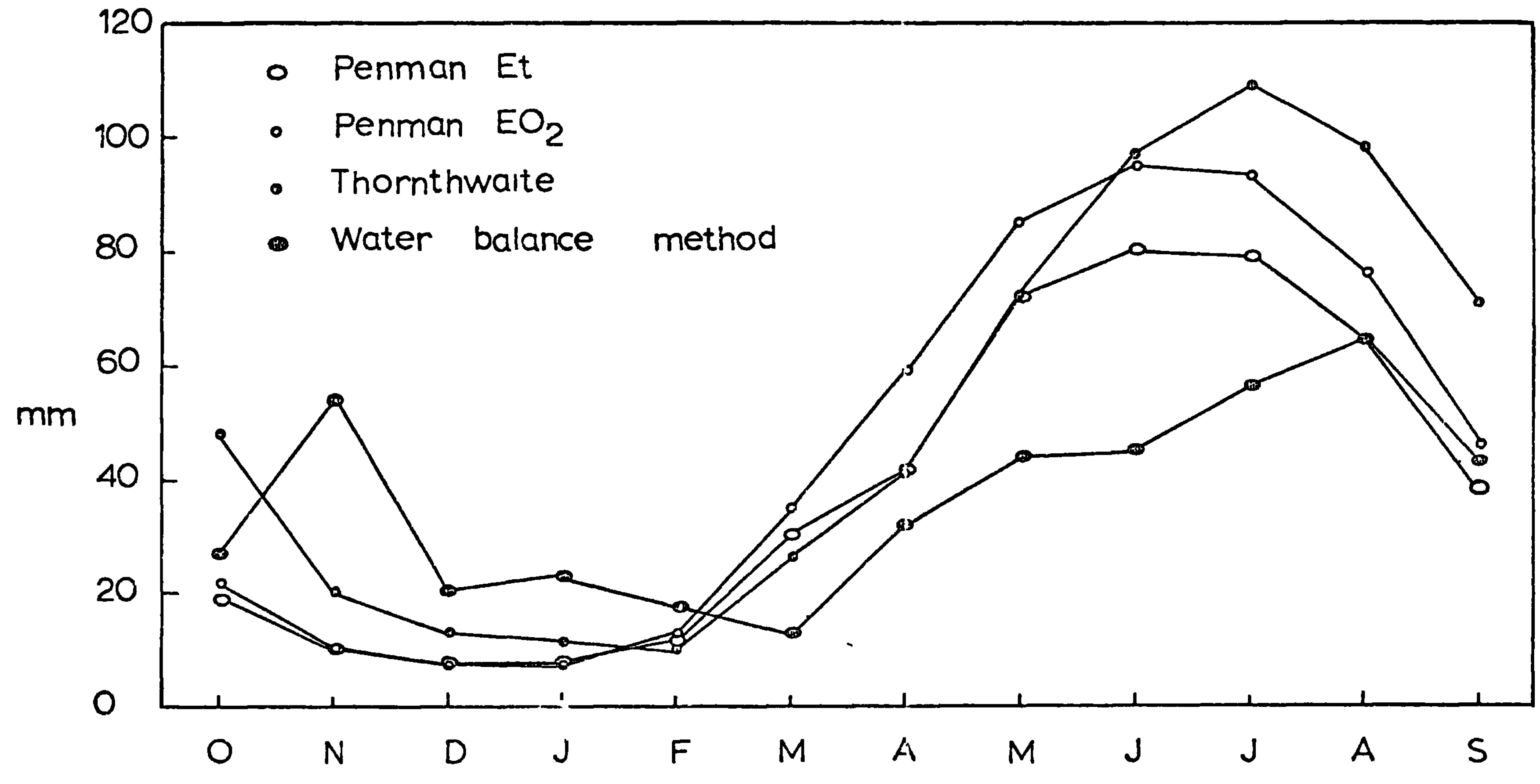


Fig 39 Average monthly evapotranspiration as determined by Penman Et, EO_2, Thornthwaite and water balance methods (1963-73)

low Thornthwaite Et value during the spring months and the high Et values during the autumn is attributed to the lag of temperature behind radiation.

The highest mean monthly value of the Thornthwaite Et is 108 0 mm in July, while those of the Penman EO_2 and Et are 95 2 mm and 80.0 mm in June. For the water balance Et, the highest mean monthly value is 64.2 mm and that occurs in the month of August

The lack of coincidence of the monthly peak values of evapotranspiration in the Thornthwaite and Penman formulae are explained by the fact that the former is dependent only on temperature, while the latter is based not only on temperature, but also sunshine, humidity and wind speed.

The highest mean monthly value of the water balance evapotranspiration during August as compared with other summer months is explained by the high rainfall occurring during this month

The monthly values of the Penman EO_2, the Thornthwaite Et, the Penman Et and the water balance Et are presented in Tables 31, 32, 33 and 34 The mean, maximum, minimum, standard deviation and coefficient of variation of the monthly values of each method may also be observed from these tables

According to these tables the highest monthly value of the Thornthwaite Et during the ten year period is 119 1 mm during July and the lowest is 1 0 mm during February Respective maximum and minimum values for the Penman EO_2, the Penman Et and the water balance Et are 113 6 mm in May and 2 3 mm in January, 99.4 mm in July and 2.0 mm in January and 106.1 mm in August and -26.2 mm in December

The negative value of -26 2 mm for evapotranspiration during December of the water year 1966 is explained by the high precipitation during the preceding month of November (184.7 mm) Much of the precipitation during November probably infiltrated into the soil and was released in December as groundwater flow Therefore, with a runoff value

Table 33· Monthly values of potential Et, (mm) by the Penman formula (Oct.63 - Sept.73)

month / water year	O	N	D	J	F	M	A	M	J	J	A	S	Total
1964	18.6	20 5	12.6	8.5	17 4	22.0	62.1	99 0	91.7	99 4	74 2	56 9	568.6
1965	19 9	16 0	14 8	14.8	20 4	25.4	59 5	85 0	96.7	70.2	73 2	39 4	535.3
1966	19 9	10 3	5 5	8.4	13 7	53.8	43 6	78 8	77.0	96.8	67 5	35 0	510.5
1967	11 0	7 4	3.6	2 0	8 3	20 0	38 2	46.9	66.4	62 4	49.9	28 0	344 0
1968	9 0	2 6	4.8	10 4	5 7	35 0	45.9	56 4	80 8	56 7	57.3	33 2	397 8
1969	19.4	6 6	8 7	5 7	7 2	19 8	54 4	54 8	77.7	87.8	70 4	40.0	452 6
1970	28.4	14 2	6.3	3 8	14 1	36.9	61 2	81 9	96 6	93 2	57 3	35.4	529 3
1971	12.8	9 2	4 8	6 6	7 2	26 4	45 2	74 8	69 9	83 0	53 3	39 4	432 5
1972	35.1	12.5	13 2	7 7	7 8	26.8	52 6	71.0	67.4	73 3	68 6	36 6	472 3
1973	16 6	11 1	5.9	9 3	14 7	33 0	47 8	70.4	76 1	68 1	68.5	40 3	461 8
Total	190 7	100 9	75 9	77 2	116 3	299 1	510 5	719.0	800 3	790 9	640.2	384 2	4704.7
Mean	19 1	10 1	7.6	7 7	11 5	29 9	51 0	71 9	80.0	79 0	64 0	38 4	470 5
S D.	7 9	3.9	3 7	3 6	5 1	10 3	8.2	15.7	11 4	15 0	8.7	7 5	68.3
C V	41 36	38 61	48 68	46.75	43 97	34 45	16 08	21 84	14 25	18.77	13 59	19 53	14.52
Max	35 0	16 0	14 8	14 8	20 4	53 8	62.1	99 0	96 7	99 4	74 2	56.9	568.6
Min	9 0	2.6	3 6	2 0	5 7	19 8	38 2	46 9	66.4	56 7	49 9	28 0	344 0

Table 34 Monthly values of Evapotranspiration (mm) by the water balance method (Oct.63 - Sept.73)

month / water year	O	N	D	J	F	M	A	M	J	J	A	S	Total
1964	30.0	62 8	4 6	0 2	10 2	37.2	27.4	13 4	67 1	30.0	58 0	32.7	373.8
1965	21 1	22 8	57 0	62 4	0.2	-2 7	40.4	35.3	34.6	79 8	48 6	87 8	487 1
1966	1 8	103 7	-26.16	12 2	15 1	-16 8	39 8	40 9	51 7	53.7	94 0	20.3	406.7
1967	54.7	32 4	8 0	8 0	35.0	10 0	27.3	78.4	27 8	66 0	81 7	51.2	480.2
1968	56.8	52 7	12.9	6 8	22 1	19.8	29.0	40.9	47.3	84.2	39 0	88.6	500 2
1969	49 6	14.9	41 8	2 9	57 3	5.6	10 7	52.0	56.4	33.5	60 7	51 0	436 4
1970	3.2	87 0	14 1	25 9	12 6	14.4	29 9	13 4	29.4	43.1	64 1	32 3	369 2
1971	23 6	66.5	43 0	14.5	0.9	32.1	34.8	34.8	60.8	43 0	106 1	5 0	465.2
1972	16 3	42.9	13 0	68 9	7 6	27.6	13.7	60.0	49 0	51.8	21.2	23.1	394.0
1973	14 2	51 3	27.0	24 9	8 1	6.7	66.8	70.4	30.3	73 2	68 2	40.4	481 4
Total	271 3	537.0	195 2	226.7	169.1	133 8	319 8	439 5	453.4	558 3	641 6	432.4	4390 2
Mean	27 1	53 7	19 5	22 7	16.9	13 4	32 0	44 0	45 3	55 8	64 2	43.2	439 0
S D	20 3	27 7	23.7	24 2	17 5	16.5	15 6	21 7	14.1	19 1	25.3	27.5	50.2
C.V	74 91	51 58	121 54	106.60	103.55	123 13	48 75	49 32	31 12	34 23	39.41	63 66	11 43
Max.	56 8	103 7	57.0	68 9	57 3	37 2	66 8	78.4	67 1	84 2	106 1	88 6	500 2
Min.	1 80	14 9	-26.2	0.2	0.2	-16 8	10.7	13 4	27.8	30 0	21 2	5 0	369 2

of 73 6 mm and low precipitation of 47 4 mm the evapotranspiration value was negative in December. On the other hand, during November, since precipitation was high and runoff was relatively low (81 0 mm), the value of evapotranspiration was therefore high as well (103 7 mm)

The scatter diagrams of the mean monthly water balance Et, the Thornthwaite Et and the Penman EO_2 values are presented in Fig 40

From a study of the correlation coefficients between the mean monthly values of the water balance Et, the Penman EO_2 and the Thornthwaite Et (Table 35) it can be observed that the water balance Et is more closely correlated with the Thornthwaite Et (correlation of 0 77) than with the Penman EO_2 value (correlation 0 64). This could possibly be due to the fact that the derivation of the Thornthwaite formula was originally based on the rainfall-runoff relationship

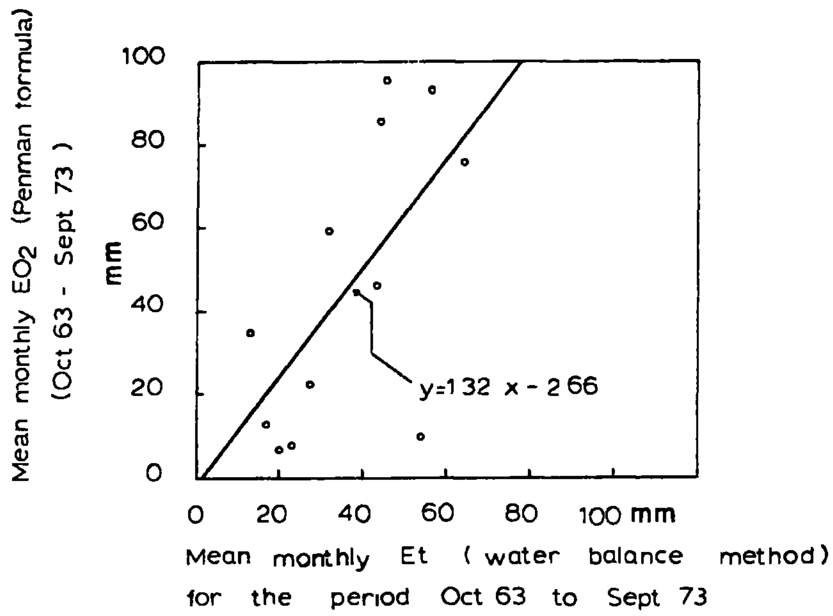
The monthly correlation coefficients of the water balance Et versus the monthly Penman EO_2 and the monthly Thornthwaite Et are both low These low correlation coefficients show that any monthly estimation of evapotranspiration by the water balance is not reliable

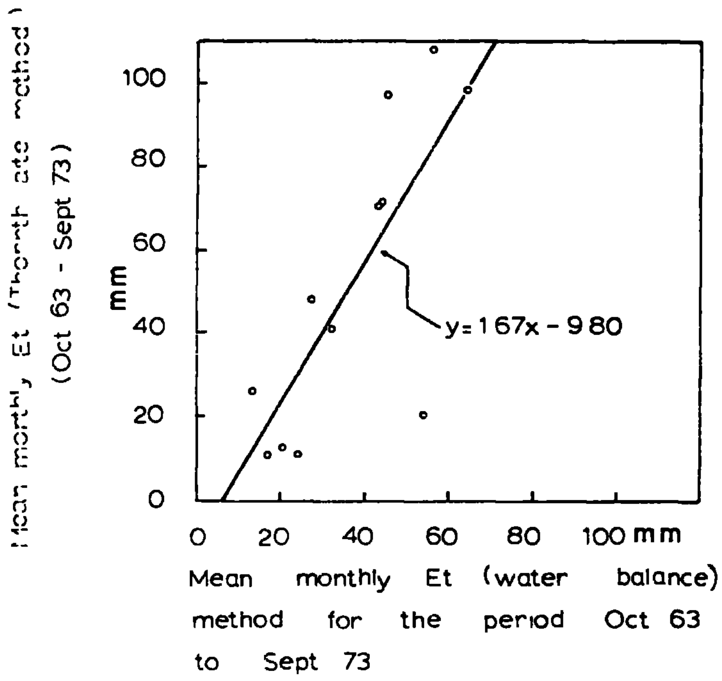
Table 35· Correlation coefficients between monthly and mean monthly values (10-year mean) of the Penman EO_2, the Thornthwaite Et and the water balance Et values, October 1963 - September 1973

Monthly Penman EO_2 versus monthly Thornthwaite Et	0 89
Monthly Penman EO_2 versus monthly water balance Et	0.31
Monthly Thornthwaite Et versus monthly water balance Et	0.43
Mean monthly Penman EO_2 versus mean monthly Thornthwaite Et	0 92
Mean monthly Penman EO_2 versus mean monthly water balance Et	0 64
Mean monthly Thornthwaite Et versus mean monthly water balance Et	0 77

Simple hydraulic lysimeters

Fig.41 represents the plot of net gain or net loss of water for the two replicates For the whole period of study, 15th May to 30th September 1973, the net loss of moisture from the two lysimeters was almost the same (64 mm and 66 mm) The trends of the two graphs also seem to be similar. However, there were periods


Mean monthly EO2 (Penman formula)
(Oct 63 - Sept 73)
mm
100
80
60
40
20
0
y=1 32 x - 2 66
0
20
40
60
80
100 mm
Mean monthly Et (water balance method)
for the period Oct 63 to Sept 73


(Oct 63 - Sept 73)
mm
100
80
60
40
20
0
y=1 67x - 9 80
0
20
40
60
80
100 mm
Mean monthly Et (water balance)
method for the period Oct 63
to Sept 73

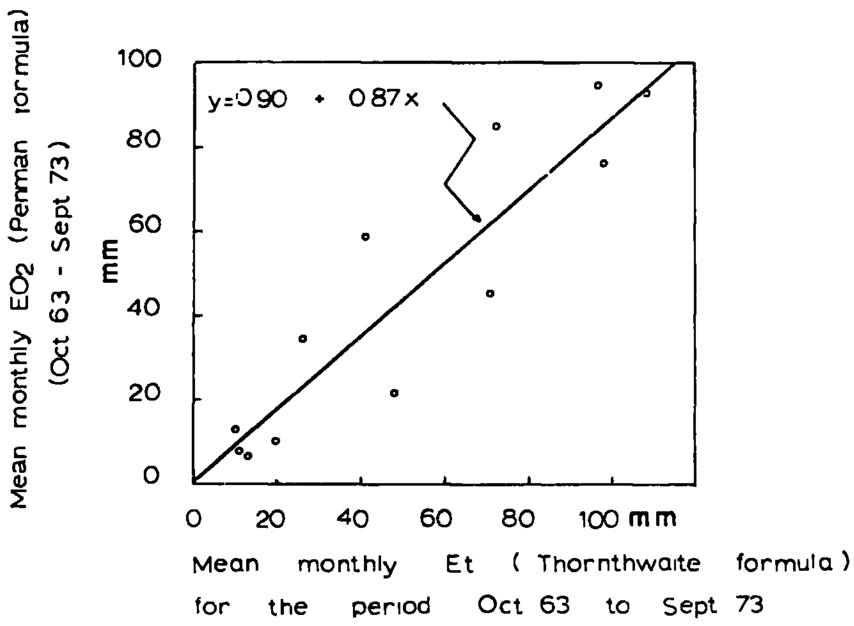

Mean monthly EO2 (Penman formula)
(Oct 63 - Sept 73)
mm
100
80
60
40
20
0
y=0 90 + 0 87x
0
20
40
60
80
100 mm
Mean monthly Et (Thornthwaite formula)
for the period Oct 63 to Sept 73

Fig. 40

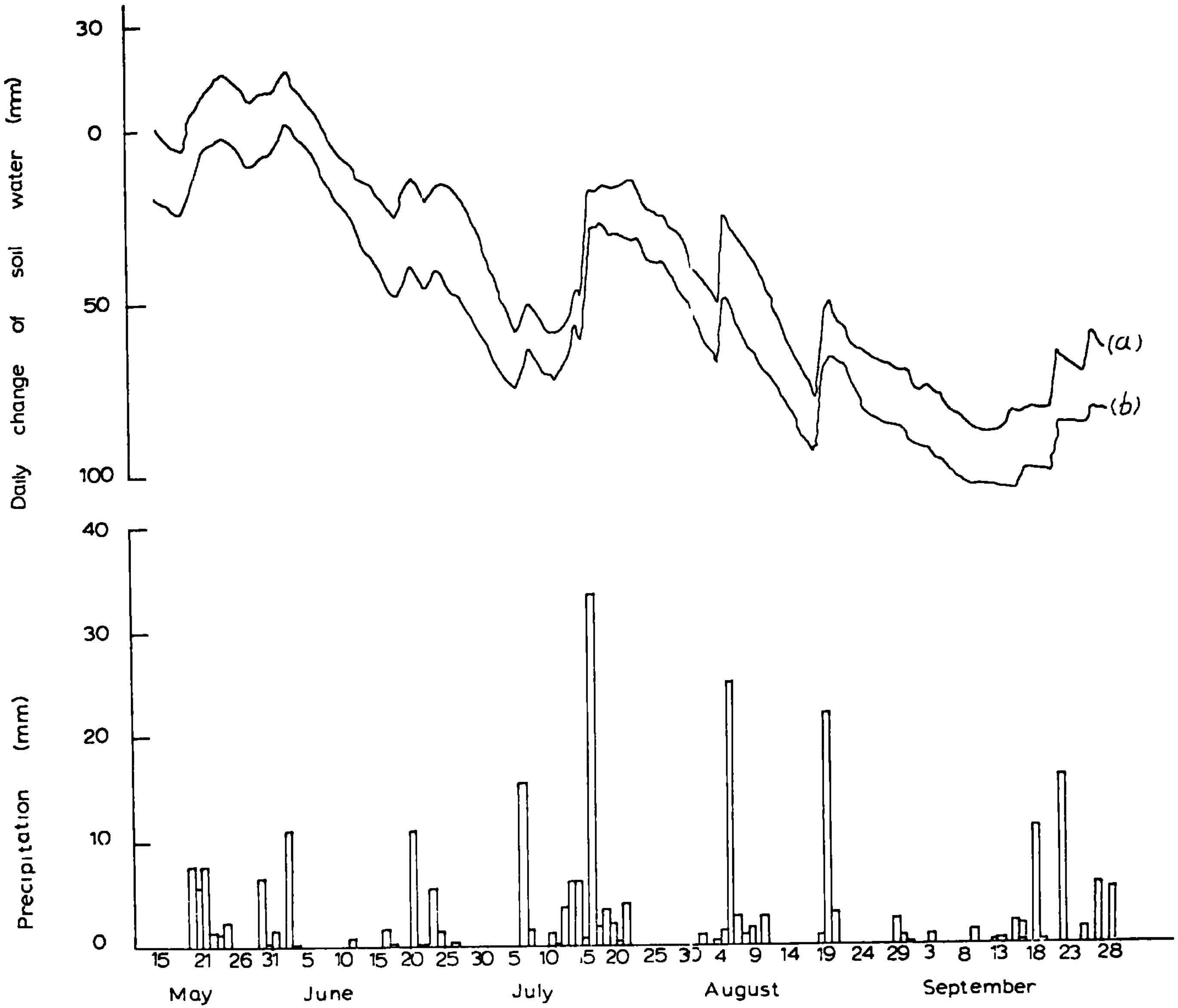


Fig 41 Plot of daily change in soil water in the two lysimeters and of daily depth of precipitation at Durham observatory during the period May 15 to September 30 1973

with a length of 20 days or so when the initial difference in the two manometer readings was more than doubled by the end of the period Taking the period starting 4th June, the depth of moisture, representing the initial difference in the two manometer readings, was about 14 mm. However, by the 23rd June, the depth of moisture equivalent to the difference in the two manometer readings was 30 mm Thus, it is observed that the daily rate of actual evapotranspiration from one of the lysimeters during the period 4th June to 23rd June was about 0.75 mm greater than the other one in the same period Since the management of the lysimeters has been similar during the whole period of this study, therefore, such a discrepancy in the results can only be attributed to errors in measurement by the lysimeters The accuracy of the lysimeters in measuring daily changes in soil moisture content may be studied by considering the recorded depths of the major rainfall amounts occurring on three different days

On 17th July, the daily raingauge recorded a rainfall depth of 33 mm. From the lysimeters, the fluctuation of one of the manometers was 49 mm and the other one was 45 mm. Thus for the former, the gain of moisture was about 33 mm, whereas for the latter, the gain was 30 mm. A difference of 3 mm in this case might not result in serious error. However, for rainless days with evapotranspiration less than 3 mm, a difference of that magnitude is very significant

Other examples are for rainy days of 5th and 19th August The depth of rainfall for 5th August was 25.2 mm and for 19th August was 22 2 mm The corresponding moisture gains for these two rainfall depths were 20.0 mm and 24 7 mm for one lysimeter and 25 3 mm and 22.7 mm for the other. Thus the gain of moisture by one of the two lysimeters has been identical to the rainfall depth measured by the raingauge, while for the other there have been discrepancies of about 20 per cent and 9 per cent in the measurement of the two rainfall depths.

Winter (1963) has suggested the use of several replicates for any measurement of evapotranspiration, so that accidental errors could be removed Using this suggestion, the average values of the loss and gain of moisture from the two lysimeters were used to plot the daily gain and loss of water for the period. The graph of the daily values of the Penman potential evaporation (EO_2) and rainfall has been plotted for comparison in Fig 42.

It can be observed from Fig 42 that the monthly Penman EO_2 during the period (15th-31st) May, June and July is higher than that from the lysimeter. During the period (15th-31st) May the lysimeter readings are lower than the Penman EO_2 value by 45 per cent. During June and July the percentage under-estimation is 14 and 9 respectively In August, on the other hand, the lysimeter value is greater than that of the Penman EO_2 by 7 9 per cent and it is almost the same during September

The relatively lower values of evapotranspiration obtained from the lysimeter in the months of May, June and July could possibly be explained by the fact that the lysimeter values are actual evapotranspiration and owing to the shortage of water during a short period, the rate has not been at its maximum However, the fact that actual evapotranspiration values during August exceed potential evapotranspiration by 7 9 per cent, can only be explained by lack of accuracy of the instrument.

The measurement of actual evapotranspiration by lysimeters was discontinued in October The reasons for this were as follows

1. The lack of accuracy of the lysimeters in the measurement of actual evapotranspiration as was discussed by the results of the two lysimeters during the period of study.

2 Owing to low evapotranspiration and frequent rain during the winter season, October to March, the lysimeters should be frequently drained This results in a disturbance of the balance of the soil container on the scooter tube and constant rebalancing of the instrument is needed

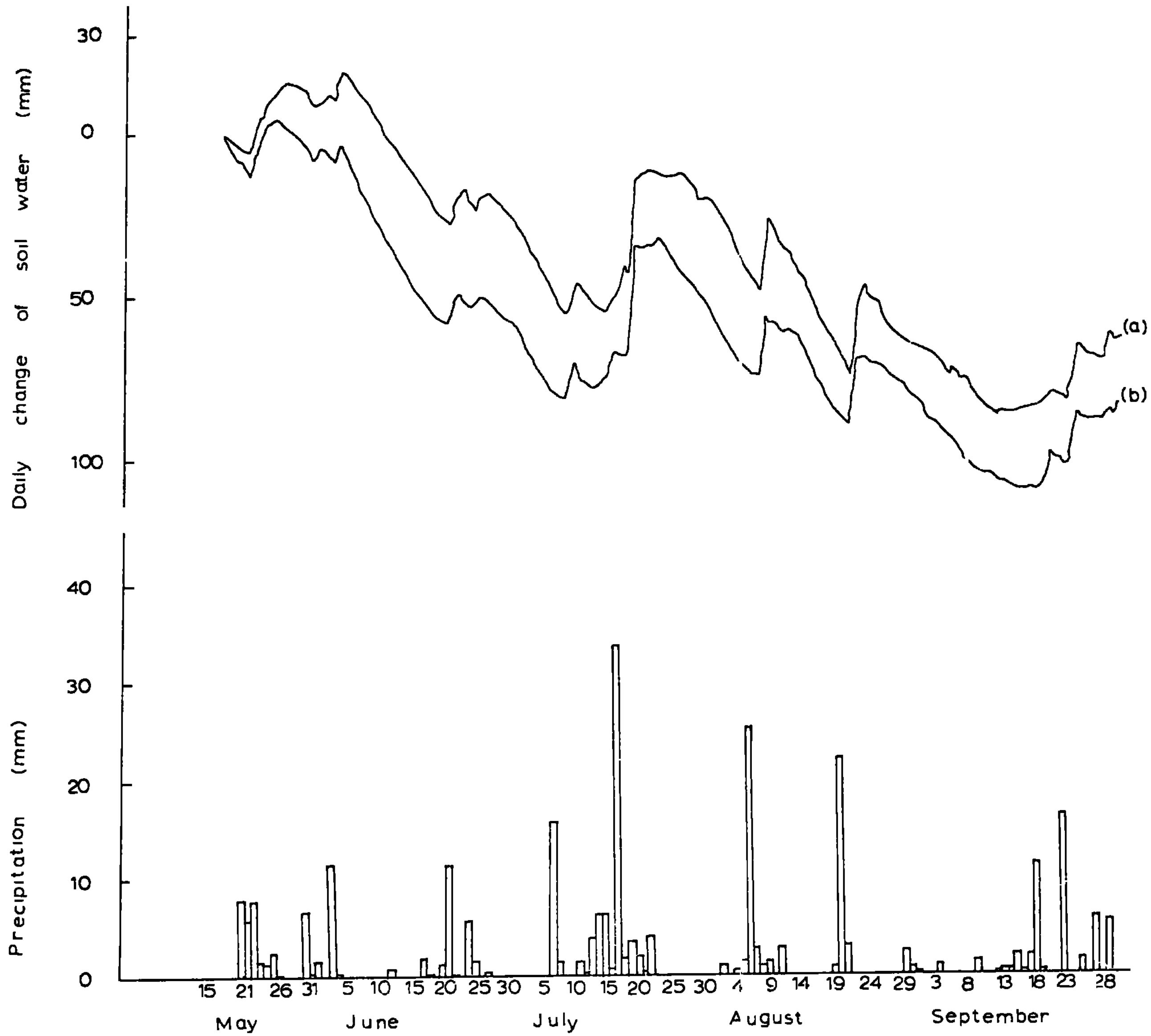


Fig 42 Plot of daily change in soil water based on average of the two lysimeters (a) and Penman EO_2 and precipitation (b) at Durham observatory during the period May 15 to September 30 1973

3. During the winter seasons the rate of actual evapotranspiration is almost equal to potential evapotranspiration and therefore measurements of actual evapotranspiration by lysimeters is not necessary

4. Any extrapolation of the results of the point measurement of the actual Et to the catchment is rather difficult This is explained by spatial variations in soil type and soil moisture content in the field The plant roots in the small lysimeter may be restricted by the container, whereas in the field such a restriction does not occur. The structure of the soil in the lysimeter might not be similar to that in the field Therefore the distribution of moisture will be different This would affect the availability of moisture, and consequently the measured actual evapotranspiration values

Areal values of actual evapotranspiration

For the study of the areal values of actual evapotranspiration, the monthly precipitation and the monthly Penman EO_2 values over the period October 1963 to September 1973 were studied. Using these data and the Penman method for the estimation of actual evapotranspiration, three water balance maps for the catchment were drawn (Figs 43, 44, and 45)

Fig 43 shows the average water balance diagram of the catchment for the period October 1963 to September 1973 The monthly precipitation and potential evapotranspiration values used (Penman EO_2) are the average of the 10 years. The calculations for the actual evapotranspiration are shown in Table 36

From the Fig 43 it is observed that according to the Penman model, actual evapotranspiration is always equal to potential evapotranspiration This is because the maximum cumulative moisture deficiency during the period does not exceed the limit of 100 mm available water for short rooted crops In fact the maximum cumulative moisture deficiency is 79.6 mm and that occurs in July The moisture deficiency starts in May and ends in October Figs 43 and 44 show the water balance diagrams of the

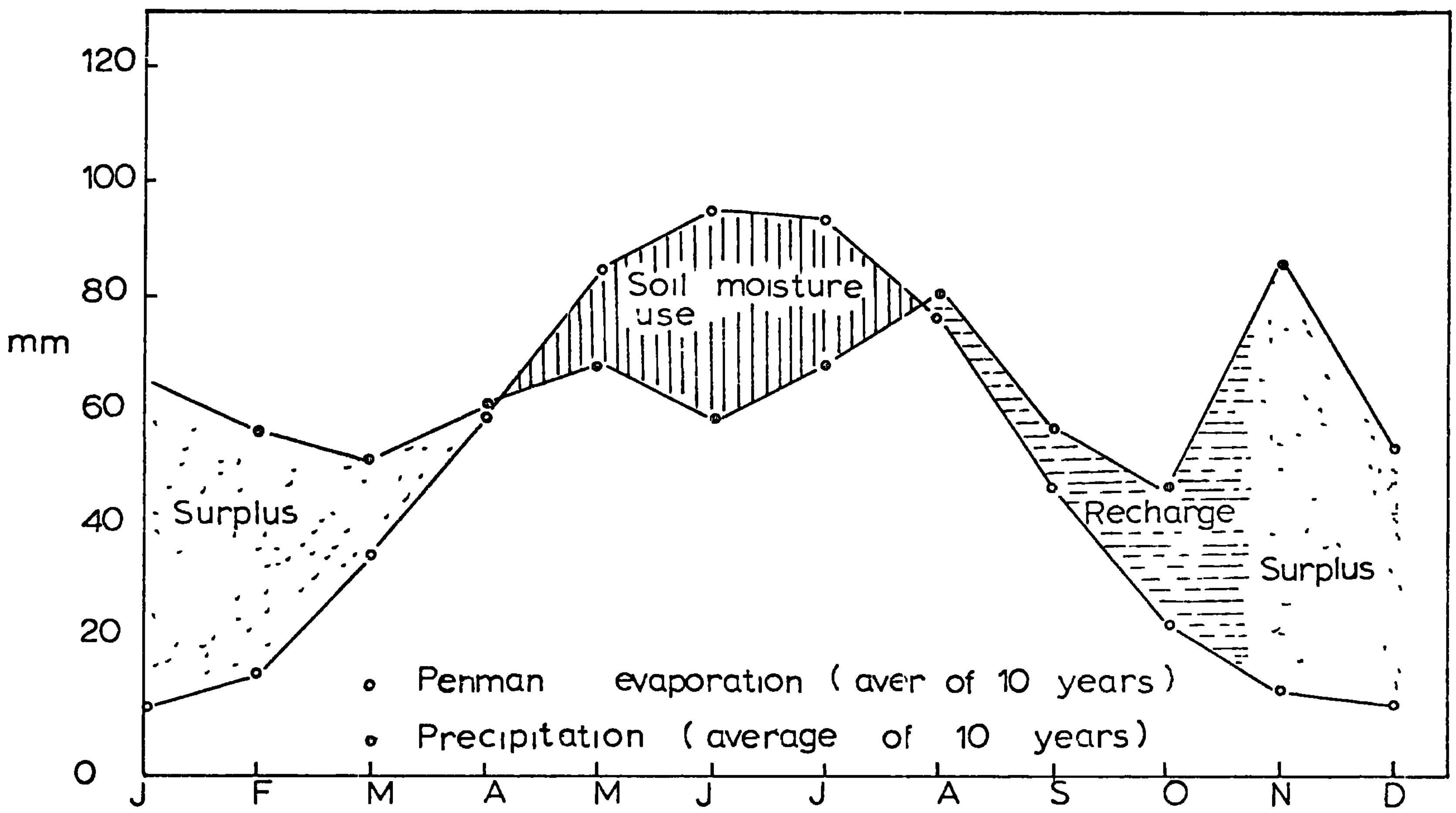


Fig 43 Diagram of average moisture balance for the period 1964-73 in the Browney basin

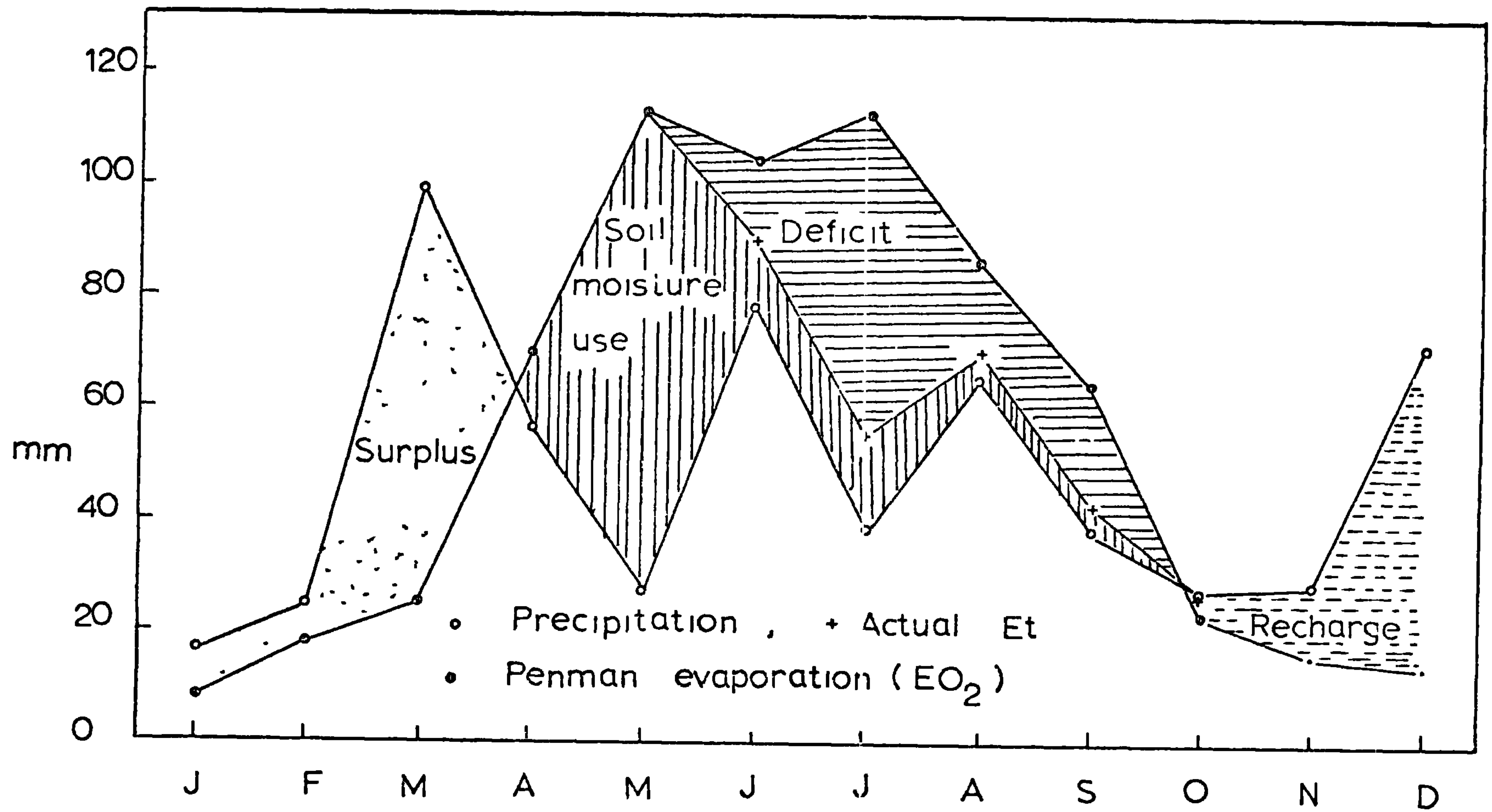


Fig 44 Diagram of moisture balance for Browney Basin (1964)

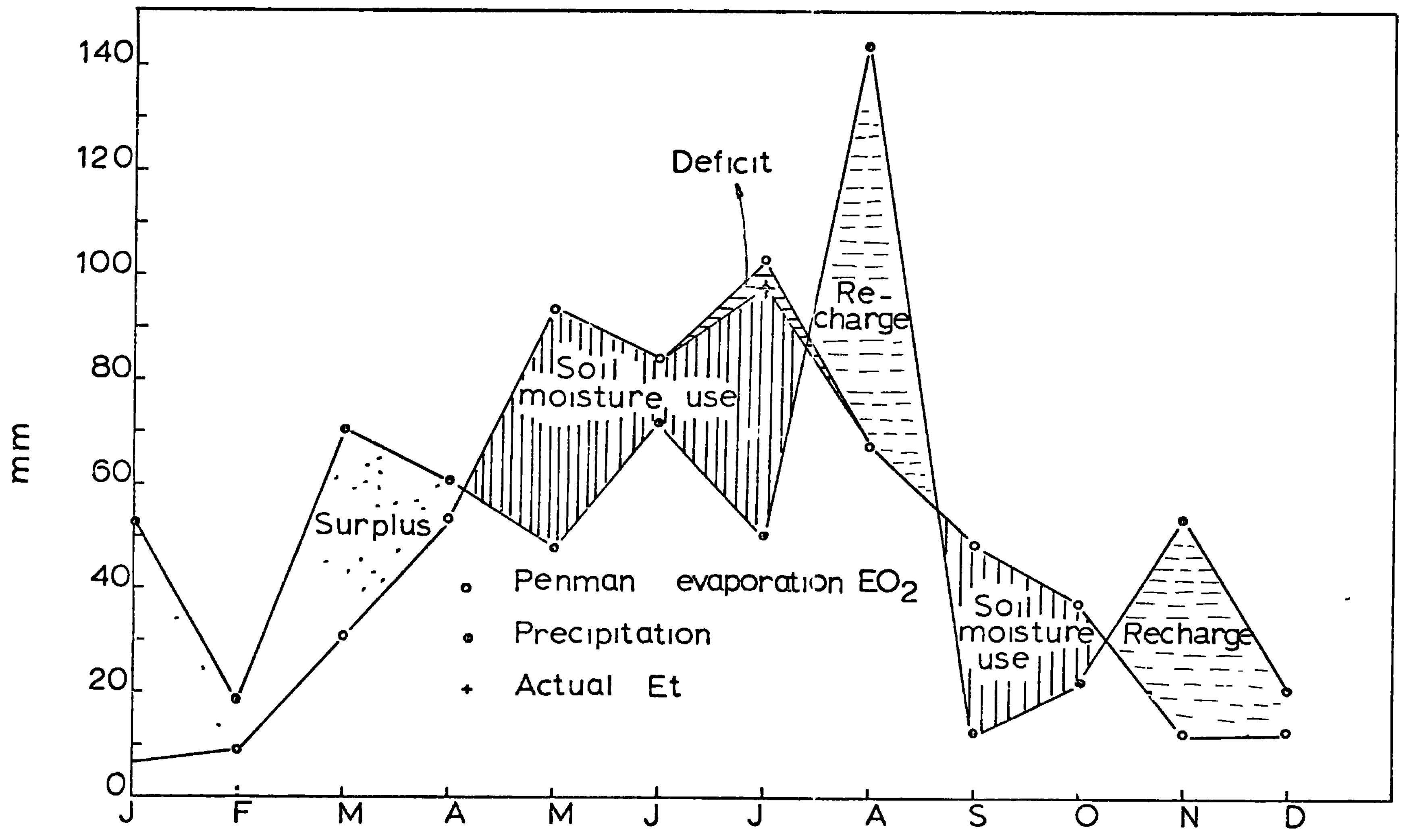


Fig 45 Diagram of moisture balance for Browney Basin (1971)

Table 36. The calculation of areal values of actual evapotranspiration by the use of the Penman method for the ten year period 1964-1973

Month	Precipitation (R) (mm)	Evapotranspiration (EO_2) (mm)	$R-EO_2$ (mm)	Cumulative Moisture Deficiency (mm)	Actual Evapotranspiration in (mm) Shallow water table zone (5%)	Deep rooted crop zone (10%)	Shallow rooted crop zone (85%)	Areal Value
January	64 5	7 5	+57 0	0	7.5	7.5	7.5	7 5
February	56 3	12 8	+43 5	0	12 8	12.8	12 8	12.8
March	50.8	34 5	+16.3	0	34 5	34 5	34.5	34 5
April	60 5	59.1	+ 1 4	0	59 1	59.1	59 1	59.1
May	68.2	85 0	-16 8	16.8	85 0	85 0	85 0	85 0
June	58.1	95.2	-37 1	53.9	95.2	95 2	95.2	95.2
July	67 6	93.3	-25 7	79.6	93 3	93 3	93.3	93.[illegible]
August	80 9	76.4	+ 4 5	75.1	76 4	76 4	76 4	76.4
September	56.0	46.1	+ 9.9	65 2	46 1	46 1	46.1	46 1
October	45.2	22.1	+23 1	42 1	22 1	22.1	22 1	22.1
November	87 2	9.9	+77 3	0	9 9	9.9	9 9	9 9
December	53 4	6.9	+46 3	0	6.9	6 9	6.9	6.9

years 1964 and 1971 The calculations are shown in Tables 21 and 37 The year 1964 is a very dry year, whereas that of 1971 is a year with appreciable rain during the summer For the year 1964 (Fig 44), actual evapotranspiration falls below potential evapotranspiration in May with a maximum of 56.8 mm in July. The total yearly difference between the actual and potential evapotranspiration is 109.2 mm, or actual evapotranspiration is less than potential by 16.8 per cent.

Considering the year 1971 (Fig.45), the actual evapotranspiration falls below the potential evapotranspiration in July by 5.1 mm only. Referring to the figure of the moisture balance for the year 1964, it is observed that the curve of potential evapotranspiration exceeds that of precipitation in April From then onward moisture for evapotranspiration is being supplied from the soil moisture reservoir at the potential rate. By May, the rate of actual evapotranspiration drops below the potential evapotranspiration rate This reaches a maximum of 56 8 mm in July. The divergence, then, decreases from July reaching 15.8 mm in August and with a slight increase in September, it reaches zero in October. From October, there is excess moisture owing to low evapotranspiration, which fills the soil moisture reservoir. The recharge of soil moisture continues until the reservoir is filled. The surplus moisture is that in excess of the deficiency of soil moisture.

For the year 1971 the period of moisture surplus continues until April. From April to July some moisture for the evapotranspiration demand is supplied from the soil moisture storage. During July, actual evapotranspiration drops below the potential evapotranspiration rate. This is because the cumulative moisture deficiency exceeds the limit of available moisture in the soil

In concluding this section, two of the main limitations of the Penman model for estimating actual Et should be discussed.

One of these limitations is that there is no provision made for

Table 37· The calculation of areal values of actual evapotranspiration by the use of the Penman method for the year 1964

Month	Precipitation (R) (mm)	Evapotranspiration(EO_2) (mm)	$R-EO_2$ (mm)	Cumulative Moisture Deficiency (mm)	Actual Evapotranspiration in (mm)			
					Shallow water table zone (5%)	Deep rooted crop zone (10%)	Shallow rooted crop zone (85%)	Areal Value
January	16.5	8 1	+ 8 4	0	8 1	8 1	8 1	8 1
February	25 4	18 0	+ 7 4	0	18 0	18 0	18 0	18 0
March	99 3	24 8	+74.5	0	24 8	24 8	24 8	24 8
April	55.6	69 4	-13 8	13 8	69.4	69 4	69 4	69 4
May	27.0	113 6	-86.6	100.4	113 6	113.6	112.6	112.8
June	78.5	104 4	-25 9	126.3	104.4	104.4	87 6	90.1
July	38 0	112 3	-74.3	201 2	112 3	112 3	45.5	55.5
August	65 0	85 6	-20 6	221 8	85 6	85 6	67 0	69 8
September	38 3	63.9	-25 6	247 4	63 9	49 3	40.3	42 4
October	26.5	22 7	+ 3 8	243 6	22 7	22 7	22.7	22.7
November	27.8	15 2	+12 6	231 0	15 2	15.2	15.2	15.2
December	71 1	13 5	+57 6	173.4	13 5	13.5	13 5	13 5

runoff in the model This limitation might result in some errors

The second limitation arises from the effect of uneven temporal distribution of rain Since the calculation of actual evapotranspiration is based on monthly rainfall and evapotranspiration, therefore if for any particular month the bulk of rainfall occurs at the end of the month following a dry spell, actual evapotranspiration might be over-estimated This is because during the dry spell, the cumulative moisture deficiency might exceed the limit of available water, and, therefore, the actual evapotranspiration rate might drop below the potential evapotranspiration rate Using the monthly values and assuming even temporal distribution of the rain, however, this effect of uneven distribution of rain will be masked

CHAPTER SIX

RUNOFF AND ITS VARIATION

Runoff is that portion of precipitation which reaches the stream channel. It could also be described as the end product of a number of physical processes which form the runoff cycle. This cycle starts with incident precipitation on a portion of land surface. Precipitation, subsequently, goes through the processes of interception, infiltration, storage in the surface depression or within the soil profiles and underlying rocks, evapotranspiration from storages of moisture, overland flow, interflow and groundwater flow, temporary storage in the stream channels, and finally movement of total flow from the upstream portion to the drainage basin outlet Total flow is then measured as discharge at the outlet of the basin.

Measurement of discharge of River Browney The discharge of the River Browney is measured at Burn Hall (259 387) Measurement is by means of a compound broad crested weir This has an overall width of 692 inches (17 58 m) and a central low level crest 215.5 inches (5 47 m) long The level of station is approximately 44 m above O.D and water level is continuously recorded Rating of the station is by weir formula and this is checked by current meter.

Yearly and seasonal runoff variations Discharge measurements were recorded in ft^3/sec between 1957 and 1967 and in m^3/sec since 1967 In this study, however, the discharge values are either converted into millimetres of runoff or used in m^3/sec.

To study the variation in yearly runoff, the water year starting October and ending September has been used. The histogram of yearly runoff, thus obtained, is shown in Fig.46. Runoff ranges between 514 4 mm in the water year 1969 to 136 0 mm in the water year 1973 The mean yearly runoff is 309 8 mm The maximum yearly runoff is 166.0 per cent of the mean and the minimum is 43 9 per cent of the mean

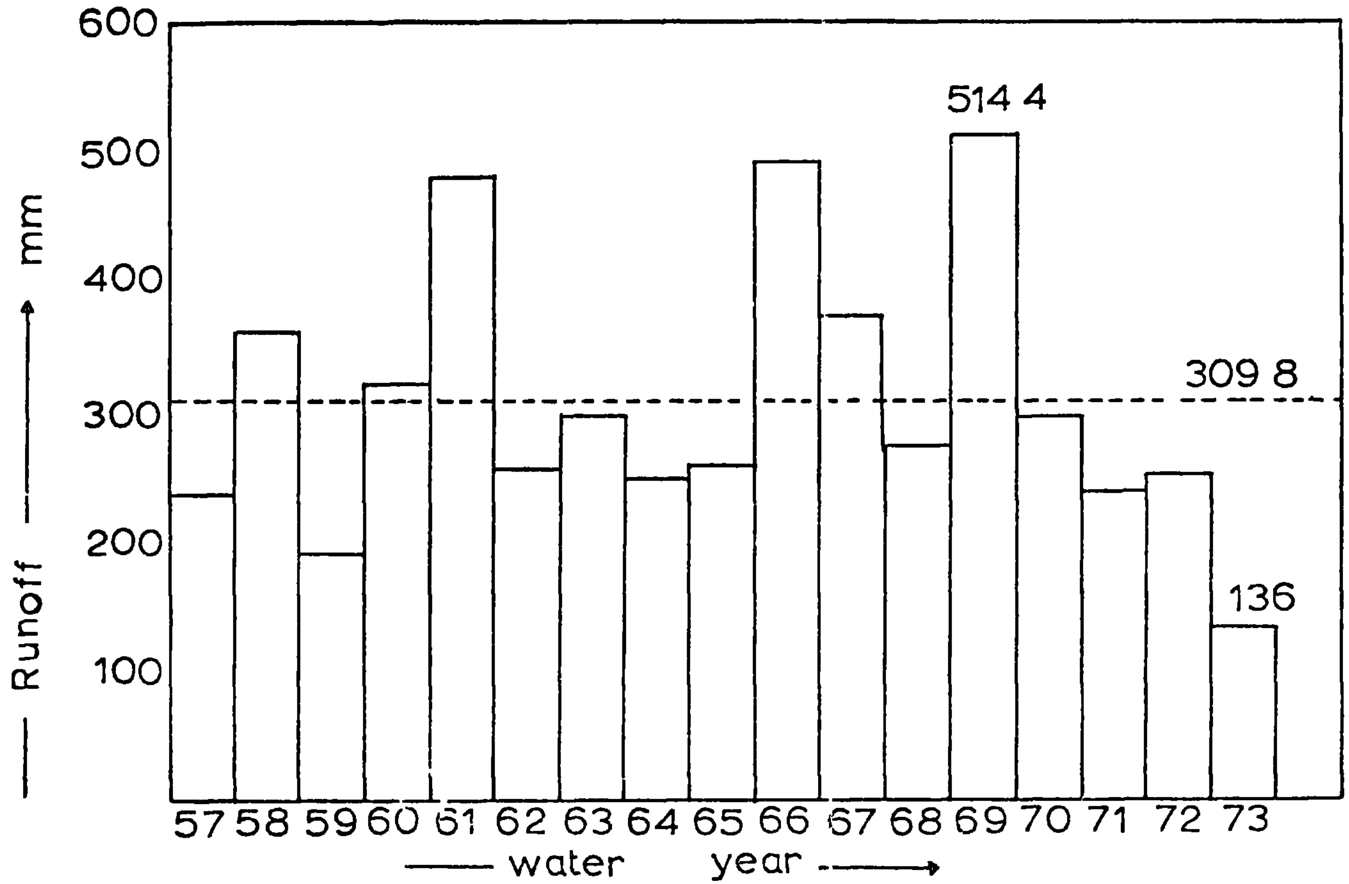


Fig 46 Histogram of yearly runoff for the River Browney at Burn Hall during the period 1957 - 73

The values of yearly runoff in six years have been above the mean and the remaining eleven years have had values less than the mean during the period 1957-1973. To explain some of the variations of yearly runoff during the period 1957-1973, the corresponding precipitation values should be considered. This is because total yearly precipitation and its distribution is the most important factor affecting runoff

Precipitation over the catchment, as mentioned earlier, has been measured by seven daily gauges since 1968. For the period 1962-1968, however, the records from four raingauges at Durham, Waskerley, Waterhouses and Satley were available Prior to 1962, the only stations whose records were accessible, were Durham Observatory and Waskerley

In order to use consistent rainfall data for the study of rainfall-runoff relationships during the 17 year period, Durham Observatory records corrected by a factor K_1 to change point rainfalls to areal rainfall was used

The parameter K_1 represents the ratio of average rainfalls over the catchment to that at Durham Observatory To find this parameter, monthly Thiessen average rainfall from the seven available raingauges throughout the catchment were summed over the period 1968-1972 and the result was divided by the value for the same period from the raingauge at Durham Observatory i.e $K_1 = \frac{3660\ 3}{3136.9} = 1\ 17$. Derivation of K_1 was based on a five year record because the data from all the gauges were not available for a longer period However the year to year variation of K_1 based on these five years of record was small i e (1.19-1 15).

Yearly precipitation values at Durham Observatory, thus, were multiplied by this factor in order to get the catchment value.

The values of yearly precipitation corresponding to the years of maximum and minimum runoff were 953 4 and 578 8 mm (Table 38). The runoff-rainfall ratio for each year was calculated For the water year 1969, which had the maximum total runoff, the ratio was 0 54, and for

Table 38 Yearly rainfall, runoff and runoff-rainfall ratio for the Browney basin during the period 1957-1973

Water year	Rainfall (mm)	Runoff (mm)	Runoff/Rainfall
1957	630.9	237.4	0.38
1958	695.0	365.3	0.53
1959	438.4	193.0	0 45
1960	828.9	323.5	0 40
1961	956.5	484.0	0.51
1962	722.6	255.0	0.35
1963	769.2	297.4	0 39
1964	658.9	251.3	0.39
1965	805 7	264.2	0 33
1966	946.5	496.0	0.52
1967	835 6	377.3	0 46
1968	837.7	278.8	0.34
1969	953.4	514.4	0.54
1970	662.7	298 9	0 46
1971	713.5	239.8	0.34
1972	658.0	252.9	0 39
1973	578.8	136.0	0.23
			Mean 0.41

the water year 1973, which had the minimum total runoff, the ratio was 0.23 This difference in the ratio of runoff-rainfall could be explained by considering the distribution of rainfall throughout the two years (Table 39)

Table 39 Rainfall distribution (mm) for the water years 1969 and 1973

month / year	O	N	D	J	F	M	A	M	J	J	A	S	Total
1969	75.8	76.4	113.5	55.9	103.4	100.4	52.9	122.1	91.3	51.7	41.9	68.1	953.4
1973	13.9	61.5	30.4	31.9	14.2	10.6	82.5	66.5	41.2	91.6	77.4	57.1	578.8

For the water year 1973, 72 per cent of the yearly precipitation occurred during the summer months a period when evapotranspiration values are high. During the water year 1969, however, 44 per cent of the precipitation occurred during the summer Thus, since the infiltration rate is generally higher during summer than winter (owing to lower moisture content and higher temperatures) much of the precipitation occurring during the summer season enters the soil zone and is subsequently used as evapotranspiration. During the winter, however, the rate of water entry into the soil is low, and since the evapotranspiration rate is also low, most of the precipitation will result in direct runoff Some precipitation which infiltrates, however, is stored in the soil zone or percolates into underground storage and is released during the summer season. The explanation, thus given, accounts for the low yearly runoff during water year 1973.

The average annual runoff-rainfall ratio for the period of 1957-1973 is 0.41 e.g. yearly runoff forms 41 per cent of the total precipitation. Considering the seasonal values of the runoff-rainfall ratio, it is observed that during the winter season, the ratio ranges from 0.79 in the water year 1958 to 0.36 in the water year 1973. This difference in

Table 40 Seasonal rainfall, runoff and runoff-rainfall ratio for the Browney basin during the period 1957-1973

Water year	Summer (April-Sept.)			Winter (Oct -March)		
	Rainfall (mm)	Runoff (mm)	Runoff/ rainfall	Rainfall (mm)	Runoff (mm)	Runoff/ rainfall
1957	372 9	78 4	0.21	258.0	158.9	0.62
1958	413.8	143.0	0.35	281 2	222.4	0.79
1959	196 3	40 3	0.20	242.1	152.6	0.64
1960	359.9	72.4	0 20	469.0	251 1	0.54
1961	434 0	110.2	0 25	522.5	373 8	0 72
1962	396 0	76 9	0 19	326 5	178 1	0.55
1963	369.1	100.5	0.27	400.0	196 9	0 50
1964	313.0	74 1	0.24	346 0	177 1	0.52
1965	468 4	123 9	0 26	337.3	140.2	0 42
1966	475.2	144.9	0 30	471 3	351 0	0.65
1967	476.2	134.6	0.28	359.4	242.8	0 68
1968	472.9	103 1	0 22	364 8	175.7	0.48
1969	428.0	168 4	0.39	537.1	346 0	0 64
1970	274 6	70 8	0.26	388.1	228.1	0.59
1971	384 1	101.9	0.26	329 4	137.9	0 42
1972	290.6	66.2	0.23	367.4	186 8	0 52
1973	416.3	80.0	0 19	162.6	56.0	0.36
	Mean for period		0 25			0.57

the runoff-rainfall ratios could be explained by the conditions of soil moisture storage due to precipitation in the last few months preceding each water year The total precipitation during the months of July, August and September 1957 and 1972 is 280 2 mm and 118.0 mm respectively. Thus the soil moisture storage at the beginning of the water year 1958 is expected to be higher than that of water year 1973 The soil during the winter season of the water year 1973, therefore, has got a greater capacity for storing moisture than during the winter season of the water year 1958 and the runoff-rainfall ratio for the winter season 1973 is,therefore, lower than that of 1958

The average runoff-rainfall ratio for the winter season during the period 1957-1973 is 0.57, or runoff during winter forms 57 per cent of the precipitation The respective value for the summer season is 0.25 and the range is 0 19 to 0 39

Rainfall-runoff equations The rainfall-runoff relationship was also studied by plotting the yearly runoff values against the corresponding rainfall amounts The equation of the line of best fit was found to be of the form $y = 0.64x - 167.68$. The correlation coefficient between the yearly rainfall values and the yearly runoff values was 0 84 (Fig 47) This value for the correlation coefficient shows that there is a high degree of association between rainfall and runoff.

The coefficient of determination, r^2, is 0.706 Therefore, about 70 6 per cent of the variance is accounted for by this regression equation. It is also clear, though, that this simple regression equation of rainfall and runoff accounts for more than 70 per cent of the variance, the extreme values scattered in the graph could only be accounted for by considering the processes of evapotranspiration, infiltration, and changes in soil moisture and groundwater storages

Two other points which might also explain the extreme values are the result of the areal rainfall assumption and the effect of mine water

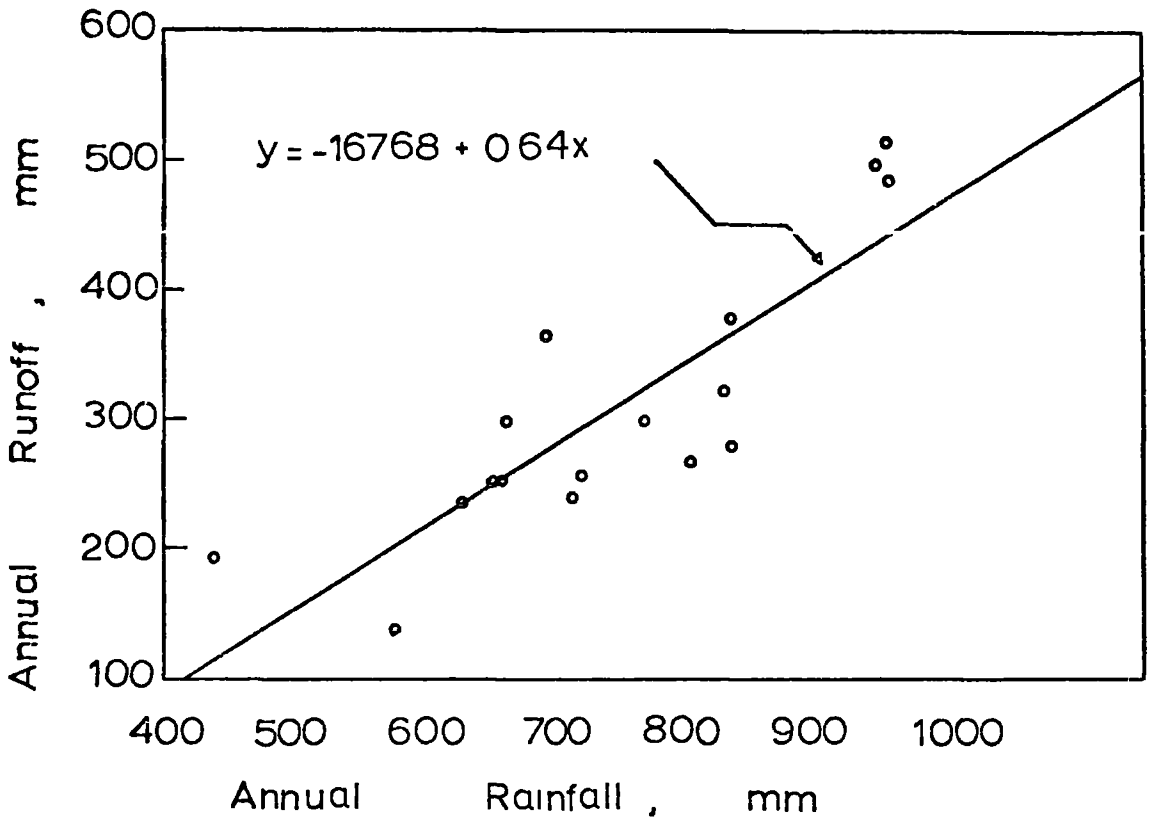


Fig 47 Relationship between rainfall and runoff for the Browney catchment using annual values during the period 1957 - 73

discharge on the total runoff.

The areal precipitation, as explained earlier, is the product of the catch of rain at Durham Observatory multiplied by a factor K_1 This factor representing the ratio of the Thiessen average of rainfall for the seven raingauges to that at Durham, varies from year to year and from storm to storm The value of this ratio, for example, is 1.19 for the year 1972, as compared with 1 15 for the year 1969 and 1 17 for the five year average (1968-1972).

The effect of mine water discharge into the river might also be important Unfortunately, no data have been available on the total discharge except for the year 1973 (calendar year), when the total yearly discharge of mine water, as given by the National Coal Board (Private Communication 1974) has been 13.6 mm or 4 4 per cent of the yearly average for the whole period

In order to investigate the seasonal rainfall-runoff relationships, seasonal runoff values were plotted against the corresponding rainfall data The line of the best fit of the scatter diagram for the winter season (October to March) had an equation of the form y = 0 75 x -60.95 while that for the summer season was y = 0.34 x -30.8. From a comparison of the seasonal scatter diagrams (Fig 48) of the rainfall-runoff relationships, it is observed that the winter diagram shows the points to be more scattered than the summer one Similar results were obtained by Smith (1964b) in his study of seasonal rainfall-runoff relationships in two Pennine catchments

Such observations are contrary to what one might expect. During the winter since evapotranspiration is low, a less scattered rainfall-runoff relationship is expected. Ward (1967) explains this anomaly by the effect of a high soil moisture deficiency at the beginning of winter season, as compared with a high moisture content (field capacity) at the beginning of summer season

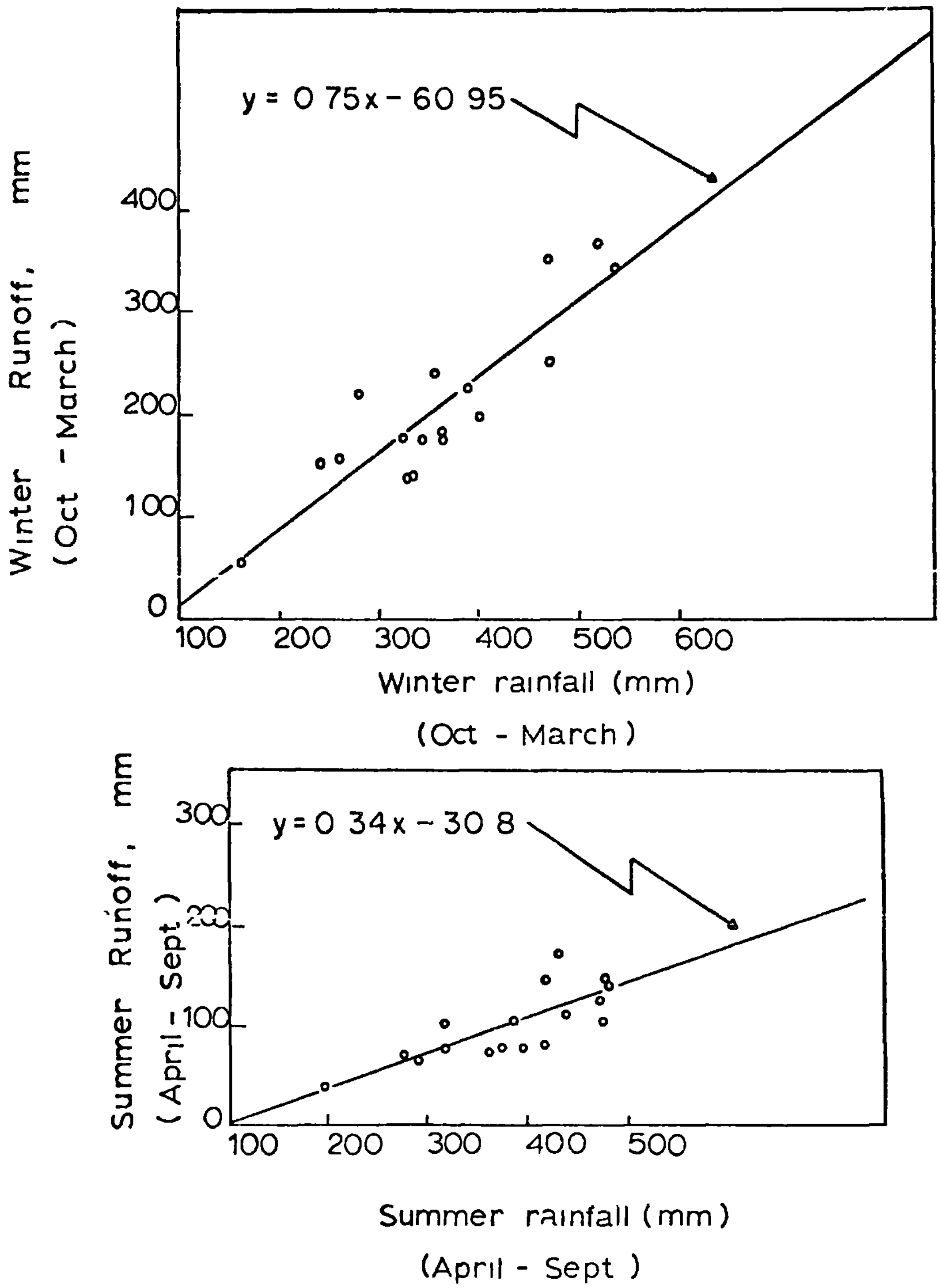


Fig 48 Relationship between rainfall and runoff for the Browney catchment using six - monthly values during the period 1957 - 73

Monthly runoff values To study the monthly values of runoff, the hydrograph of monthly streamflow for the period October 1956 to September 1973 is shown in Fig.49 From this graph it is seen that the highest monthly runoff is in October 1960. The value of runoff for this month is 249 m^3/sec or 120 mm The minimum monthly value occurs in October 1959 with a value of 6 m^3/sec or about 3 00 mm. Comparing these values with the mean monthly value for the whole period e.g 25.8 mm, it is observed that the maximum monthly runoff value is 4.65 times the mean while the minimum is 0.12 times the mean. Considering the maximum monthly runoff in each year, it is observed that January, February and March are the months with the highest frequency of occurrence of maximum monthly runoff per year. Each has been recorded to have the maximum monthly flow in four out of seventeen years This might be explained by the low infiltration capacity of the soil (high moisture content) and low evapotranspiration during these months.

In only one year out of the seventeen years of runoff study has the monthly maximum occurred in a summer month This was May 1973. 1973, as mentioned earlier, was an exceptionally dry year with 72 per cent of the precipitation occurring during the summer months. The depth of runoff for May 1973 was 22 mm, the lowest among all the monthly maximum values.

As for the driest months of the year, September is the month with the lowest flow in seven out of the seventeen years of study. Lowest monthly flows have also occurred during the months of June to November (Table 41)

Considering the mean monthly runoff values, there is an increase in runoff from October to January and a decrease from January to July (Fig 50). There is a slight increase during July and August. This is most probably due to the occurrence of relatively intense convectional rainfall during these months

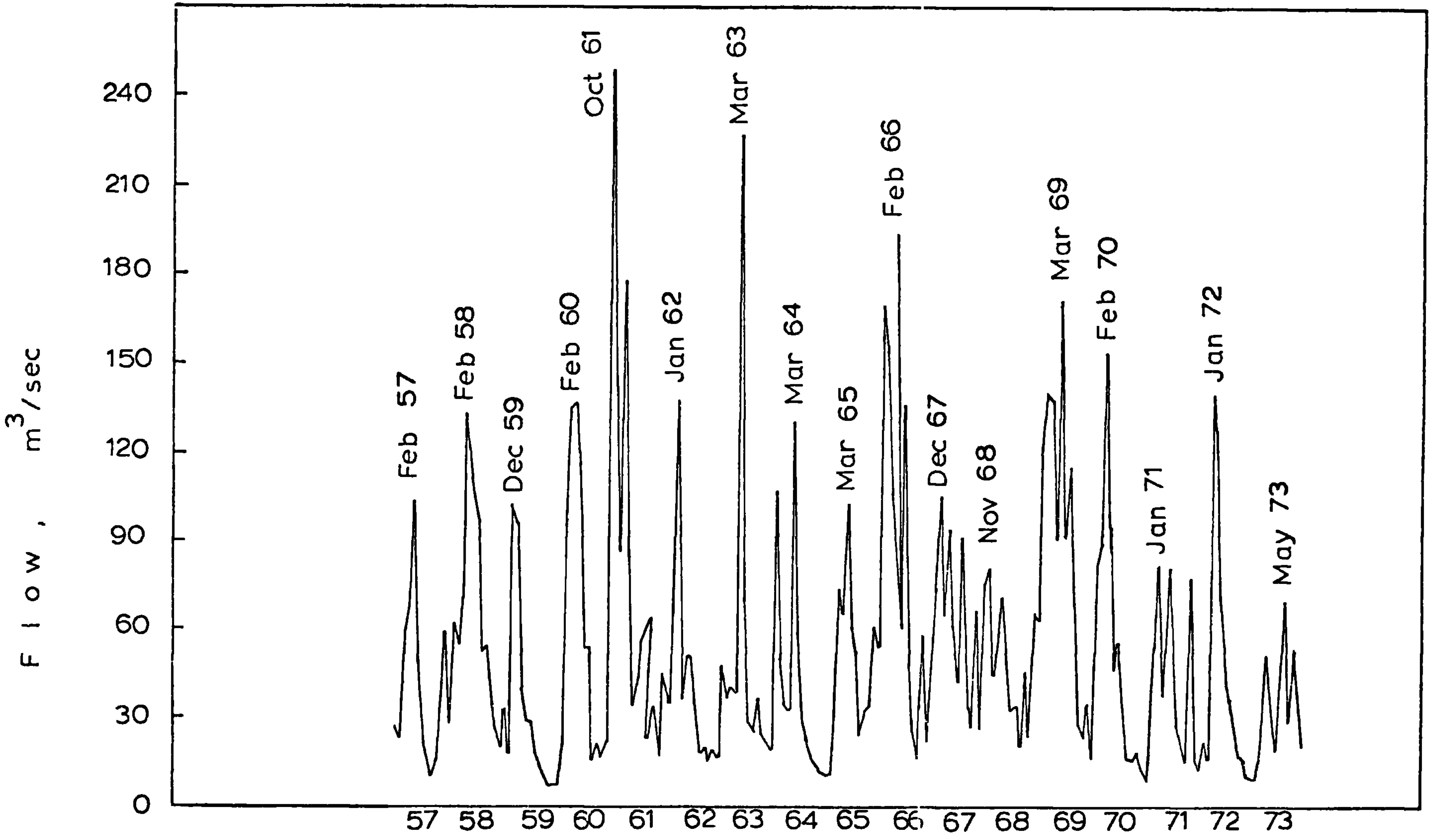


Fig 49. Hydrograph of monthly streamflow for the period October 1956 to September 1973 for the Browney River at Burn Hall

Table 41. Monthly values of maximum and minimum runoff (mm) for each water year during the period 1957-1973

Water Year	57	58	59	60	61	62	63	64	65	66	67	68	69	70	71	72	73
Max	Feb. 50 0	Feb. 64.3	Dec 49.3	Feb. 66.2	Oct 120.0	Jan. 66.3	Mar 110 0	Mar. 63.3	Mar 49 7	Feb. 93.8	Dec. 49.9	Nov. 38.8	Mar. 82.1	Jan. 72.9	Jan. 38 8	Jan. 68 0	May 22.0
Min	June 5.1	Sept. 9 70	Sept. 3 30	Oct 3 20	Sept. 8 3	Aug 7.10	Oct 8.0	Sept. 5.7	Nov 5.1	July 7.7	Sept. 11.9	June 8.9	Aug. 11.2	Sept. 6.2	July 6 9	Sept 4.8	Oct. 4.0

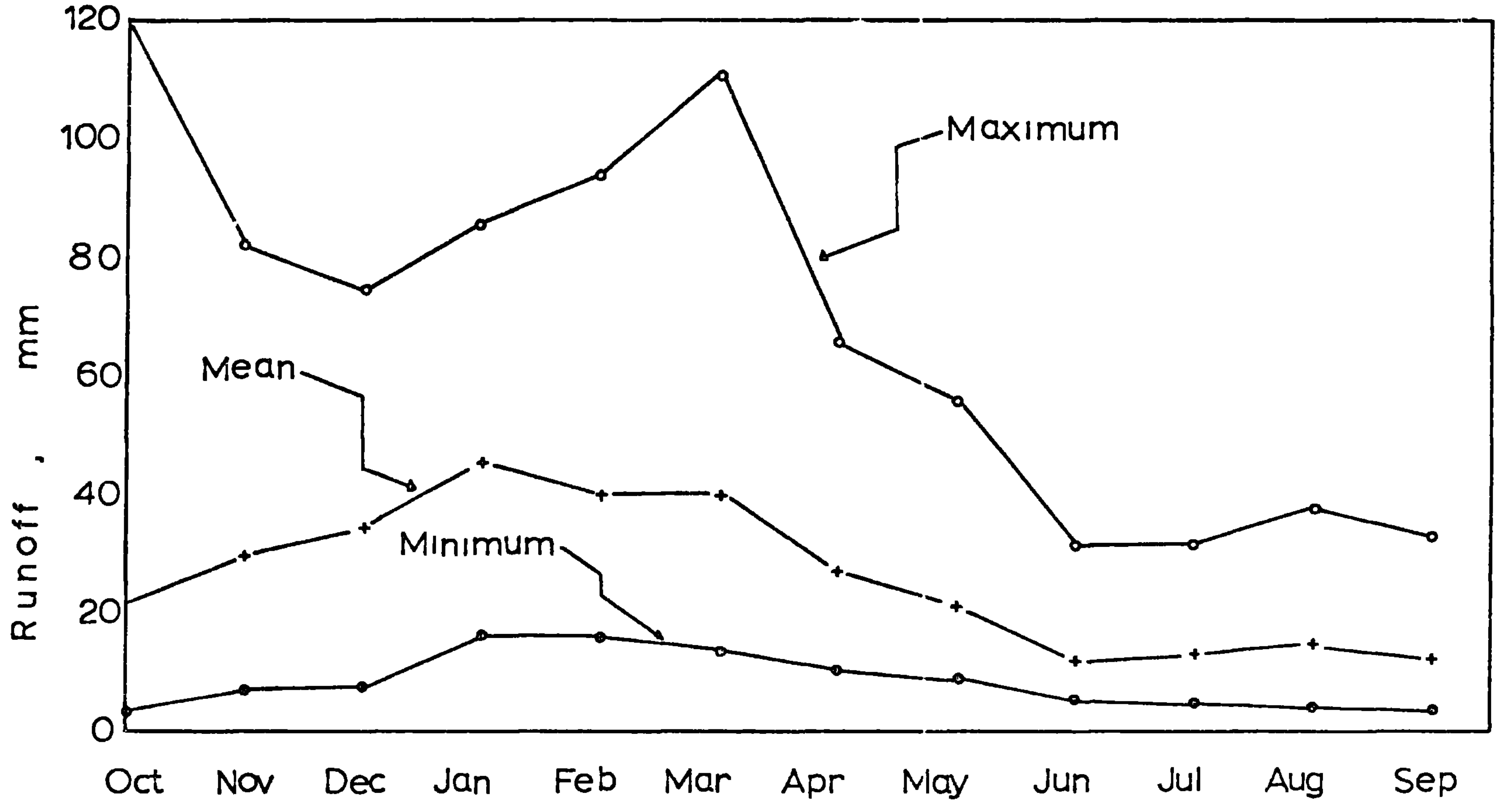


Fig 50 Mean, minimum and maximum monthly runoff for the River Browney at Burn Hall during the period October1956 to September 1973

Thus September and June on average have the lowest monthly values of runoff, while January has the highest monthly flow. The lowest and highest values of runoff in each month are also plotted in Fig 50. It is observed that the curve of lowest monthly runoff is almost parallel to that of the mean However this is not the case with that of the highest monthly runoff.

The ratio of mean monthly runoff to mean monthly precipitation is calculated in Table 42 and is shown in Fig 51 The value of the ratio is lowest in July, the month with the highest value of evapotranspiration From July the ratio increases until it reaches a peak of 0.85 in March The high ratio in March might be explained by the following reasons

1. Soil moisture storage is high during this month of the year, therefore, the infiltration rate is low.
2. Some of the precipitation falling during the months of November to February is used for recharging the soil moisture and groundwater storage. Therefore, runoff for these months is lower and thus the ratios are lower.
3. Some of the moisture which had percolated into groundwater storage during the months preceding March might have contributed to total runoff in March as the result of groundwater flow, thus resulting in a high runoff-rainfall ratio
4. Rising temperature during March results in the thawing of snow This snow-melt accompanied by rainfall might cause the ratio of mean monthly runoff to mean monthly precipitation to be higher in March than in other months.

Duration curves A duration curve is defined as the cumulative frequency diagram of a continuous time series The duration curve displays the frequency of various magnitudes which are equalled or exceeded. The yearly and monthly duration curves for the Browney River are obtained by

Table 42 Mean monthly rainfall and runoff (mm) and runoff-rainfall ratios during the period 1957-1973

Month	O	N	D	J	F	M	A	M	J	J	A	S
Rainfall	55.1	76 2	61.8	68 7	53 2	46 8	53.6	62 2	56 2	74.3	79 0	59.7
Runoff	22 1	29 5	34 2	45 3	39 4	39.8	26 7	21 1	12 2	12 9	14 4	12 2
Runoff/ Rainfall	0 40	0 39	0 55	0.66	0.74	0.85	0.50	0 34	0 22	0 17	0.18	0 20

arranging the runoff in descending order of magnitude. For the yearly duration curve for the period 1957-1973, there are 17 years of data, therefore each year covers 5.88 per cent of the time Thus the highest runoff value (514.4 mm) covers the range from zero to 5 88 per cent and the lowest from 94 12 to 100 per cent From the yearly duration curve (Fig.52), it is observed that fifty per cent of the time the runoff is over 280 mm, 25 per cent of time the runoff is over 365 mm and 75 per cent of time it is over 250 mm

The duration curve for the monthly runoff values are shown in Fig.53 From this duration curve it is observed that 50 per cent of the time monthly flow is greater than 39 m^3/sec (18 9 mm). Twenty-five per cent of the time flow is above 70 m^3/sec (34 0 mm) and 75 per cent of the time it is over 21 m^3/sec (10.2 mm). In view of the fact that the seasonal cycle of monthly flows in Fig 53 is disguised, separate duration curves for the winter and summer seasons are drawn in Fig.54 The data are grouped into different class divisions and the discharge is expressed in terms of the mean. A logarithmic scale is used in order to show the two end parts of the curve more clearly The seasonal cycle of runoff, thus, is very clear. The cause of the seasonal cycle, as mentioned earlier, is primarily due to the difference between rainfall and evapotranspiration. It is clear that the high monthly flows occur in the winter months. The highest monthly flow during winter is 4.65 times the

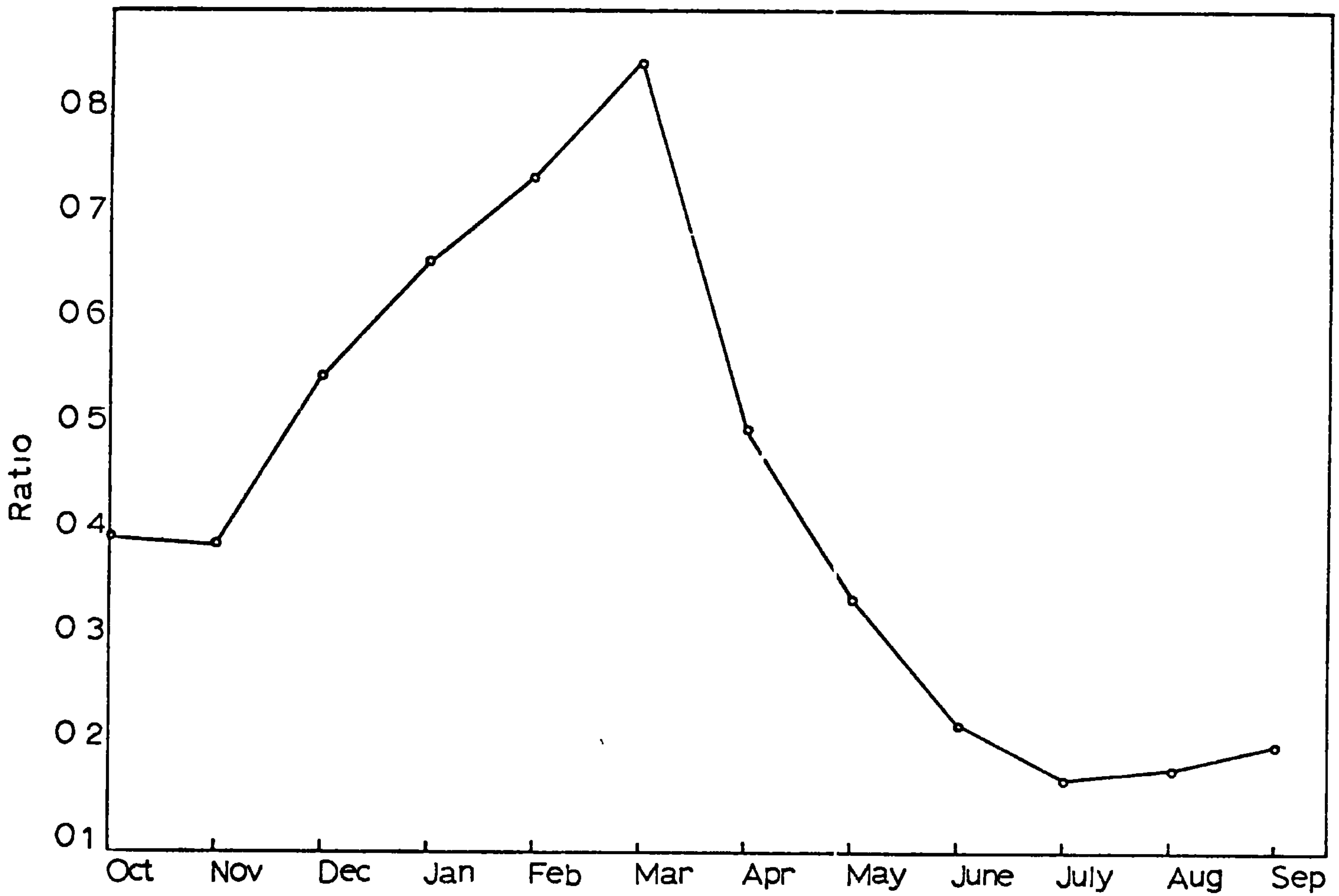


Fig 51 Ratio of mean monthly runoff to mean monthly rainfall for the Browney Basin at Burn Hall during the period Oct 1957 - Sep 1973

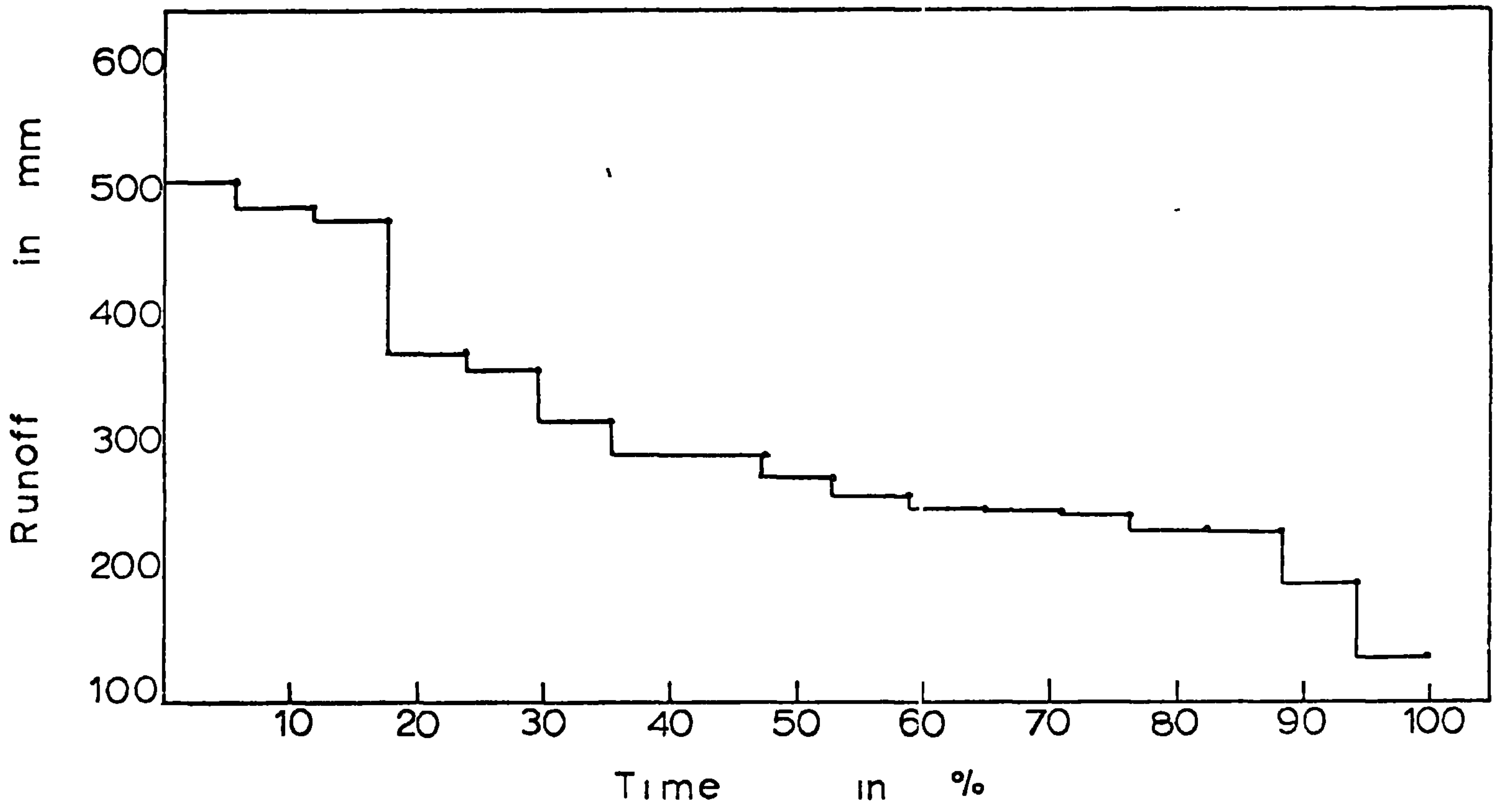


Fig 52. Yearly duration curve for the Browney River at Burn. Hall during the period 1957 -73

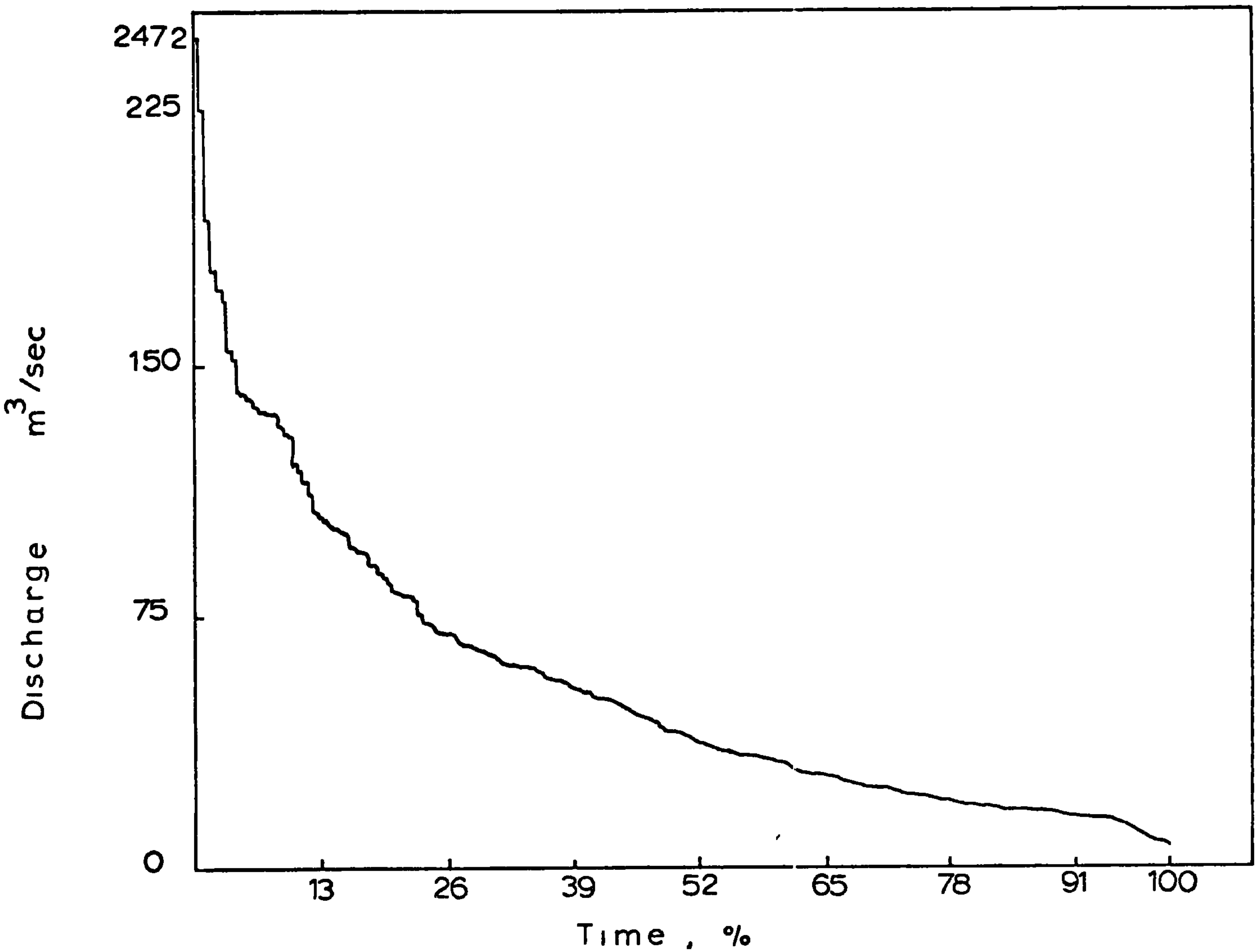


Fig 53. Monthly duration curve for the Browney River at Burn Hall during the period October 1956 to September 1973

mean, while during the summer it is 2 48 times During the winter season about 55 per cent of the monthly flows are above the mean, while the corresponding figure for the summer is 17 per cent and for the whole year is 36 per cent.

Another observation from the duration curves of winter and summer season is that the lowest flows have got the same chance of occurrence in the summer as in the winter months.

For the daily values of discharge, duration curves of three periods are derived by grouping the mean daily discharge during each period into class intervals (narrower for the lower values and wider for the higher values) in order that the number of observations in different intervals would be well distributed. The three periods for which duration curves are derived are

1. The water year 1969 This year was the year with highest annual runoff (514.4 mm). It also included the highest mean daily runoff of 43.3 m^3/sec (20 99 mm) throughout the period of 1957-1973.
2. The water year 1973 During this year the depth of runoff was the lowest of all the 17 years studied e.g. 136 00 mm.
3. The five year period from October 1969 to September 1973. The average yearly runoff during this period was 289 mm This is about 6 per cent less than the 17 year average This five year period, included the wettest as well as the driest years of the period 1957-1973 considered.

The curves derived (Fig 55) indicate that for the dry year of 1973, only 20 per cent of the total daily discharges are equal or over 1 m^3/sec (0 5 mm) whereas for the wet year of 1969, 20 per cent of the total daily discharges are less than 1 m^3/sec (0 5 mm) or 80 per cent of the flow are equal or over 1 m^3/sec. For the five year period the corresponding value over 1 m^3/sec is 44 per cent. The highest daily discharge in 1969 (43 3 m^3/sec or 20 99 mm) is almost eight times that of the water year 1973. About 5 per cent of the daily discharges during

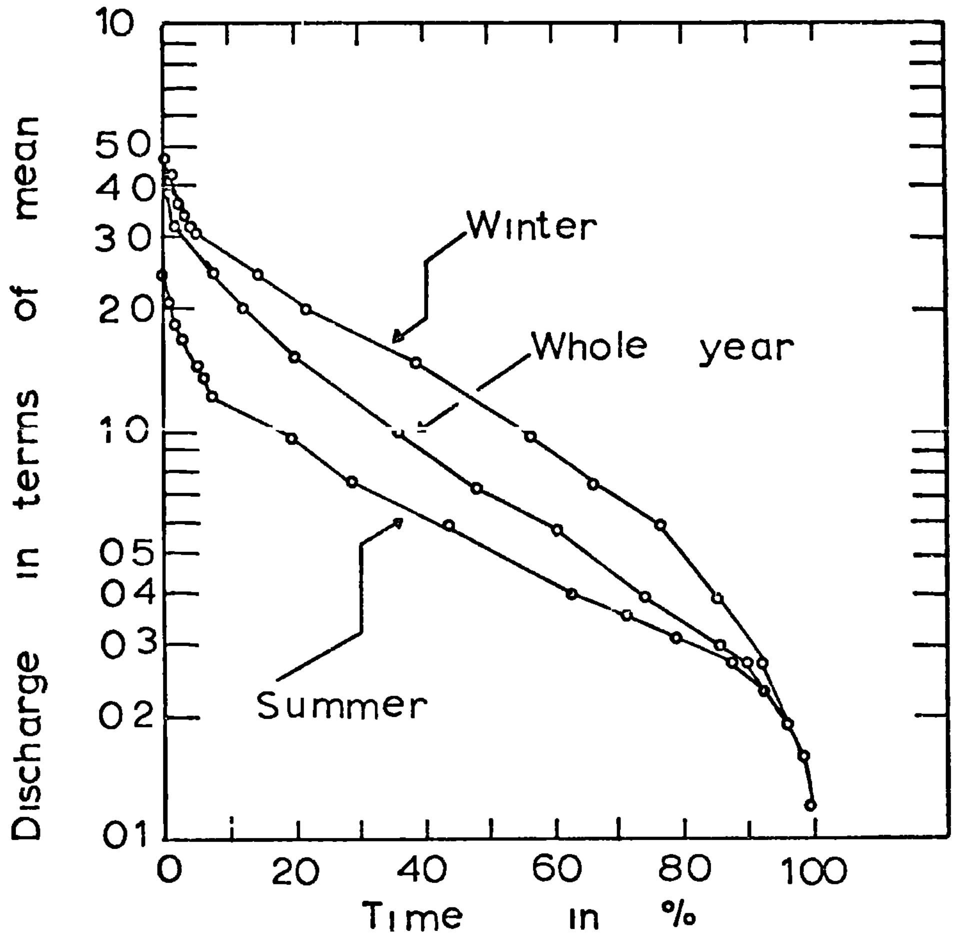


Fig 54 Monthly duration curve (winter, summer & whole year) for the Browney River at Burn Hall during the period October 1956 to September 1973

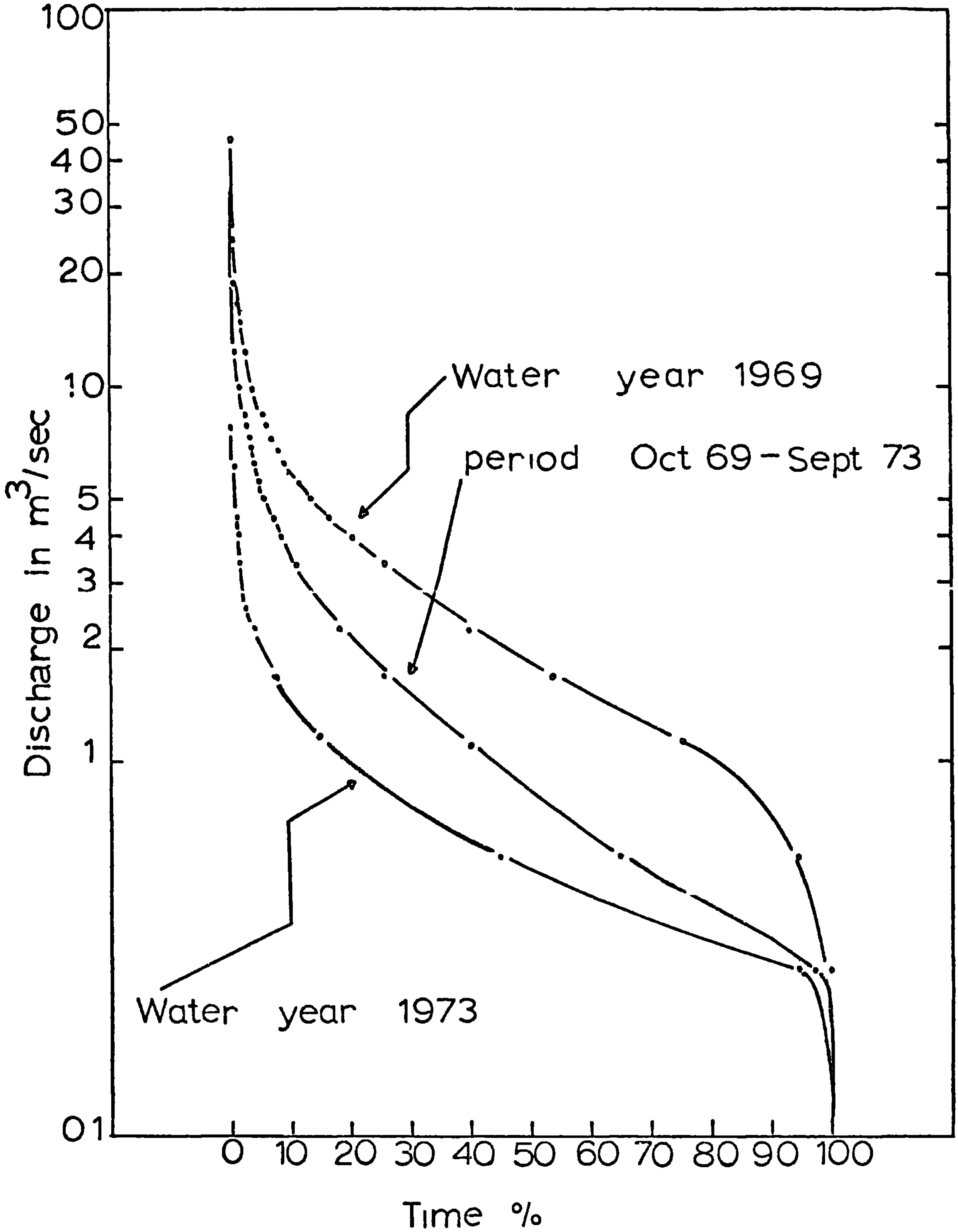


Fig 55 Duration curves of mean daily discharge for the Browney River at Burn Hall during the water years 69,73 and for the period Oct 69 - Sept 73

1973 are less than 0 28 m^3/sec (0 14 mm), whereas during 1969, the percentage of mean daily flows below 0 28 m^3/sec is nil On the other hand, about 6 per cent of the flows are equal to or greater than 8 m^3/sec (3.9 mm) during 1969, while for 1973 the percentage is zero.

Extreme values. Extreme values of runoff are those of floods and droughts Floods and droughts may be defined in several ways. Floods, for example, can be referred to as the highest mean daily discharge within a period of a year It could also designate the high streamflow which exceeds the capacity of the normal channel, (Foster, 1949). Drought, on the other hand, refers to the lowest mean rate of flow over a short period of time in a calendar year e g lowest monthly flow per year Drought in relation to precipitation also has been defined as a period, of at least 15 days with daily rainfall less than or equal to 0 2 mm (Meteorological Office, 1963). This definition, of course, does not take into account the influence of evapotranspiration, and the amount of water needed by plants Furthermore the effect of a shortage of rainfall depends on whether the soil is moist or dry at the beginning of the period (Thornthwaite, 1955)

Floods generally are caused as a result of excessive precipitation or melting of a thick blanket of snow Snow-melt can cause floods only in areas where the heavy snow cover may be preserved until late in spring A frozen ground will impede the rate of infiltration, thus all the precipitation will drain as surface runoff

Within the Browney River, the cause of the floods is excessive rainfall or snow-melt accompanied by rainfall The duration of snow cover within the catchment increases from about 17 days at Durham (elevation 102 m), 21 days at Ushaw (elevation 181 m) (Smith, 1970) to perhaps 30 days in the extreme western portion In spite of the fact that there is no seasonality in terms of rainfall depth in the basin throughout the year, almost all the peak daily flows have occurred in the winter season (with the exception of 14th August 1971 and 17th July 1973.

Droughts within the Browney basin occur during the period June to November as shown by the study of the 17 years of runoff. The month with the highest frequency of low flow is September, as was mentioned earlier.

In order to forecast the occurrence and magnitude of floods and droughts for the Browney basin, the 17 years of runoff records were used. To assemble the data for either flood or drought study, two methods are available namely, the annual maximum series and the annual exceedance series (chapter 2) In the annual maximum series for floods the highest mean daily flow (or maximum monthly flow) is selected. Therefore, there is one flood value for every year. With this method the second highest daily discharge in a year which might be higher than peaks in other years is ignored.

Thus the data were assembled by the annual series method For the flood studies the data for highest mean daily flow per year, as well as the yearly maximum monthly flow were used. For the drought study, the minimum monthly flow in each year was considered.

Floods The purpose for any flood study is to find the period, on average, between two values of flow which equal or exceed a particular magnitude. This period is called the return period or the recurrence interval Therefore an N years flow is one which is expected to be equalled or exceeded, on average every N years and it has a recurrence interval (Tr) of N years. There are several formulae used for the calculation of the recurrence interval and Chow (1964) refers to them in detail. The most common is described by Weibull (1939) which expresses Tr by the formula $Tr = \frac{n+1}{m}$, where m is the event ranking and n is the number of events (floods).

For the Browney River the highest mean daily flow per year for the 17 years are arranged in decreasing order. Using the formula - $Tr = \frac{n+1}{m}$, the return period for each flood is obtained (Table 43) Upon calculation of return periods, each flood is plotted versus its return period on probability paper. A straight line is then fitted through the points and extended (Fig 56).

Based on this graph, the 100-year flood (0 99 probability) is 74 m^3/sec (36 0 mm). This value for the 100-year flood might not be accurate because the extrapolation of data was based on only 17 years of record However, for the estimation of floods with shorter return periods, the accuracy increases

To estimate the occurrence and magnitude of monthly floods, the maximum monthly value of runoff for each year is selected The data thus obtained, are arranged in decreasing order, as discussed earlier for calculation of daily floods The recurrence interval for each runoff event is calculated by the equation of $Tr = \frac{n+1}{m}$, and each value of runoff is plotted against its return period on a log-probability paper A line representing a distribution with the same mean and standard deviation as the series to which the points are assumed to belong, is passed through them For the mathematical solution of this line the following procedure is adopted after Bannerman (1966)

1. The monthly values of flow are arranged in decreasing order
2. The logarithm of these values are found.
3. The difference between the mean of the logarithms and each of them is found
4. The value of the standard deviation is calculated
5. The antilog of the mean of logarithms of the flow is found. This represents the mean of the log series.
6. The 100-year flood was then the summation of m (mean of logarithms of flows) and 2 33 (99 per cent probability) x standard deviation (S), or $m + 2.33 \times S$

Table 43 Calculations for the studies of daily floods for the Browney basin at Burn Hall

Date	Runoff mm	flow m^3/sec	Rank (m)	Probability ($\frac{m}{n+1}$)
18 Dec.1969	20 99	43 27	1	5.6
7 Mar.1963	20.86	43 02	2	11.1
19 Nov.1966	17.84	36.79	3	16.7
14 Aug 1971	15.25	31.44	4	22.2
20 Oct.1961	13.49	27 82	5	27.8
3 Feb.1972	11 79	24 30	6	33.3
25 Mar 1964	10 60	21.85	7	38.9
5 Nov.1968	9 92	20.46	8	44.4
9 Oct.1967	8 99	18.54	9	50.0
26 Feb.1960	7 99	16 47	10	55 6
18 Dec 1959	7 93	16.36	11	61.1
23 Mar.1965	7.79	16.05	12	66.7
18 Jan.1970	7.72	15.92	13	72.2
29 Mar 1958	7.32	15.08	14	77.8
8 Jan.1962	5.50	11 35	15	83.3
14 Feb 1957	4.38	9.03	16	88.9
17 July 1973	4 05	8.35	17	94.4

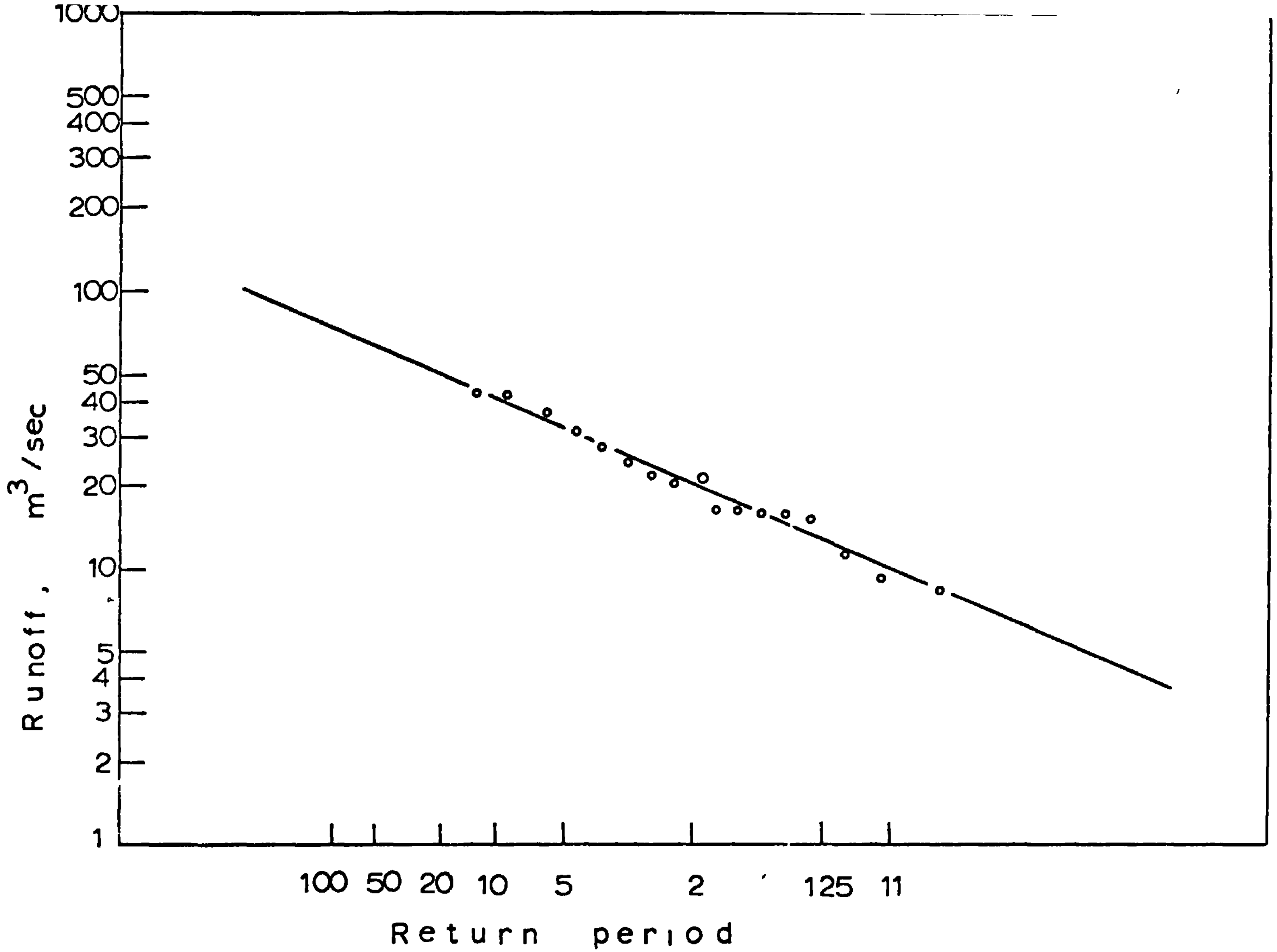


Fig 56. Annual 24-hr maximum flow for the Browney River during the period 1957-1973

7. The antilog of m + 2 33 S was found.
8. A straight line was then passed through the mean of the log series and the point of the 100 year flood on the graph, (Fig 57 and Table 44).

The 100-year monthly flood thus calculated is 154 15 mm.

Droughts To estimate the occurrence and magnitude of drought, the minimum monthly value of runoff for each year is selected The data are then arranged in increasing order. The return period for each monthly flow is calculated, and each value of flow is plotted against its corresponding return period on log probability paper A line is then fitted to these points (Fig.57). The procedure being similar to that for finding the 100-year flood, except that the 100-year drought was considered by subtracting 2 33 x standard deviation from the mean of the log series. The calculations are shown in Table 45

Thus the 100-year drought for the Browney basin based on 17 years of record was obtained to be 2.52 mm The line of best fit joined the mean (6.29 mm) to this point.

Unit hydrograph concept This concept, used by Harvey (1971) to investigate the seasonal flood behaviour of the River Ter in central Essex, was adopted to study the short term distribution of runoff in response to intense storms within the Browney basin. A unit hydrograph (defined as a hydrograph of unit depth of runoff usually 1 inch (25 4 mm)) and its modified version, the distribution graph are useful in studying the runoff patterns of watersheds of different sizes.

In order to derive the unit hydrograph of the Browney basin, autographic runoff records for the period 1956 to 1968 were considered and from them isolated hydrographs of substantial runoff volume were selected. However, since there were no recording raingauge charts available to study the intensity pattern of the storms prior to 1962, the runoff records studied were limited to the period 1962-1968 A study of the hydrographs of runoff due to major storms revealed that no isolated

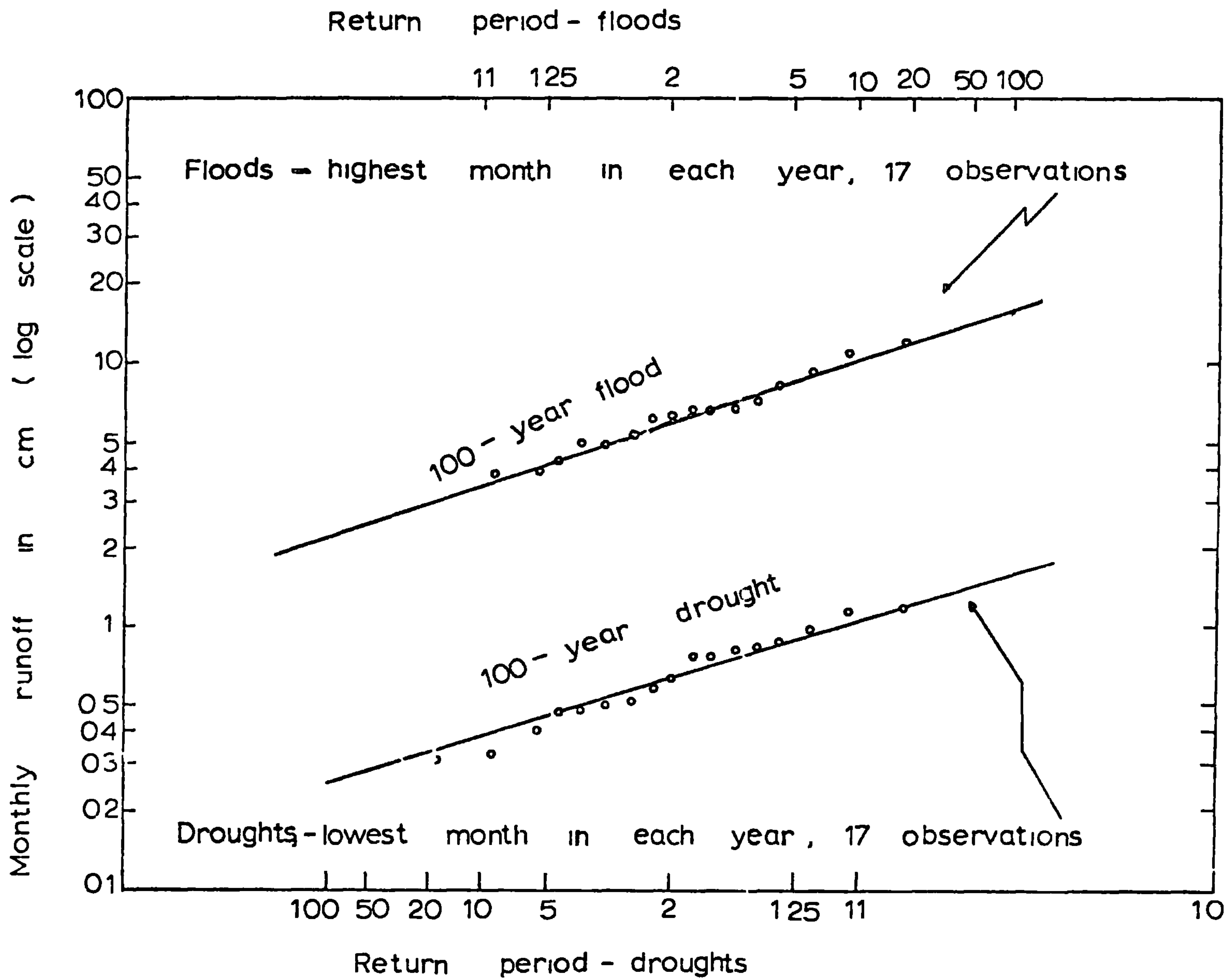


Fig 57 Derivation of 100-year drought and 100-year flood for River Browney

Table 44 100-year flood calculations (monthly values) for the Browney basin at Burn Hall

Date	Rank (m)	Return period	Monthly runoff mm	Log of runoff	Mean - log of runoff (d)	d^2
Oct.1961	1	5.6	120 00	2.079	0.301	0.091
Mar 1963	2	11 1	110 01	2 041	0 263	0.069
Feb 1963	3	16 7	92.01	1 964	0.186	0.035
Mar 1969	4	22.2	82.07	1.914	0.136	0.018
Jan.1970	5	27.8	72.90	1 863	0.085	0.007
Jan.1972	6	33 3	68.05	1 833	0.055	0.003
Jan.1962	7	38 9	66 31	1 822	0.044	0.002
Feb 1960	8	44.4	66.16	1.821	0 043	0 002
Feb.1958	9	50 0	64 34	1.808	0 030	0.001
Mar.1964	10	55 6	63 33	1 802	0.024	0.001
Oct 1967	11	61.1	53 92	1 732	-0.046	0.002
Feb 1957	12	66.7	50 02	1.699	-0.079	0 006
Mar.1965	13	72.2	49.70	1.696	-0.082	0.007
Dec 1959	14	77 8	42 29	1.626	-0 152	0.023
Jan.1971	15	83.3	38 79	1.589	-0 189	0.036
Nov.1968	16	88.9	38.77	1.588	-0.190	0 036
May 1973	17	94 4	22.00	1 342	-0.436	0.190
Total			1100.67	30.219		0.527
Mean			64 74	1.778		0.031

Antilog = 59 9 St Dev. 0.176

Log 100-year flood = 1.778 + 2 33 x 0.176 = 2.188

Antilog 2 188 = 154 15 mm

Table 45 100-year drought calculations (monthly values) for the Browney basin at Burn Hall

Date	Rank (m)	Return period	Monthly runoff mm	Log of runoff	Mean - log of runoff (d)	d^2
Oct.1960	1	5.6	3 20	0 505	-0.294	0 086
Dec.1959	2	11 1	3.26	0.513	-0 286	0 082
Oct.1973	3	16 7	4 00	0.602	-0.197	0.039
Jan 1971	4	22.2	4.65	0.667	-0.132	0.017
Jan.1972	5	27.8	4 83	0 684	-0.115	0.013
June 1957	6	33.3	5.11	0 708	-0.091	0.008
Mar.1965	7	38.9	5.14	0.711	-0 088	0.008
Sept 1964	8	44.4	5.72	0.757	-0 042	0.002
Sept.1970	9	50 0	6.19	0.792	-0.007	0.0001
July 1966	10	55.6	7.51	0.876	0 077	0.006
Aug.1962	11	61.1	7.60	0 881	0.082	0.007
Oct.1963	12	66 7	8.04	0.905	0.106	0.011
Sept.1961	13	72.2	8.33	0 921	0.122	0.015
June 1968	14	77.8	8 79	0 944	0 145	0 021
Sept 1958	15	83.3	9 73	0.988	0 189	0.036
Aug.1969	16	88 9	11 20	1.049	0.250	0.062
Sept 1967	17	94.4	11.92	1 076	0 277	0.077
Total			115.22	12.579		0.490
Mean			6.78	0.7988		0.0288

Antilog = 6.29 St. Dev. = 0.170

log 100-year drought = mean of log series - 2.33 x St.Dev

= 0.7988 - 2 33 x 0.17 = 0.402

Antilog 0.402 = 2.52 mm

hydrograph had a runoff depth of 1 inch (25 4 mm) or more In fact the two hydrographs which were ultimately chosen for the study had direct runoff depths of 0 61 inches (15 5 mm) and 0 24 inches (6.1 mm). The two storms producing these two hydrographs had depths of 2.76 inches (70.1 mm) and 2 48 inches (63.0 mm) respectively. The mass diagrams of these two storms are shown in Fig 58 As expected there is some variation in the mass diagrams of these two storms. This is because uniform rainfall rates over an extended period of time are uncommon Such variation in the rainfall intensity, however, is smoothed out in the course of surface detention and during direct runoff.

These two storms were those of 5th November and 8th and 9th August, 1967. The two-hourly ordinate hydrographs of runoff of these two storms were drawn. The baseflow was separated from total flow by prolonging the recession part of the preceding flood up to the peak. From this point a straight line was drawn joining it to a point two days after the peak. The value of 0.2 was derived by using the formula $N = A^{0\ 2}$ where N is the number of days after the peak and A is the area of the basin.

For each storm, the two-hourly values of baseflow were subtracted from those of the total discharge to get the corresponding direct runoff value. Each ordinate was then expressed as a percentage of the sum of the ordinates to get the distribution graph. The two distribution graphs were then plotted in Fig.58. A point which should be mentioned is that due to minor secondary rain, a slight deviation from the normal recession curve occurred in the hydrograph of 5th November. Therefore, in deriving the distribution graph, the upper portion of the recession limb has been prolonged so as to bypass this slight deviation Some of the observations made from the study of these two distribution graphs are

1. The total direct runoff produced due to the storm in November was 0.61 inches (15.5 mm) while that of August was 0.24 inches (6.1 mm). The runoff was measured by planimetering the area under the graphs of

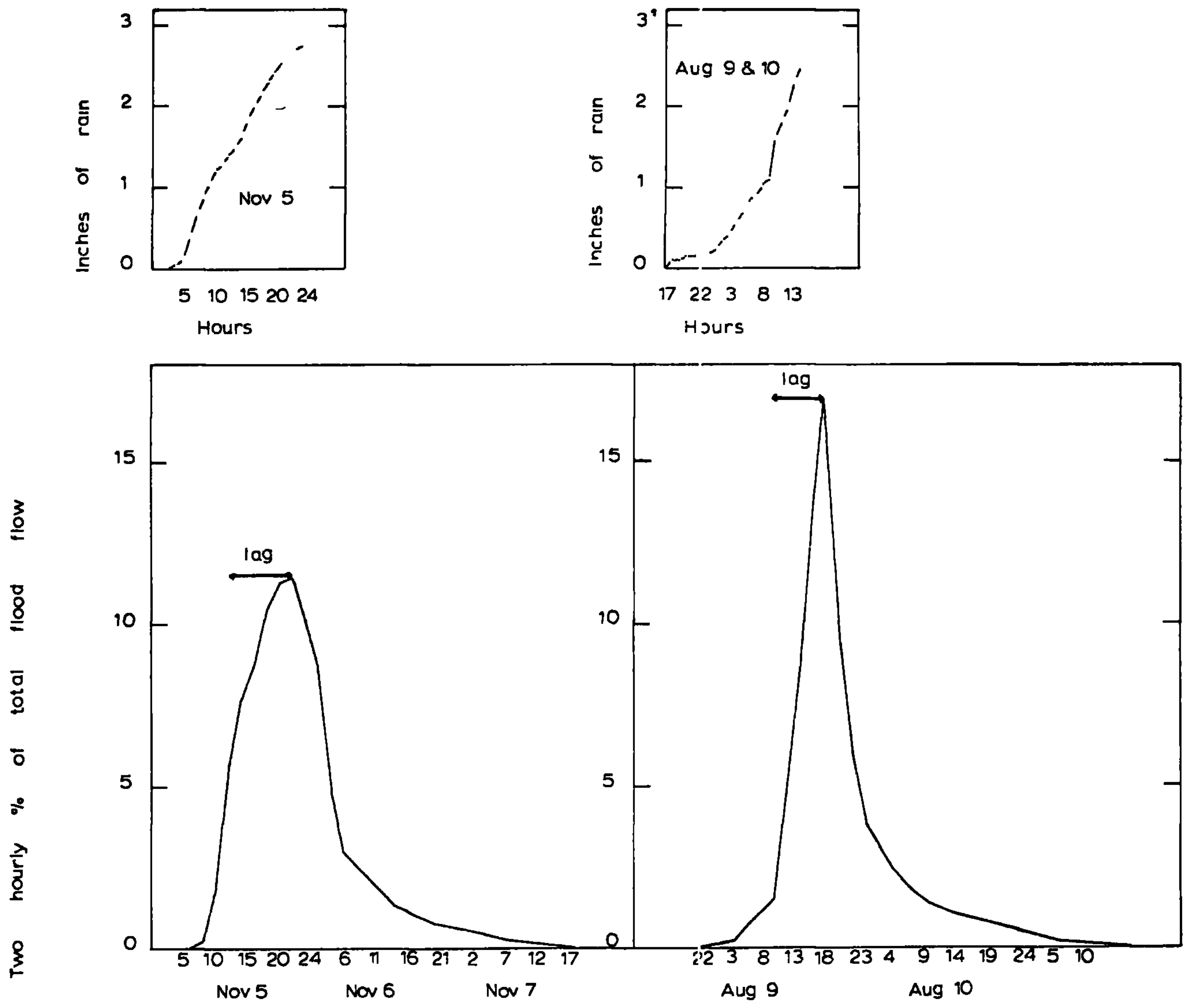


Fig 58 Mass diagrams of two rainstorms and the distribution graphs of their respective floods in the Browney basin during November and August 1967

direct runoff. The rainfall producing these two runoff hydrographs, as mentioned earlier, were 2.76 inches (70.1 mm) and 2 48 inches (63 0 mm) and this fell in 22 and 21 hours respectively

2. The period of rise in the hydrograph of November was 16 hours and that of August was 19 hours.

3 The peak ordinate due to the November precipitation was 11.45 per cent as compared with 16.8 per cent for that of August.

4. The basin lag time for the storm of November was 10 hours as compared with 8 hours for the August storm.

Discussion There is a marked difference in the distribution graphs of the two storms The depth of direct runoff due to the November storm was two and a half times that of August The mass diagrams of the storms show, however, that the variation in the pattern of these storms is slight. This observation plus the fact that the summer peak flow is percentage-wise higher than that of November are explained by two main factors

1. The soil moisture content preceding the August flood was lower than that preceding the November flood. The soil, therefore, had a greater capacity for storing moisture during August than November and the runoff volume is thus lower during the August flow

2. The infiltration rate at the initial stages of the flood was probably higher in August, because of lower soil moisture content and the higher temperature. Therefore, during August, the initial ordinates of the runoff hydrograph were lower and the percentage of the peak runoff value was higher than those of November

Thus the study of the two distribution graphs show that during the summer, only rainfall intensities greater than the infiltration rate might cause runoff from the very beginning of rainfall. During the winter, however, appreciable runoff may occur immediately after the start of precipitation which gradually rises to a peak In such cases the percentage of peak flow in the winter season is lower.

One more possible explanation, suggested by Harvey (1971), that might be given to account for the discrepancy between the distribution graphs of winter and summer is that, during the summer, runoff occurs mainly from the poorly drained portion of the catchment or steep valley sides, whereas during the winter all parts of the catchment might be contributing to runoff.

CHAPTER SEVEN

THE ESTIMATION AND PREDICTION OF RUNOFF

Prediction of runoff is quite often required for the construction of hydraulic structures, in the preparation of river forecasts and for the evaluation of the effect of changes in land use. Studies of flood magnitude and frequency, drought magnitude and the period they last are the core of many hydrological research projects Sometimes such studies are undertaken either because the catchments are ungauged or because insufficient data are available

Due to differences in objectives of runoff estimation, different methods have been developed and applied Some of these methods were developed when the science of hydrology was in its infancy, therefore, they were purely empirical and without any sound hydrologic principle There are, on the other hand, examples of mathematical models which simulate the whole runoff cycle according to the established findings of hydrologic research The development of such models were the result of application of digital computers to hydrology These models are continuously in a state of change in order to incorporate the latest findings of research.

Methods of prediction

These can be divided into empirical formulae, infiltration method and infiltration indices, regression and graphical methods, the unit hydrograph method, moisture accounting methods, simulation models and statistical methods.

Empirical formulae The relationship for the peak rate of flow has been presented by the formulae of the form $Q=CA^n$ in which

Q - is the peak flow

A - is the drainage area

C - is a function of land use or topography

n - is a constant that has a range between 0 2 and 0 9 (Rodda 1971)

An example of such formulae is the one derived in a study of the mean annual flood in a number of basins in England and Wales where the discharge is given as $Qm=CA^{0.85}$ (Rodda,1971).

Another group of empirical formulae are those which include a rainfall intensity term. An example of such formulae is the Rational formula, often referred to as Lloyd-Davis method (Chow,1964). The peak discharge in this formula is expressed by an equation of the form $Q=CIA$ where

Q - is the peak flow in ft^3/sec

C - is runoff coefficient which depends on drainage basin

I - is the rainfall intensity in inches per hour

A - is the area of the basin in acres

Infiltration approach Horton in 1939 presented his theory of infiltration and suggested that runoff represented the difference between rainfall intensity i, and the infiltration capacity fp, (the maximum rate at which rain can enter the soil, Kohler,1963). To use this method a standard infiltration capacity curve is drawn on an intensity-time graph of the rain, and the area between the two curves is estimated. This method, however, has a very limited application due to the following reasons (Bhatnagar,1969).

1. Infiltration capacity is normally highest at the beginning of a rainfall event, while rainfall generally begins at moderate rates and a substantial period may elapse before rainfall intensity exceeds the infiltration capacity and runoff begins.

2. The rainfall might occur intermittently and not always at rates greater than the infiltration capacity

3. This method does not account for interflow.

As a result of these limitations several infiltration indices have been developed which are semi-empirical and more practical. One of these is the "Ø index" defined as the average rainfall intensity above which the rainfall volume equals runoff volume(Wilson, 1969), (Fig.59). This index assumes that the losses due to interception, depression storage and soil moisture deficiency are part of infiltration Another index which is a more refined version of the former is called "W-index" (Linsley et al, 1958) which is the average infiltration rate during the time interval rain intensity exceeds the infiltration capacity, i e. $W = \frac{P-Q-S}{t}$. In this formula

P - is precipitation

Q - is the total amount of runoff from the storm

S - is the summation of interception and depression storages.

Linsley et al (1958) mention that these methods can be applied to small areas of homogeneous characteristics with a unique infiltration capacity curve, and that they are not accurate for applying to conditions of varied rainfall (amount, intensity and duration), and varied infiltration characteristics

Least square methods These methods are based on the development of relationships between rainfall and runoff Such equations result in the estimation of the discharge volume over extended periods i.e a year Equations of this type were derived for the Browney basin in the preceding chapter. As can be observed (p 160) these equations have the form runoff = a x precipitation - b in which

a - is a constant

b - is the threshold of precipitation below which there is no runoff.

Chow (1964) refers to an example where inclusion of a mean annual temperature term resulted in a satisfactory relationship between rainfall and runoff being derived.

The dependence of runoff on other important factors, however, precludes any accurate short period estimation by rainfall alone Therefore, attempts have been made to relate runoff to several other important factors by means of multiple regression equations The factors considered

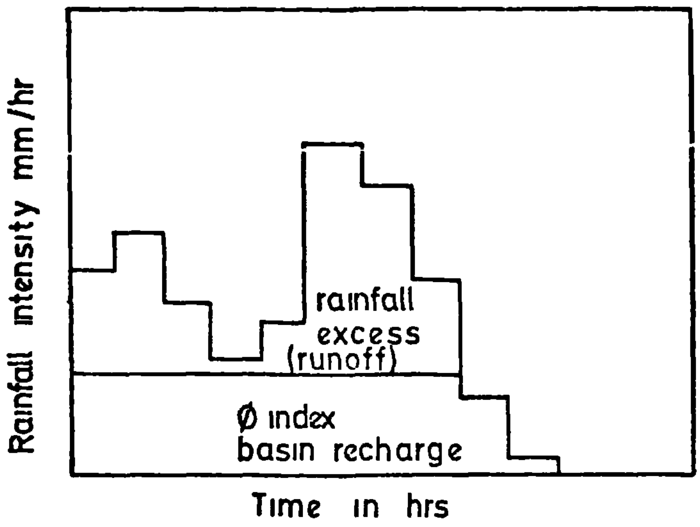


Fig 59 Diagram illustrating the derivation of ∅ index
(Redrawn from an original by Wilson (1969)

were rainfall characteristics (intensity and duration), soil moisture conditions prior to rainfall, days to last rain, initial flow at the time of beginning of the rain event and baseflow in the stream prior to the storm.

Out of these different factors, rainfall intensity, duration and soil moisture conditions prior to the storm were chosen to relate to runoff in multiple regression equations. The intensity and duration of rainfall could easily be found from autographic raincharts. Determination of soil moisture conditions, however, was complicated. This complication was the result of spatial variations of soil moisture within a watershed and consequently empirical indices were employed The index commonly used in the U S A. is called the antecedent precipitation index. This index is defined by Kohler et al (1951) as $API = A_1 P_1 + A_2 P_2 + \ldots A_n P_n$ (Chow,1964) In this formula

P_n - is the precipitation n days before the storm

A_n - is a constant

Since the calculation of the index on a day by day basis was difficult, the index was calculated by assuming that depletion of soil moisture is proportional to the available storage i e $I = IoK^n$ where

Io - is the initial value of the index (mm)

I - is the index value n days later

K - is a recession constant varying between 0 85 and 0.98 with an average value about 0.92 (Chow,1964).

According to this formula if n=1, then the value of any day is K times that of the preceding one If any precipitation occurs, the index is increased by the amount of precipitation which is recorded.

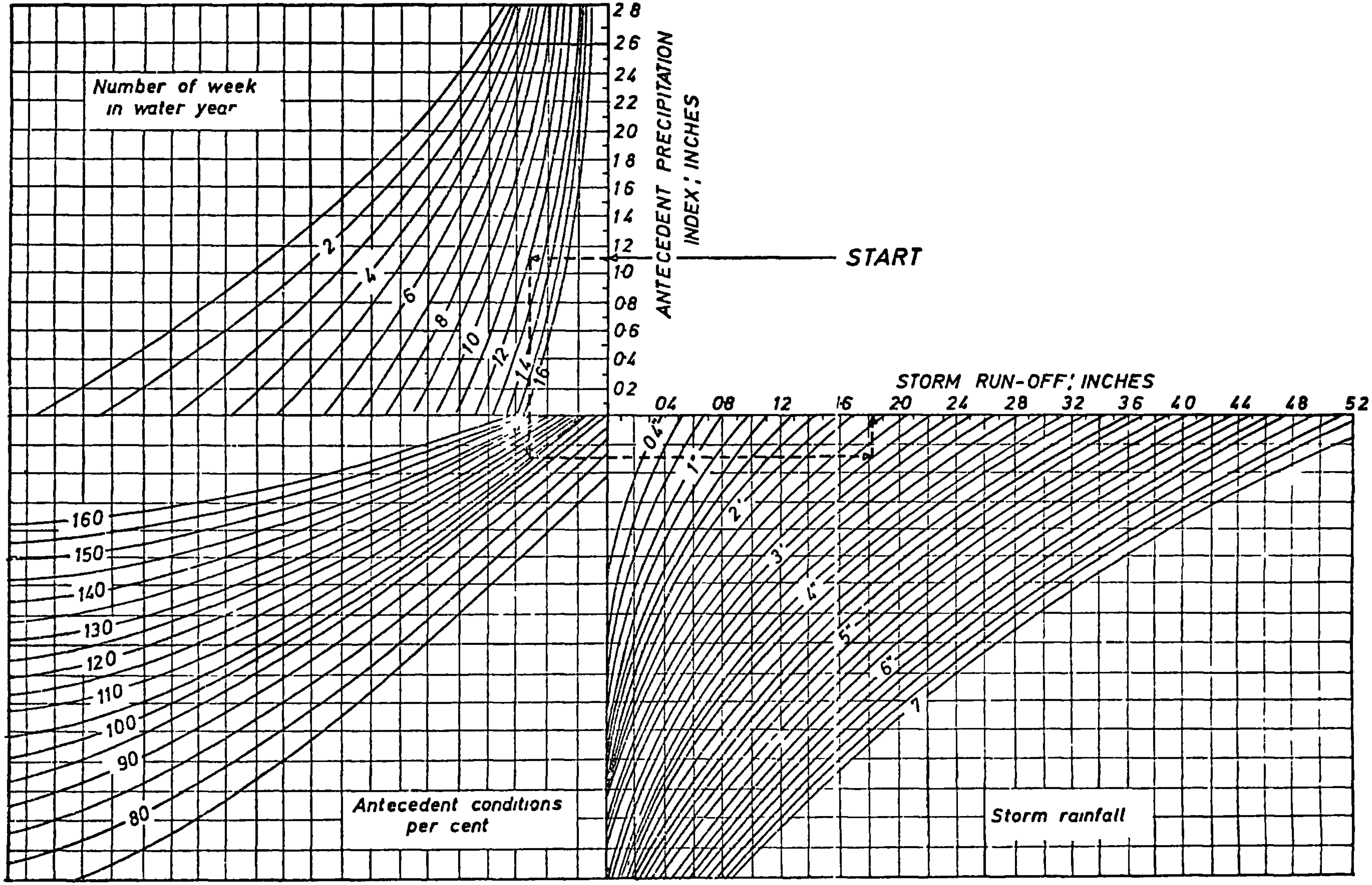
One of the limitations of this index is that by assuming K to be constant, no account is taken of evapotranspiration variations This defect has, however, been corrected by inclusion of a calendar date as an independent variable upon which runoff depends. The measure of calendar date was considered to be the week of the year. Kohler et al

(1951) used A.P I , week of year and storm duration and developed relationships between storm runoff and precipitation This relationship was presented by a graphical method of co-axial form. An example of this method for computing storm runoff in the River Thames at Teddington is given in Fig.60 To estimate storm runoff from such a relationship, the value of the antecedent precipitation index is determined and entered in chart A A horizontal line is drawn to intersect the week number A vertical line from the week number curve is then drawn to intersect the storm duration line and from there a horizontal line is drawn to intersect the storm precipitation curve A vertical line from this last intersection point indicates the storm runoff

Unit hydrograph method The basic theory behind the unit hydrograph (unit graph) method was outlined by Sherman in 1932 This theory has been one of the most important contributions to the science of hydrology and is used to study rainfall-runoff relationships. The results of the application of the unit graph method in runoff studies have often been used in the design of hydraulic and flood protection structures

A unit hydrograph was defined in the preceding chapter as a hydrograph of unit depth of runoff (usually 1 inch or 25 4 mm). The storm of effective rainfall producing runoff is assumed to be uniformly distributed over the catchment during a specific period of time. In the original paper of Sherman surface runoff was considered to be the only water contributing to the unit graph. This was because the concept of interflow was not known at that time After the recognition of this concept, the term direct runoff was used instead (Barnes,1959) The use of the word "unit" in the definition has resulted in some confusion Chow (1964) states that Sherman's original use of the word "unit" was to denote a "unit of time" of effective precipitation, which later was interpreted as unit depth of effective rainfall The assumptions of the unit hydrograph are given by Barnes (1959)

Fig.60. Coaxial relationship for computing storm runoff for the River Thames at Teddington (After Andrews,1962)



(a) Hydrographs of rainfall of the same duration have the same time base.

(b) Hydrographs of runoff produced by rainfalls of similar duration, but different intensities, have ordinates proportional to their rainfall intensities. In other words, if one rainfall is m times the other, the ordinates of its runoff hydrograph is m times the other one as well (Fig 61).

(c) The rainfall intensities are uniform both in time and space.

(d) The runoff following the storm is not affected by the melting of snow or ice.

(e) The storm period is comparatively isolated in the record It follows a period of low streamflow and there is no further rainfall until the peak is well passed

Barnes (1959) discussing the validity of the assumptions, mentions that assumption (a) is in no way essential This, he explains, is because a large flow will take longer to recede. Assumption (b) is known as the principle of superposition or proportionality and is the basic element of the conventional unit hydrograph The effect of minor differences in duration is not significant and a tolerance of 25 per cent from the established duration is acceptable (Linsley et al,1958).

The validity of assumption (c) is by far the most important for the application of the unit graph method This is because storms with different intensities over different periods of time produce different unit hydrographs Basin size is very important in this respect Over a large basin the assumption of uniformity of rainfall will not hold and in such cases if a hydrograph of runoff has resulted from a rainfall moving downstream, a high peak would result whereas the peak would be much lower if it resulted from rainfall moving upstream. Linsley et al (1958) put a limitation of 2000 mi^2 (5108 km^2) as the upper size limit for basins on which the assumptions of the unit graph might hold.

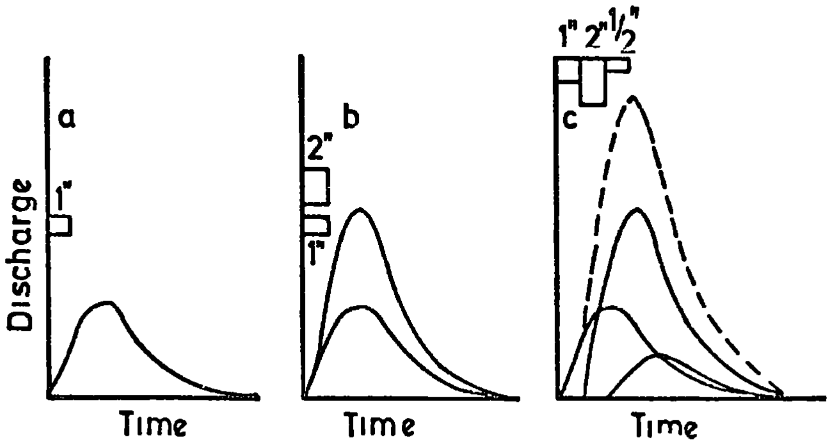


Fig 61 a Unit hydrograph b Unit graph applied to 2" of rainfall excess c Unit graph procedure applied to a storm
(Redrawn from an original by Morgan & Johnson (1962))

Synthetic unit graph The lack of actual data in many watersheds and the utility of the unit hydrograph in flood studies for the prediction of hydrographs of actual storms, resulted in some investigation to derive synthetic unit graphs These were derived by a consideration of measurable basin characteristics and correlation of them with unit graph elements The basin features considered were basin area, shape, topography, channel slopes, stream density and channel storage The elements of the hydrograph which were considered necessary for the synthesis of unit graphs were three

1 - The time interval between the centre of mass of the excess rainfall and the peak discharge (basin lag).

2 - The peak discharge

3 - The time base of the unit graph

Examples of research on synthetic unit graphs are those Snyder (1938), Linsley (1943) and Taylor et al (1952) produced The relationships developed by Snyder (1938) for unit graphs of basins in the Appalachian Mountains as the result of a rainfall of duration $t_p/5.5$, t_p being the time lag in hours, is given by the following equations

$$t_p = C_t\ (LL_{Ca})^{0\ 3}$$

$$Q_p = 640\ C_p\ A/t_p$$

$$B = 3 + 3^{t}p/24$$

t_p - time lag in hours

C_t - a coefficient with an average value of 2.0

L - length of main streams from the outlet to the upstream divide

L_{Ca} - length in miles along the main stream from the gauging station to a point opposite (nearest) to the centroid

Q_p - unit hydrograph peak discharge in cfs

C_p - a coefficient of 0 625

B - time base of unit graph in days

A - drainage area in square miles

For other durations of rainfall excess (T), the adjusted lag is $T_{P.A} = 0.25\ (T - t_P/5\ 5)$ which is to be substituted for t_p.

Knowing these quantities and the fact that the volume under the curve represents one inch of runoff, the unit graph can be drawn.

Morgan (1962) mentions that synthetic unit graphs may differ widely from the actual unit graph From an analysis of synthetic unit graphs, he found that peak flow variations were between 198 per cent above and 69 per cent below the actual observed peak flows for twelve selected basins and he suggested that unit graphs should be obtained from actual data when records permit.

Distribution graph If the ordinates of a unit graph are presented as a percentage of the sum of all ordinates, a series of dimensionless numbers are obtained which define the runoff characteristics of a basin The plot of these values represents the distribution graph which was originally suggested by Bernard (1935) Such a graph was used to study the runoff patterns of the Browney basin during the summer and winter seasons (chapter 6), (Fig 62)

Moisture accounting methods Runoff has been frequently determined by the moisture accounting technique The simplest technique is that which assumes one fixed moisture holding capacity for the soil From this reservoir, water is depleted by evapotranspiration and is recharged by precipitation Runoff would then occur when precipitation is in excess of the moisture deficiency, i e R=P-Et-D or runoff is equal to the algebraic summation of precipitation, evapotranspiration and soil moisture deficiency prior to the storm Examples of this type of approach are those of Penman (1949) and Thornthwaite and Mather (1955). There are several limitations associated with this approach One of these limitations is explained by the fact that there are areas with appreciable deficiencies from which substantial runoff occurs without satisfying the soil moisture demand. Another limitation is due to the uncertainty about

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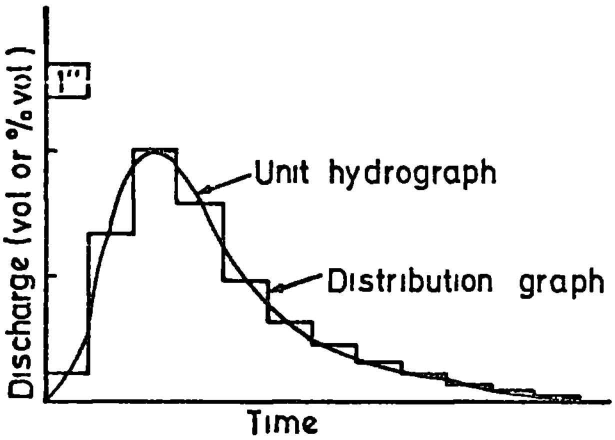


Fig 62 Relationship between unit hydrograph and the distribution graph
(Redrawn from an original by Morgan & Johnson (1962))

the relationship between potential and actual evapotranspiration when the soil moisture content is below field capacity The assumption of a single value for soil moisture storage in the Thornthwaite and Mather method is not valid because of spatial variations in soil moisture content from point to point within a basin.

In 1957 Kohler suggested the concept of the two capacity storage technique (Kohler and Richard, 1962). According to this concept the total soil moisture capacity consists of two portions i e. the upper storage and the lower storage. The capacity of the upper zone is about one inch (25 4mm), while that of lower zone is several inches He then assumed that evapotranspiration depletes the upper level at the potential rate and any subsequent demand for Et is satisfied from the lower zone proportional to the moisture available i e depletion of the lower zone occurs when the upper zone storage is exhausted Precipitation would first satisfy the upper layer storage and then contribute to that of the lower zone Runoff, therefore, would be the precipitation in excess of the deficiency of the two storage levels.

Comparing this two storage capacity concept with the single storage concept, the former has the advantage of accounting for increased evapotranspiration following a dry period (Kohler and Richard, 1962), while in the latter method since the ratio of actual to potential Et is assumed in proportion to the available moisture, no storm, unless big enough to saturate the basin, can account for increased evapotranspiration following a dry weather period.

Following the concept of the two capacity storage, Kohler et al (1962) suggested the multi-capacity storage technique According to this concept the distribution curve of soil moisture storage is represented by several values i e 2 inches (5 1 mm), 5 inches (127.0 mm), 10 inches (254 0 mm) and 20 inches (508 0 mm) This representation was in consideration of the spatial variations of soil moisture content throughout the basin and is in accordance with the Penman method applied to the study of the water balance of the Stour catchment (Penman, 1950).

Day by day moisture deficiency is computed for each of these storages independently The assumption is made that the depletion of soil moisture occurs by evapotranspiration at the potential rate as long as the deficiency is less than the assumed capacity The average basin moisture deficiency is then considered to be the weighted mean of deficiencies of the capacities. The weights used by Kohler et al (1962) were determined by correlation analysis, so that the best index of storm runoff was obtained

To estimate runoff from the deficiency index thus obtained, he suggested a relationship of the form $Q_t = (p^n + d^n)^{1/n} - d$ should be adopted for the derivation of storm runoff, where

Q_t - is the total runoff in inches

P - is the precipitation in inches

d - is the mean weighted deficiency

n - is a derived function of soil moisture always greater than one given by the regression equation $n = C + Kd$ with C a constant and K the regression coefficient

The initial values of C and K were suggested to be 2 0 and 0.5 which could then be changed if the runoff thus calculated did not compare closely with the observed value during a period of time This equation is in fact that of a series of curves which approach deficiency asymptotically, (Fig 63) and differ from the threshold concept of Thornthwaite. This equation assumes that runoff is negligible at the beginning of precipitation and that precipitation minus runoff approaches the deficiency asymptotically as precipitation continues

Simulation methods Simulation is defined as the indirect investigation of the response or behaviour of a system (Crawford and Linsley, 1966). Simulation methods are divided into physical, analog and digital types

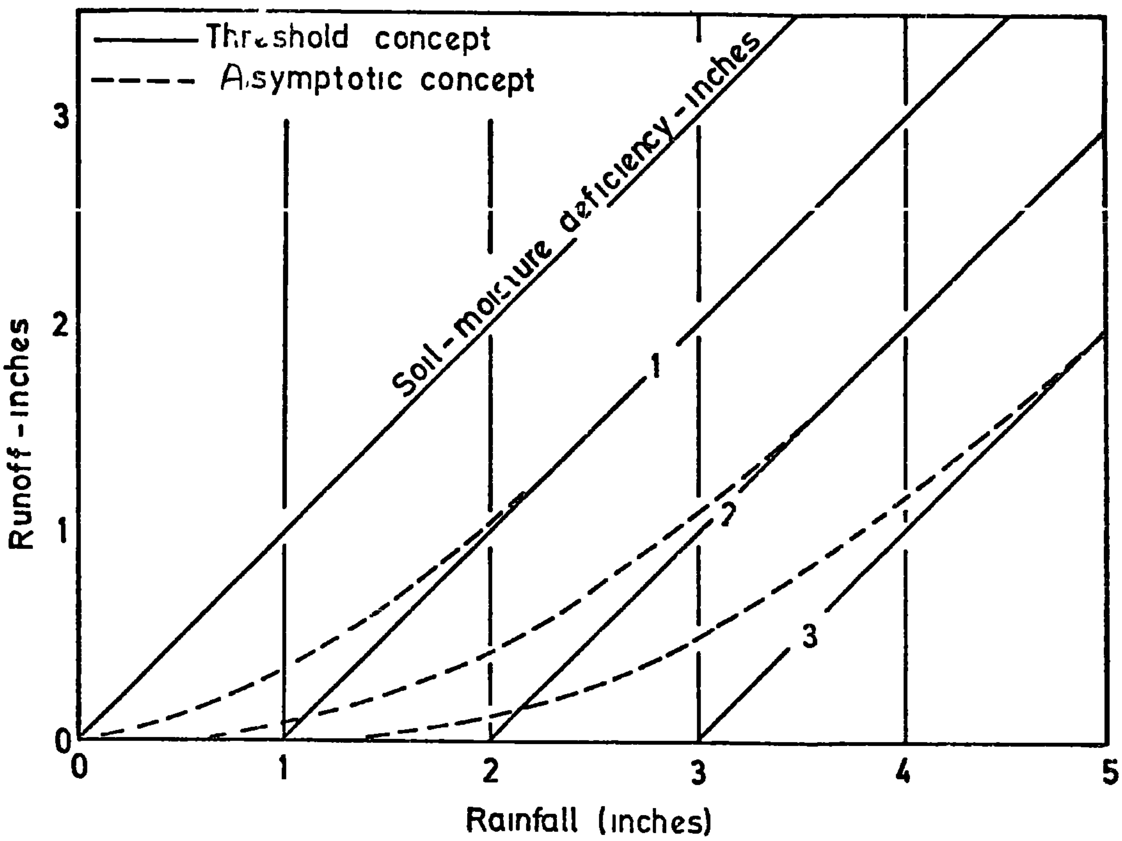


Fig 63 Rainfall – Runoff relation (Redrawn from an original by Kohler & Richard 1962)

Out of these different methods of simulation, digital ones appear to have been developed most for the estimation of continuous flow data. Within the category of digital simulation, the Stanford Watershed Model, is the first functioning model and by far the most widely applied one (James, 1972).

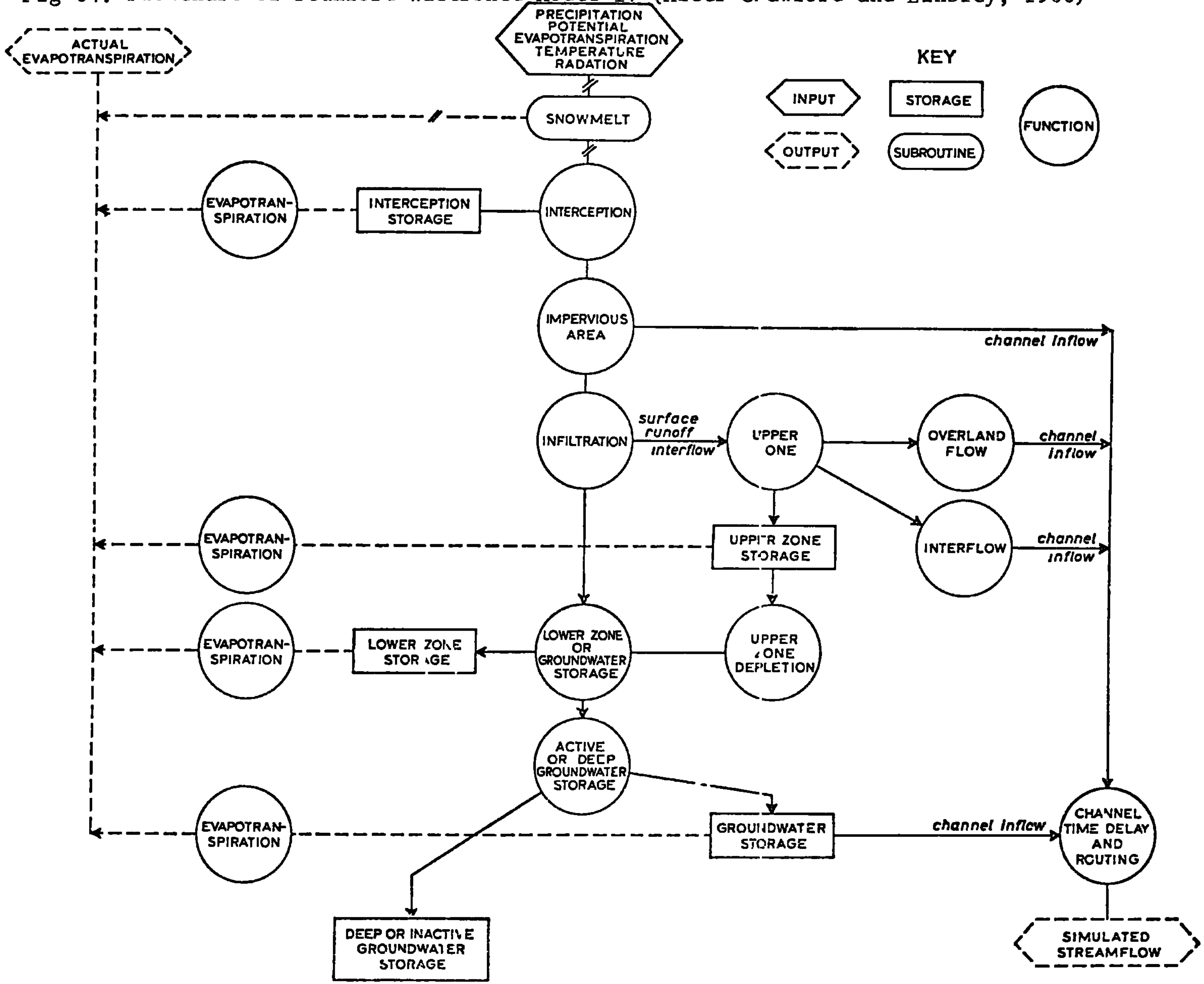
The Stanford Watershed Model is a deterministic conceptual non-linear mathematical model which has been progressively developed during the period of 1959-1966 at the Civil Engineering Department of Stanford University The original reason for undertaking the work was to develop a model which could be used as a device to synthesize riverflow data to supplement short records for the estimation and prediction of peak flow frequencies as well as runoff volume and water balance calculations. The model has, however, been found useful in the evaluation of the effect of artificial changes in the hydrological regimes of catchments and as a means of exploration of the runoff process for an improved understanding of hydrology.

The first model when developed, was reported by Linsley and Crawford in 1960. In this model daily rainfall and evapotranspiration values were the main hydrological data fed as input and daily runoff was the output of the model. The results of runoff simulated from this model were not particularly accurate. The reason could possibly be explained by the use of daily rainfall values which were used to divide the major storms between two days, even though the total duration of such storms may have been considerably less than 24 hours

The Stanford Watershed Model II was reported in 1963 (Crawford and Linsley,1963). The main input hydrological data in the model were the hourly rainfall and daily evapotranspiration rates from the catchment The output from the model was a continuous hydrograph. Some of the variables in this model were fitted semi-automatically by the insertion of a subroutine When this subroutine was in use, riverflow data and selected values of daily groundwater flow had to be provided as additional input data The computer starts the calculation with a set of values for variables which had been assumed by the operator on the basis of experience and when a month's value had been computed, the machine compared the computed monthly total river flow and the groundwater flow with the observed If these values were not within a pre-set tolerance, the machine automatically selected new values for variables following instructions given in the subroutine and repeated the computation for the month The constants determined by the machine for each month were not completely consistent and some smoothing was required to give a single set of values for use in river flow synthesis After volumetric adjustments were made, the hydrograph of the major storms were compared with the computed ones and adjustment was made to give the best fit The model was tested on eight watersheds ranging in size from 22-88 mi^2 (57-228 km^2) and with mean annual rainfalls of between 23 (589 mm) and 54 (1373 mm) inches. Errors in the values of peak flows on two of the flood hydrographs shown in the paper were approximately 20 per cent and the mean daily flow and time of flood in some cases exceeded 100 per cent Nevertheless the errors were said to be random and the frequency distribution characteristics of the desired series agreed well with the observed data

The final results of the pioneering research on watershed modelling at Stanford were given in 1966 in a detailed report by Crawford and Linsley The flow chart of the Stanford Watershed Model IV is shown in Fig 64. Incident rainfall, according to this chart, is first subject to the operation of interception storage This is done by diverting the moisture for interception until the storage is full Thereafter, all of the rainfall (minus evapotranspiration loss from interception storage) is subject to the operation of impervious area which immediately results in runoff and channel inflow Simultaneously the rainfall is subject

Fig 64. Flowchart of Stanford Watershed Model IV (After Crawford and Linsley, 1966)



to infiltration in pervious parts of the catchment which determines the amount of moisture penetrating the soil surface. The remainder is subject to diversion into depression storage or might result in overland flow, interflow and delayed infiltration to the lower zone storage.

The part of the rainfall which has infiltrated down the soil profile to the groundwater level, plus the fraction of delayed infiltration from the upper zone is subject to operation of lower zone Some of this moisture would add up to the lower zone storage, from which evapotranspiration demand, not satisfied from the upper zone and interception storage, will occur at a rate proportional to the ratio of current storage and nominal storage. Excess moisture would then be subject to active groundwater storage or inactive storage. Active storage will contribute to channel inflow and will satisfy any left over evapotranspiration demand Channel inflow due to the presence of impervious areas, surface runoff, interflow and groundwater flow are all summed up and they become subject to channel time delay and routing, the result of which would be the simulated streamflow Crawford and Linsley (1966) represented each of the above mentioned processes by an equation, which is the same for all the watersheds They also used two sets of factors i.e. measured parameters, which are determined by each researcher using the information from the hydrometeorological maps and data and the fitted parameters which are to be established by trial and error methods i.e. simulating a period of runoff with one set of values, and comparing the results with the actual runoff during the period. Those sets of values would be selected that give the closest fit between the actual and simulated runoff. Dawdy and O'Donnell (1965) mention that the Stanford Watershed Model seems to point the way to a new approach to an old problem. The model since its development in 1966, has been applied to many watersheds in the U S A

The Stanford Watershed Model has gone through a number of changes since 1966 by Crawford The result of the latest development published is a new model called Hydrocomp Simulation Programming (Hydrocomp International, 1970), incorporating a much more sophisticated routing technique. This new routing technique called the "Kinematic Wave" approach uses the actual channel dimensions, roughness coefficients and calculates continuous stage and flow velocities throughout the system

Statistical methods These methods are based on the application of mathematical principles to available data to predict frequency of occurrence of floods and droughts of different magnitudes An example of these methods was used in chapter 6 to predict 100-year flood and 100-year drought for the Browney basin

CHAPTER EIGHT

THE APPLICATION OF THE STANFORD WATERSHED MODEL IV TO THE BROWNEY BASIN

The Stanford Watershed Model has been applied to many watersheds in the U S.A. since its development in 1966. However, its application to catchments in other countries has been rather limited. In the United Kingdom there have been some works done on simulation of the River Clyde in Scotland by the Stanford Watershed Model IV (Fleming,1970 and Bunny, 1973). In these studies the period of simulation has been limited to two years, one year for optimization and the other for testing the accuracy of optimized parameters.

Crawford and Linsley (1966), however, suggest the use of two to three years of record for optimization of parameters and a similar length of time to test the validity of optimized parameters Therefore, to simulate runoff from a watershed by this model, a minimum of four years of record should be available

One of the reasons for the limited application of the model to British rivers could be the difficulty which is often experienced in collecting and processing the data which are required for the simulation. For example to obtain one year's data of hourly rainfall, a total of 8760 values have to be read visually from rain charts. Considerable time has also to be spent in recording and processing meteorological data for the derivation of the Penman estimate of evapotranspiration.

In order to apply the model to the Browney basin a search was made of the available raingauge charts of the recording Casella gauge at the Durham University Observatory to decide on the number of years for which the model could be run. Unfortunately many of the charts prior to January 1969 were missing and records were faulty owing to the malfunctioning of the recorder's pen. Therefore, the period of run selected started on 1st January, 1969 and continued until 30th September,1973.

Hourly rainfall for this period was read visually from the charts and punched onto computer cards.

Daily evapotranspiration data for the same period were calculated by the Penman formula using daily climatological data. In view of the fact that the Penman EO_2 values were shown to be more representative of catchment evapotranspiration than Penman EO_1 and Penman Et estimates, (chapter 5), the Penman EO_2 was taken as the potential evapotranspiration input data in the final optimization of parameters and simulation of the catchment. During the initial testing of the model, however, the results of evapotranspiration measurement study were not available and so the input evapotranspiration data were those of Penman Et. The daily values for evapotranspiration used were the average daily values of every third of a month (10 days). This procedure was followed because the Penman formula is more accurate for estimation of Et over longer periods.

<u>Model operation and description of parameters</u> The Fortran IV version of the Stanford Model, used in this study, has been translated from Sub Algol (original language of the model) by E.A.Anderson of U.S Weather Bureau. Operation of the model in the Fortran version starts by accepting input from a number of recording gauges (max. five) and it then continuously calculates riverflow values at several points along the river channel (max. ten) called flowpoints. The flowpoints are usually situated at the river gauging stations The runoff contributing area of each flow point is divided into segments so that there are one or more segments for each recording gauge. The boundaries of these segments are decided upon by topographic consideration or by constructing a Thiessen network, (Thiessen,1911)

For the basin under study there is only one recording raingauge The whole area has, therefore been considered to represent one segment and the riverflow has been simulated at one flow point which coincides with the river gauging station at Burn Hall

Figs. 65 and 66 show the items of input and their corresponding formats required for running the model. The first figure shows that all the parameters (measured and fitted) and supplementary information (e.g. name of the basin) are listed on 15 cards. The input data on these cards are explained in the following paragraphs. Throughout the program 1 means yes and 0 means no.

1st Card - Information concerning the run (INFRO)

2nd Card - Basin name. (BASIN)

3rd Card - (1) First month of the first year. (MO1)

(2) Last two digits of the first year. (YR1)

(3) Last month of the second year. (MO2)

(4) Last two digits of the second year. (YR2)

(5) Number of flow points in the run. (NPTS)

(6) Number of gauges in the precipitation input stream. (PXIN)

(7) Number of precipitation gauges used in the run. (NGAGES)

(8) The gauges in the input stream to be used in the order they are to be assigned e.g. PXIN = 10, NGAGES = 3, RGIN = 8,2,6. Then the 8th gauge in the input stream would be gauge 1 for the run etc (RGIN)

4th Card - (1) Output hourly flows above preset base if yes. (OUTHER)

(2) Exponent of the infiltration curve equation

$$B = \frac{CB}{\left(LZS/LZSN\right)^{POWER}}$$ (POWER)

(3) Wt. factor in equation UZSNT = UZSN + UZSNWF x AEPI (UZSNWF)

5th Card - Monthly potential evapotranspiration adjustment factor. (PEADJ)

6th Card - Raingauge PE adjustment factor. (GAGEPE)

7th Card - (1) Raingauge name. (RGN)

(2) Ratio of average segment rainfall to average gauge rainfall. (K1)

Fig.65.

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C
          STANFORD WATERSHED MODEL IV INPUT FORMAT
               TRIAL CALIBRATION RUN STANFORD WATERSHED MODEL      K.SHAHLAEE
               RIVER BROWNEY NORTHEAST ENGLAND
   10    72     9    73     1     1     1  1
    1                2.0                                                        0.0
  1.0   1.0   1.0   1.0   1.0   1.0   1.0   1.0   1.0   1.0   1.0   1.0
  1.0
CASELLA                1.17 0.05 0.10 0.00 4.00 0.10 0.00 0.24 0.00 0.20
                        936 0.06 0.30 0.700.990 1.00
                       0.00 8.10 0.33 0.14 0.00 0.00 0.42
BURN HALL                             69.7  0.0    0    0    6    8
                                   0    1    1    0       500 2500
                                0.00 0.29 0.33 0.15 0.06 0.04 0.03 0.01
                                   1    1    1    1    1    1    1    1
                                   0    0    0    0    0    0    0    0
                                   0    0    0    0    0    0    0    0
C
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Fig.66

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          STANFORD WATERSHED MODEL IV INPUT FORMAT
           72101 0.026 0.026 0.026 0.026 0.026 0.026 0.026 0.026 0.026 0.026
           72102 0.020 0.020 0.020 0.020 0.020 0.020 0.020 0.020 0.020 0.020
           72103 0.031 0.031 0.031 0.031 0.031 0.031 0.031 0.031 0.031 0.031 0.031
           72111 ......................................................
           72112 ......................................................
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           ............................................................
           73011 0.008 0.008 0.008 0.008 0.008 0.008 0.008 0.008 0.008 0.008
           ............................................................
           73003.......................................................
           72101     9     9     9     9     9    11    13    12    15    17
           72102    12    11    10    11     9     9     9     9     9     9
           72103    11    10    10     9    10    11    12    13    11     9    10
           ............................................................
           73103    13    23    23    16    14    13    17    19    23    20
        A     0     0
        212325 72 10  1 1
        212325 72 10 09 2                                              .16  .03  .00
        212325 72 10 11 1      0.02 0.02
        212325 72 10 11 1 0.01                                 0.02 0.02 0.03
        212325 72 10 20 2                  0.03
    99
        212325 72 11 03 1                     0.06 0.07 0.11 0.02                 0.05
        ......................................................................
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	(3)	Impervious area (fraction).	(IMPV)
	(4)	Interception storage maximum value (inches).	(EXPM)
	(5)	Nominal upper zone storage (inches)	(UZSN)
	(6)	Nominal lower zone storage (inches).	(LZSN)
	(7)	Infiltration index	(CB)
	(8)	Interflow index	(CC)
	(9)	Actual evapotranspiration loss index.	(K3)
	(10)	Portion of groundwater recharge, assigned to deep percolation (fraction).	(K24L)
	(11)	Evapotranspiration from groundwater	(K24EL)
8th Card -	(1)	Overland flow length.	(L)
	(2)	Overland flow slope.	(SS)
	(3)	Manning's n for overland flow.	(NN)
	(4)	Interflow recession rate.	(IRC)
	(5)	Groundwater recession rate	(KK24)
	(6)	Groundwater recession variable output	(KV)
9th Card -	(1)	Initial upper zone storage (inches).	(UZS)
	(2)	Initial lower zone storage (inches)	(LZS)
	(3)	Groundwater storage (inches)	(SGW)
	(4)	Groundwater slope.	(GWS)
	(5)	Initial surface detention storage (inches).	(RES)
	(6)	Initial interflow detention storage (inches).	(SRGX)
	(7)	Initial interception storage (inches).	(SCEP)
	(8)	Antecedent potential evapotranspiration index.	(AEPI)
10th Card -	(1)	Flow point name.	(FPN)
	(2)	Area, square miles.	(AREA)
	(3)	Channel attenuation parameter constant K	(KS1)
	(4)	Variable K of ycs.	(VARK)

(5) Vary lag if yes. (VARL)

(6) Routing internal-Hours. (RT.EINT)

(7) Number of elements in the time delay. (ELEMTS)

11th Card - (1) = 1 if observed six hour flow
= 2 if observed hourly flow
= 0 if neither (CHECK)

(2) Observed mean daily flow if yes (COMPAR)

(3) Plot mean daily if yes (PLOT)

(4) Plot hourly or six hour flows if yes. (PLOTHR)

(5) Preset base for output hourly flow. (MINFW)

(6) Maximum ordinate mean daily flow plot. (PLOTMX)

(7) Maximum ordinate hourly or 6 hour-plot. (PHRMX)

12th Card - Channel time delay histogram (fraction of flow reaching the channel which is delayed before reaching the segment outlet by the TIMEAR of the subscript. (TIMEAR)

13th Card - Number of raingauges to be used for each segment of TIMEAR. (GAGEAR)

14th Card - Additional flow point entering each segment of TIMEAR. (ADD FLOW1)

15th Card - Additional flow point entering each segment of TIMEAR. (ADD FLOW2)

Among all these parameters (measured or fitted), those appearing on cards seven to fifteen are discussed in more detail in the following pages.

$\underline{K_1}$. (Ratio of average segment rainfall to average gauge rainfall) The derivation of this parameter was discussed in chapter 6. The value of this parameter is 1 17.

<u>IMPV</u> (Fraction of impervious area). The fraction of the total area which is impervious alters runoff volumes and affects the whole processes of overland flow and interflow

To determine the fraction of impervious area of the Browney basin, measurements were made of total area of buildings and roads on a 1 25000 O.S. map. This area amounted to 7 per cent of the total area Crawford and Linsley (1966) suggested the adjustment of measured and estimated impervious area to the effective impervious area by referring to the relationship presented in Fig.67. Based on this relationship the effective impervious area in the Browney basin was estimated to be about 5 per cent of the total catchment.

EPXM (Interception storage parameter) The vegetation cover of the catchment governs the value of this parameter The effect of the vegetation cover in reducing the amount of precipitation at any time reaching the ground surface also depends on its current storage level

According to the model, all incoming moisture enters interception storage until a pre-assigned volume EXPM is filled. However, the process of evaporation which takes place at the potential rate, depletes the moisture from interception storage, and consequently precipitation will continue to replace the water from interception storage lost by evaporation

Crawford and Linsley (1966) give some estimates for interception storages. According to these values the EXPM of a grassland watershed is 0.10 inches (2 5 mm) while those for moderate forest cover and heavy forest cover are 0.15 inches (3 8 mm) and 0 20 inches (5 1 mm) respectively. For the Browney basin a value of 0.1 inches (2.5 mm) was considered to be a good estimate of interception storage.

UZSN (Nominal upper zone storage) This parameter simulates the diversion of overland flow into depression storage, soil fissures and disturbed or dry soil surfaces. The parameter is, therefore, independent of rainfall intensity Evapotranspiration and percolation to the lower zone continuously removes water from the upper zone. The potential addition of moisture to the upper zone storage is given by the relationship shown in Fig 68 which expresses any increase in upper zone storage in terms of the ratio of current upper zone storage (UZS) to nominal upper zone storage (UZSN)

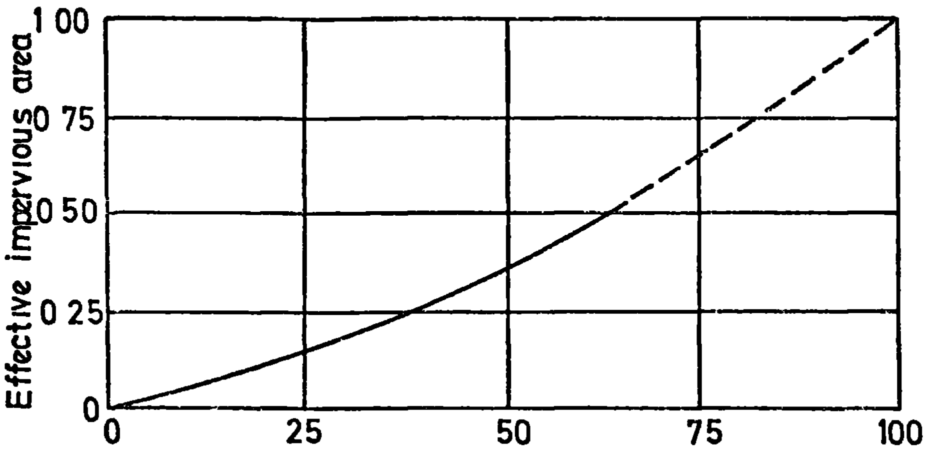


Fig 67 Total impervious area (%)
(After Crawford & Lindsley 1966)

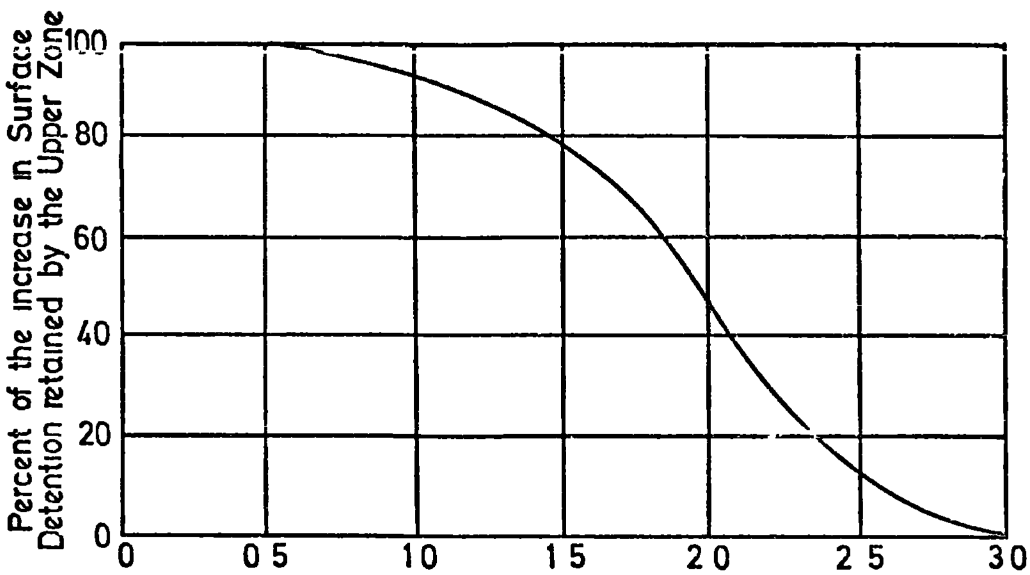


Fig 68. Upper zone soil moisture ratio
(After Crawford & Linsley 1966)

This relationship shows how upper zone storage may prevent overland flow from a portion of the watershed. Evapotranspiration from the upper zone occurs at the potential rate

The nominal value for upper zone storage can initially be estimated from Table 46 However, it may need to be optimized later. The initial value of UZSN in this study was taken to be 0 6 inches (15.2 mm)

Table 46 Estimation of UZSN. After Crawford and Linsley (1966).

Watershed	UZSN
Steep slopes Limited vegetation Low depression storage	0.06 LZSN
Moderate slopes Moderate vegetation Moderate depression storage	0.08 LZSN
Heavy vegetal or forest cover Soils subject to cracking, high depression storage, very mild slopes	0 14 LZSN

LZSN (Nominal lower zone storage) This parameter models the storage of the infiltrated moisture in the soil profile and is approximately equal to the median value of lower zone storage In other words LZSN is the storage level at which 50 per cent of all incoming moisture moves into groundwater storage This parameter is one of the most important of all and is the one on which the long term volume of runoff depends The value of LZSN should also be determined by several runs However, for watersheds with rainfall distributed uniformly throughout the year, an initial estimate of $4 + \frac{1}{8}$ x (Mean Annual Rainfall) inches is given. The initial estimate of this parameter for the Browney basin was 7 5 inches (190 5 mm).

CB (Infiltration index) This is an input parameter that governs the over all level of net infiltration and is also used to calculate deep percolation In view of the fact that the infiltration rate is highly dependent on soil moisture conditions within the soil profile, a close relationship exists between the value of LZSN and CB, which in turn affects the runoff process. Crawford and Linsley (1966) suggest a range of 0.3 to 1.2 for values of CB under different climatic and soil regimes Based on their suggestion an initial value of 0 6 was used

<u>CC</u> (Interflow parameter) This parameter is an index of the ratio of the quantity of moisture added to interflow detention, to that added to surface runoff detention. Therefore, it governs the hydrograph shape and timing This is because with an increase in the quantity of moisture joining interflow detention, there would be a delay for the flow to reach the channel and thus the hydrograph would be less peaked than when a higher proportion of moisture is joining the surface runoff detention. The range of values of CC is given by Crawford and Linsley (1966) to be between 0.5 to 3.0. The initial value, based on this guideline, was adopted to be 1 5

<u>K3</u> (Actual evapotranspiration loss index). Evapotranspiration, according to this model, occurs at the potential rate from the interception and upper zone storages. Any excess evapotranspiration (Ep) which is not thus satisfied, will draw moisture from the lower zone based on the concept of evapotranspiration opportunity. According to this concept the areal variation of evapotranspiration opportunity is represented by a cumulative frequency distribution curve shown in Fig.69.

Using this figure, the water loss for evapotranspiration from the lower zone when Ep is less than r is

$$E \text{ (actual Et)} = Ep - \frac{(Ep)^2}{2r}$$

The variable r (evapotranspiration opportunity) is given by $r = K3 \frac{LZS}{LZSN}$, where K3 is an input parameter. Simulated values of K3 for different vegetation covers are shown in Table 47 From this table a value of K3 = 0.23 was chosen for the watershed.

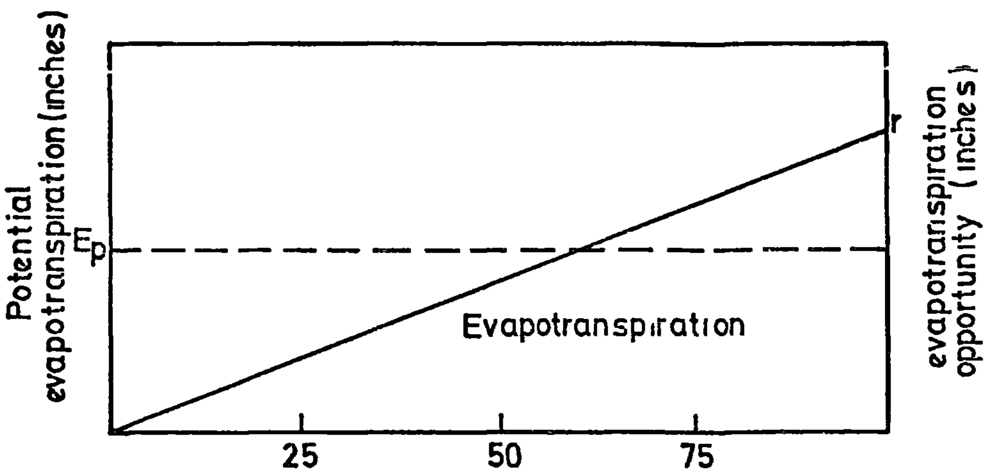


Fig. 69 Percentage area with a daily evapotranspiration opportunity equal to or less than the indicated value (After Crawford & Linsley (1966))

Table 47. Values of K3 for different watershed cover. After Crawford and Linsley(1966)

Watershed cover	K3
Open land	0.20
Grass land	0 23
Light forest	0 28
Heavy forest	0.30

K24L (Index to inactive groundwater recharge). This parameter governs the fraction of recharge from active groundwater to the inactive section. Estimation of this parameter for catchments with such losses should be based on observed changes in deep groundwater levels or estimates of subsurface outflow from the basin Alternatively K24L may be approximated by trial.

For the Browney basin an initial estimate of 0 2 was applied.

K24EL (Fraction of basin with shallow groundwater). This parameter simulates the loss of groundwater due to evapotranspiration by vegetation. During periods of dry weather, the shallow groundwater zone would be the only section of the basin for which actual evapotranspiration, equal to the potential values, would be occurring. Therefore, this results in exhaustion of groundwater storage. For the Browney basin this value was assumed to be 0 05.

L (Length of overland flow) This parameter is a measure of stream spacing and as mentioned in the first chapter is determined from the formula

$$L = \frac{1}{2D}\sqrt{1-\left(\frac{S_c}{S_y}\right)^2}$$

and is equal to 936 feet (285 m) in the Browney catchment.

SS (Overland flow slope). Following the procedure recommended by Linsley et al (1949) and Nash (1966), the mean overland flow slope was found to be 0 06

NN (Manning's n for overland flows). This parameter was estimated, using Table 48 given by Crawford and Linsley (1966), and its value is 0.30.

IRC (Daily interflow recession rate). This parameter is important in watersheds in which interflow forms an important portion of direct runoff

Table 48 Values of Manning's n for different watershed cover After Crawford and Linsley (1966)

Watershed cover	Manning's n for overland flow
Smooth Asphalt	0.012
Asphalt for concrete paving	0.014
Packed clay	0.030
Light turf	0.200
Dense turf	0.350
Dense shrubbery and forest litter	0.40

It is defined as the ratio of interflow discharge on any day to that of the preceding one To determine this parameter either of two methods could be used (1) the trial and error method and (2) the Barnes (1939) method The former method can be applied by matching the shape of the simulated and the observed hydrographs.

To calculate IRC by the Barnes method, the hydrograph of a storm on 14th August 1971 was plotted on semi-logarithmic paper, (Fig.70). The groundwater recession limb of the hydrograph was extended back under the hydrograph. The residual ordinates above the groundwater which represented the combined surface runoff and interflow was replotted, and a straight line was fitted to the recession part. Extension of this line separated interflow from surface runoff, from which IRC was calculated to be 0 6.

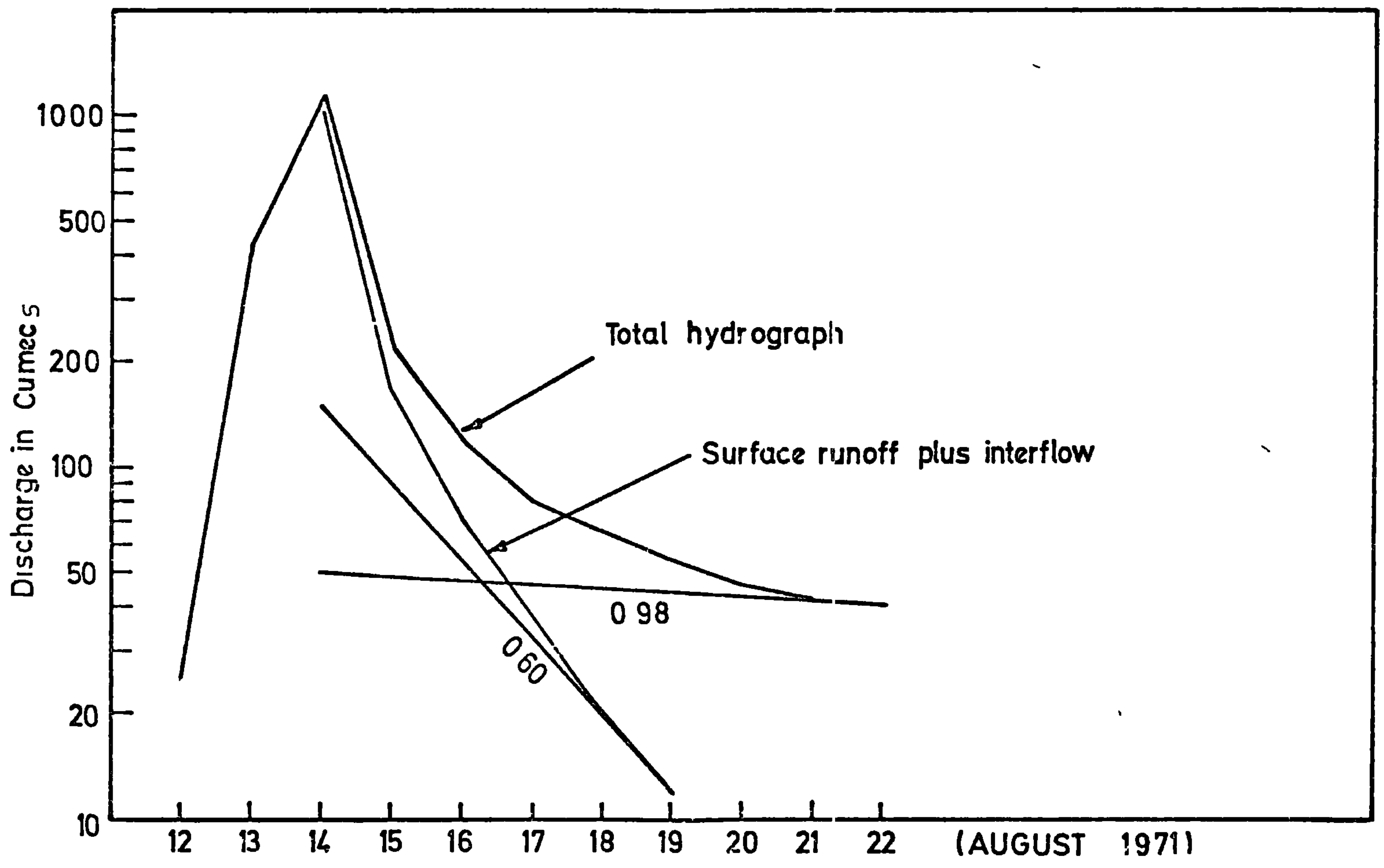


Fig. 70. Hydrograph analysis showing the derivation of daily interflow and ground water recession rates.

KK24 (Daily groundwater discharge recession rate). This parameter is defined as the ratio of groundwater discharge at any time to that of twenty four hours earlier. The value of this parameter was calculated to be about 0.98.

KV (Parameter for variation of groundwater recession rates). This parameter has been introduced in the groundwater outflow equation, in order to add flexibility in the simulation of groundwater outflow. Groundwater outflow GWF at any time is given by,

$$GWF = LKK4 \times (1.0 + KV \times GWS) \times SGW$$

where GWS is groundwater slope and SGW is groundwater storage. The parameter LKK4 is defined as $LKK4 = 1.0 - (KK24)^{\frac{1}{96}}$. When KV is zero and there is no inflow to SGW, the groundwater outflow equation reproduces the commonly used logarithmic depletion curve, i.e. the flow after a period of n days decreases by $(KK24)^n$ and a semi-logarithmic plot of discharge against time is a straight line. Crawford and Linsley (1966), justify the use of KV by reasoning that "... if the typical daily dry season recession rate in a stream is 0.99 and a recession of 0.98 is more typical, when the groundwater storages are being recharged, the value of KK24 can be set to 0.99 and the value of the parameter KV can be adjusted so that 1.0 + KV x GWS will reduce the effective recession rate to 0.98 during recharge period"

For the Browney basin a KV value of 1.0 was adopted.

POWER (Exponent of the infiltration curve equation). The infiltration curve equation is in the form of

$$B = \frac{CB}{\left(LZS/LZSN\right)^{POWER}}$$

in which B is the infiltration volume. Black (1973) reported that numerous trials with H.S.P. (Hydrocomp Simulation Programming), the successor to the Stanford model, have shown that a value of 2.0 is applicable over a wide range of watershed conditions. Consequently this value of POWER was adopted for the Browney basin

UZSNWF. (Weight factor in equation UZSNT = UZSN + UZSNWF x AEPI). The value of this parameter is assumed to be zero By this assumption, the seasonal variation in the surface storage due to changes in vegetation and farming practices is not considered significant.

UZS, LZS, SGW, GWS, RES, SRGX, SCEP and AEPI (Initial storages). These parameters simulate the land moisture conditions at the beginning of the run and they are defined as

UZS (Initial upper zone storage)

LZS (Initial lower zone storage)

SGW (Initial groundwater storage)

GWS (Initial groundwater slope)

RES (Initial surface detention storage)

SRGX (Initial interflow detention storage)

SCEP (Initial interception storage)

AEPI (Antecedent potential Et index)

KS1 (Hourly stream channel storage recession parameter) This parameter is a time distribution factor, determined by the formula

$$KS1 = \frac{\text{Discharge in hr } (t+1)}{\text{Discharge in hr } (t)}$$

or by graphical techniques suggested by Barnes (1939) and described by Linsley et al (1949, 1958) In the model KS1 is a part of the equation $O_2 = I - KS1 (I - O_1)$, which is adopted to route the outflow hydrograph produced by channel translation calculations through a storage system to simulate attenuation in the channel system O_2 is outflow at time t_2, O_1 at time t_1, and I is average inflow at times t_2 and t_1

Channel system To simulate the channel system in the model, a modification of Clark's (1945) empirical routing method is applied

In this simple routing method, called a time-area curve, Clark assumed that a plot of time of flow from any point in the watershed to the outlet represents an outflow hydrograph, when it is caused by a short rainfall. In this method, attenuation due to channel storage is neglected

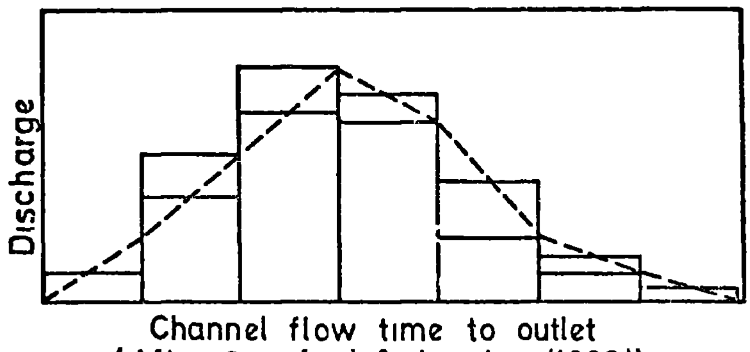
In the Stanford model, since the land surface is modelled separately, Clark's method is modified and called the channel-time-delay curve. It represents an outflow hydrograph which is a response to an instantaneous surge of inflow to the system Attenuation due to storage is represented by reservoir routing A channel-time-delay curve and histogram are shown in Fig 71 To construct such a curve for a watershed the Manning equation may be used to find an estimate of the time of flow from any point in the channel system to the outlet for any assumed discharge.

Alternatively for a basin of a size equal to or greater than that of the Browney, the time-delay histogram could be calculated from typical hydrographs of the watershed This is because for such basins the shape of the outflow hydrograph is primarily dependent on the channel system.

For the Browney basin the hydrograph resulting from a storm on 5th November, 1967 was chosen. This hydrograph, was divided into eight elements each of six hours duration The fraction of the area under each element was obtained by planimetry The corresponding elements in the hydrograph, then, represented the elements of the channel-time-delay histogram adopted for the basin, and are presented in the following Table

Table 49 Elements of channel-time-delay histogram

1	2	3	4	5	6	7	8
0.09	0.29	0 33	0 15	0.06	0 04	0 03	0 01



(After Crawford & Linsley (1966))

Fig.71 CHANNEL-TIME-DELAY CURVE and HISTOGRAM

CHAPTER NINE

SENSITIVITY TESTS AND OPTIMIZATION OF THE FITTED PARAMETERS

Fitted parameters, used in the Stanford Watershed Model, are those that cannot be measured easily. To decide on their values for a particular catchment, therefore, their sensitivities should be tested before they are optimized.

The sensitivity of a parameter in a hydrological model, refers to the degree of variation of simulated flow as a result of changing the value of that parameter Among the fitted parameters in the model, some are more sensitive than others, while there are some which do not affect the output significantly The sensitivity of a parameter, of course, depends to a large extent on the conditions of the watershed to which the model is applied.

Optimization, as referred to hydrological models, is the selection of fitted parameters by closely matching the simulated flow to the recorded one. If the parameters of the model are not correctly optimized, the model would be unfit for any forecasting or synthesizing of streamflow The success of a watershed model, therefore, depends to some extent on the realistic optimization of the parameters and that in turn is dependent on the researcher's judgement and his knowledge of the hydrologic cycle

Dawdy and O'Donnell (1965) refer to other requirements for closely fitting the simulated flow to recorded flow by a hydrologic model These requirements, they mention, are the hydrologic validity of the mathematical model, the accuracy of the data used, the method of fitting the model to the data and the criteria used for closeness of fit

The more refined a hydrologic model is, the more fitted parameters it would contain which need to be optimized The increased number of such parameters results in extra complexity in the model and manual optimization would then be difficult. For this reason several attempts have been made to devise some automatic means of optimization Examples of such methods are given by Dawdy and O'Donnell (1965), Ibbitt and O'Donnell (1971) and James (1972).

In this study manual optimization has been followed. The guidelines given in the published report of Crawford and Linsley (1966), which were briefly explained in the earlier chapter, were used initially. However some of these initial values were changed after many unsuccessful trials The criteria of closeness of fit and decision on the final values of the optimized parameters were based on the total yearly and monthly flows as well as the shape of daily hydrographs over the whole year.

James (1972), in optimization of the parameters of the Stanford Watershed Model by an automatic method, noticed that several combinations of parameters could produce nearly equally good results. One combination simulated better peak flows while another produced low flows more accurately

In the trial and error method adopted in this study, an attempt was made to find that combination which could best simulate the peak flows, as well as low flows, simultaneously. This criteria of closeness of fit was examined by comparing the frequency distribution of the actual and simulated daily flows. The choice of this criteria was in accordance with the objectives in this study

1. To test the ability of the model to predict runoff from the Browney catchment without any particular emphasis on the high or low flows

2. To get an understanding of those components of the water balance equation within the catchment which are not easily studied in other ways e g actual evapotranspiration and groundwater flow.

The water year 1972 was used for the initial sensitivity tests and optimization Normally for any optimization run, depending on the purpose of the application of the model, one or more representative years of data would be employed. For example, if the model is to be used for forecasting peak rates of flow, a water year with high peak flows would be employed. Under some other conditions, for example in reservoir construction, the purpose of the model might be to predict drought periods. In such a case water years with high peak flows would not be of such importance

With the objectives of the present study in mind, no particular emphasis was given to the selection of the period for testing and optimization and the available data for the water year 1972 were used for the purpose of initial testing and optimization. It should be mentioned that throughout this study the hydrographs of mean daily flow were drawn by the computer graph-plotter at Durham University The program written for this purpose is shown in Appendix IV.

Results of the sensitivity tests

The results of the sensitivity tests with the major parameters are given below These tests were carried out by running the model, and varying one parameter at a time The results were then used for the selection of the sensitive parameters in the trial and error optimization. The parameters initially considered were CB, LZSN, UZSN, CC, K3, K24L, LZS, UZS, KK24, IRC, and KV

CB (Infiltration index) The effect of CB under the climatic conditions of the Browney catchment may be observed by referring to Figs. 72 and 73 and Table A (Appendix V). Fig.72 and Table A (Appendix V) present the monthly simulated flow for CB values of 0.01, 0.1 and 0 6, while the hydrographs of mean daily flows of CB 0.1 and 0.01 are plotted in Fig 73 In this figure the hydrograph of CB with a value of 0.6 was omitted for reason of clarity. From Fig.73 it is clear that decreasing the value of CB, increases the peak rate of flow by increasing the surface runoff and decreasing the groundwater flow.

During the months of January, February and March, rainfall and

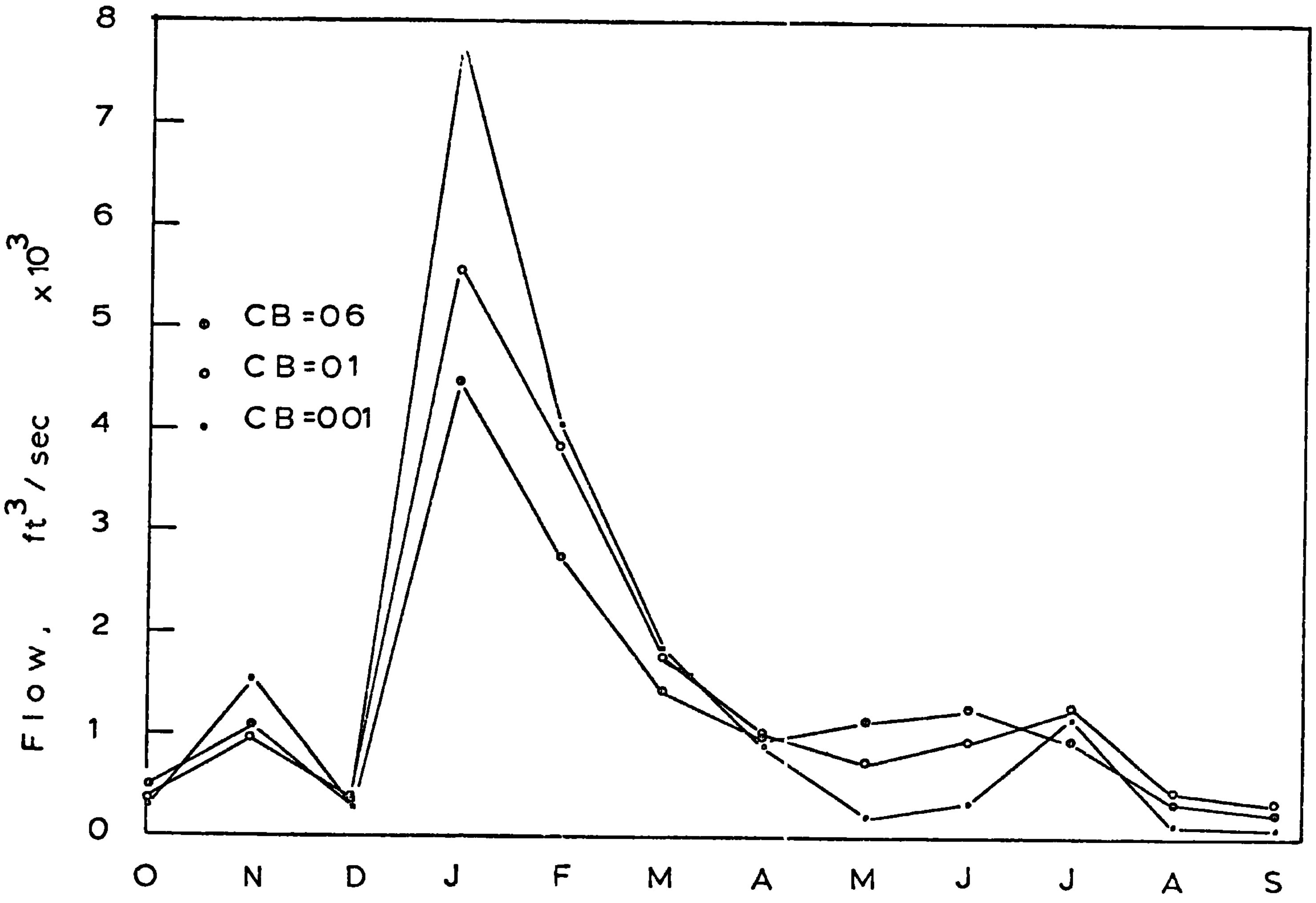
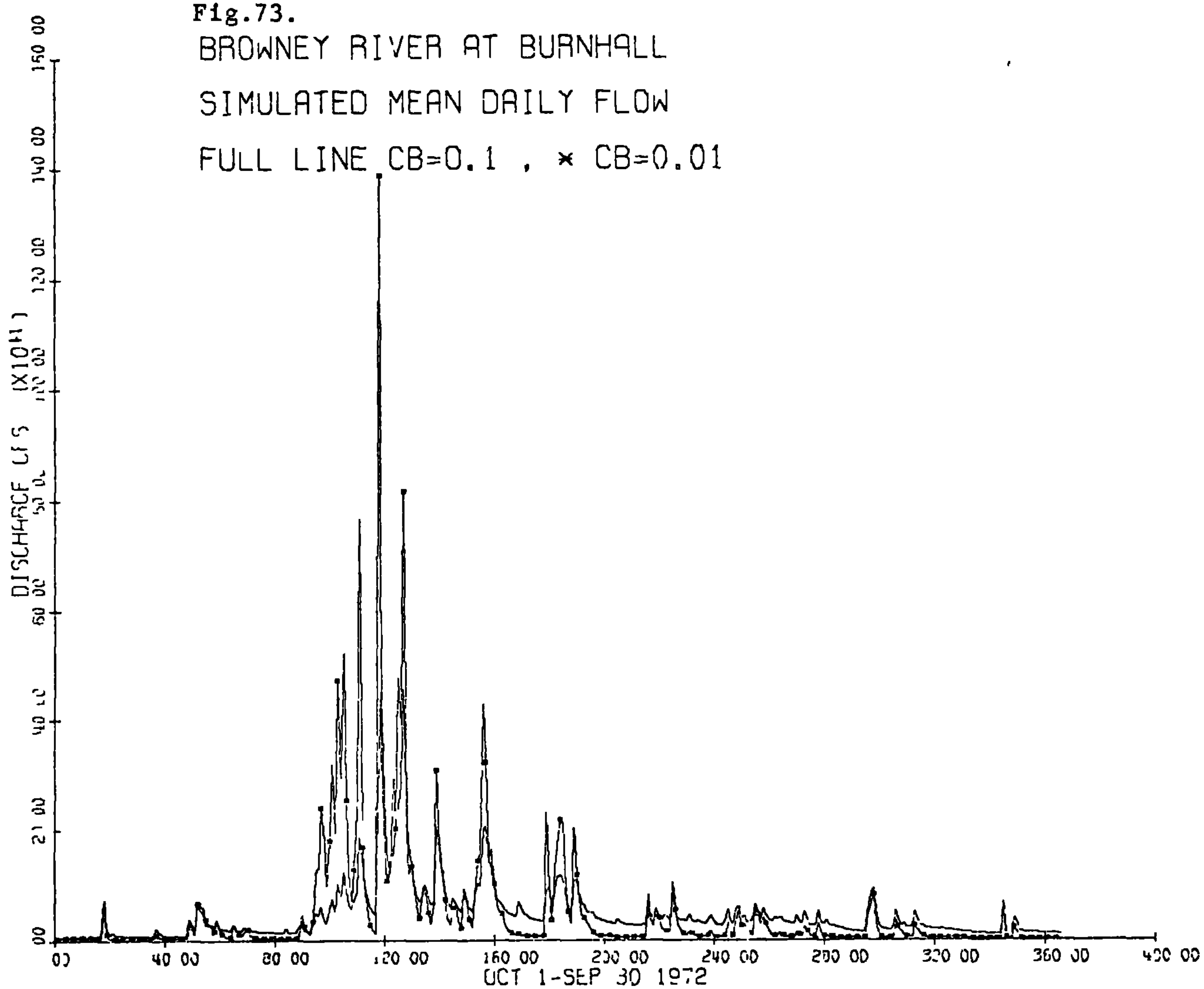


Fig 72 Browney River at Burn Hall Simulated monthly flow for three values of CB

Fig.73.
BROWNEY RIVER AT BURNHALL
SIMULATED MEAN DAILY FLOW
FULL LINE CB=0.1 , * CB=0.01



low infiltration have resulted in large amounts of surface runoff. Under higher CB values, however, the amount of direct runoff has been low and much of the rainfall has infiltrated into the soil to be stored and subsequently released as groundwater flow This is very clearly shown by the comparison of simulation results of a CB value of 0.01 with either the 0.1 or 0.6 (Fig.72 and Table A (Appendix V) This shows that monthly flows during the summer season are much lower for low CB values than for higher ones. An exception is the flow during July 1972 which was higher for lower CB values. This was because of a rainfall depth of 2.84 inches (72.0 mm) occurring during that month The low CB value of 0.01 resulted in large amounts of surface runoff which compensated for the low groundwater flow

Another important observation of the effect of CB can be made by comparing the monthly simulated values of April, July, August and September During these months the flow is higher for the CB value of 0 1 as compared with the simulated flows of 0 01 and 0.6. From these results, it is inferred that there is a threshold value of CB above which the lower direct runoff component results in lower flows than of the threshold value, and below which the low infiltration, with a subsequent lower base flow, produces lower total flow.

To show the variation of high mean daily flows with CB, two examples from 26th January and 4th February, 1972 are cited For 26th January, the peak flow due to a CB value of 0.01 is about three times that of a CB value of 0.1 and for 4th February, the peak due to a CB value of 0.01 is almost twice as much as that due to a CB value of 0 1

The effect of CB upon the volume of flow for the year 1972 is shown in Table A (Appendix V). For this year, which is a year of average flow, decreasing the CB value has resulted in increasing the simulated flow This effect, however, was reversed when the data for the year 1973, a dry year, were employed, (i e decreasing CB values reduced the total flow) Table B (Appendix V) The reason may be explained by the fact that a low CB value under a low rainfall condition would decrease the base-flow and increase evapotranspiration In such a case any surface runoff would be stored in the upper zone which serves for satisfying evapotranspiration demand occurring at the potential rate.

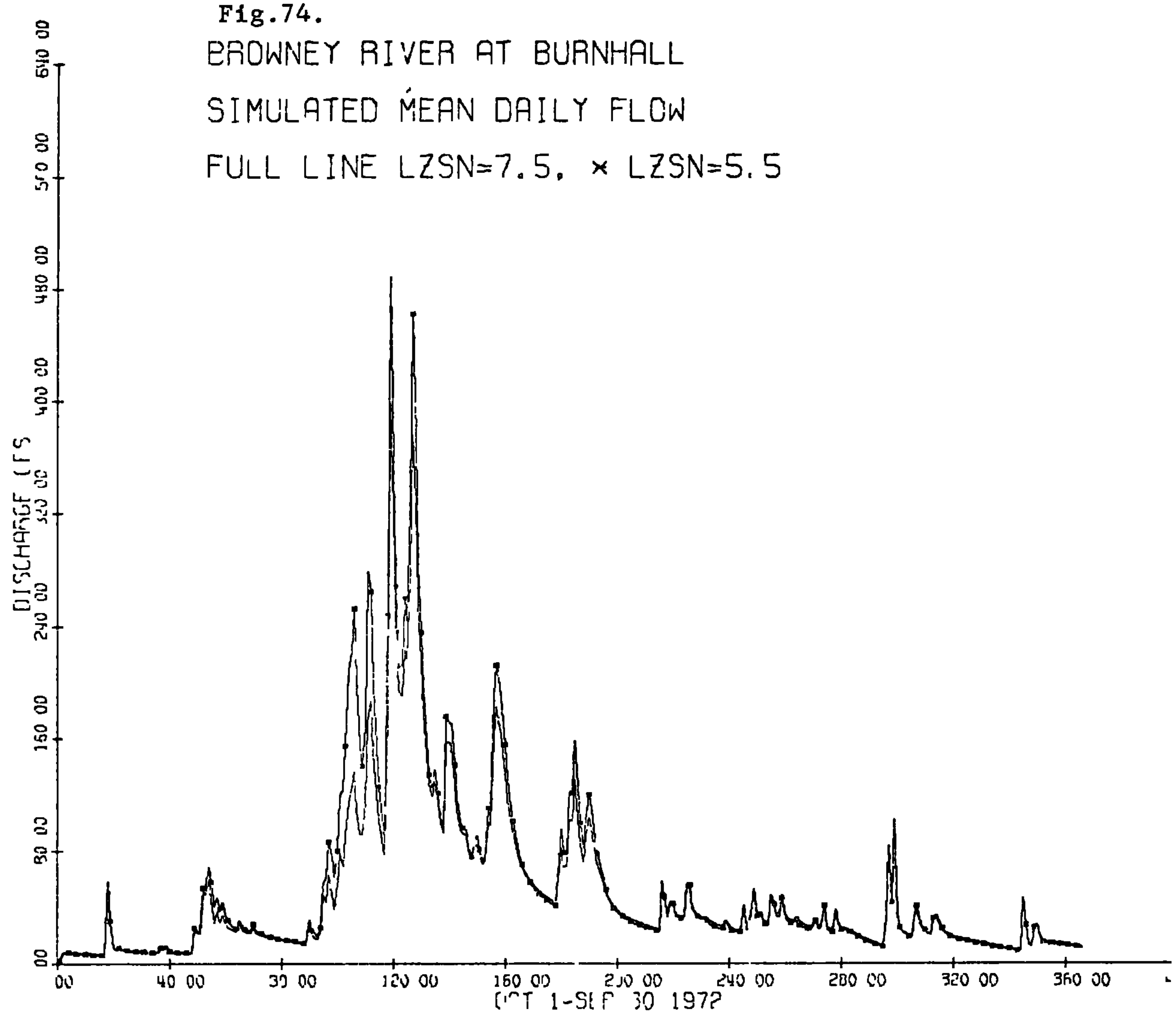
LZSN (Nominal lower zone storage) The effect of LZSN on simulated flow can be described by referring to Figs. 74 and 75 and Table C (Appendix V). Fig.74 shows that decreasing the value of LZSN from 7 5 inches (190 5 mm) to 5 5 inches (140 0 mm) has increased the peak rates of flow Taking the mean daily value of 27th January, for instance, there is an increase of approximately 20 per cent with a decrease of about 27 per cent in the value of LZSN.

From Table C (Appendix V), it is observed that decreasing the LZSN value by 27 per cent has increased the yearly value by about 13 per cent However, this yearly increase is mostly due to increases of flow during the winter months In fact there is a slight decrease in the monthly flows of the summer months

These observations could be accounted for by considering the physical basis and formulation of the model The fact that a lower value of LZSN increases runoff is because a shallow soil with a low moisture holding capacity will not be able to retain as much moisture as a deep soil. Any precipitation in excess of the moisture holding capacity, therefore, results in runoff.

The slightly lower summer monthly flows associated with a LZSN value of 5.5 inches (140 0 mm) as compared with 7 5 inches (190.5 mm), is however explained by referring to the modelling of the actual evapotranspiration process. Since the rate of actual evapotranspiration from the lower zone is assumed to be proportional to the ratio of $^{LZS}/_{LZSN}$, it is to be expected that with a constant value of LZS, a lower LZSN will result in higher evapotranspiration This effect, on a long term basis,

Fig.74.
BROWNEY RIVER AT BURNHALL
SIMULATED MEAN DAILY FLOW
FULL LINE LZSN=7.5, × LZSN=5.5



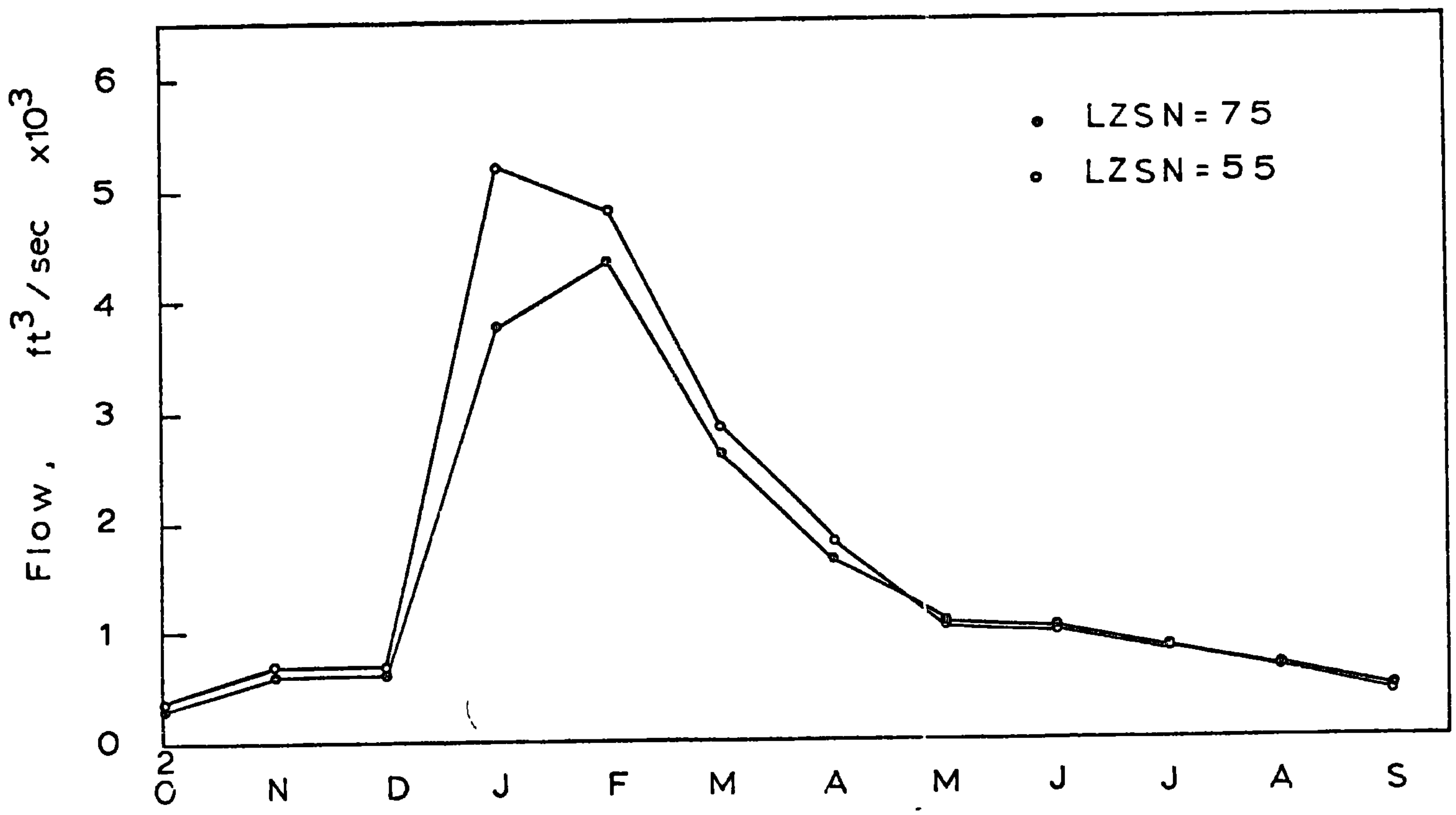


Fig 75 Browney River at Burn Hall Simulated monthly flow for two values of LZSN

depletes the lower zone and hence reduces runoff

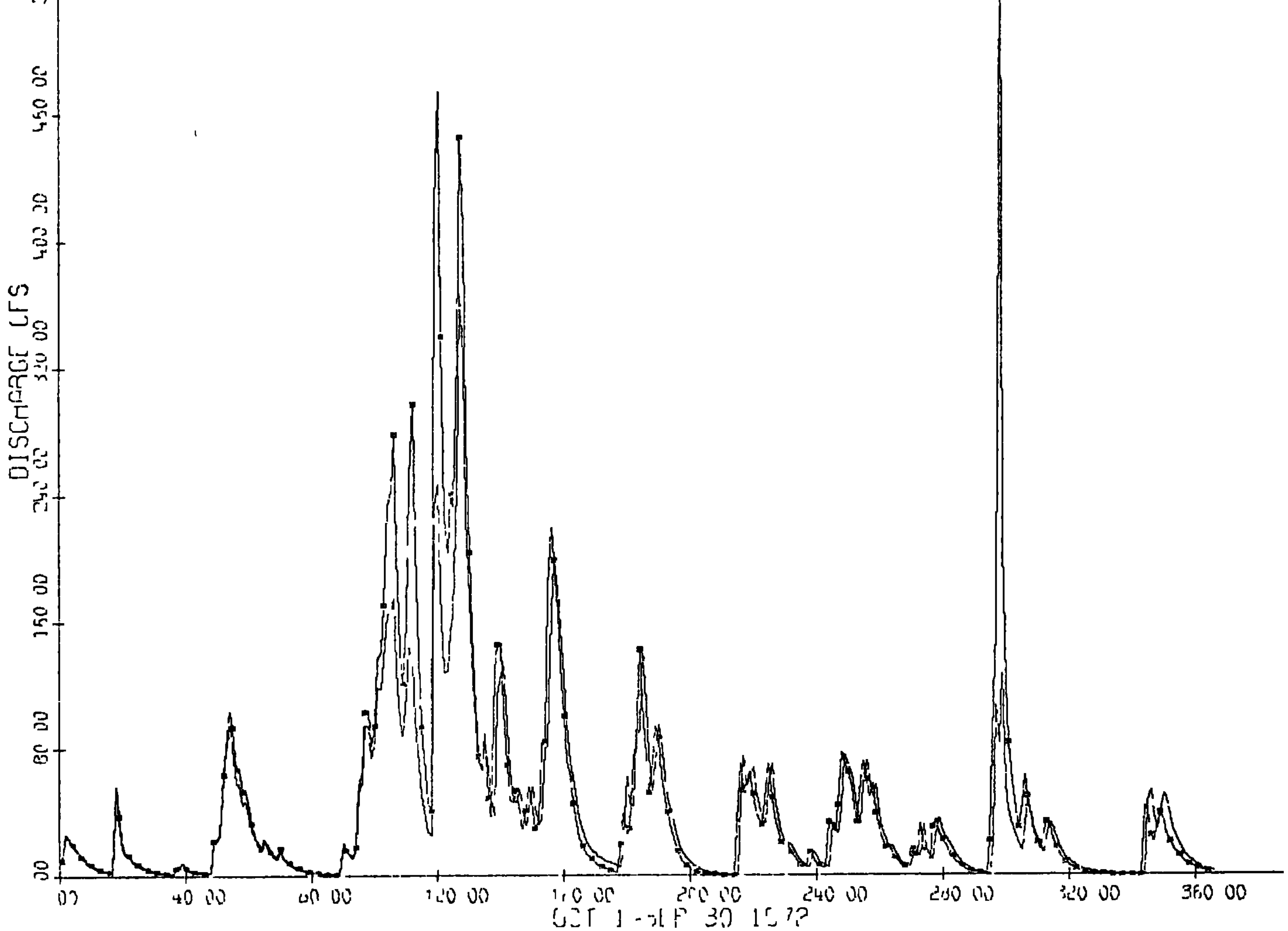
UZSN (Nominal upper zone storage). This parameter is by far one of the most important ones under the hydrological regime prevailing in the Browney catchment and in those areas where precipitation is uniform throughout the year. Its importance is partly due to its retarding effect on direct runoff.

UZSN is also important because of its effect on actual evapotranspiration. In the Stanford Watershed Model, actual evapotranspiration is assumed to be equal to the potential value provided that moisture is available in the upper zone It is, therefore, concluded that if the value of UZSN is under-estimated, incorrect peaks could result during the summer seasons of high evapotranspiration demand. Conversely if UZSN is over-estimated, the runoff values might be under-estimated. The discussion presented in the above paragraphs can be verified by referring to Figs 76 and 77 and Table D (Appendix V), which show the mean daily and monthly flows of two runs of simulation obtained with two different values of UZSN i.e. 1 0 inches (25 4 mm) and 0.3 inches (7.6 mm).

Thus, it is observed that the simulated flows resulting from a UZSN value of 0.3 inches (7 6 mm) have produced peaks which are much higher than those of a UZSN value of 1.0 inches (25.4 mm). An example is the artificial peak which is produced by about 1 1 inches (28 0 mm) on 23rd and 24th July,1972. Since UZSN is lower than what it should be i.e. 0.3 inches (7.6 mm) not much of the rainfall received has served for evapotranspiration or delayed infiltration This however, is not the case with a UZSN value of 1 0 inches (25 4 mm) which by virtue of its high storage capacity, has retarded much of surface runoff and thus, has produced a much lower peak (Fig 76). On a yearly basis increasing UZSN from 0 3 to 1 0 has reduced the total flow by some 20 per cent.

CC (Interflow index). The effect of this parameter on simulated flows was studied by the results of two runs during the initial stage of the

Fig.76.
BROWNEY RIVER AT BURNHALL
SIMULATED MEAN DAILY FLOW
FULL LINE UZSN=1.0, *UZSN=.3



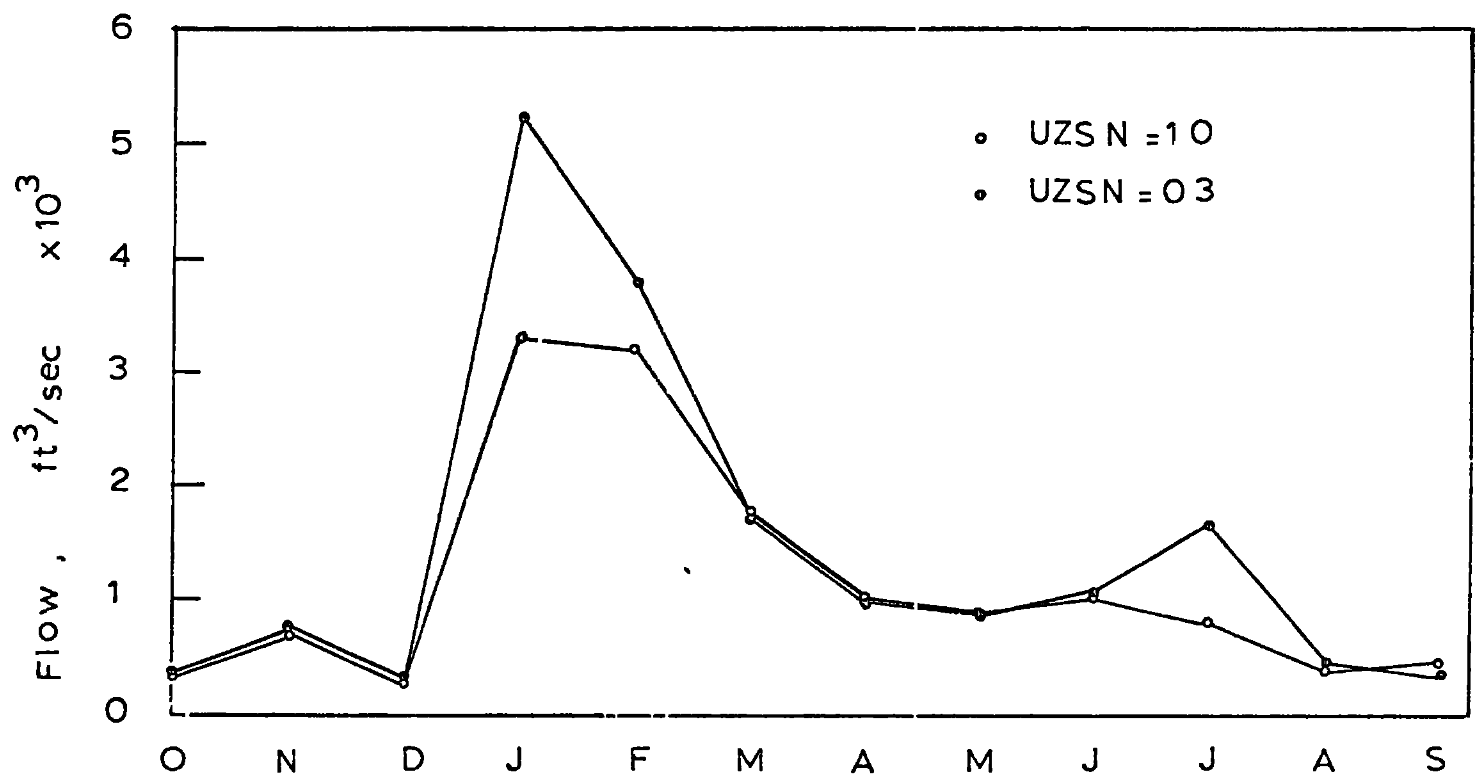


Fig 77. Browney River at Burn Hall Simulated monthly flow for two values of UZSN

sensitivity testing. These results are presented in Table E (Appendix V) The immediate conclusion from studying the yearly and monthly values in this table was that the interflow index does not affect the simulation of flows significantly

This conclusion, though valid under the above circumstances, (Table E (Appendix V), high infiltration value of 0 6), was rejected later when an optimized CB value of 0 1 was used In two subsequent trials with CC values of 0.5 and 3 0, it was found that peak values of runoff were increased as a result of lowering CC values from 3 0 to 0 5 (Fig 78) The increase in the peak flows, thus obtained, is clearly reflected in the monthly and yearly flows, (Table F (Appendix V)).

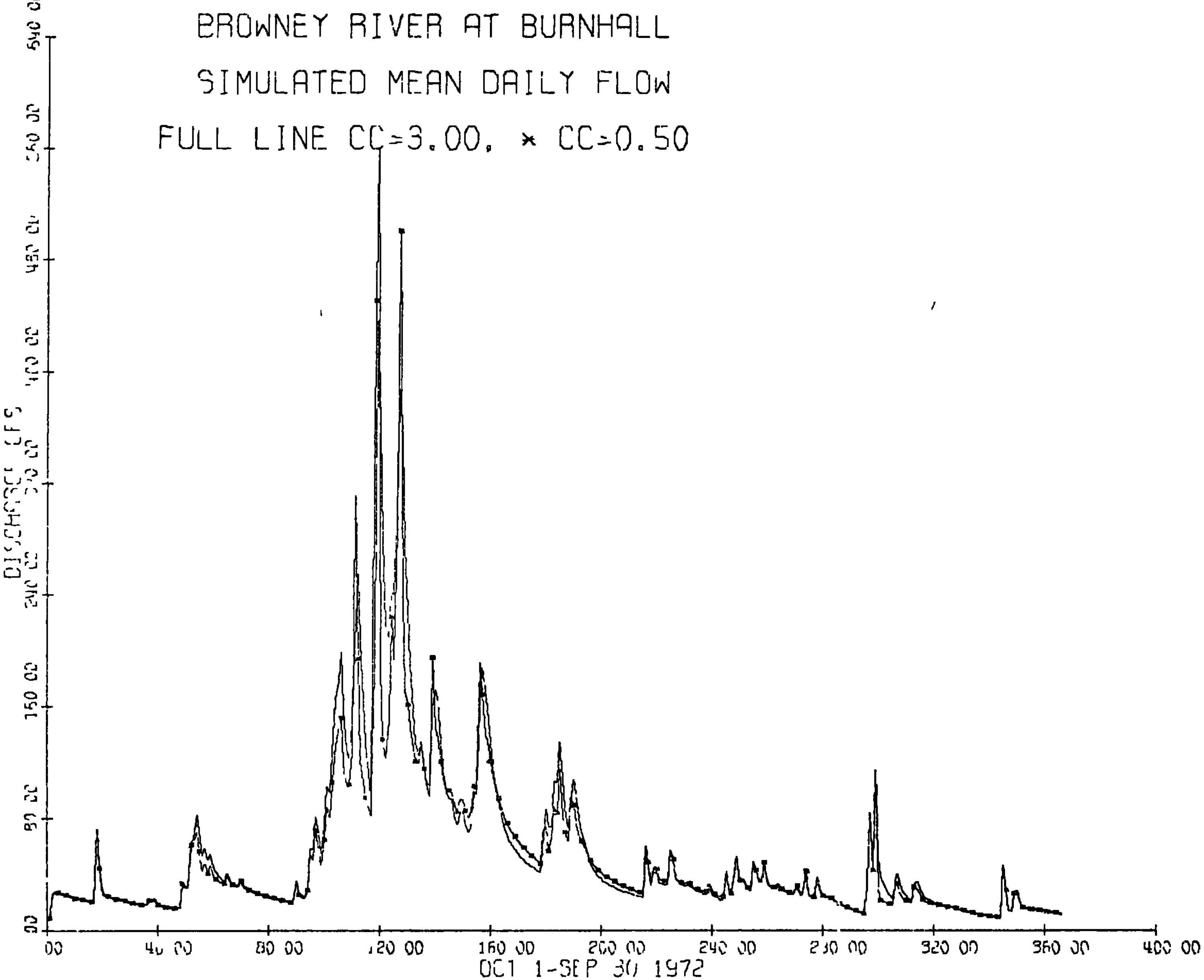
The results obtained from these two sets of values with CC can be explained by considering the role of CC (assigning the level of interflow relative to overland flow) and the infiltration index CB

For the first two runs since the value of CB was very high, almost all the precipitation had infiltrated without significant surface runoff occurring. Therefore, under this condition, the CC parameter was not of any significance in altering the level of interflow relative to overland flow.

The results obtained from the latter observation, however, reflected the significance of the CC value, because a CB value, representative of catchment infiltration, had resulted in proper distribution of direct runoff and baseflow. Under this condition, therefore, the lower CC value of 0.5 had increased overland flow relative to interflow, and so had increased the hydrograph peak.

<u>K3</u> (Actual evapotranspiration loss index). This parameter has been one of the least sensitive in this study Increasing its value from 0.20 to 0 25 (25 per cent increase) did not result in any change of flow (Table G (Appendix V)) Fig 79 shows the hydrographs of mean daily flow for two runs with K3 values of 0 20 and 0 25. Though the perfect

Fig.78.


BROWNEY RIVER AT BURNHALL
SIMULATED MEAN DAILY FLOW
FULL LINE CC=3.00, * CC=0.50
DISCHARGE CFS
OCT 1-SEP 30 1972
00
40 00
80 00
120 00
160 00
200 00
240 00
280 00
320 00
360 00
400 00
80 00
160 00
240 00
320 00
400 00
480 00
560 00
640 00

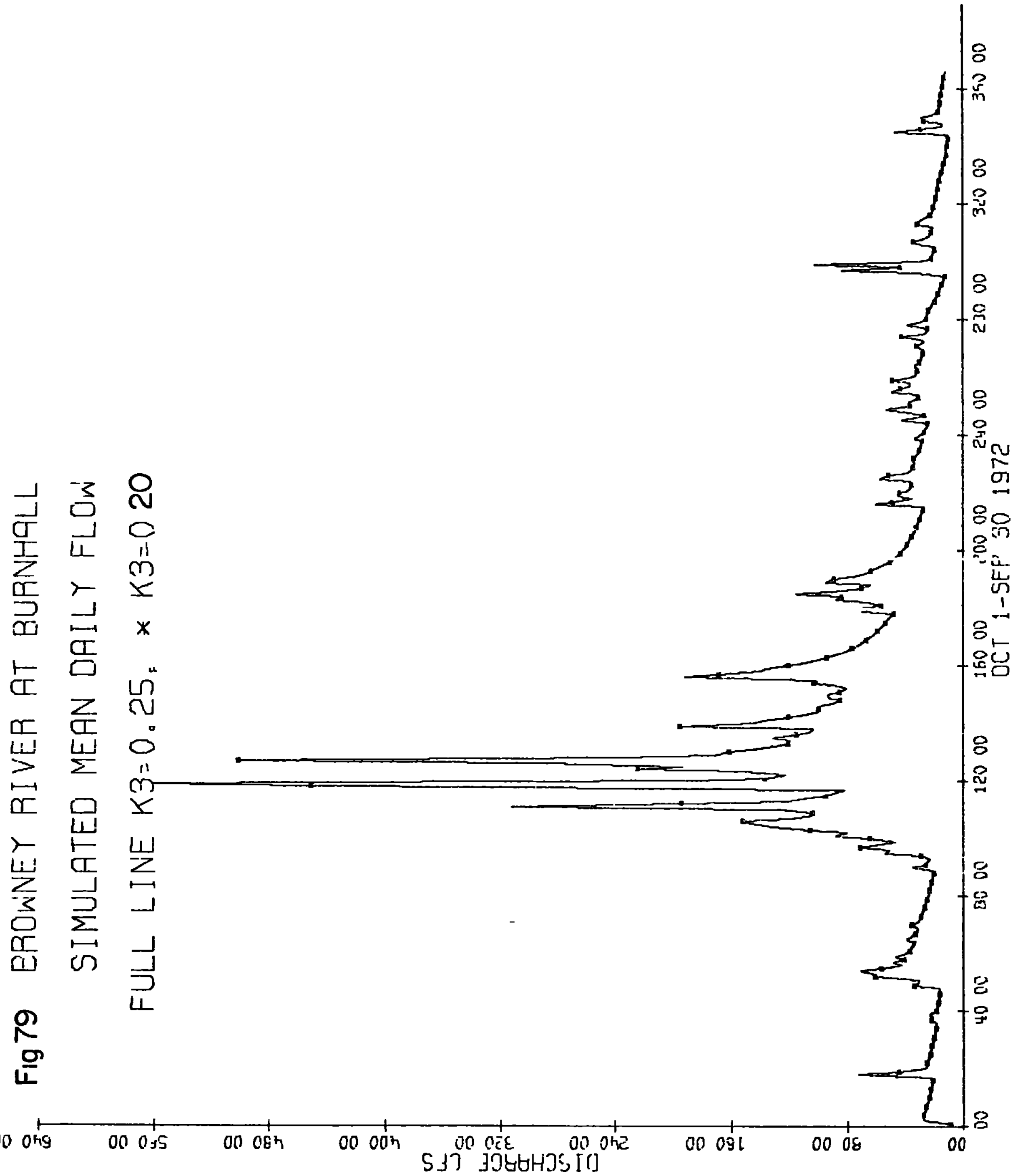


Fig 79 BROWNEY RIVER AT BURNHALL

SIMULATED MEAN DAILY FLOW

FULL LINE K3=0.25, * K3=0.20

coincidence of the two hydrographs proves the insensitivity of K3 under the hydrological conditions prevailing in the Browney basin, it was thought worthwhile to test the sensitivity of K3 for another water year For this purpose the dry year of 1973 was used. The results of runs with this year also verified the insensitivity of K3 (Table H (Appendix V))

From Table H (Appendix V) it is noticed that increasing K3 by 25 per cent has reduced the yearly flow by 1 3 per cent

K24L (Deep percolation index) In order to get an estimate of the sensitivity of this parameter, the model was run using two different values of K24L, i e. 0 25 and 0 00 The hydrographs of daily flows and monthly flows for these two runs are presented in Figs 80 and 81. The monthly and yearly values are also tabulated in Table I (Appendix V)).

The total yearly percolation of water to the inactive groundwater zone i.e leakage, was about 17 per cent Since deep percolation is a fixed portion of the inflow to groundwater, and since inflow to groundwater is a function of the infiltration index, the water loss for any specific K24L value therefore increases with increasing CB values

K24EL. (Groundwater evapotranspiration parameter) Decreasing the value of K24EL increases the total runoff. Under conditions of no moisture deficiency, actual Et would equal the potential value and, therefore, this parameter will not affect the simulated flow The effect of this parameter for a dry year however, is very much more pronounced

For the sensitivity test of this parameter, the dry year of 1973 was used The monthly simulated flows for this year for two runs with K24L values of 0 1 and 0 2 are presented in Fig 82 and Table J (Appendix V) The yearly flow has decreased by 7 per cent as a result of raising the K24EL value from 0 1 to 0 2. From the plot of monthly flows, it is observed, that the yearly increase of 7 per cent occurred during the summer months, when low moisture availability prevented evapotranspiration

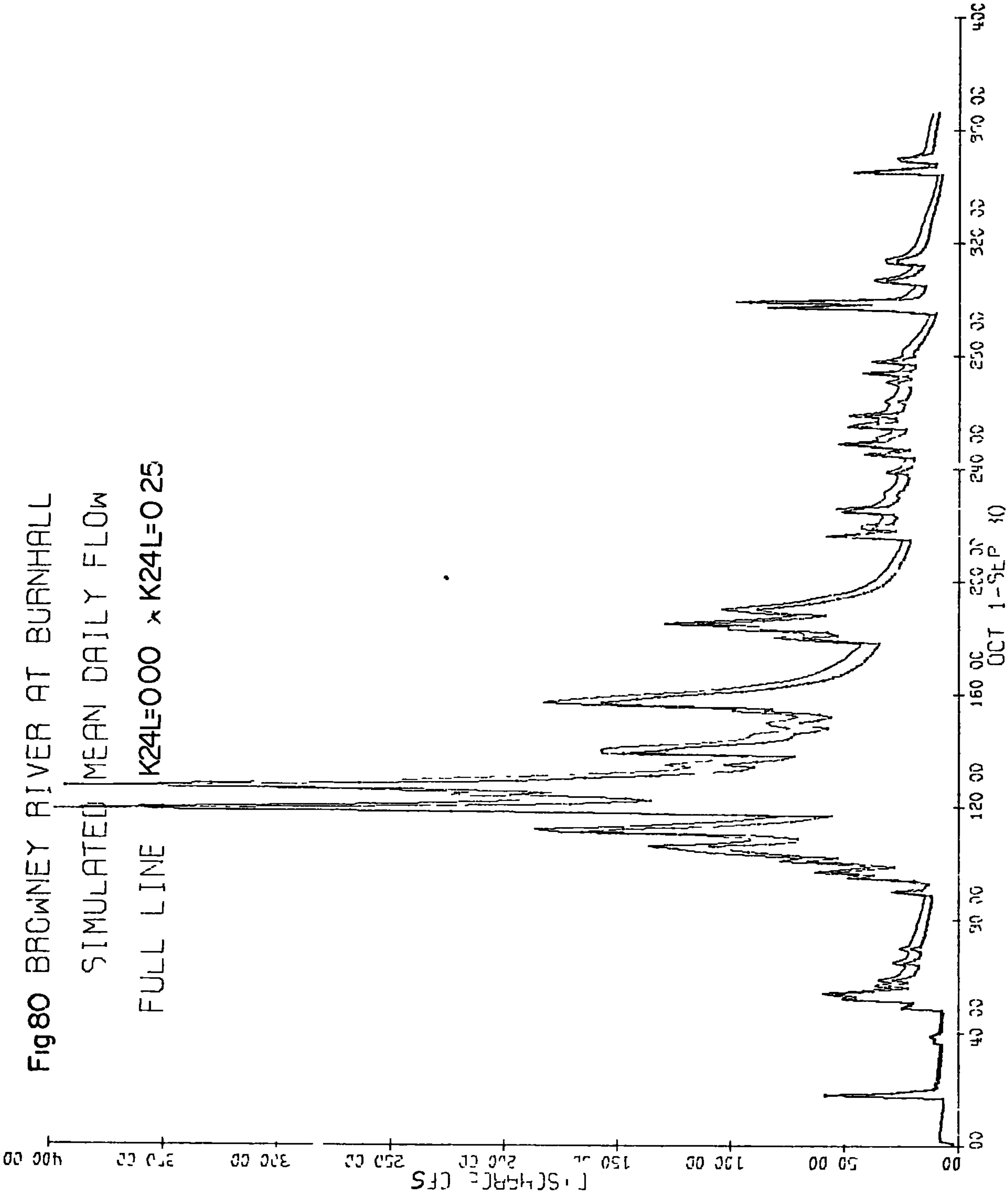


Fig80 BROWNEY RIVER AT BURNHALL

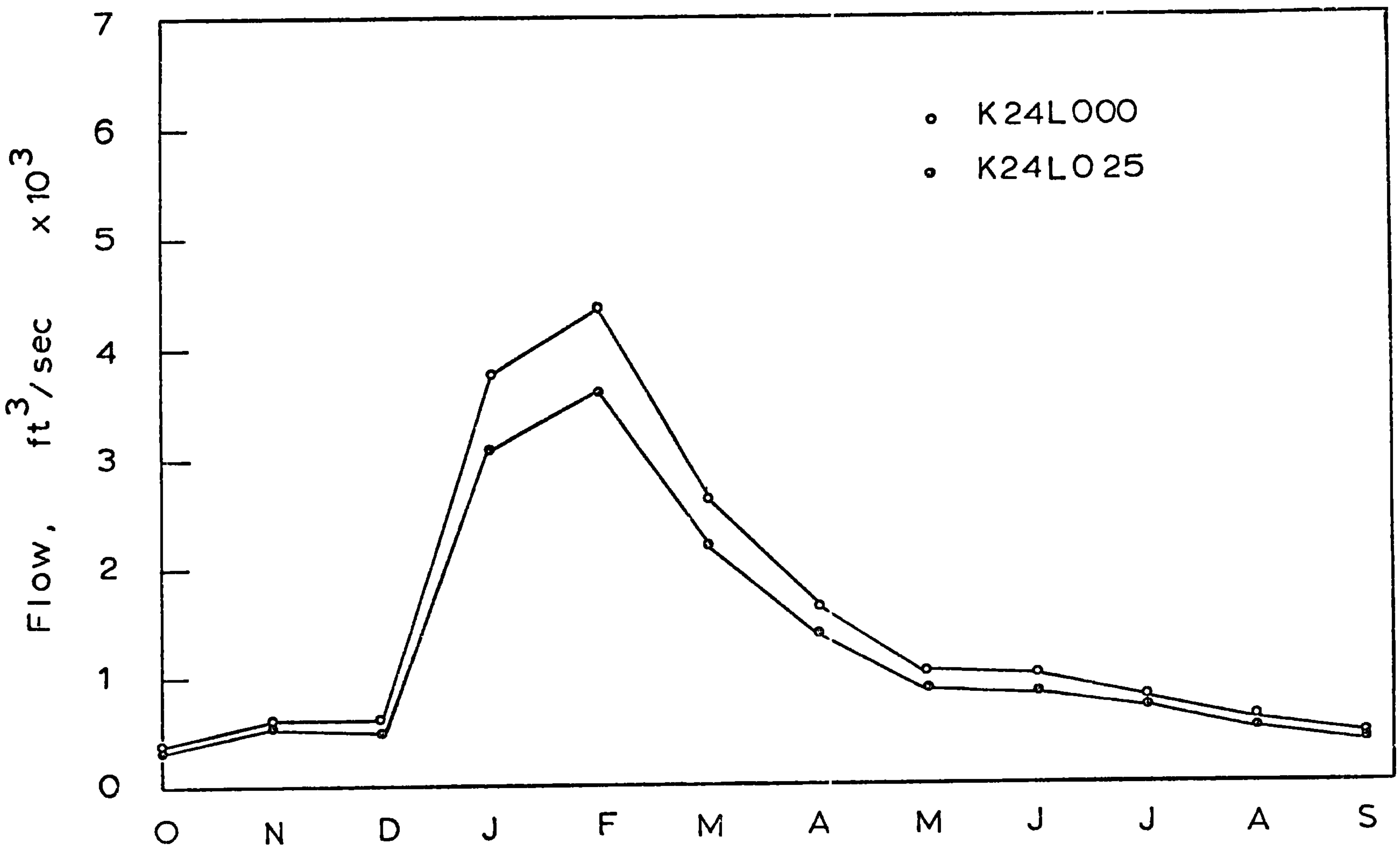


Fig 81. Browney River at Burn Hall Simulated monthly flow for two values of K24L

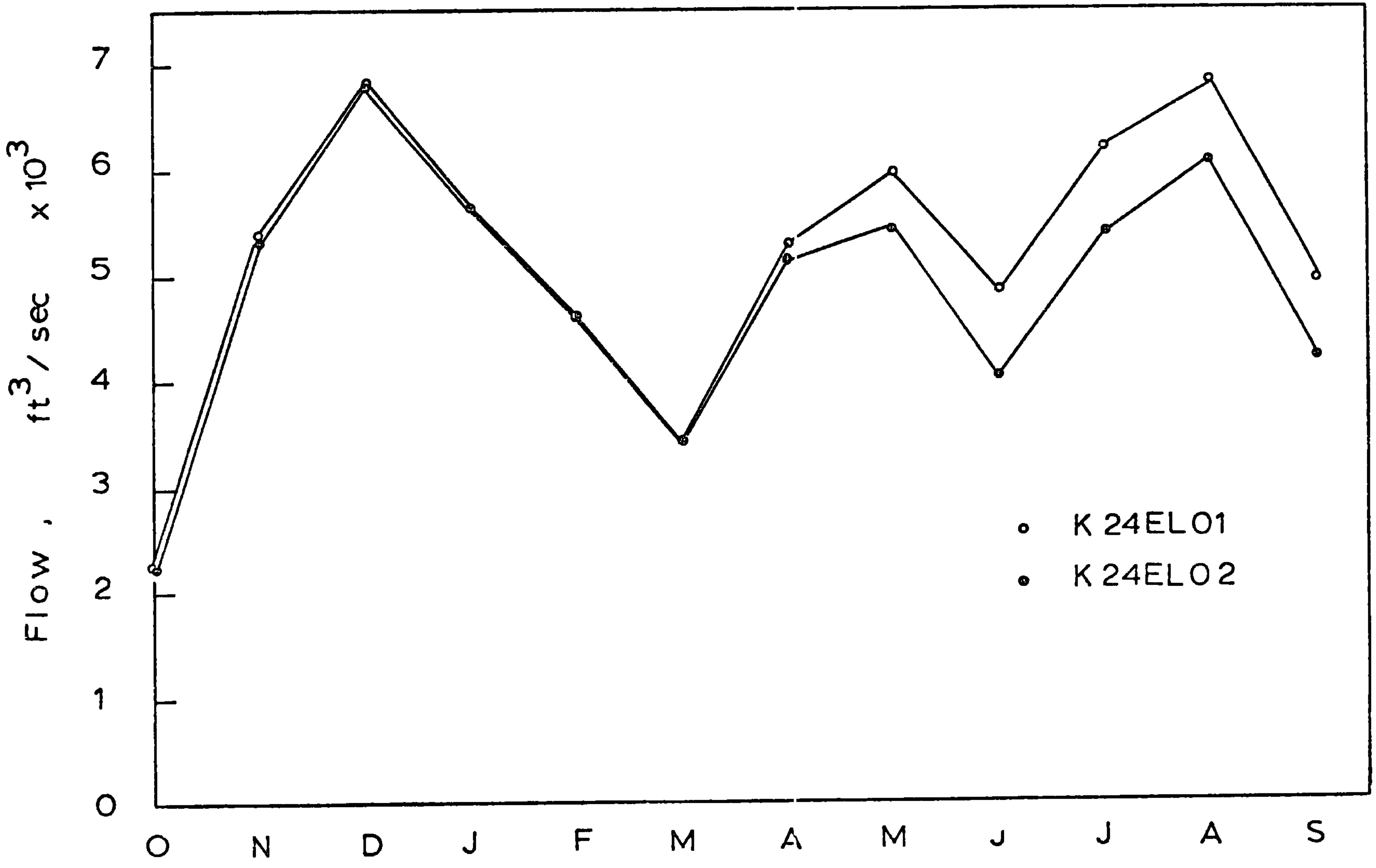


Fig 82 Browney River at Burn Hall Simulated monthly flows for two values of K24EL

from taking place at the potential rate

UZS (Initial upper zone storage) The effect of this parameter is very pronounced during the first few months of the run. Fig 83 shows the hydrographs of two runs with UZS values of 1.00 inches (25 4 mm) and 0 00. The monthly values are also presented in Table K (Appendix V). The simulated flow during October and November due to a UZS value of 1 00 inches (25 4 mm) are higher than those due to a UZS value of 0.00 by some 118 per cent and 26 per cent respectively The initial high UZS value of 1 0 inches (25.4 mm) has also increased the simulated flow during the months following November However, this effect has been slight.

The explanation for this observation is that when the storage is high, it can not prevent any overland flow due to precipitation, and moreover some moisture from upper zone storage serves as delayed infiltration, thus increasing the groundwater flow The combined effects of these two factors, therefore, result in high runoff volumes, until the storage is depleted to its normal capacity.

The per cent increase in the total yearly flow due to increasing the UZS value has been about 7 per cent.

LZS (Initial lower zone storage). The initial lower zone storage is also very important in affecting the runoff volume The effect of this parameter, like that of UZS, is well pronounced at the start of the run Fig.84 shows the plot of two simulated hydrographs due to LZS values of 5.00 inches (127 mm) and 8 00 inches (103.2 mm). The monthly values are also presented in Table L (Appendix V)

The change in the yearly value of runoff due to the increase of the LZS value is about 22 per cent. The table of monthly values shows that the runoff volume of almost every month has been increased

IRC (Interflow recession rate). To investigate the sensitivity of IRC, the parameter whose value was determined initially by Barnes' method of

Fig.83.
BROWNEY RIVER AT BURNHALL
SIMULATED MEAN DAILY FLOW
FULL LINE UZS=0.00,*UZS=1.00

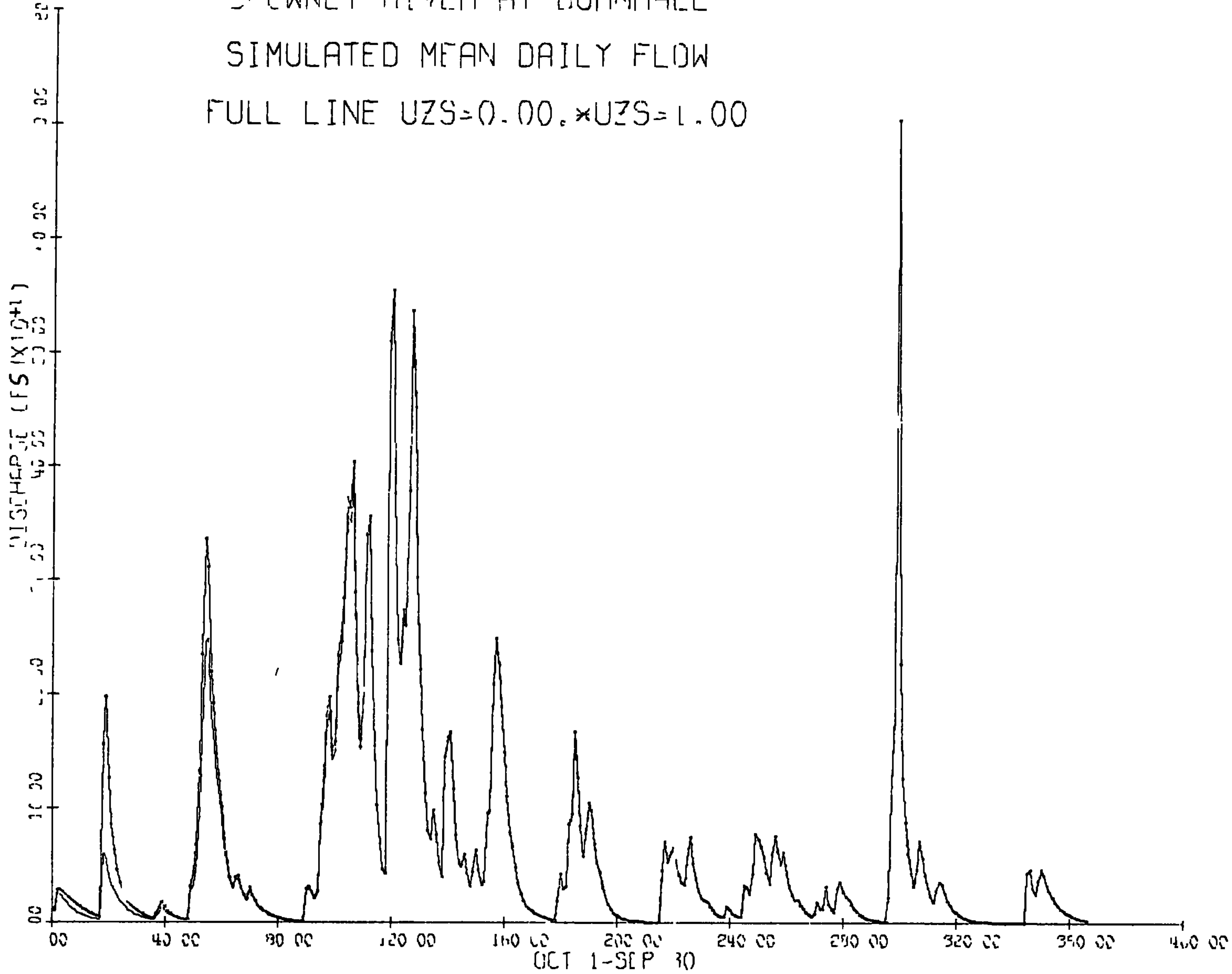
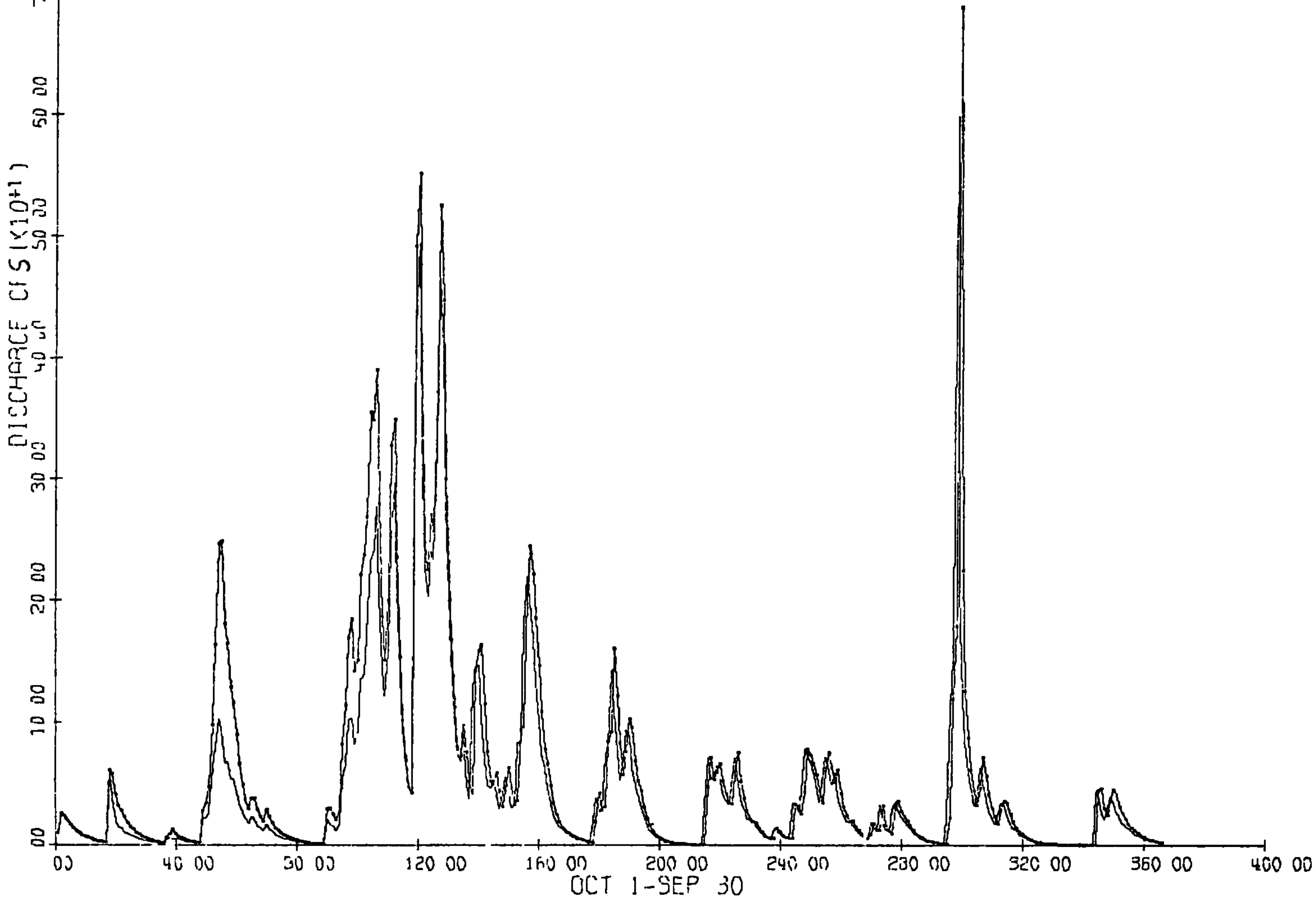


Fig.84.

BROWNEY RIVER AT BURNHALL

SIMULATED MEAN DAILY FLOW

FULL LINE LZS=5.00, ×LZS=8.00



hydrograph analysis, the model was run twice using IRC values of 0.70 and 0 50 The monthly values of these two runs are presented in Table M (Appendix V), and the corresponding daily hydrographs are shown in Fig 85 From these results it is observed that the yearly totals have not been affected by the change in IRC value i e the yearly totals for both runs are 9.8 inches. The variation of monthly flows and daily flows due to this change in IRC values has also been slight.

KK24 (Groundwater recession rate) The value of this parameter like IRC, was determined by Barnes' method of hydrograph analysis However, since minor trial and error adjustment was thought to be necessary, a sensitivity test was carried out by running the model with KK24 values of 0.99 and 0 90 respectively Fig 86 shows how a 10 per cent decrease of this parameter has increased the peak values at the expense of the dry flows The yearly total was also affected by this variation of KK24 i.e the 10 per cent decrease in KK24 resulted in an approximately 10 per cent increase in the value of the yearly flow. The trend of monthly flow variation due to KK24, Table N (Appendix V), is explained by the fact that when the KK24 value is lower than a certain amount, the slope of its recession curve would be higher, thus any inflow to the ground-water storage will be depleted at a faster rate than that with the higher KK24 value. This explanation, therefore, accounts for lower flows (with a KK24 value of 0.9 as compared to 0 99) during the summer months excluding August. During this month, a precipitation depth of 6.4 inches (163 mm), contributed to groundwater storage which was depleted subsequently

KV (Groundwater recession variable output) To study the effect of this parameter on the simulated flows, the model was run twice, using KV values of 1.00 and 0.00 The hydrographs of daily flows obtained are presented in Fig.87 and their monthly values are given in Table O (Appendix V) From a study of these results it is noticed that a KV value of 1 00, has increased the monthly flows during the winter months, the period when the

Fig.85.

BROWNEY RIVER AT BURNHALL

SIMULATED MEAN DAILY FIOW

FULL LINE IRC=0.70, ×IRC=0.50

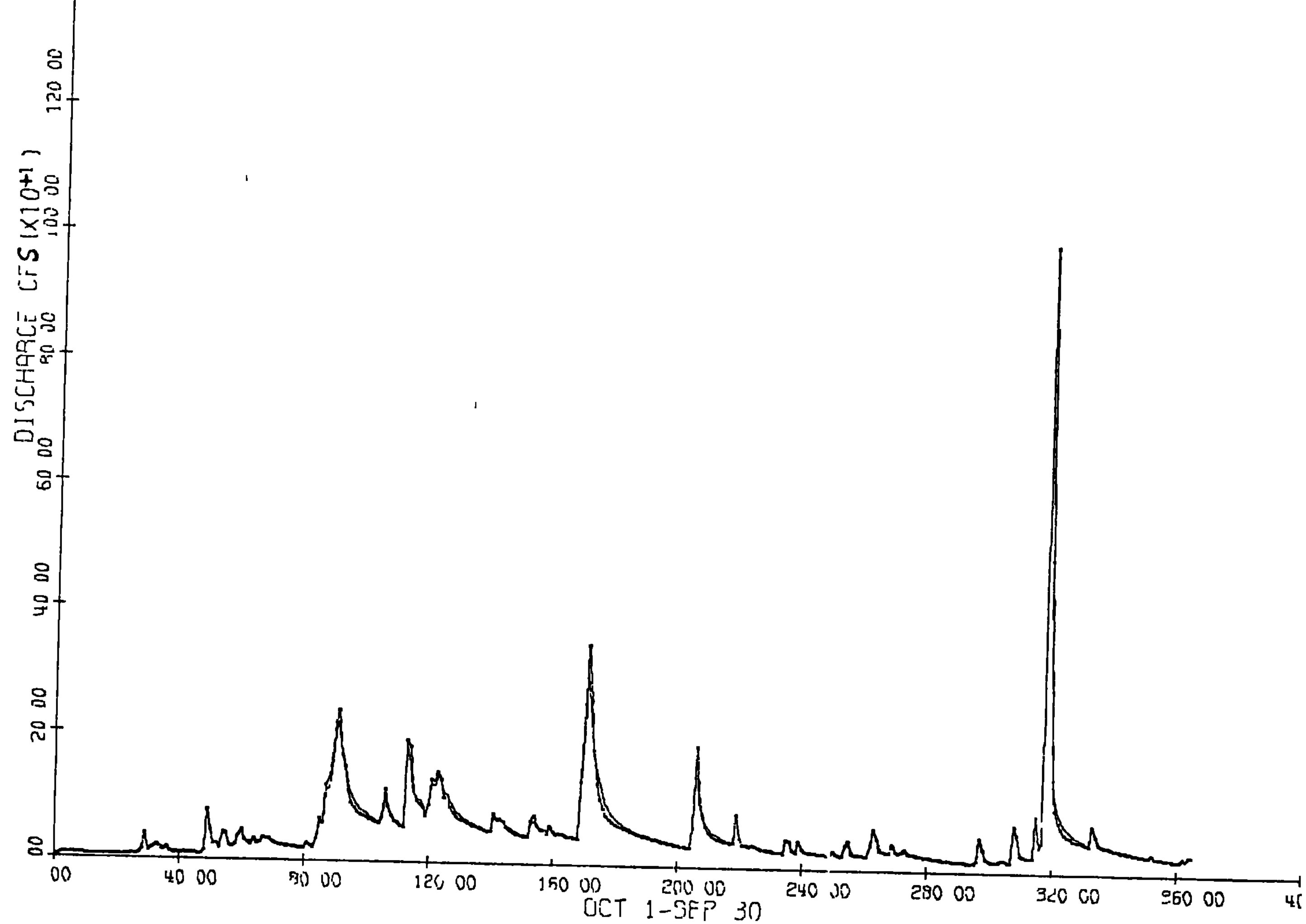


Fig.86.

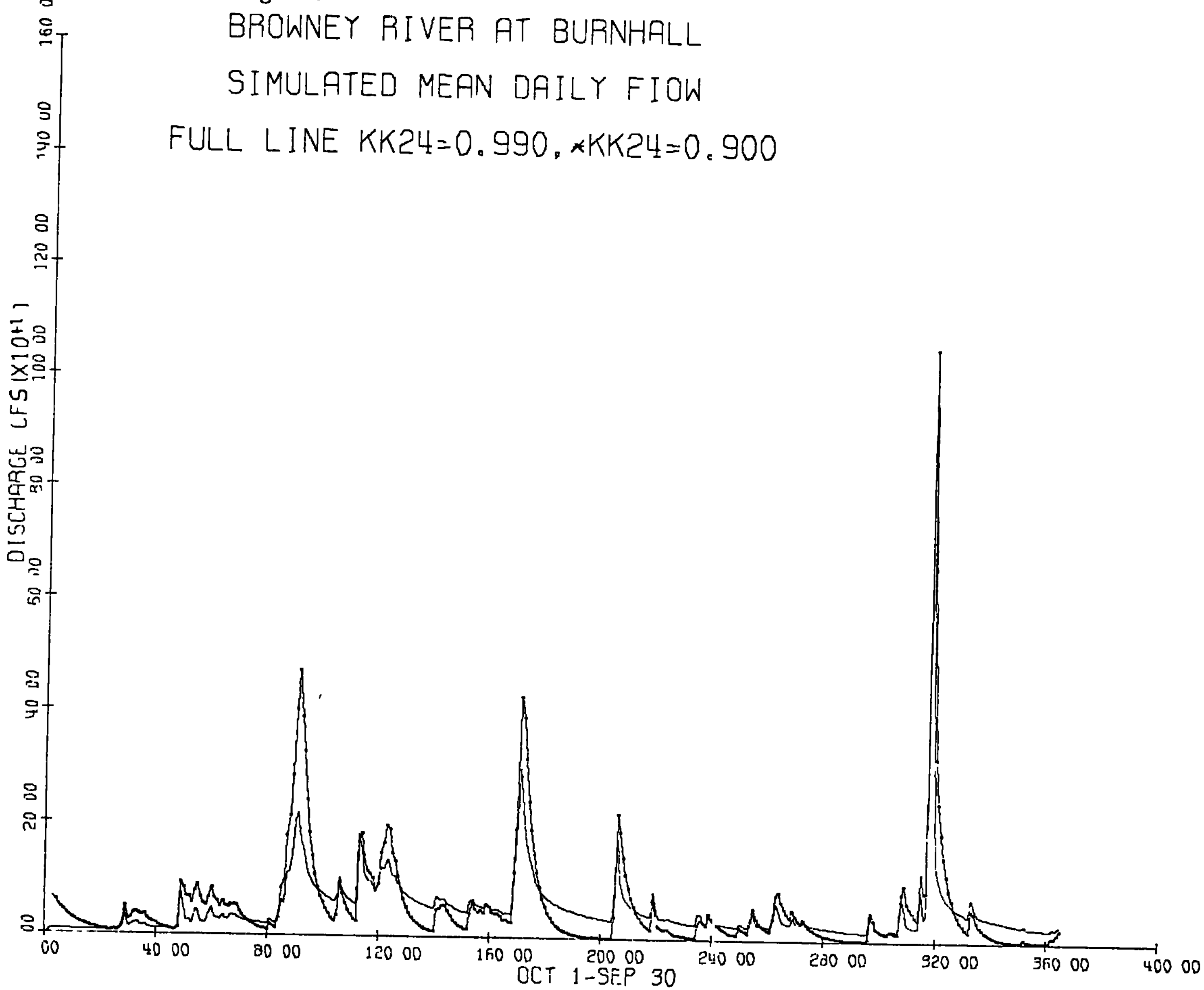
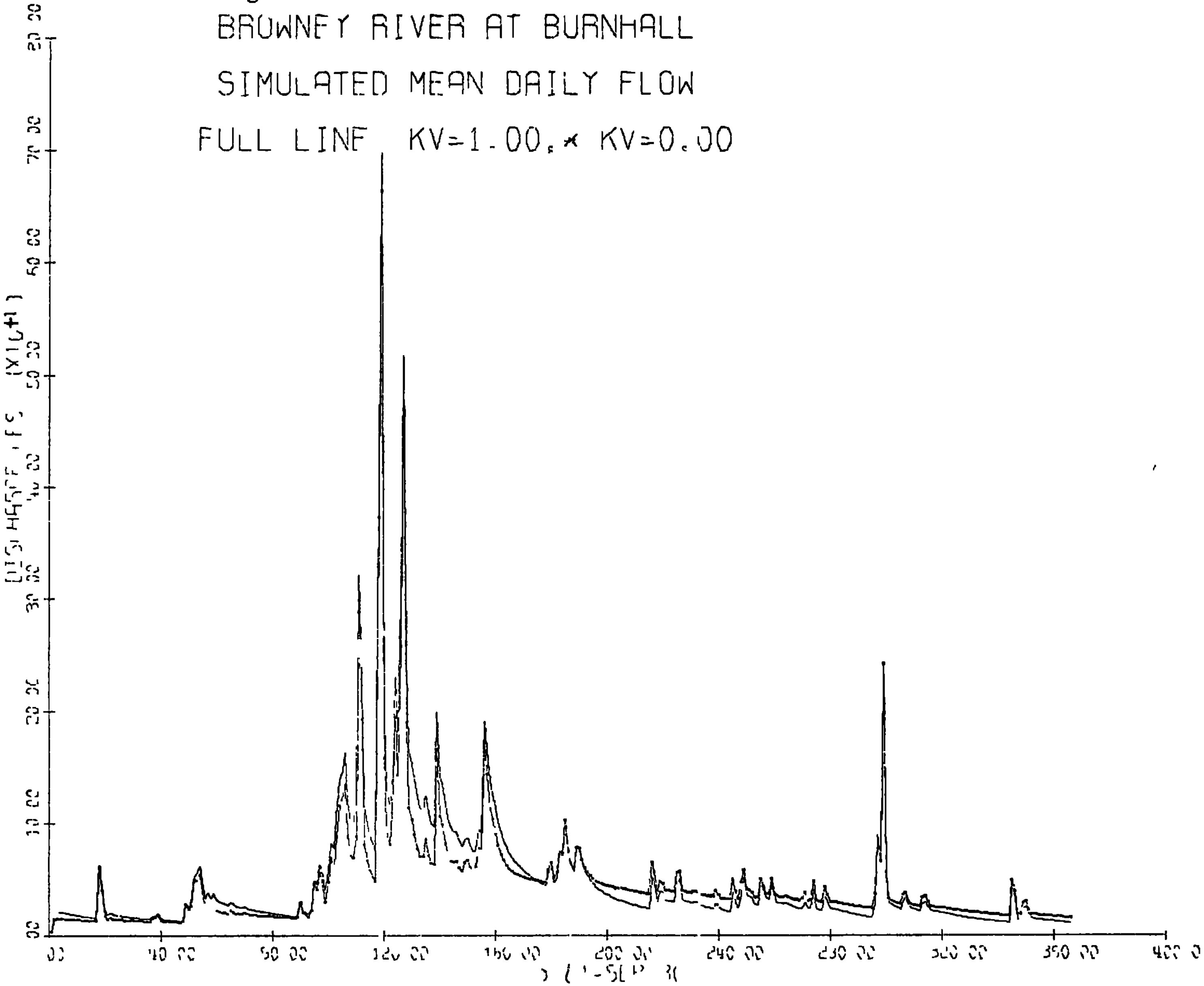

BROWNEY RIVER AT BURNHALL
SIMULATED MEAN DAILY FIOW
FULL LINE KK24=0.990, *KK24=0.900
DISCHARGE CFS(X10+1)
00
20 00
40 00
60 00
80 00
100 00
120 00
140 00
160 00
00
40 00
80 00
120 00
160 00
200 00
240 00
280 00
320 00
360 00
400 00
OCT 1-SEP 30

Fig.87.
BROWNEY RIVER AT BURNHALL
SIMULATED MEAN DAILY FLOW
FULL LINE KV=1.00, * KV=0.00



volume of moisture in groundwater storage is high, whereas the flow during the summer months has been decreased

The increase in the yearly value, using a KV value of 1 0, has been about 5 per cent

The explanation for this effect of KV can be given by referring to its inclusion in the groundwater outflow equation

$$GWF = LKK4 \times (1\ 0 + KV \times GWS) \times SGW$$

In this equation, KV behaves in such a way that when inflow to groundwater storage is taking place, the recession rate is lower, whereas during dry periods, the recession rate is steeper.

Optimization results

Numerous trials were run while information was being collected concerning the working of the model and the sensitivity of the parameters Almost all the parameters were varied, one at a time, during these trial runs The parameters, however, which were not tested were those controlling the monthly and raingauge potential evapotranspiration. The reason for not varying these was that not enough measured evapotranspiration data were collected from the evapotranspirometric studies which were being carried out concurrently with this study Lack of accurate potential evapotranspiration values, therefore, did not allow precise estimation of Et parameters. It was thought that any trial and error variation of these optional parameters would add to the complexity of the calibration of the model and thus would mask the effects of the other parameters controlling the runoff volumes. Penman Et values (albedo 0.25), were, therefore, thought to be the most representative for the catchment during the initial runs.

The initial objective for the calibration of the model was set to be the total yearly and seasonal distribution of the flow In fact using the guidelines provided by Crawford and Linsley (1966) it was possible to produce a simulated yearly value which closely matched the recorded one i.e 10.09 inches (256 3 mm) simulated and 10 02 inches (254 5 mm) recorded. However, when the seasonal values of simulated and recorded flows were considered, it was noticed that the simulated flows during the summer months were higher than the recorded flows while during the months preceding the summer season they were lower Referring to the values of the main parameters in the run of the model i.e LZSN, CB and UZSN which were 7.5 inches (190.5 mm) 0 6 and 0.6 inches (17 4 mm) respectively and considering the results of the sensitivity tests, outlined earlier, it was concluded that the CB value was rather high. It was the high CB value which had resulted in too much baseflow at the expense of low direct runoff during February, March and April Another reason for the high runoff values during the summer months, which was not recognised then, was probably the result of low evapotranspiration values (Penman Et) used, which in fact were later found to be too low for the summer period

A third source of error or rather uncertainty in the simulation was in the estimation of the initial soil moisture conditions in the upper zone, lower zone and groundwater storage As was shown in the results of the sensitivity tests, these parameters affect the initial conditions significantly High values of these parameters could increase the flows during the first few months and thus affect the yearly values. Indeed, difficulties in the selection of these parameters explain partly why the calibration of the model should preferably be based on two or more years of data.

The parameters estimated for the initial runs of the water year 1972, were thus changed in order to more closely match the yearly total as well as the seasonal distribution of simulated and recorded flows When one set of optimized parameters was obtained, it was used for other years. In one case the application of the calibrated parameters to other years produced results which under-estimated the yearly flows by some 30 to 35 per cent for two years, while it over-estimated another year by just 2 per cent.

A study of these results and the parameters used, showed that the main runoff parameters i.e LZSN, CB, UZSN were quite close to their optimized values. However, since the initial moisture conditions used in the simulation were different from those at the end of the preceding water year, some error had occurred. This source of error, therefore, was corrected by using LZS, UZS, SGW and GWS values for each year derived from those at the end of the preceding year Thus the results turned out to be more reasonable.

Application of the Penman EO_2 values, obtained from correlation with measured Et better matched the yearly and seasonal values. The effect of the evapotranspiration input data on the simulated flow will be presented in the following chapter.

After the calibration of the volumetric parameters, attempts were made to fit the shape of the hydrograph of simulated flow to that of the recorded flow. For this reason, the recession parameter KK24 was changed slightly.

The results of the optimization of the parameters are given in Table 50 These results were used to simulate the flow of the years 1969, 1970, 1971 and 1973. Together with the year of 1972, therefore, five years of flow have been simulated and the results will be discussed in the following chapter

Table 50 The values of parameters used for the simulation of the flow in the Browney catchment by Stanford Watershed Model IV, starting 1st January,1969

Parameter		
K_1	1.17	Calculated
IMPV	0.05	Measured
EPXM	0.10 inches (2 5 mm)	Estimated
UZSN	0.45 inches (11 4 mm)	Optimized
LZSN	6.50 inches (165 1 mm)	Optimized
CB	0.10	Optimized
CC	0.60	Optimized
K3	0.23	Estimated
K24L	0 00	Estimated
K24EL	0.05	Estimated
L	936 ft (285 m)	Calculated
SS	0.06	Calculated
NN	0.30	Estimated
IRC	0.70	Calculated
KK24	0.990	Optimized
KV	1 00	Optimized
UZS	2.20 inches (55.9 mm)	Optimized
LZS	13.10 inches (332 7 mm)	Optimized
SGW	1 70 inches (43.2 mm)	Optimized
GWS	1.70 inches (43.2 mm)	Optimized
RES	0.00	Estimated
SRGX	0 01	Estimated
SCEP	0.10	Estimated
AEPI	0 10	Estimated

CHAPTER TEN

SIMULATION RESULTS AND DISCUSSION

Simulation results of water year 1972

The water year 1972 was used for optimization of the fitted parameters. The criteria of optimization, were the yearly total flow, monthly total flow and the hydrograph shape The hydrographs of the monthly flow and the mean daily flow for this year are presented in Figs. 88 and 89 The yearly total, mean monthly values, standard deviation of the monthly values and the correlation coefficient between simulated and recorded monthly values are shown in Table 51 The difference in the yearly totals of the actual and simulated flows is less than 3 per cent of the actual flow The correlation coefficient of 0 99 for monthly flow shows that the monthly values are highly correlated. The correspondence between the monthly totals is shown in Fig.88

Table 51 Yearly total flows, mean monthly flow, standard deviation of the monthly flow and the correlation coefficient between monthly recorded and simulated flows, water year 1972

	Yearly total ft^3/sec (m^3/sec)	Mean monthly ft^3/sec (m^3/sec)	Standard Deviation	Monthly correlation coefficient
Recorded flow	18788 0 (533 06)	1567 7 (44.5)	1619.2	0.99
Simulated flow	18229.0 (517 2)	1518 8 (43 1)	1344.6	

Fig 89 shows how closely the hydrographs of actual and simulated mean daily flows match each other There are, as expected, some peaks in the hydrograph of the recorded flows which are not reproduced by the model. As an example, the highest peak of actual flow which occurred in early February, is not simulated by the model. In fact this specific peak was not simulated in any of the numerous trials of the model Only

Fig 86.

SIMULATED AND RECORDED MONTHLY FLOW AT BURNHALL, WATER YEAR 1972

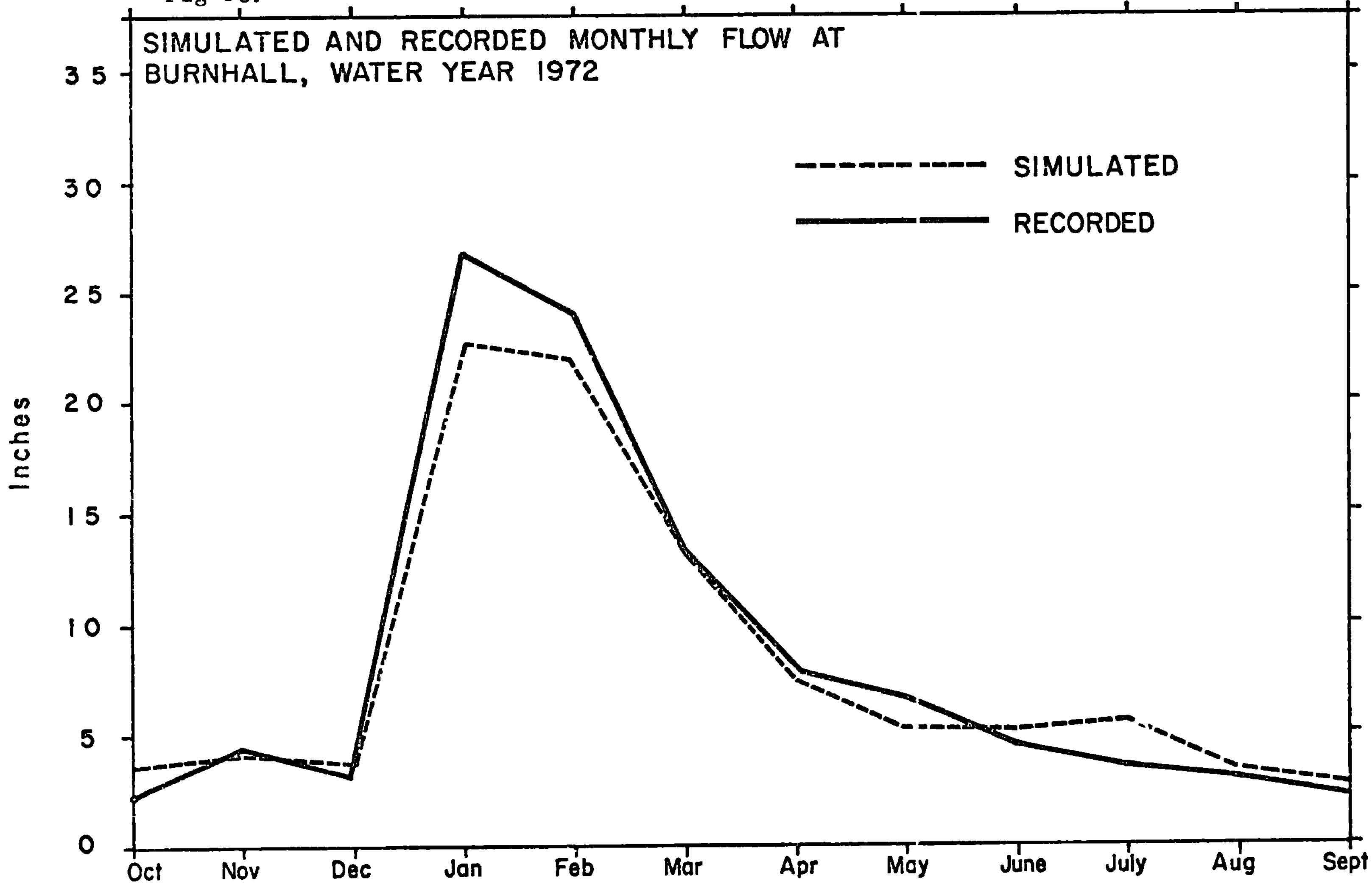
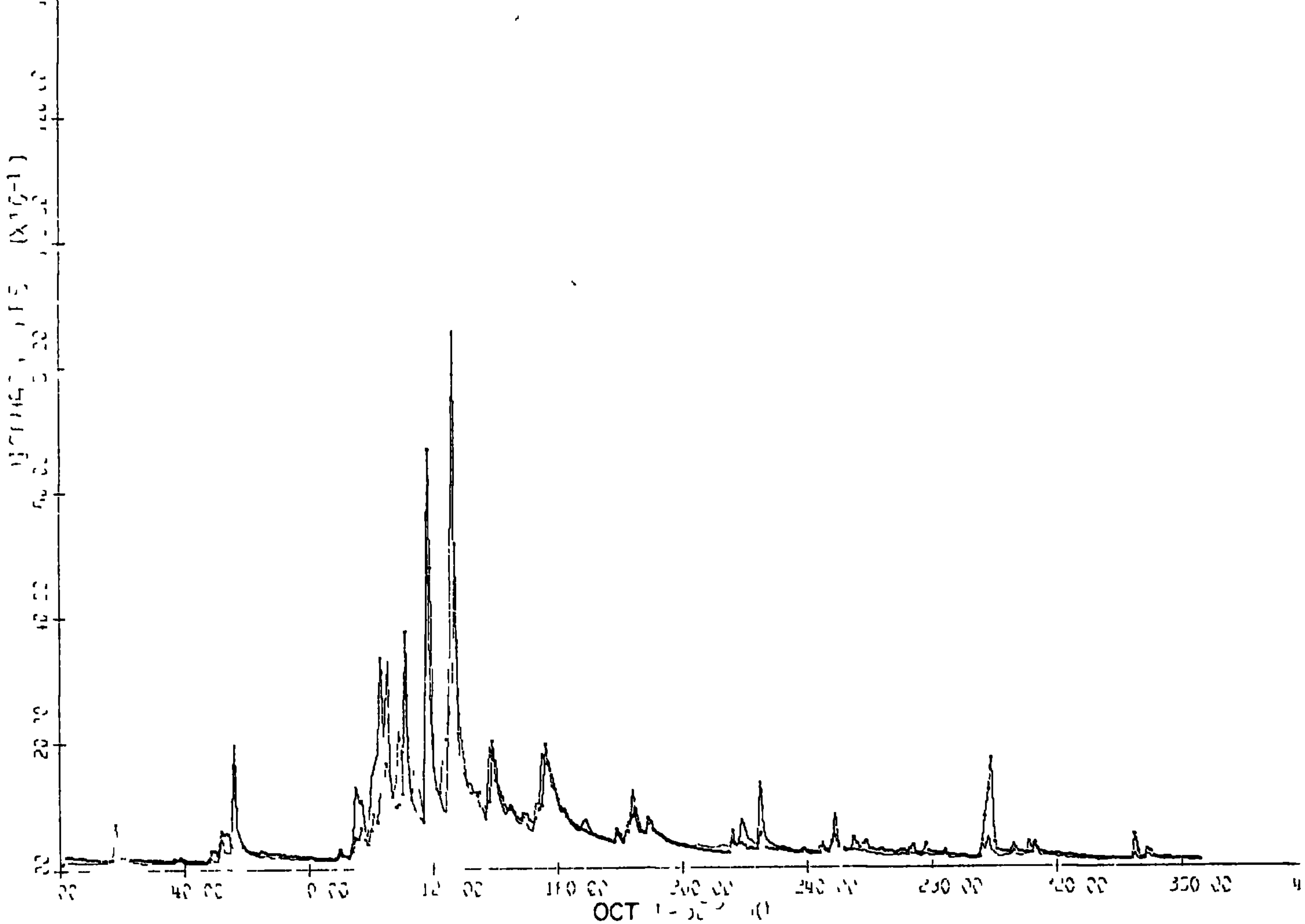


Fig.89.
SHOWNEY RIVER AT PURNHALL
MEAN DAILY FLOW, WATER YEAR 1972
FULL LINE RECORDED, × SIMULATED



a CB (infiltration index) of 0 01 which increased some other peaks by 300 per cent, produced a mean daily flow approximately matching the actual flow Under this very low CB value, the baseflow was, however, too low, Fig 72.

There are on the other hand some flows in the simulated daily hydrograph which over-estimate those in the actual hydrograph As an example, the minor peak in late July could be cited The two main possible sources of errors causing such discrepancies are

1. Inaccurate input data
2. Empirical representation of some of the processes involved

Of the input data, the hourly rainfall is the most significant in the simulation results. For the Browney basin the recording Casella gauge is located close to the lower part of the basin The value of the catch of rain by this gauge times K_1 (parameter expressing the ratio of mean rainfall over the catchment to that at Durham Observatory), might not have been representative of the actual rain depth over the catchment at all times. Owing to the spatial variations of rainfall over the catchment, K_1 varies during each rain Malfunctioning of the pen of the gauge could also result in erroneous rainfall data River gauging stations, no matter how accurately they measure the streamflow, could also result in up to a ten to fifteen per cent error

The second source of errors is due to temporal and spatial variations of the individual hydrologic processes occurring within the catchment As an example, the estimation of the infiltration process within a catchment by an average value is far from being accurate This is because infiltration is a function of soil surface cover, water content of the soil, hydraulic conductivity of the soil, soil texture, soil structure and temperature. Each of these factors mentioned depends on other parameters which vary from location to location and from time to time From this discussion, it follows that there are some elements of empiricism in the mathematical representations of these hydrologic processes Of all the processes in the Stanford Watershed Model IV, the channel system is the most empirical This is because of the adoption of a single time-delay-histogram for all conditions of flow, without any consideration of the dimensions of the channel system This weakness of the Stanford Watershed Model had been recognised by its authors and in H S.P. (Hydrocomp Simulation Programming), which is the successor to the Stanford Watershed Model, the "Kinematic Wave" routing has been substituted for the Clarks' channel-time-delay histogram (Black,1973). Kinematic wave theory refers to the movement of a flood wave in the downstream section of a river channel The kinematic movement of the wave is governed by the continuity equation $\frac{Q}{x} + B\frac{y}{t} = o$, which reflects the storage mechanism (Fleming and Fahmy,1973) In this formula

Q - is discharge at the 'cross section'

x - is the distance in the direction of flow

y - is the vertical depth of flow and

t - is the time

In this approach the actual channel dimensions and roughness coefficients are used and continuous stage and discharge throughout the system are calculated

<u>Results of application of the model to other years</u>

Table 52 shows the five year total flow, mean monthly flow, standard deviation of mean monthly flow and the correlation coefficient between the recorded and simulated monthly flows for the period January 1969 to September 1973.

For the individual years, similar tables are presented in the following paragraphs (Tables53 to 56). The graphs of monthly variations are shown in Figs 90 to 93

Considering the five year results, the recorded mean monthly value exceeds the simulated value by some 2 per cent (Table 52).

Table 52 Five year total flow, mean monthly flow, standard deviation of the monthly flow and the correlation coefficient between monthly recorded and simulated flows, January 1969 to September 1973

	5 year total ft^3/sec (m^3/sec)	Mean monthly ft^3/sec (m^3/sec)	Standard Deviation	Monthly Correlation
Recorded flow	97078 85 (2757 0)	1668 05 (47.37)	1451 86	0.94
Simulated flow	93054 21 (2642 7)	1632 53 (46.36)	1537.83	

Table 53 Yearly total flow, mean monthly flow, standard deviation of the monthly flow and the correlation coefficient between monthly recorded and simulated flows, January to September 1969.

	Yearly total ft^3/sec (m^3/sec)	Mean monthly ft^3/sec (m^3/sec)	Standard deviation	Correlation Coefficient
Recorded flow	26509 5 (752 9)	2945 5 (83 6)	1833 1	0 90
Simulated flow	26000 1 (738 4)	2888.9 (82 0)	1997.7	

A monthly correlation coefficient of 0 94 shows that the recorded and simulated monthly flows are highly correlated For the individual years of 1969 (Jan-Sept) (Table 53), 1970 (Table 54) and 1971 (Table 55), the correlation coefficients are high and the mean monthly values or yearly totals of the recorded and simulated flows vary by only 1 9 to 8 7 per cent. The results for the water year 1973, however, show that simulated mean monthly flow under-estimates the recorded flow by some 38 per cent (Table 56). A correlation coefficient of 0.57 shows that there is no correlation between the monthly values

Due to this appreciable difference between recorded and simulated

Table 54 Yearly total flow, mean monthly flow, standard deviation of the monthly flow and the correlation coefficient between monthly recorded and simulated flows,water year 1970

	Yearly total ft^3/sec (m^3/sec)	Mean monthly ft^3/sec (m^3/sec)	Standard Deviation	Correlation coefficient
Recorded flow	22076 0 (625 5)	1835 5 (52.1)	1561.8	0.97
Simulated flow	23936 4 (679.8)	1994.7 (56 6)	1902 4	

Table 55 Yearly total flow, mean monthly flow, standard deviation of the monthly flow and the correlation coefficient between monthly recorded and simulated flows, water year 1971

	Yearly total ft^3/sec (m^3/sec)	Mean monthly ft^3/sec (m^3/sec)	Standard deviation	Correlation coefficient
Recorded flow	17593 2 (499 6)	1466.1 (41.6)	953 8	0.96
Simulated flow	18368 4 (521 7)	1530 7 (43 5)	948 4	

Table 56 Yearly total flow, mean monthly flow, standard deviation of the monthly flow and correlation coefficient between monthly recorded and simulated flows, water year 1973

	Yearly total ft^3/sec (m^3/sec)	Mean monthly ft^3/sec (m^3/sec)	Standard Deviation	Correlation coefficient
Recorded flow	10160 4 (288 6)	846 7 (24 0)	394 4	0 57
Simulated flow	6388 4 (181 4)	523 4 (14 9)	119 8	

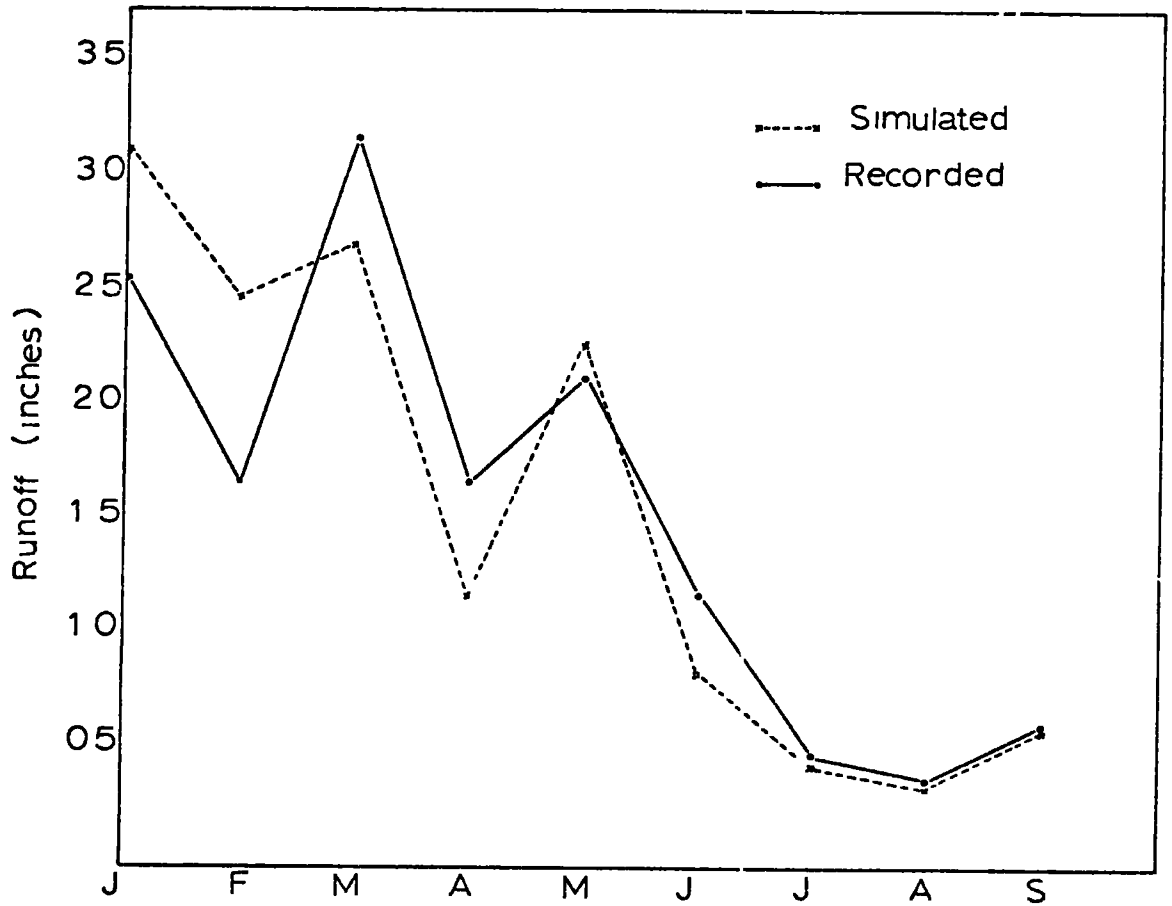


Fig 90 Browney river at Burn Hall
Simulated and recorded monthly runoff,
Jan - Sept 1969

Fig 91

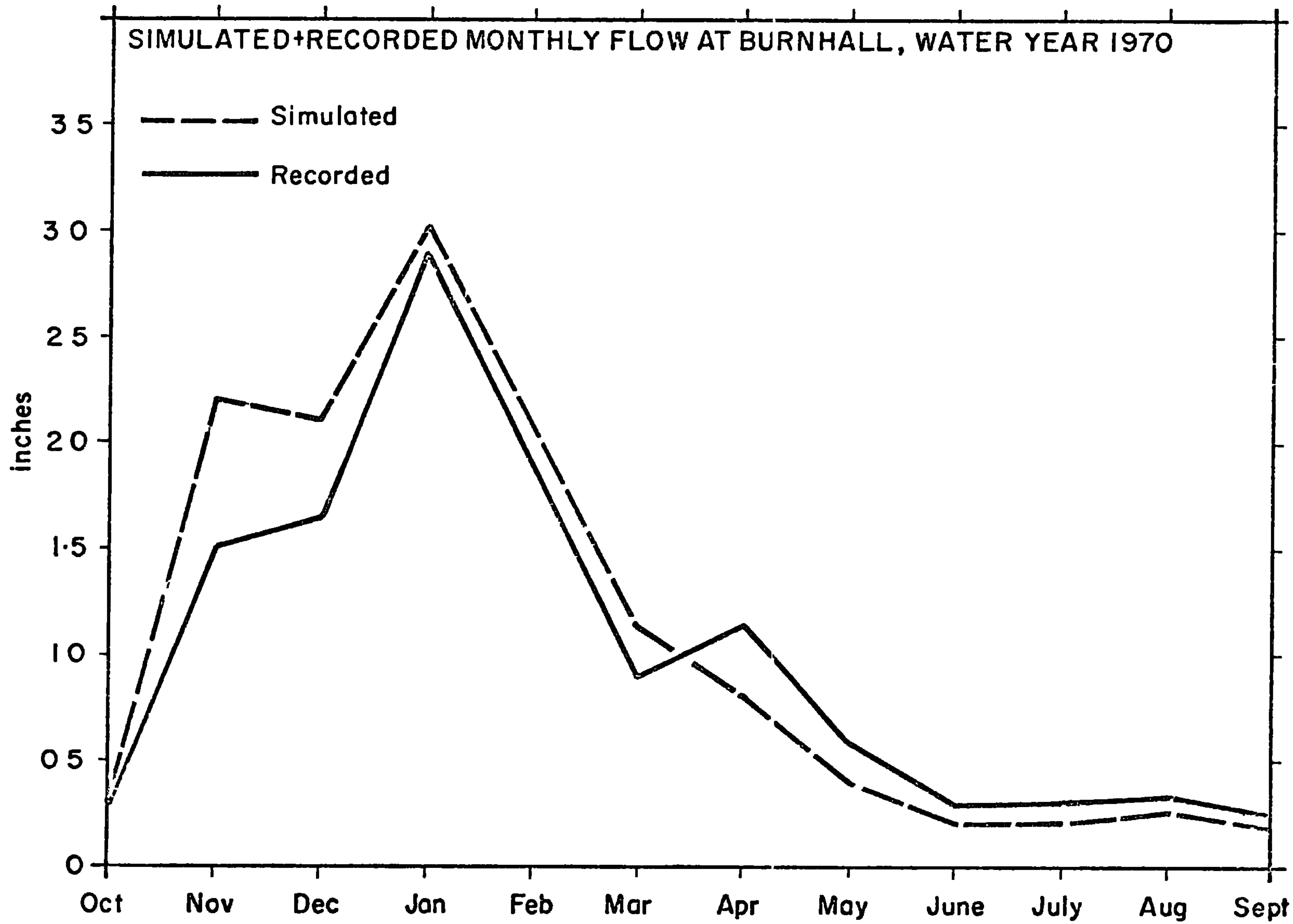
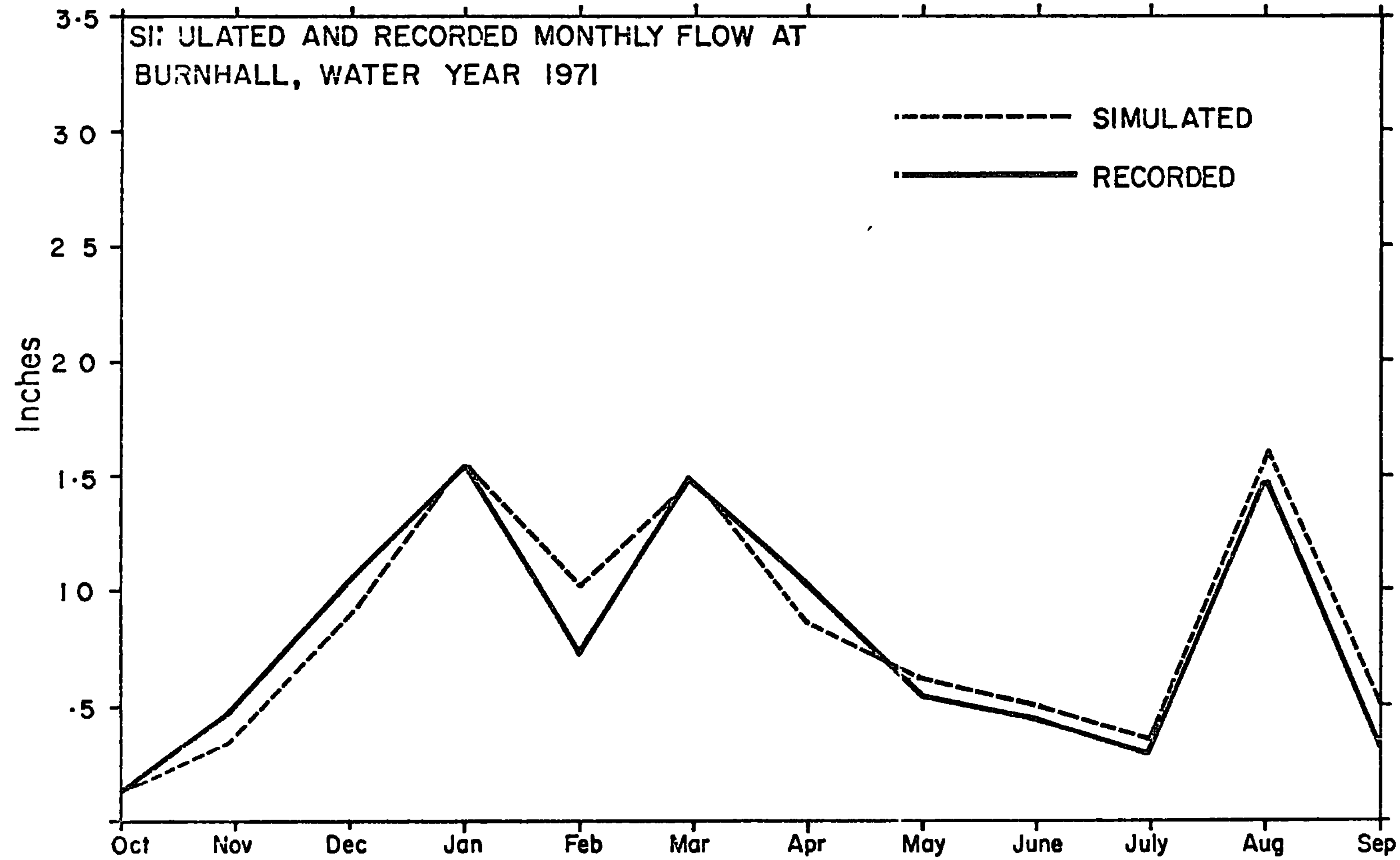

SIMULATED+RECORDED MONTHLY FLOW AT BURNHALL, WATER YEAR 1970
Simulated
Recorded
inches
3 5
3 0
2 5
2 0
1·5
1 0
0 5
0
Oct
Nov
Dec
Jan
Feb
Mar
Apr
May
June
July
Aug
Sept

Fig 92


SI: ULATED AND RECORDED MONTHLY FLOW AT
BURNHALL, WATER YEAR 1971
SIMULATED
RECORDED
Inches
3·5
3 0
2 5
2 0
1·5
1 0
·5
Oct
Nov
Dec
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Feb
Mar
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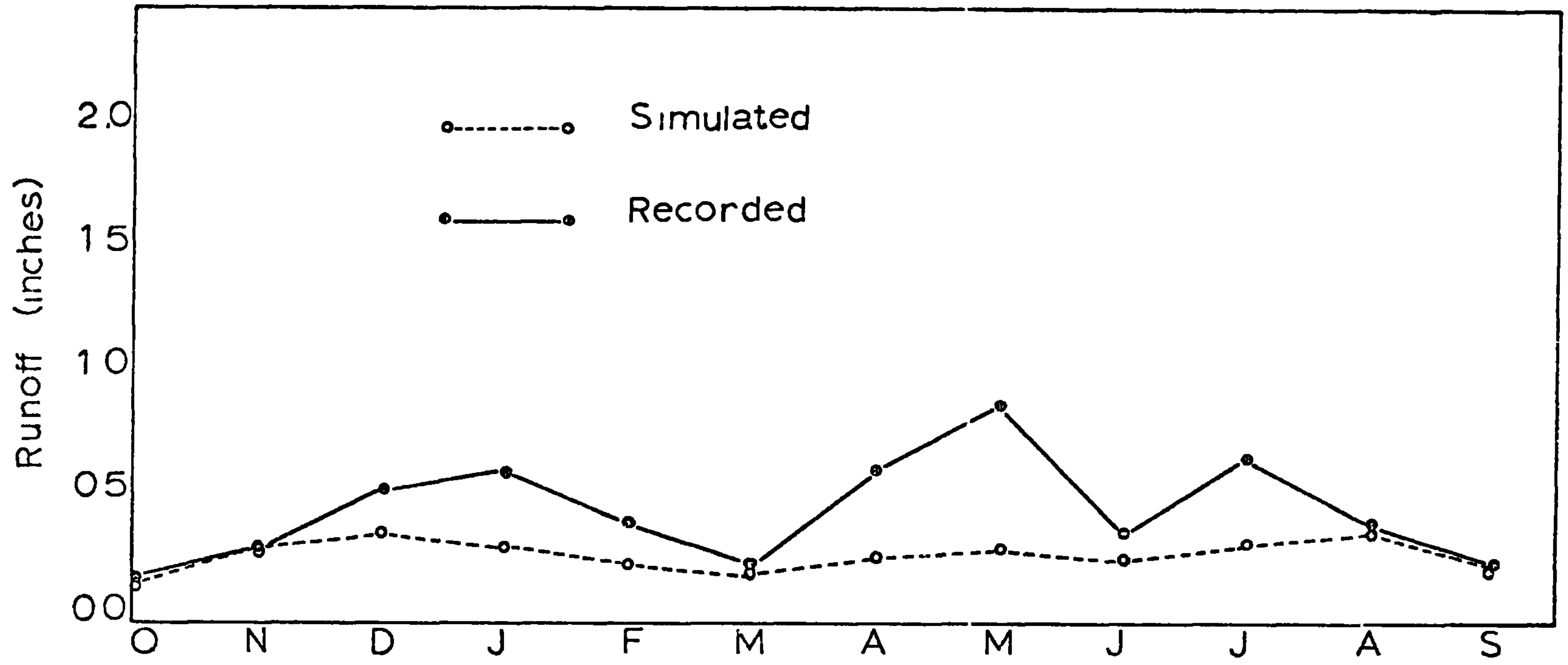


Fig 93 Browney River at Burn Hall
Simulated and recorded monthly runoff, water year 1973

flow for the water year 1973, the flow for this year was simulated using lower values of LZSN and CB It was thought that if the optimized parameters used were not representative, thus resulting in low simulated flow for 1973, then, lower values of LZSN and CB, (5 5 and 0 05 respectively) should increase the simulation results for this year This however proved not to be the case, i e. the total flow obtained was only 2 per cent higher than the one obtained with the LZSN of 6 5 and CB of 0 1 (Fig 94).

Of course decreasing the values of LZSN and CB affect the simulated flows for the other years as well This point was discussed earlier in connection with the sensitivity tests,of parameters as applied to the Browney catchment Yet another example could be given for the water year 1971, when lowering the CB value from 0 1 to 0 06 increased the yearly runoff by more than 18 per cent and the runoff during the month of August by more than 44 per cent (Table 57) August has been the month with 7 32 inches (186 mm) of rain and thus it is clear how lowering the CB value could result in such an increase in the simulated flow of this month and that of the whole water year.

Considering this effect of CB upon the simulated flow during the water year 1971, it is concluded that lowering the CB value not only has not corrected the under-estimation of flow during water year 1973, but it could also have resulted in increasing the flow for other years significantly

It should also be mentioned that if the CB value is too low, peak flows will be over-estimated and the baseflow would be under-estimated (Crawford et al, 1966), and conversely that if the CB value is high, peaks would be too low while baseflow is too high.

Since for the water year 1973, the baseflow has been properly simulated (Fig 98), therefore it can be stated that the value of CB which has been correct in simulating the flows during 1969 to 1972 is not in

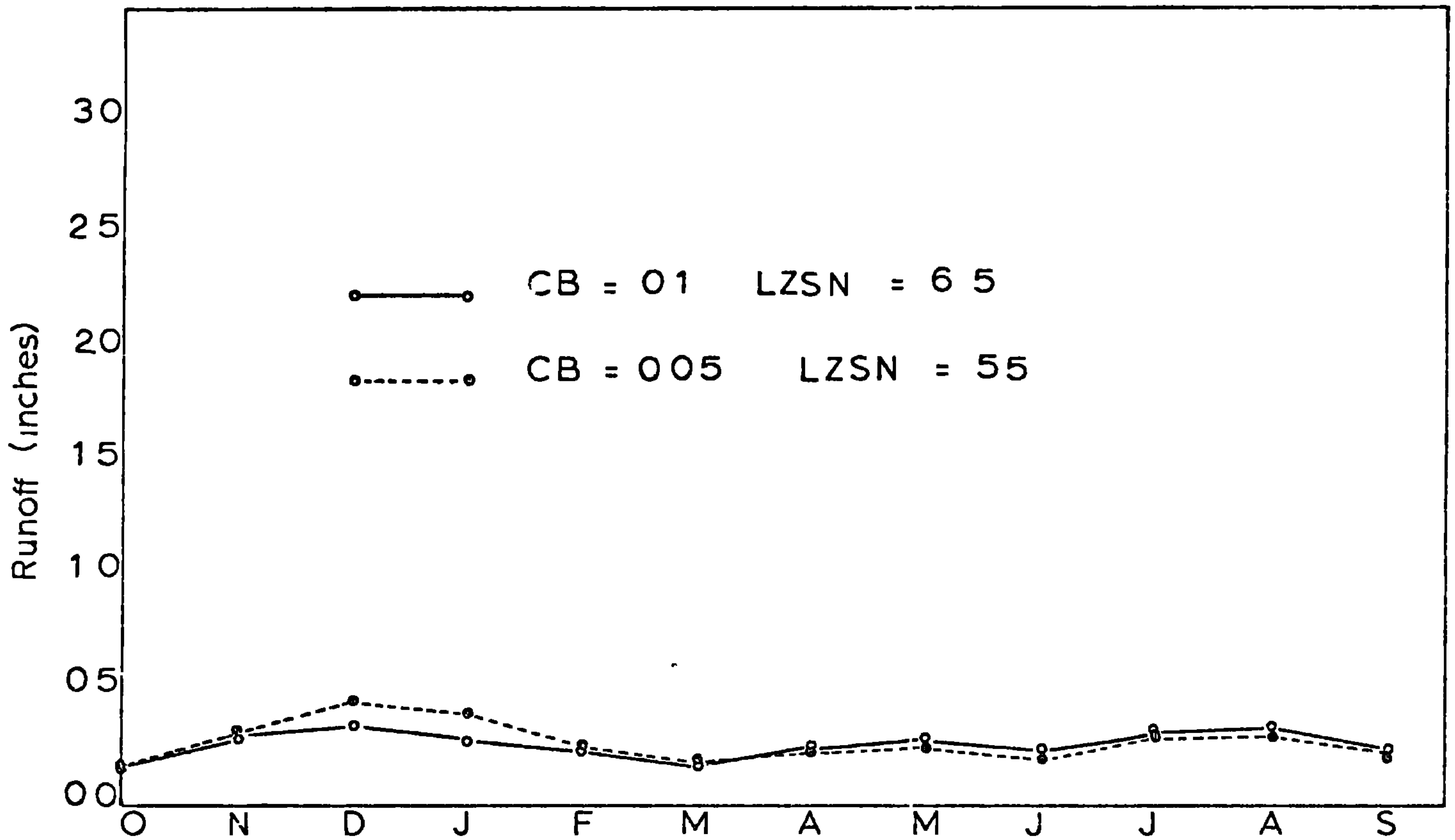


Fig 94 Browney River at Burn Hall
Simulated monthly runoff for two values of CB & LZSN, water year 1973

<u>Table 57</u> Monthly simulated flow in ft^3/sec (m^3/sec) for two C3 values during the water year 1971

CB	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL
0.10	258 (7.3)	670 (19 0)	1653 (46.9)	2928 (83.2)	1892 (53.7)	2752 (78.2)	1584 (45.0)	1120 (31.8)	967 (27.5)	649 (18 4)	3028 (86.0)	927 (26.3)	18369 (521 7)
0.06	282 (8 0)	1023 (29.0)	3252 (92 4)	3429 (97 4)	1694 (48 1)	3465 (98 4)	1580 (44.9)	917 (26.0)	722 (20.5)	450 (12 8)	4329 (122.9)	689 (19 6)	21791 (618.9)

error for 1973 as well To explain the probable reasons for under-estimation of the simulated flow during the water year 1973, the following statements can be made.

1. The formulation of the process of actual evapotranspiration might have resulted in such under-estimation of simulated flow According to the model, actual evapotranspiration occurs at the potential rate if there is moisture available in the upper zone storage (UZSN). During the water year 1973, an exceptional year with the lowest total runoff in the 17 years of study, 72 per cent of the precipitation occurred during the summer months, (Table 39) The distribution of precipitation, the high evapotranspiration demand during the summer and the modelling of the actual evapotranspiration process, therefore, might have caused the loss of too much moisture through evapotranspiration and thus produced low simulated runoff.

2. The ratio of the average segment rainfall to that of the gauge rainfall (K_1) might have been under-estimated for the storms occurring during this water year. Thus this factor could partly account for the low simulated flow during this year.

3. The amount of mine water pumped into the river during the year 1973 was about 0.54 inches (13.6 mm) Mine water, thus accounts for about 10 per cent of the total recorded flow during this year, and it, therefore, partly explains the low simulated flow during water year 1973.

For the study of mean daily flows, the duration curves of the simulated and recorded flows are presented in Fig 95 The two duration curves closely match each other except at the lower end where the simulated mean daily flows below 10 ft^3/sec (0 028 m^3/sec) are 9 45 per cent as compared with 2 2 per cent for the recorded flows.

As far as the reproduction of peak daily flows are concerned, the highest recorded mean daily flow is 1109 0 ft^3/sec (31 03 m^3/sec) in August 1971 The corresponding value of the simulated flow is

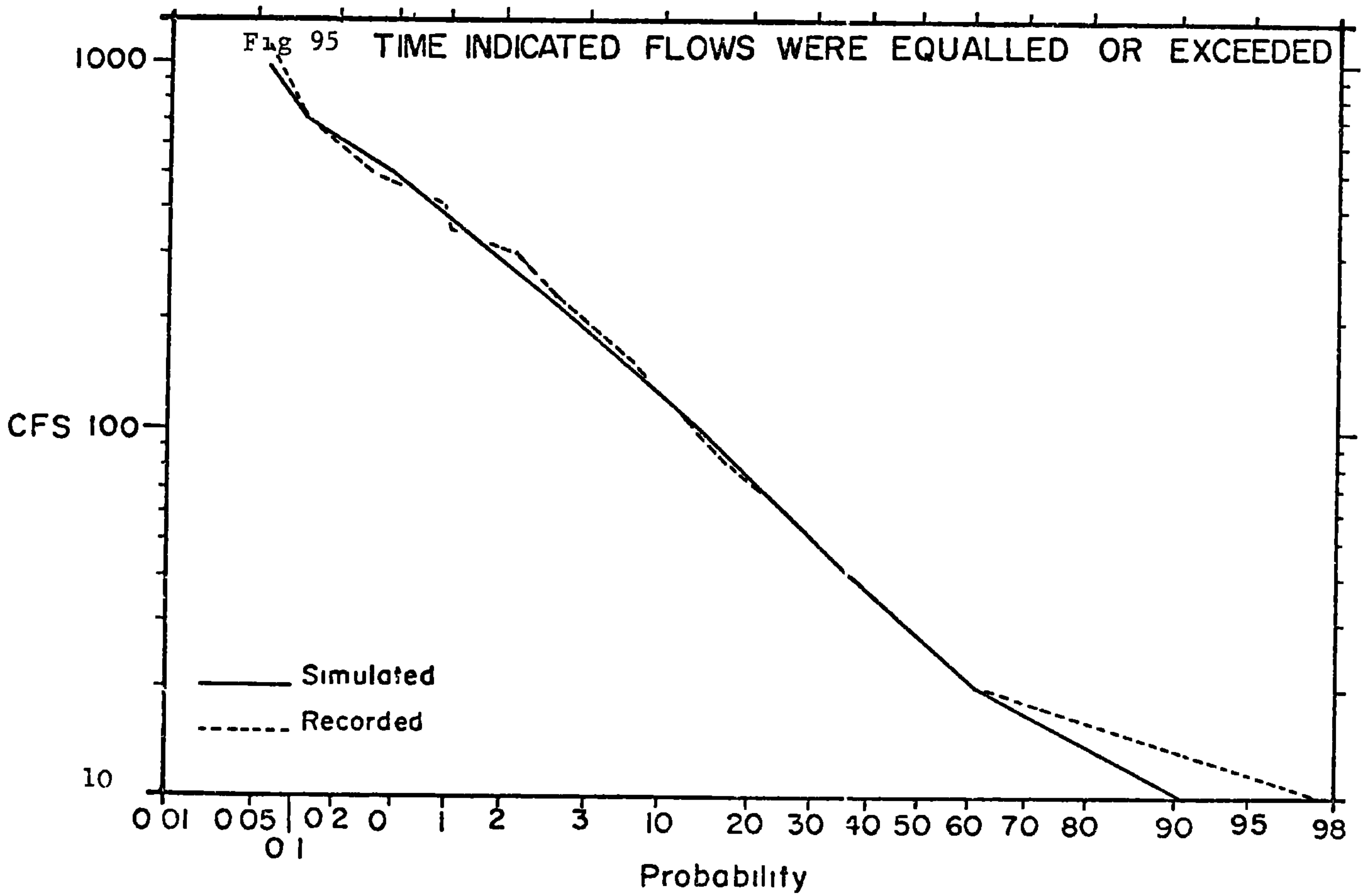


Fig 95 TIME INDICATED FLOWS WERE EQUALLED OR EXCEEDED

984 3 ft^3Sec (27 6 m^3/sec). The variation of simulated and actual mean daily flows for each year can also be observed from either of Figs. 96 to 99

Baseflow contribution of total runoff

One of the objectives of simulation of the hydrologic regime of a watershed is the estimation of baseflow. The amount of baseflow also called groundwater flow or dry weather flow and its hydrograph is an indication of the infiltration capacity of a basin and the permeabilities and storage capacities of the underlying rocks

Kunkle (1962) divided groundwater discharge into two components bank and basin storage discharge Bank storage is due to storage of runoff during high river stages This component of groundwater is not of much significance for a basin like the Browney due to the rather impermeable nature of the mainly boulder clay deposits adjacent to the river Basin storage discharge is that portion of precipitation which has infiltrated through the surface layers, and upon reaching the main groundwater zone has been delayed and subsequently released into the streams

It is therefore concluded that the distribution, amount and intensity of precipitation affects this component of runoff If the precipitation falls during a period of high evapotranspiration demand, less water would infiltrate into the soil layer, thus groundwater flow would decrease.

There are several methods commonly used for the separation and estimation of groundwater flow from total flow (Linsley and others ,1958) For example, baseflow can be estimated by drawing a line joining the point of rise to a point on the hydrograph N days after the peak, and measuring the area under this line. Alternatively, the recession curve prior to the storm can be extended under the peak and then joined to a point on the hydrograph corresponding to N days after the peak The assumption made in adopting these methods is that the time base of direct

Fig.96.

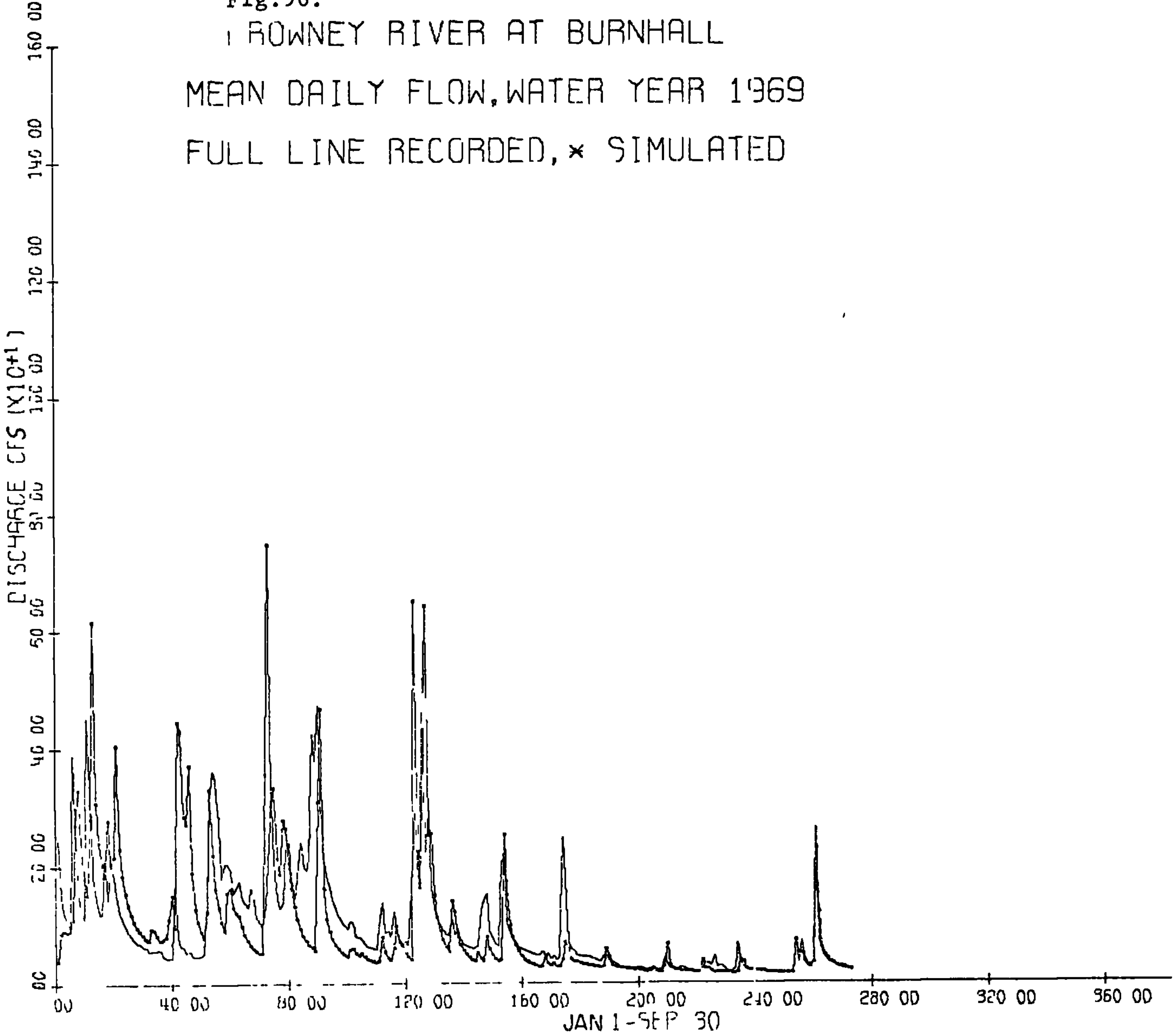

ROWNEY RIVER AT BURNHALL
MEAN DAILY FLOW, WATER YEAR 1969
FULL LINE RECORDED, × SIMULATED
DISCHARGE CFS (X10+1)
00
20 00
40 00
60 00
80 00
100 00
120 00
140 00
160 00
40 00
80 00
120 00
160 00
200 00
240 00
280 00
320 00
360 00
JAN 1-SEP 30

Fig.97.
BROWNEY RIVER AT BURNHALL
MEAN DAILY FLOW, WATER YEAR 1970
FULL LINE RECORDED, * SIMULATED

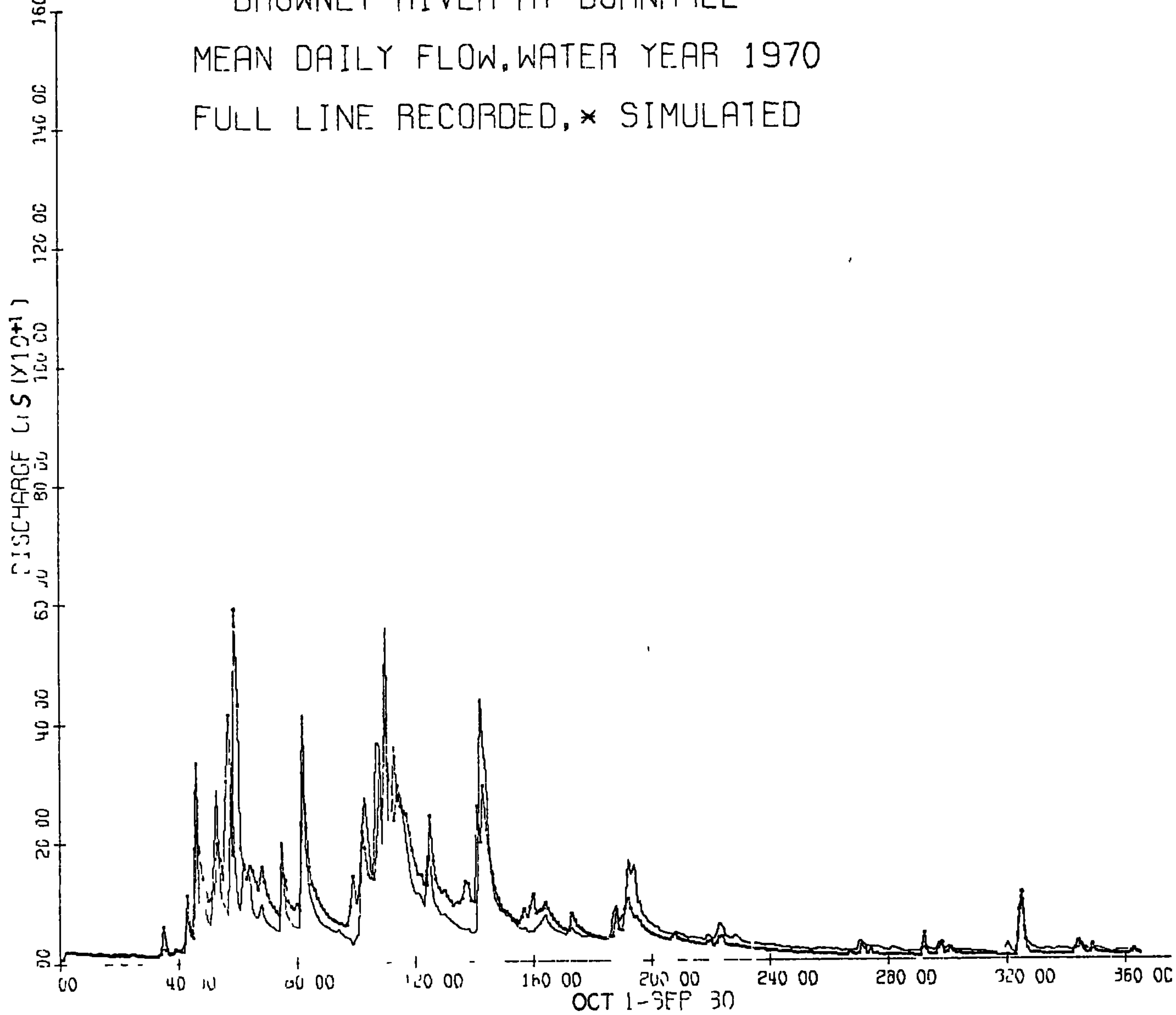
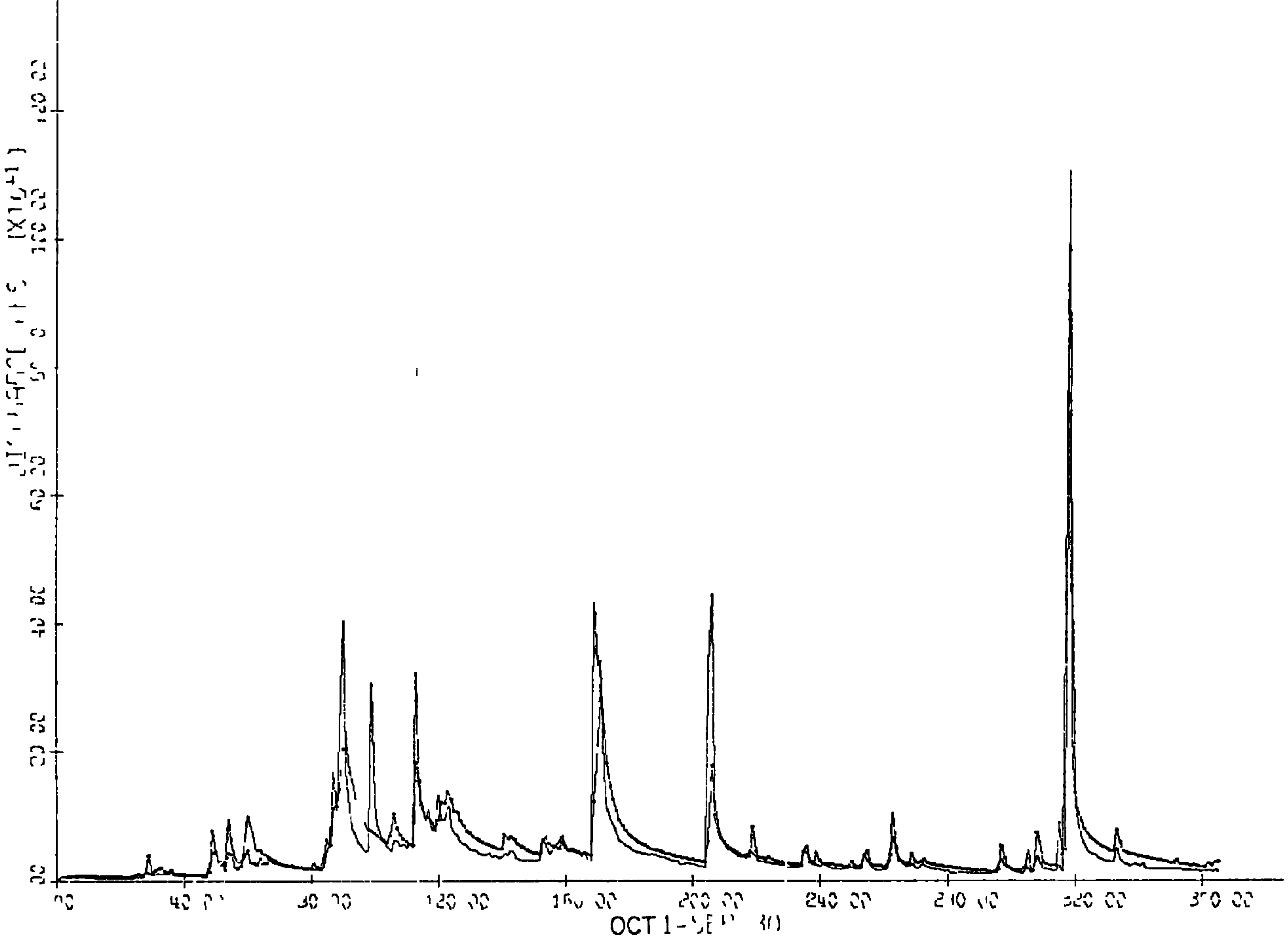


Fig.98

BROWNEY RIVER AT BURNHALL

MEAN DAILY FLOW, WATER YEAR 1971

FULL LINE RECORDED, × SIMULATED



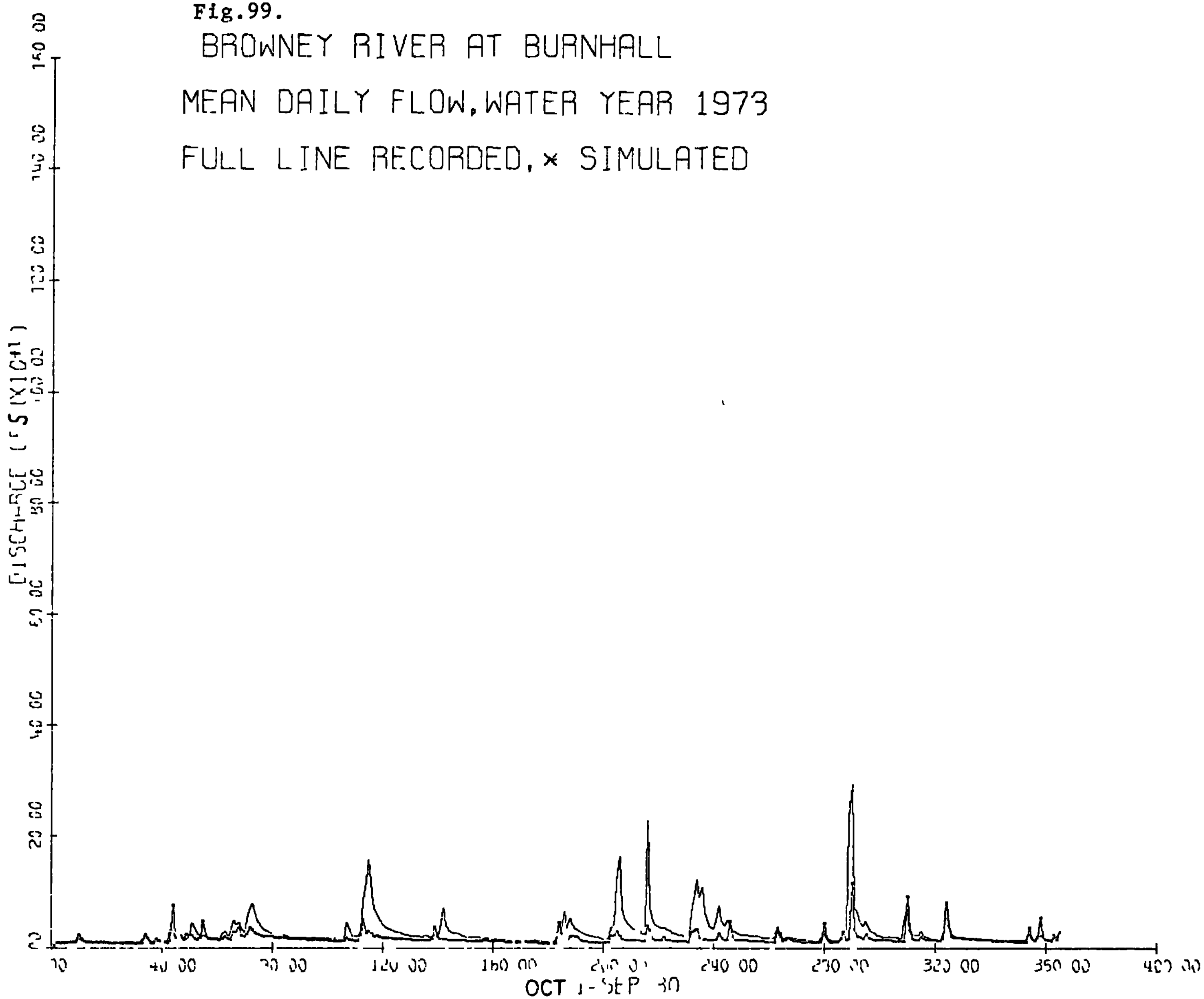


Fig.99.
BROWNEY RIVER AT BURNHALL
MEAN DAILY FLOW, WATER YEAR 1973
FULL LINE RECORDED, × SIMULATED

runoff is constant from storm to storm and the value of N is estimated by the size of the basin

Kunkle (1962) mentioned that these methods are difficult to use and are hard to employ consistently Therefore, he suggested a method which could consistently be used for comparative study while giving reasonable precision. He assumed that the basin storage discharge could be represented graphically by a straight line joining the lowest values at the end of the year. For the bank storage discharge, he suggested a series of recession lines having the same slope These recession lines are connected by a series of straight lines joining the minimum discharge on semi-log paper

In the Stanford Watershed Model IV, the groundwater outflow is represented by an equation of the form

$$GWF = LKK4 \times (1.0 + KV \times GWS) \times SGW$$

where GWS is groundwater slope and SGW is groundwater storage In this equation $LKK4 = 1.0 - (KK24)^{1/96}$ where KK24 is the ratio of current groundwater discharge to the discharge 24 hours earlier

To show the amount of groundwater flow by the Stanford Watershed Model, Table 58 has been prepared. This table shows the amount of groundwater as a percentage of the total precipitation and as a percentage of the recorded and simulated flows for the period January 1969 to September 1973

The percentage of the groundwater flow to total precipitation varies from 26 03 (corresponding to an annual precipitation of 28 39 inches (721 mm)) to 11 64 (an annual precipitation of 22 50 inches (572 mm)). The low percentage of 11 64 belongs to year 1973 Though the depth of precipitation during this year was lower than the other years, the temporal distribution of precipitation was the major reason for the low value of 11 64 This is shown by looking at the percentage distribution of precipitation during the winter and summer seasons of

Table 58 Yearly precipitation, runoff (recorded and simulated), and groundwater flow in inches (mm) and groundwater flow as a percentage of the total precipitation and total runoff

Year	Precip	Runoff		G W flow	% G W flow / ppt	% G W.flow/runoff	
		Recorded	Simulated			Recorded	Simulated
1969 Jan -Sept	25.78 (654 8)	14 12 (358 6)	14.04 (356 6)	6 13 (155 7)	23 78	43.41	43.69
1970 Oct -Sept	28 39 (721 1)	11.75 (298 4)	12.77 (324 4)	7.39 (187.7)	26 03	62.89	57 91
1971 Oct.-Sept.	30.47 (773 9)	9.38 (238 2)	9.80 (248 9)	6 69 (164 9)	21.95	71.32	68 26
1972 Oct.-Sept.	27.20 (690 9)	10 02 (254 5)	9.73 (247 1)	6.48 (164.6)	23.82	64 67	66.60
1973 Oct -Sept	22.50 (571 5)	5.42 (137 7)	3.39 (86 1)	2.62 (66 6)	11.64	48 34	77.29

each year (Table 59). During the water year 1973, 72 per cent of the total precipitation occurred during the summer This figure far exceeds the corresponding values for other years. Therefore it can be concluded that much of the precipitation during this year was used for the evapotranspiration during the summer season This then has resulted in less moisture infiltration and less groundwater flow for the water year 1973

Table 59 Percentage distribution of precipitation during the winter and summer seasons of water years 1970-1973

Water year	1970	1971	1972	1973
Winter (Oct -Mar.)	59 1	46 5	53.9	28 0
Summer (April-Sept)	40 9	53.5	46.1	72.0

A similar explanation could be given for the relatively low percentage of groundwater flow to precipitation during the water year 1971 The total precipitation during this year was 30 47 inches (774 mm) which is more than those of other years. However, the percentage of groundwater flow to precipitation during this year has been the lowest (with the exception of year 1973) Studying the monthly distribution of rainfall during the water year 1971 showed that less than 25 per cent of the yearly rainfall (7 32 inches (186 mm)) had occurred in late August and September, a period when soil moisture storage is low owing to high evapotranspiration during the preceding summer months. Thus much of this rainfall probably had been used in replenishing soil moisture storage.

Groundwater flow expressed as a percentage of the total recorded flow varies from 43 4 per cent (corresponding to an annual precipitation of 25.78 inches (655 mm)) to 71 32 per cent (corresponding to an annual precipitation of 30.47 inches (774 mm)) The average value for the five year period is 58.13 per cent

As a percentage of simulated flow, the range is from 43.69 to 77.29 per cent The high value of 77.29 is due to year 1973, where simulated flow is below the recorded flow by some 38 per cent. The yearly average for the five year period is 62 75 per cent.

As for seasonal variations, the average summer baseflow during the five year period is 69 69 per cent of total flow while the average baseflow during the winter period it is 61 12 per cent.

The percentage of baseflow during the summer could be compared with those values reported by Smith (1969) for two catchments in the drainage basins of the rivers Wear and Tees, north-east England

Using Kunkle (1962) in his study of baseflow, he compared the contribution of baseflow to runoff in two upland catchments of a relatively impermeable nature. The total baseflow, thus obtained, averaged 65 per cent of the summer discharge for one catchment, as compared to 32 per cent for the other. He attributed this difference to variations in the characteristics of catchments.

In a similar study of actual groundwater discharge to the total river flow for the river Stour, Ineson and Downing (1964) found that the baseflow from this permeable sandstone catchment ranged between 65 per cent and 78 per cent of the total river flow For the river Itchen, Hampshire, which has a chalk aquifer, the corresponding values were between 77 and 90 per cent during the period October 1958 to September 1962.

As far as the average monthly variation of baseflow is concerned, January is the month of maximum groundwater flow (five year average of 1 1 inches (28 mm)), and October is the month of least groundwater i e 0.21 inches (5 mm) During the five year period, the average monthly groundwater flow decreased from January till October (with the exception of a slight rise in August which is usually caused by high rainfall during July and August) Groundwater flow increased during the period October to January The monthly variations of baseflow for individual years are shown in Fig 100

Actual evapotranspiration

The actual evapotranspiration within the Browney catchment can also be studied from the results of simulation by the use of the Stanford Watershed Model. These results are shown in Table 60 and Figs 101 to 105 Table 60 shows the yearly variation in the values of potential and actual evapotranspiration

Table 60 Values of potential and actual evapotranspiration and their respective differences in inches (mm) during the water years 1969-1973

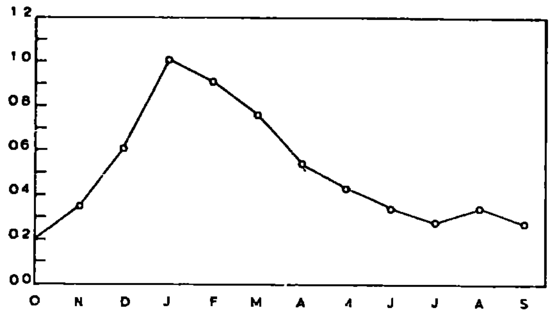
Water year	1969 (Jan -Sept.)	1970	1971	1972	1973
Potential Et	19.03	24.32	20 80	21 52	21 33
	(483 4)	(617 8)	(528 3)	(546.7)	(541 8)
Actual Et	16 93	20 63	18 58	19 44	18 41
	(430 0)	(524 0)	(471 9)	(493 8)	(467 6)
Difference	2.10	3.69	2 22	2.08	2.92
	(53.4)	(93 8)	(56 4)	(52 9)	(74.2)

Differences in the yearly values of actual and potential evapotranspiration range from 3.69 inches (93.8 mm) to 2 08 inches (52.9 mm), corresponding to the water year 1970 and 1972 respectively These differences are due to the amount and distribution of precipitation and potential evapotranspiration. The relatively high evapotranspiration demand of 24.32 inches (617 8 mm) and low precipitation during the summer of 1970 (Fig 102), resulted in the large deficit of 3 69 inches (93 8 mm) The highest monthly deficit is about 1 30 inches (33 mm) in June 1970 During this year, about 4 45 inches (113 mm) of the evapotranspiration demand was satisfied from soil moisture storage.

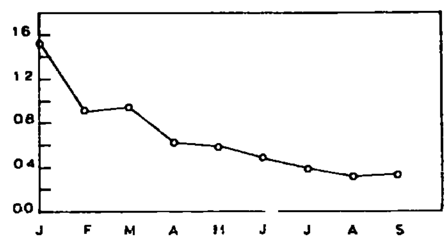
Monthly variations of actual evapotranspiration can be studied from the water balance diagrams of the five year period presented in

Fig 100

GROUND WATER RUNOFF (INCHES)

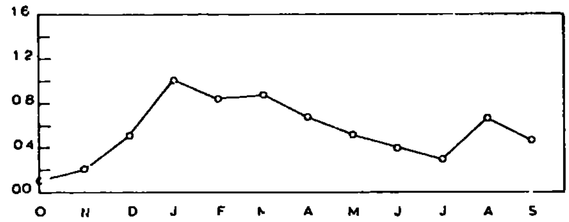


Variation of average monthly ground water runoff in the Browney basin during the period Jan 1969 – Sept 1973

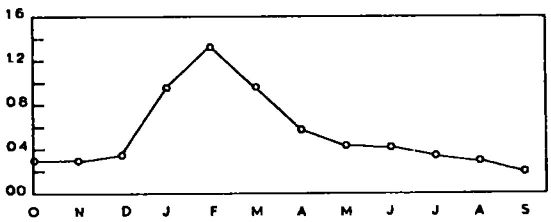


Variation of monthly ground water runoff in the Browney basin during the period Jan – Sept 1969

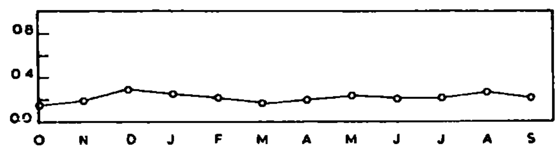
Variation of monthly ground water runoff in the Browney basin during water year 1970



Variation of monthly ground water runoff in the Browney basin during water year 1971



Variation of monthly ground water runoff in the Browney basin during water year 1972



Variation of monthly ground water runoff in the Browney basin during water year 1973

Figs 101 to 105. In each figure the monthly values of rainfall, potential evapotranspiration and actual evapotranspiration, as estimated by the Stanford Watershed Model, are presented.

In these figures the surplus zones show the period and the corresponding amount by which precipitation exceeds evapotranspiration

The zones labelled 'soil moisture use', show the period and the corresponding amount of moisture drawn from the soil moisture reserves This is the period during which evapotranspiration is in excess of precipitation.

The deficit zones show the period and the respective amount by which actual evapotranspiration drops below the potential values. From a general study of these figures, it is shown that actual evapotranspiration falls below potential evapotranspiration in March and reaches its maximum in June or July.

An important observation from these figures is that there are months during which the total monthly precipitation was in excess of the monthly evapotranspiration, yet actual Et has been below that of potential Et e.g. April 1973 or September 1969. This is explained by the effects of an uneven temporal distribution of rain during those months During a period of evapotranspiration demand, for example, no moisture might have been available, whereas on some other days moisture might have been in excess of evapotranspiration. Since in the Stanford Watershed Model IV, runoff and other components of the hydrologic cycle (e g. actual evapotranspiration) are calculated every 15 minutes, the model therefore accounts for the effect of any short dry period on evapotranspiration This short term moisture accounting by the model is an advantage when compared with other models such as that of Penman which was used previously for the estimation of actual evapotranspiration

Moisture surplus, according to Figs 101 to 105 occurs mostly during November to February, the period of least evapotranspiration demand.

Fig. 101 Moisture budget of Browney basin, Jan–Sept 1969

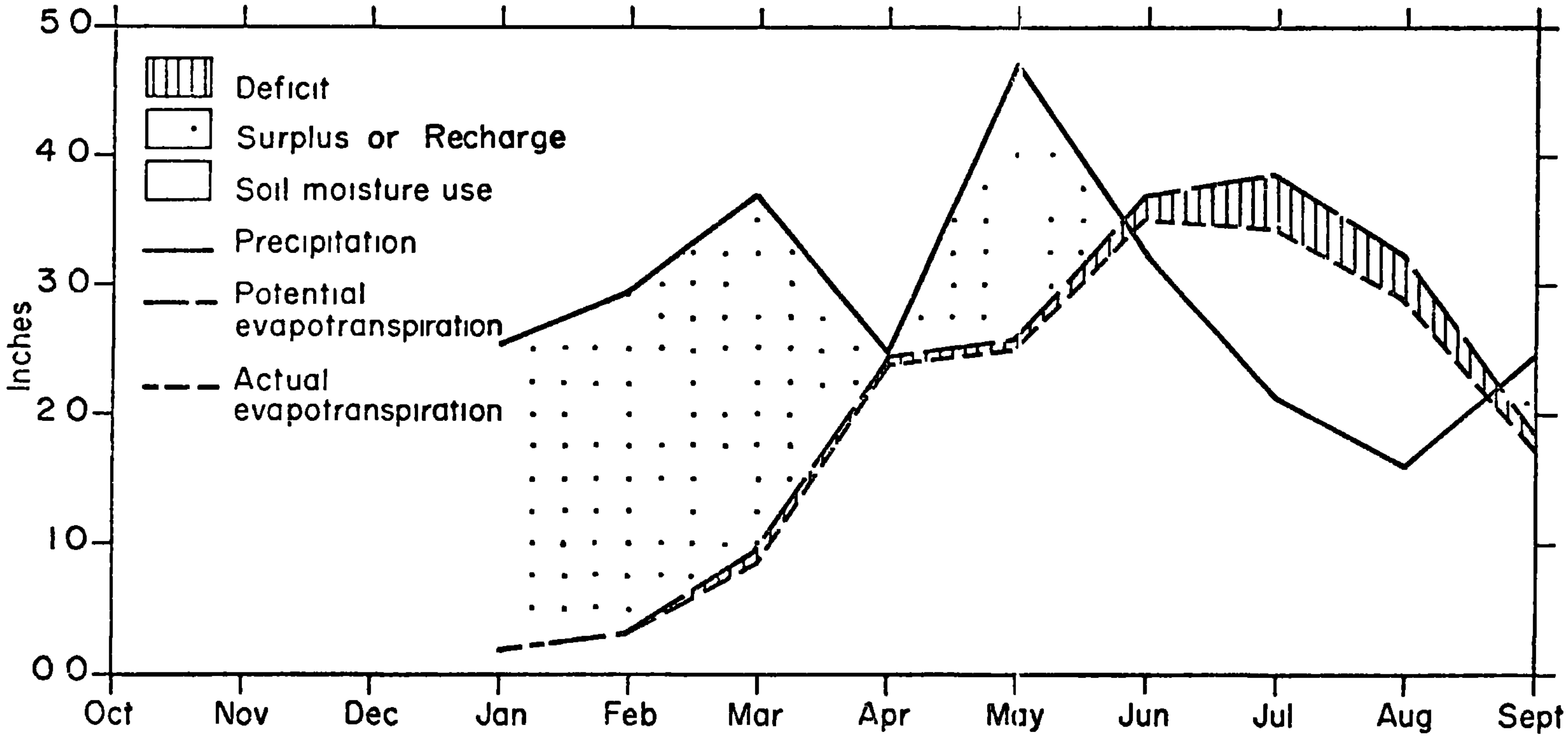


Fig .102 Moisture budget of Browney basin, water year 1970

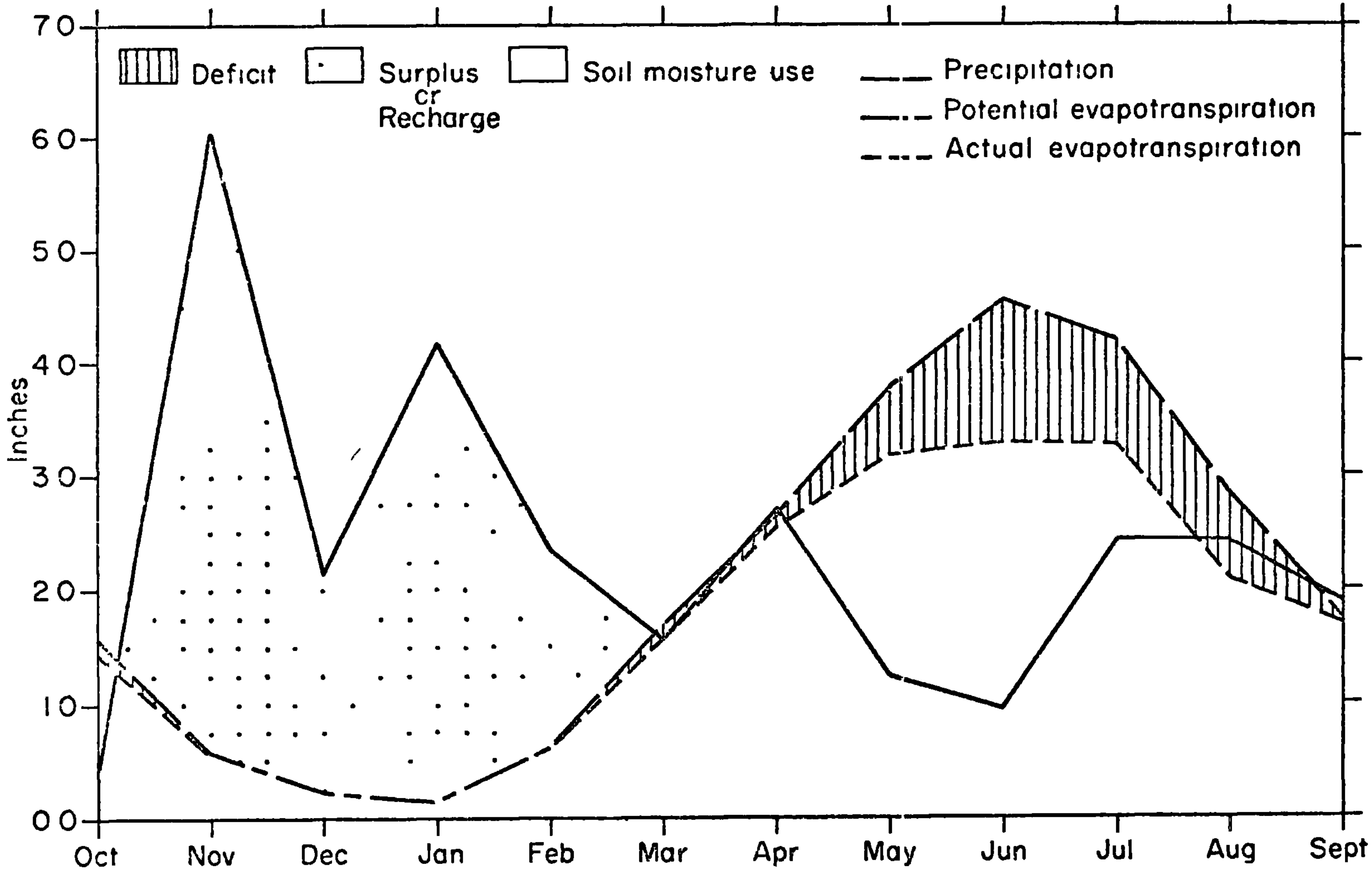


Fig 103 Moisture budget of Browney basin, water year 1971

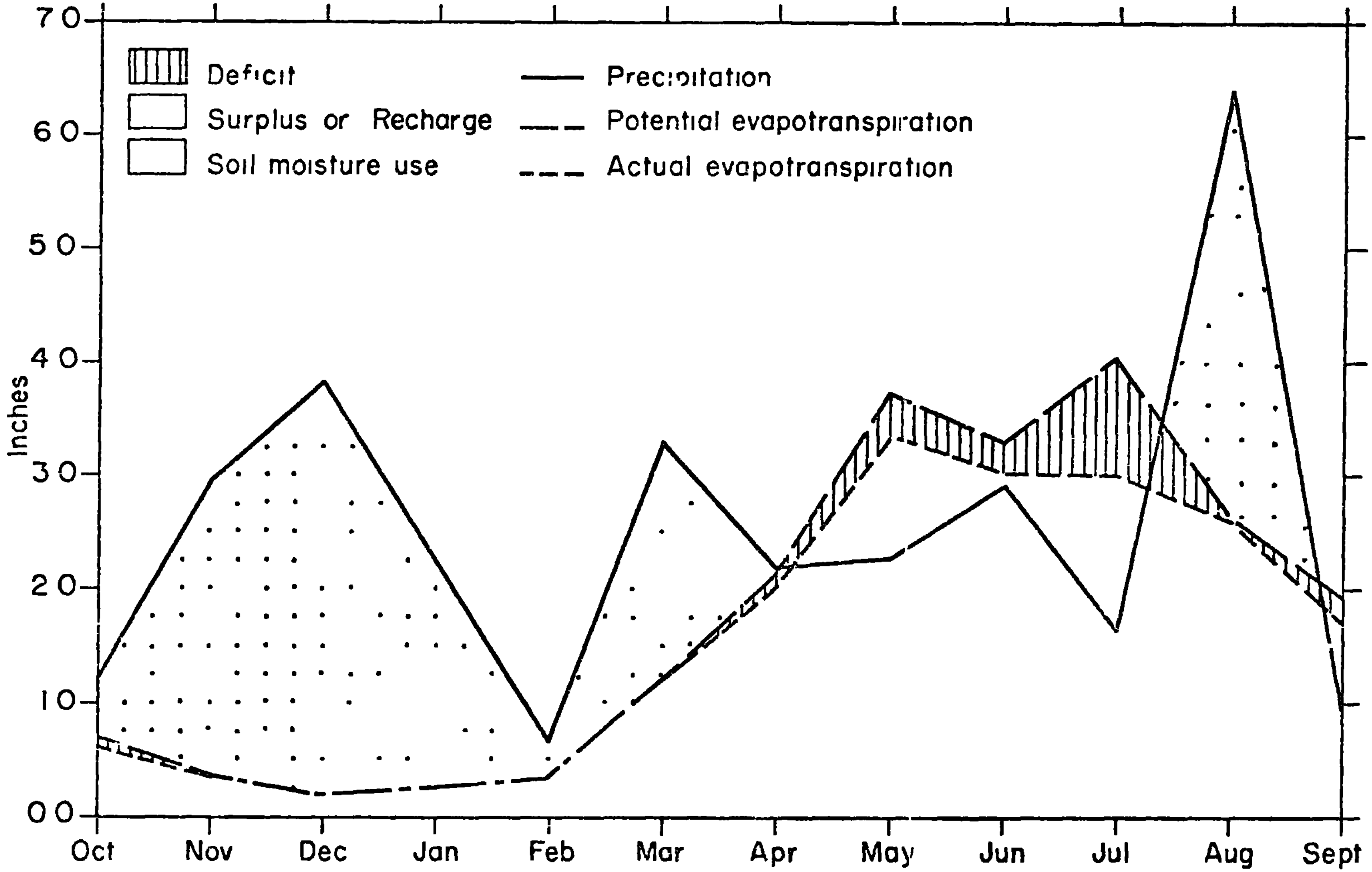


Fig 104 Moisture budget of Browney basin, water year 1972

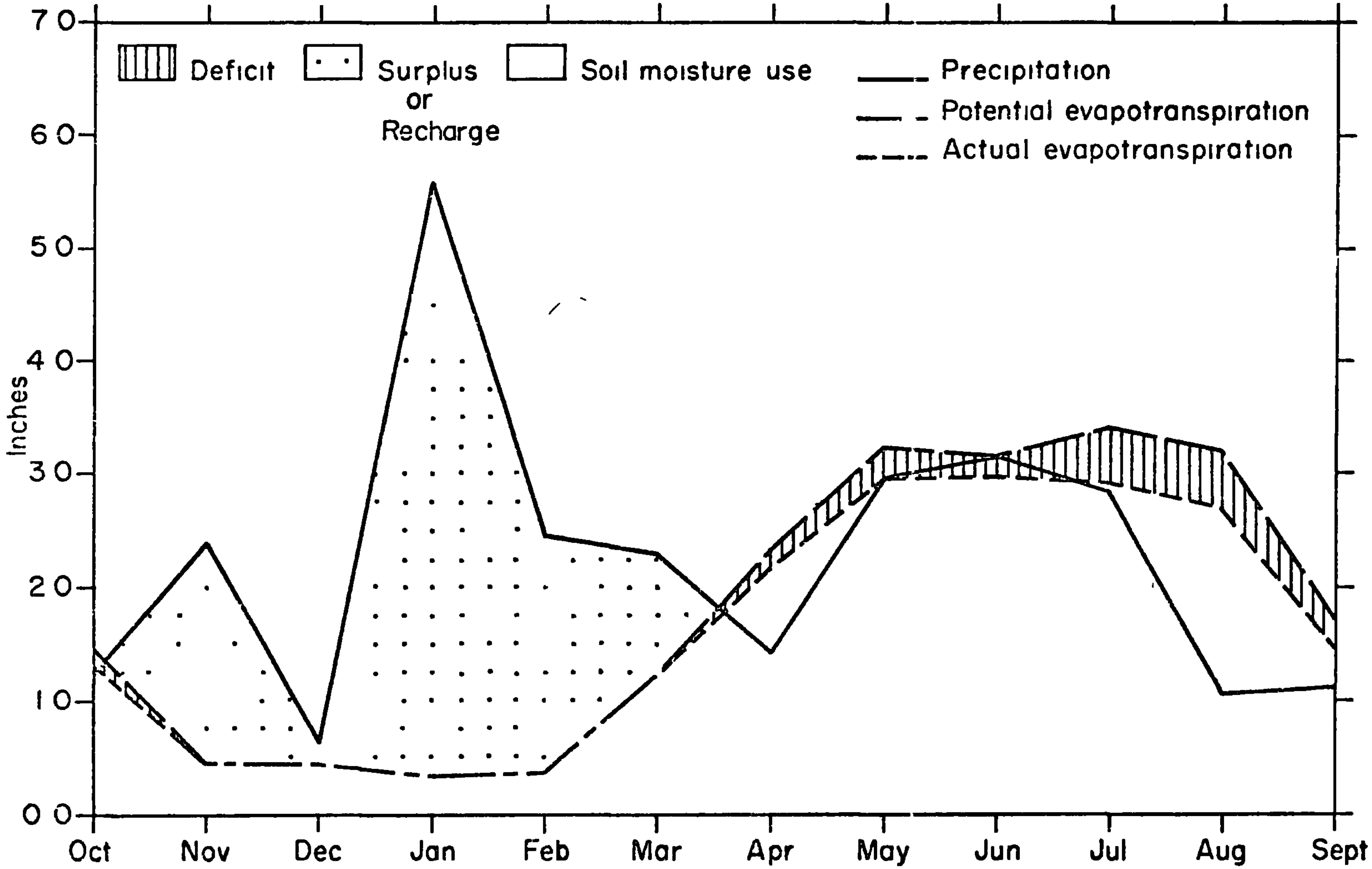
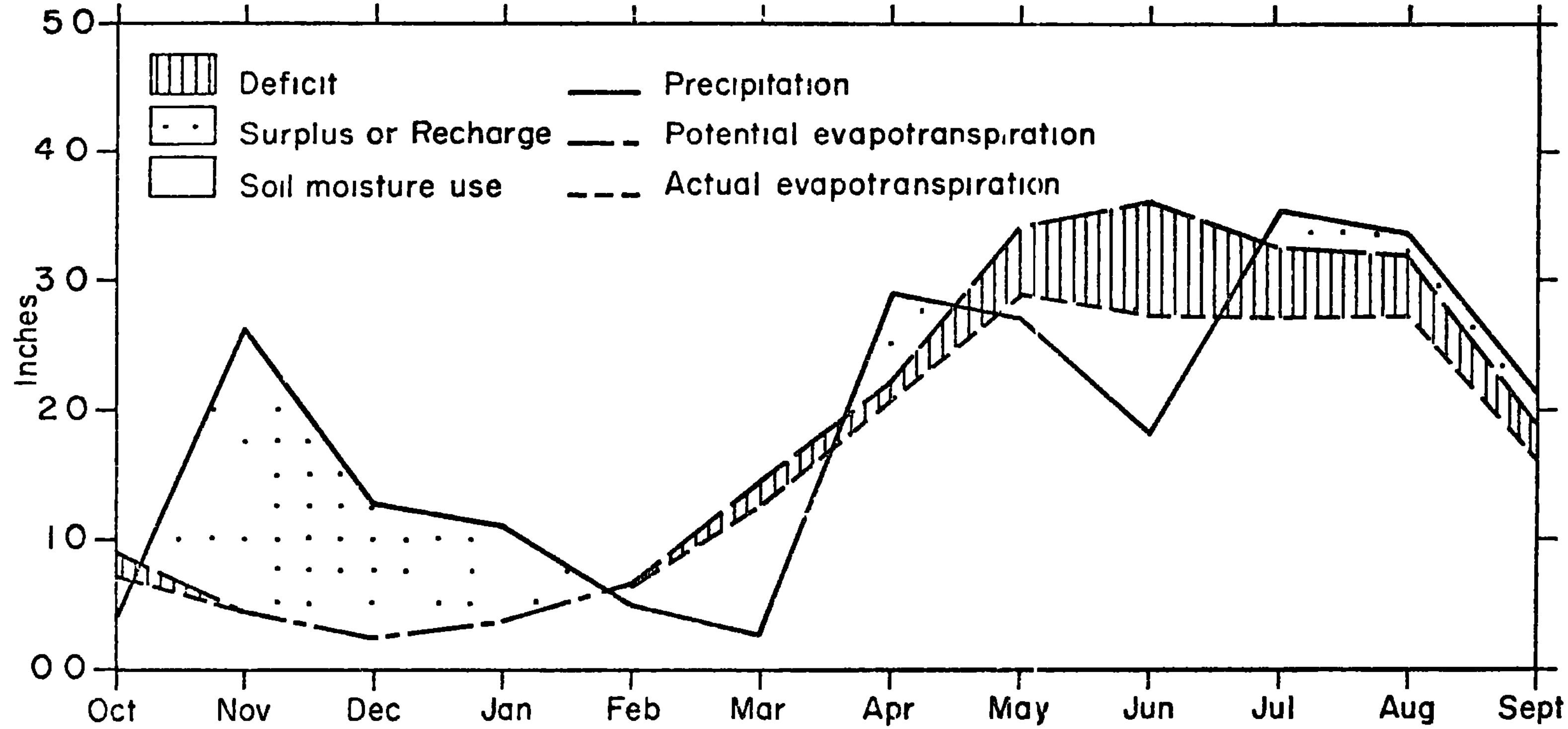


Fig 105 Moisture budget of Browney basin, water year 1973



During this period some of the surplus moisture will contribute to the soil moisture reservoir which has been depleted during the preceding summer.

Finally, the method of actual evapotranspiration estimation by the Stanford Watershed Model can be compared with that of Penman (1949), which was used in the earlier section on evapotranspiration For this comparison, the yearly values of actual evapotranspiration for the water year 1971 obtained by the two methods are considered The total actual evapotranspiration estimated by the Stanford Watershed Model during this year was lower than the potential Et by about 56 4 mm, Table 60, while the actual Et calculated by the Penman method is lower than potential value by only 5 1 mm (Table 22) Such a difference in the estimation of actual evapotranspiration by these models can be explained by two main factors

1 - In the Stanford Watershed Model IV provision is made for direct run-off which quite often occurs, especially in upland areas due to intense thundery rain. Thus during a high intensity rainfall, some moisture might be lost as runoff, while there is still some moisture deficiency in the soil In the Penman model, however, no runoff is assumed to occur, until the whole soil moisture deficiency is removed

2 - As mentioned earlier actual evapotranspiration is calculated every 15 minutes by the Stanford Watershed Model In the Penman model, on the other hand, the estimation of actual evapotranspiration was based on monthly rainfall and potential evapotranspiration data. Thus in the Penman model, the effect of uneven temporal distribution of rain on evapotranspiration is not considered This effect is very important during one month when there is a dry period of three to four weeks followed by a high rainfall amount.

<u>Effect of the potential evapotranspiration input data on the results of streamflow simulation</u>

Potential evapotranspiration data used in this study to simulate runoff were those of the Penman EO_2 which were shown to be most representative of the catchment potential evapotranspiration (See chapter five) Under circumstances when the accuracy of the evapotranspiration data is unknown, optional parameters available in the Stanford Watershed Model could be varied in order to optimize the evapotranspiration data This method, however, adds to the complexity of the optimization procedure

As part of this study, the Penman potential evapotranspiration values were used to investigate the effect of differences in input potential evapotranspiration data on the simulation results and also to test the representativeness of the Penman EO_2 values as an indication of the average potential evapotranspiration within the catchment.

The results of the yearly simulated flows using the Penman Et values for the five year period under study are shown in Fig.106. On the same figure the recorded yearly flows and simulated flows using the Penman EO_2 value are plotted These results indicate that the simulated flows with Penman Et, exceed those with Penman EO_2, during each year of the study The simulated flows with Penman Et also exceed the recorded flows during the first four years of the study For the water year 1973, however, the simulated flows with Penman Et are below those of the recorded flow, though the difference is smaller than that observed with Penman EO_2 This difference is about 26 per cent (1 44 inches (37 mm)).

From a study of Fig.106 it is also noticed that the effect of differences in input evapotranspiration has been more pronounced on the water year 1971 (a relatively wet year), than any other year For this year, the increase in total simulated flow owing to differences in evapotranspiration has been 1 39 inches (35 mm) while the corresponding value for the water year 1970 has been 0 41 inches (10 mm) and for the five year average has been 0 84 inches (21 mm) The results can be ex-

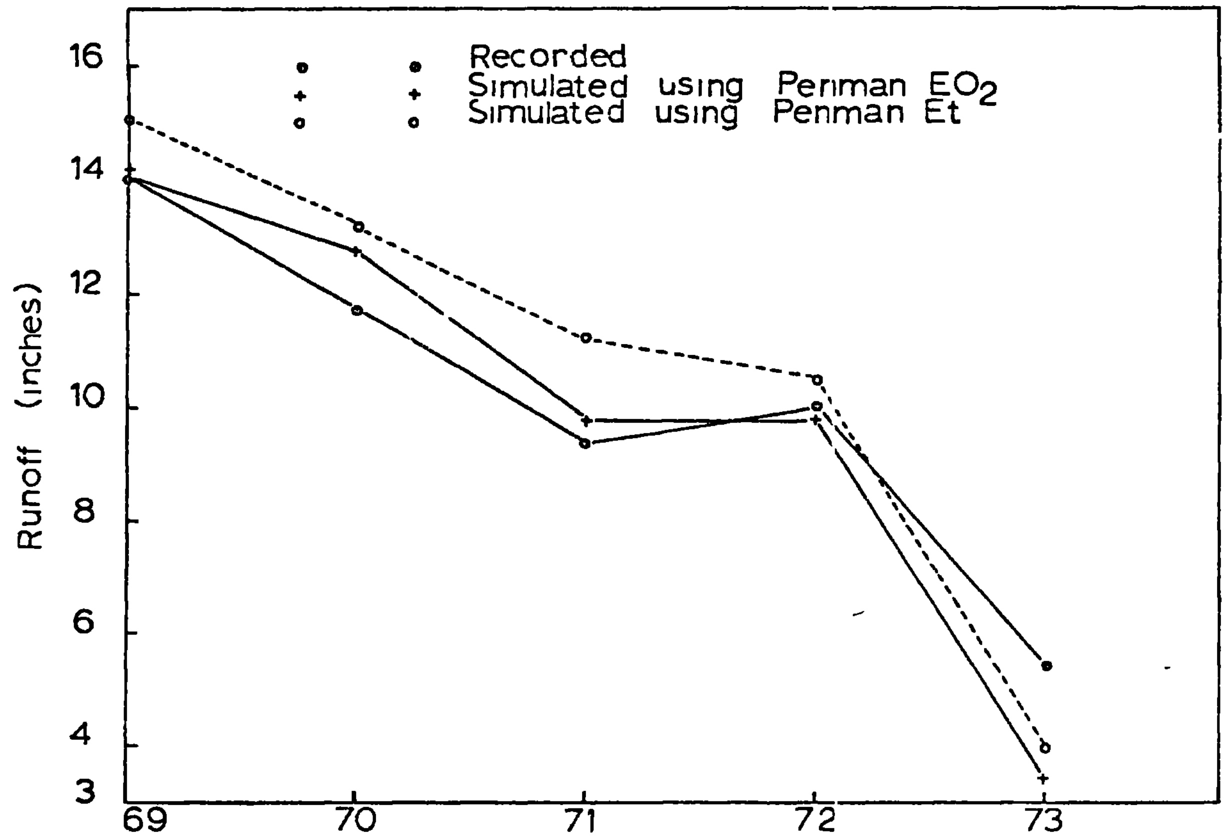


Fig 106 Yearly recorded and simulated runoff values using Penman EO_2 and Penman Et for the Browney river at Burn Hall

plained by the fact that the effect of differences in the potential evapotranspiration value on runoff is greater under wet than dry conditions. When a lot of moisture is available increasing potential evapotranspiration values decreases runoff by the simplified water balance formula i.e runoff = precipitation - evapotranspiration However under dry conditions the rate of evapotranspiration is controlled by the actual state of moisture, therefore, the effects of the differences in evapotranspiration data upon runoff would depend upon the level of moisture availability. The lower the moisture level, the less would be the effects.

This point can be observed by referring to Table 61. This table shows that decreasing potential evapotranspiration during August 1971 (a relatively wet month with 7.32 inches (186 mm) of rain) by 0.47 inches (12 mm) has resulted in a similar reduction in actual evapotranspiration i.e. 0 46 inches (12 mm). However during August 1970 (a relatively dry month with 3.0 inches (76 mm) of rain), decreasing potential evapotranspiration by 0.59 inches (15 mm) has reduced the actual evapotranspiration by only 0 21 inches (5 mm), and thus the effect of lowering potential evapotranspiration has been modified

Table 61· The effects of differences in potential Et on actual Et under two different moisture levels

	Penman EO_2 inches(mm)	Actual Et inches(mm)	Penman Et inches(mm)	Actual Et inches(mm)
August(1971)	2 63	2.56	2.16	2.10
	(66.8)	(65.0)	(54.9)	(53 3)
August(1970)	2.85	2.09	2 26	1 88
	(72.4)	(53.1)	(57.4)	(47 8)

Figs 107 to 109 show that streamflow is almost unaffected during the winter months This is because the Penman Et and EO_2 values do not vary significantly during the winter period

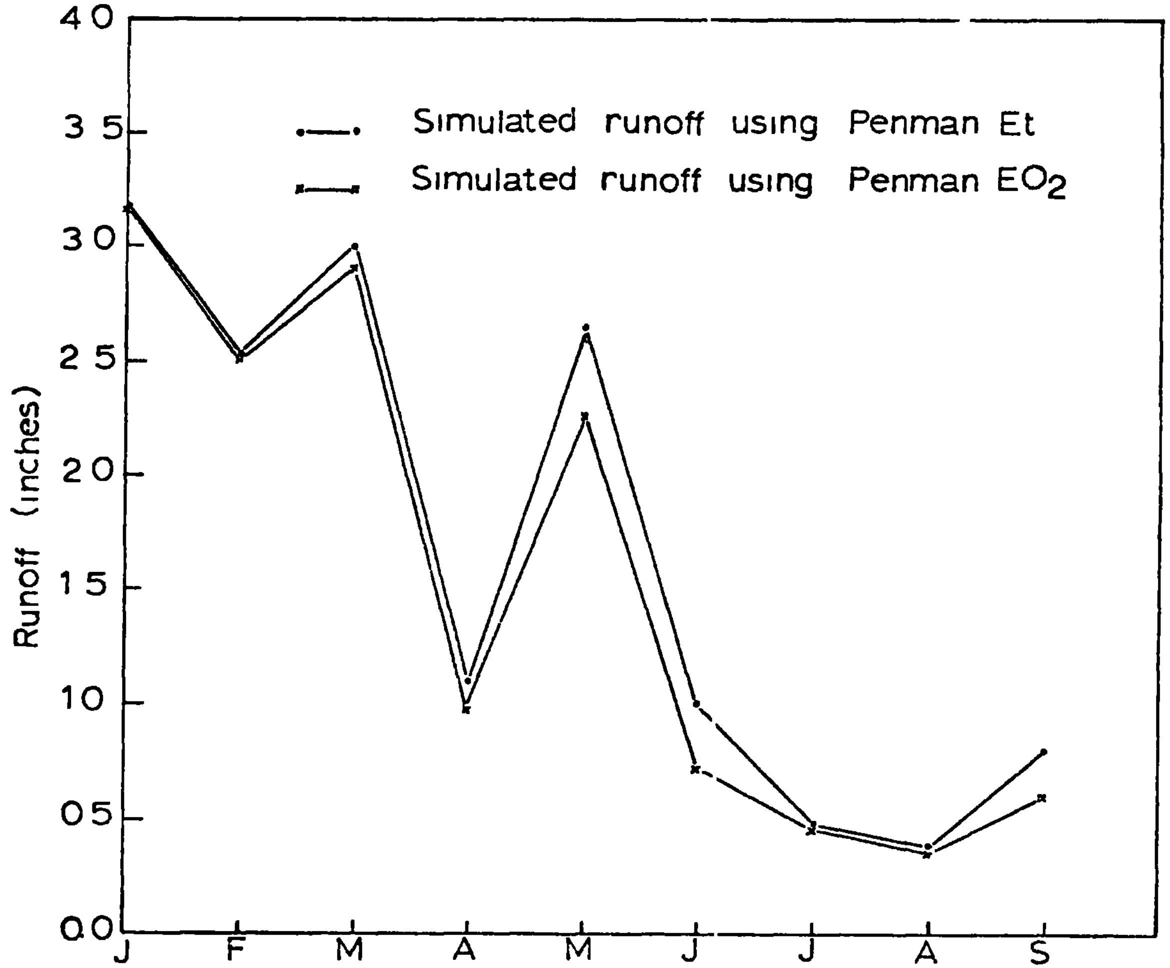


Fig 107 The effect of Penman Et input data as compared with Penman EO_2 on the monthly simulated runoff for the Browney river at Burn Hall, year 1969

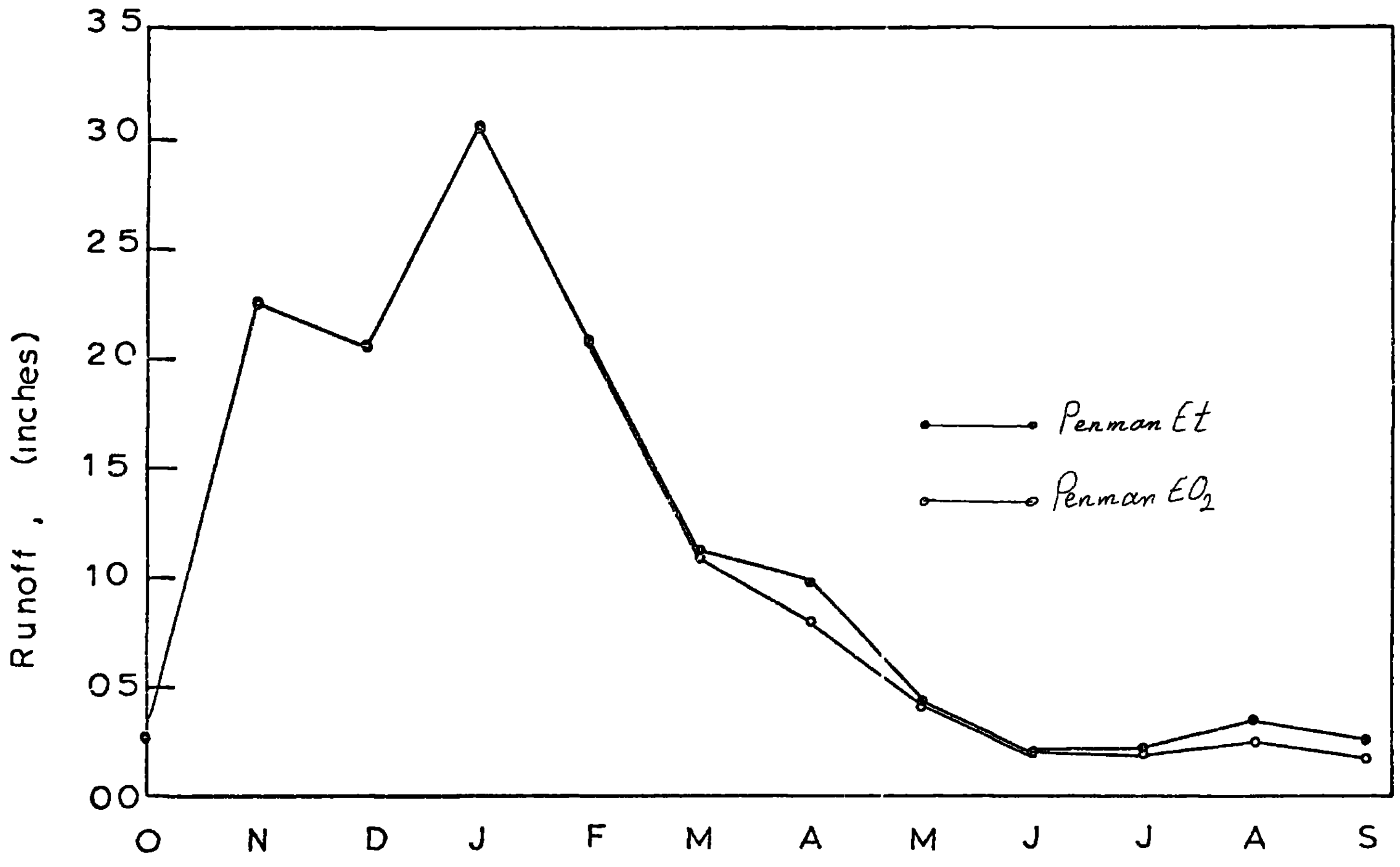


Fig 108 The effect of Penman Et input data as compared with Penman EO_2 on the monthly simulated runoff for the Browney River at Burn Hall, water year 1970

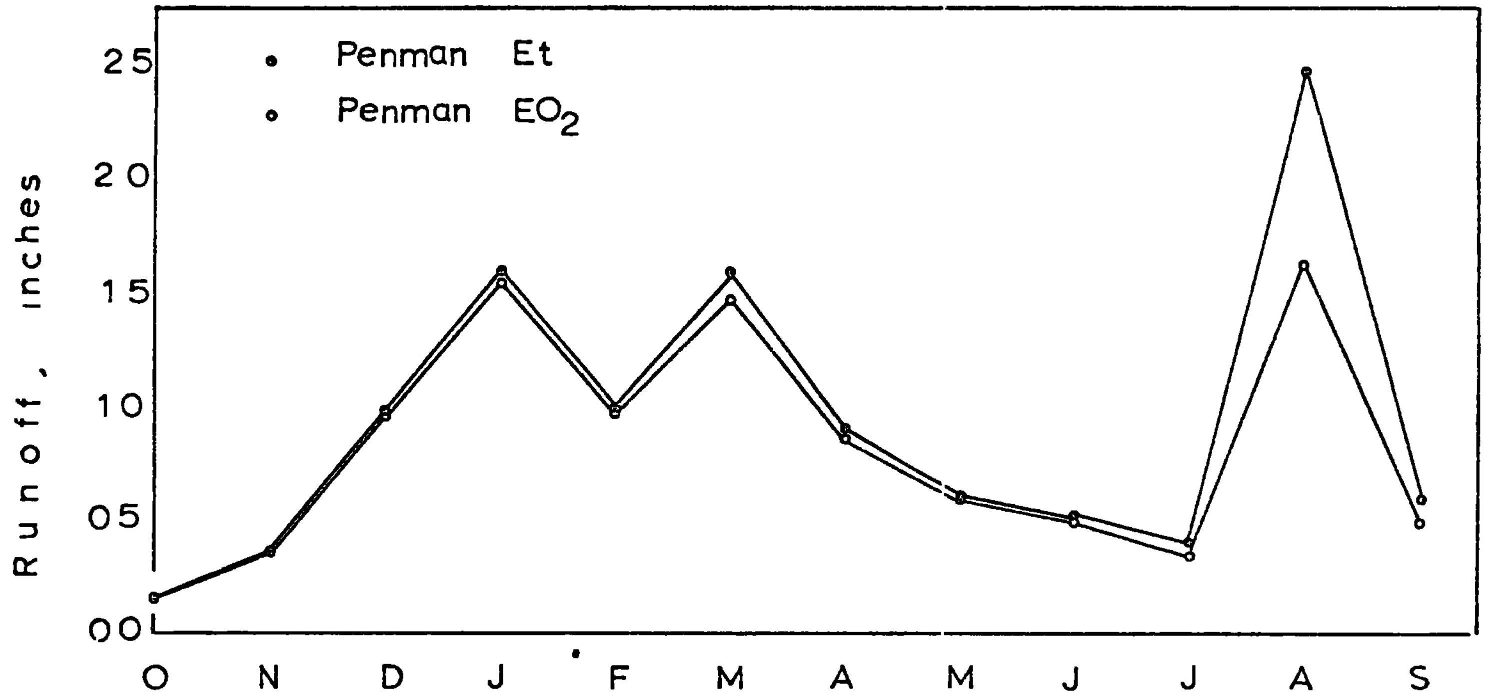


Fig 109 The effect of Penman Et input data, as compared with Penman EO_2 on the simulated runoff for the Browney River at Burn Hall, water year 1971

To show the effect of evapotranspiration on daily flows, Fig 110 represents the simulated hydrographs of daily flows for August 1971, using the Penman EO_2 and Penman Et values. From this figure it is observed that there has been a significant increase in the peak daily flows of 14th and 15th August with the Penman Et values Such an increase in the peak flows could possibly be explained by the effect of low Et on increasing soil moisture content and subsequently by decreasing the infiltration rate. As a result runoff was increased

In the study of the effects of differing evapotranspiration input data on the simulated flows of each year, all the parameters, including the initial soil moisture conditions, were kept constant However, using the Penman Et values, the initial soil moisture conditions for any year would be higher owing to lower depletion of soil moisture storage in the preceding water year Therefore, it can be concluded that in such a study, if the initial moisture conditions for any year were chosen from those at the end of the preceding year, the streamflow for that year would have been affected more by the difference in evapotranspiration.

To give an example the value of LZSN at the end of year 1969 was 9.34 inches (237 mm) using the Penman EO_2 values, and 10.55 inches (268 mm) using the Penman Et values To compare the effect of evapotranspiration input data on simulated flow for the year 1970, the same value of 9.34 inches (237 mm) which was used for the original simulation was adopted (instead of 10 55 inches (268 mm)) for simulation with Penman Et However, if the value of 10 55 inches (268 mm) had been used, simulated flow would have been higher owing to this increase in initial lower zone storage and, therefore, the effect of the difference in potential evapotranspiration would have been more pronounced Thus the simulated flow with Penman Et would have exceeded the recorded flow by a greater amount than was shown in Fig.106 This leads to the conclusion that the Penman Et values are

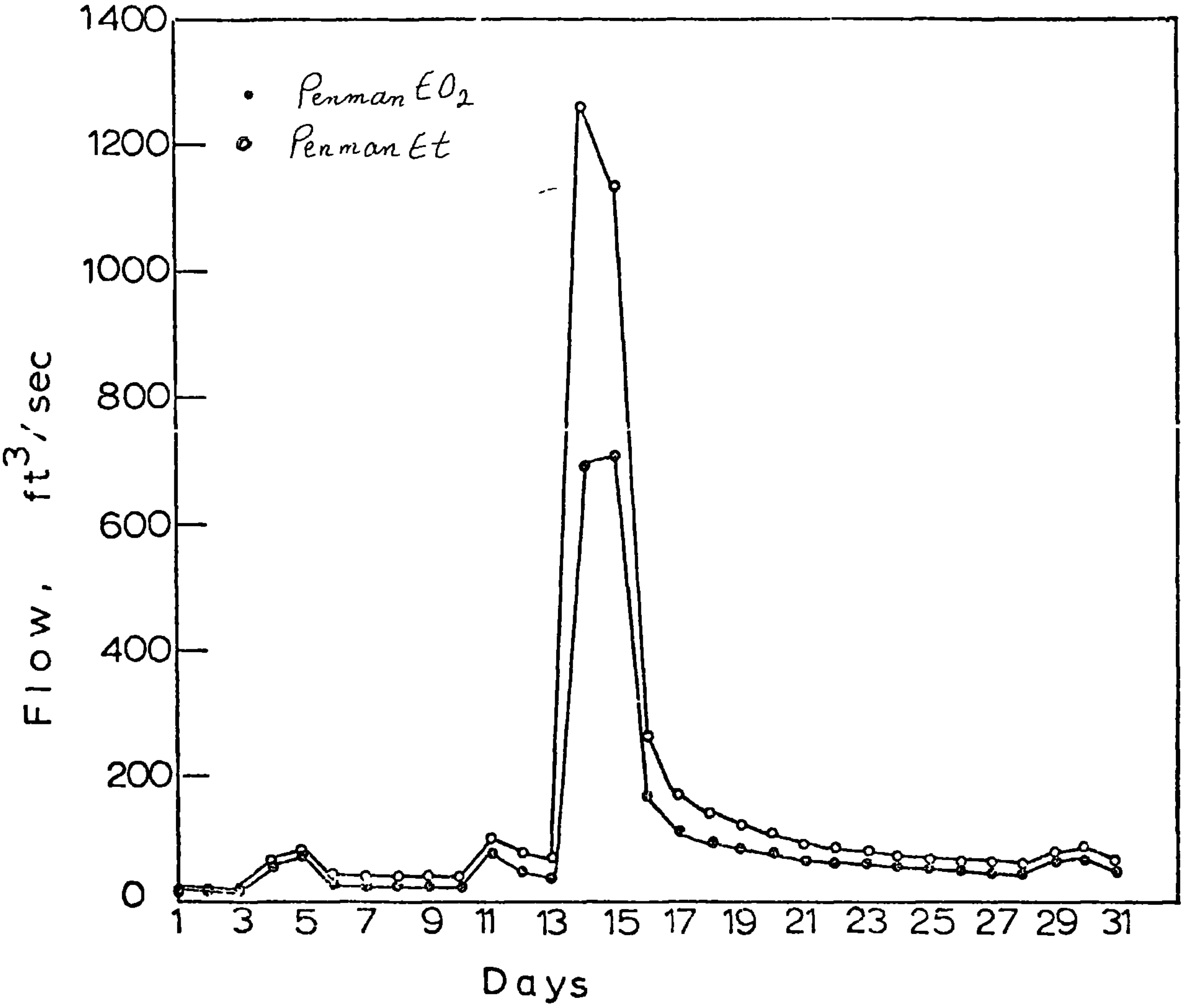


Fig 110 Simulated Hydrographs of daily flow for the Browney River at Burn Hall during August 71, using Penman EO_2 and Penman Et values

too low and potential evapotranspiration by the Penman EO_2 is more representative of the average evapotranspiration of the catchment, thus the results obtained in the evapotranspiration studies are confirmed

SUMMARY AND CONCLUSIONS

The aims of this thesis were to study aspects of the hydrology of the Browney basin, north-east England. For this purpose different hydrological methods and techniques were employed to investigate precipitation, evapotranspiration and runoff in the catchment. The overall hydrological performance of the catchment, then, was studied by the simulation of river discharge using the Stanford Watershed Model IV.

Initially the catchment characteristics were studied These characteristics were geology, soil, land use, shape, elevation, slope and drainage network, and they were discussed in chapter one The hydrological importance of each of these factors was also mentioned

In chapter two, the spatial and temporal variations of precipitation were investigated. For this purpose a description of the raingauges whose data were employed in this study were given The different methods for the estimation of mean areal rainfall were later discussed. Arithmetic, Thiessen and isohyetal methods were used. Based on these methods, the mean areal rainfall for the years 1968 to 1972 was determined It was observed that the mathematical method of estimating mean areal rainfall gave the highest value for the mean areal precipitation, where as the Thiessen polygon gave the lowest This difference was explained by the difference in weight given to the raingauge located at the western end of the catchment.

From the study of the yearly isohyetal maps, the catchment was divided into three different rainfall zones In the eastern and western parts of the catchment, yearly precipitation increased in an upstream direction The rate of increase, however, was greater in the western zone owing to the higher elevations and greater slopes In the central zone yearly rainfall was rather uniform

Considering the correlation coefficients between yearly precipitation and the elevations of the raingauge stations, it was shown that the association between these two factors was high and that the regression equation between yearly precipitation and altitude during 1968 to 1972 explained between 71 to 97 per cent (corresponding to years 1972 and 1970) of the areal variation in yearly precipitation

To study temporal variations of precipitation over a long period, the daily precipitation data at Durham Observatory during the 35-year period 1939 to 1973 were used Frequency curves of yearly rainfall, yearly 24-hr maximum rainfall and monthly rainfall were prepared. From these curves the return period for the occurrence of a rainfall amount greater than or equal to a certain value was derived Thus, the frequency curves calculated from data from the Durham Observatory, showed that there was a chance of getting 630 mm or more rainfall every other year and 812 mm or more every 30 years Similarly annual 24-hr maximum rainfall of 29 8 mm and 51 4 mm had return periods of 2 years and 30 years respectively.

Daily rainfall amounts were also grouped into 4 classes equal to or greater than 0 2 mm, 1 mm, 2 mm, 5 mm and 10 mm Based on these data the average frequency (in per cent) of days per month with precipitation greater than or equal to the stated amount during the 35 year period was determined. Separate frequency curves of the number of rain days with precipitation greater than or equal to 0 2 mm, 1 mm, 2 mm, 5 mm and 10 mm for the months of August, November and April were drawn.

From the results of the frequency studies of daily rainfall amounts it was observed that the average frequency (in percentage per month) of daily rainfall equal to or greater than 0 2 mm, 1 mm, 2 mm and 5 mm was highest during November. For the rainfall amounts greater than or equal to 10 mm, however, August had the highest monthly frequency. The daily rainfall amounts greater than or equal to 10 mm, also had high frequencies during July and September The relatively high frequency of intense rainfall during the late summer months gave evidence of the occurrence of convective precipitation during this time of the year. Further evidence for the occurrence of convective precipitation was obtained from the study of hourly rainfall during the period 1969 to 1974 From the six year study of hourly rainfall, it was observed that August had the highest frequency (per cent per month) of hours with precipitation equal to or greater than 2 mm, 5 mm and 10 mm, while November had thehighest frequency of hours with precipitation greater than or equal to 0 2 mm and 1 mm

From the study of hourly precipitation, it was observed that intensities of hourly rainfall in the Browney basin were generally low The highest hourly rainfall occurred in July with a depth of 20 3 mm The longest run of dry days per month during the 35 year period was 30 and that occurred in the year 1953.

The section on evapotranspiration was divided into three parts In the first part (chapter three) a review of literature on the theory and methods of measurement of evapotranspiration was presented. The theoretical formulae were discussed and the advantages and limitations of the empirical formulae of Penman and Thornthwaite were mentioned. Other methods for the measurement of evapotranspiration were studied subsequently Thus, the principles, advantages and disadvantages of the catchment water balance, evapotranspirometers, evaporation pans, lysimeters and empirical methods of estimation of actual evapotranspiration were discussed.

In the second part of the section on evapotranspiration (chapter four), the methods which were applied to measure or estimate evapotranspiration in the Browney basin were outlined and the procedures followed were discussed in detail These methods were evapotranspirometers, catchment water balance, the empirical formulae of Penman and Thornthwaite, simple hydraulic lysimeters and the Penman drying curve method for estimating actual Et Thus two sets of evapotranspirometers were set at Durham (elevation 102 m) and Honey Hill (elevation 334 m). The evapotranspirometer set at Durham was run for five months (May to September 1973) and that at Honey Hill was run for a 12 month period (July 1973 to July 1974). Potential evapotranspiration was also determined by the Penman and Thornthwaite formulae For this purpose, the meteorological data from Durham Observatory were used. Thus daily potential evaporation (albedo 0 05) and potential evapotranspiration (albedo 0 25) were determined by the Penman formula over a 10 year period (1963 to 1973) The calculations were facilitated by the application of a computer program Penman potential evaporation was estimated by applying two different empirical wind functions.

For the Thornthwaite formula, a simple program was written by which the monthly values of Thornthwaite potential evapotranspiration over the 10 year period 1963 to 1973 were determined

In order to estimate the average evapotranspiration over the catchment, the values calculated from the different methods explained earlier were compared with that of the water balance method. The average yearly water balance Et was obtained by deducting the average yearly runoff from the average yearly precipitation during the period 1963 to 1973.

For the measurement of actual evapotranspiration, two different methods were used. The first method employed simple hydraulic lysimeters These lysimeters were tested during a period of four and a half months at Durham Observatory.

The second method was the application of the Penman model for the estimation of actual evapotranspiration Using this latter procedure, diagrams of average moisture budget for the 10 year period 1963 to 1973 and for the years 1964 and 1971 were drawn

The results of the evapotranspiration study were given in the last part of the section on evapotranspiration (chapter five). Thus by the comparison of the measured evapotranspiration values at Durham and Honey Hill it was observed that evapotranspiration values were about 10 per cent greater at the higher elevations. No definite conclusion, however, was drawn from these results The variation at the sites of measurement and possible observational errors in the measurement of input and output moisture might have resulted in greater evapotranspiration at the higher elevation.

Comparing the Penman potential evaporation and evapotranspiration values with measured evapotranspiration, it was concluded that the Penman evapotranspiration values were too low. The explanation for the low values of Penman Et, was thought to be the lack of measured radiation data

From the study of average yearly evapotranspiration calculated by the water balance method and the results of measured evapotranspiration, it was concluded that, Penman EO_1 and Thornthwaite Et were probably too high to represent areal evapotranspiration within the catchment. Therefore, the Penman EO_2 was assumed to be the most representative estimate of potential evapotranspiration

The results of actual evapotranspiration studies obtained from the two simple hydraulic lysimeters showed some significant differences during the short periods of observation Another problem in the use of such small lysimeters was in the extrapolation of the actual evapotranspiration values obtained at a point to values representative of the whole catchment. This was because the volume of soil in the lysimeters was limited and therefore the soil-plant-water relationships existing in the field were different from that of the lysimeters

Application of the Penman method for the estimation of actual evapotranspiration showed that on the basis of average moisture budget for the 10 year 1963 to 1973, there was no moisture deficiency in the Browney basin However, for the year 1964, the driest year during the 10 year period, the total depth of actual evapotranspiration was below that of potential evapotranspiration by 109.2 mm or 16.8 per cent

In chapter six, runoff variations during the period 1957 to 1973 were studied Initially the yearly and seasonal variations of runoff were investigated and the corresponding rainfall-runoff equations were derived The results obtained showed that on average, yearly runoff was 41 per cent of total precipitation The average ratios of runoff-rainfall, however, were 0 57 during the winter season and 0.25 during the summer season

The range of monthly values of runoff during the 17 year period was between 120 mm and 3 mm Both of these extreme values occurred in October. The mean value of monthly runoff was 25.8 mm From the study of monthly runoff values, it was observed that January, February and March were generally the months with highest flow, while September was the month with lowest flow.

The duration curves of yearly, monthly and mean daily flows were presented later in the chapter. From the study of these duration curves, it was concluded that fifty per cent of the time yearly runoff was greater than or equal to 280 mm, and fifty per cent of the time monthly runoff was greater than or equal to 18.9 mm. As for the mean daily flows, the highest was 20.99 mm Fifty per cent of the time mean daily flows were greater than or equal to 0 45 mm

The results of the study of extreme runoff values were also presented in chapter six According to these results, the 100-year flood for the Browney basin was 154 mm (monthly flow), while the 100-year drought was 2.5 mm (monthly flow) The runoff pattern of the catchment was studied by considering the hydrographs of two rather large similar storms, one of the storms had occurred during the summer, while the other had occurred during the winter season From the study of the distribution graphs of these two storms, it was concluded that there was a marked difference in the runoff pattern between the summer and winter seasons During the winter the rise in the hydrograph is gradual, while during the summer the rise is sharp. The peak flow (as a percentage of the total runoff) during the summer is greater than that during the winter The lag during the summer season was higher than that during the winter season

Chapter seven of this thesis was used to review some of the methods of runoff prediction These methods were the empirical formulae, infiltration methods and infiltration indices, regression and graphical methods, the moisture accounting method and statistical methods This chapter was concluded by a definition of simulation and an explanation of the progressive development of the Stanford Watershed Model IV for the prediction of runoff and the indirect investigation of the hydrology of a catchment. In this chapter the implication was made that while the Stanford Watershed Model IV was the least empirical of all the runoff prediction methods, it was not the ultimate answer to the problem of runoff prediction

In chapter eight the application of the Stanford Watershed Model IV to the Browney basin was explained. There were two main objectives in the application of the Stanford Watershed Model IV to the Browney basin These were

1 - To test the ability of the model in the prediction of runoff from a British catchment, using the Browney basin as a case study.

2 - To study the overall hydrological performance of the Browney basin and, in particular, the components of the water balance equation within the catchment which are not easily studied otherwise e g. groundwater flow and actual evapotranspiration.

The operation of the Stanford Watershed Model and the input data required for its application to the Browney basin were explained in chapter eight The important parameters were discussed

In chapter nine the results of the sensitivity tests of optimized parameters were given A computer program was written to plot the hydrographs of daily flows for this purpose. The process of optimization was defined and the values of the optimized parameters used in the simulation of daily flows for the period January 1969 to September 1973 were given. These values were obtained after numerous unsuccessful trials of the model It was concluded that the infiltration index, upper zone storage and lower zone storage parameters were the most sensitive in the hydrological regime of the Browney basin while the actual evapotranspiration parameter was the least sensitive The importance of the initial moisture condition parameters in the simulation of yearly hydrographs by the Stanford Watershed Model IV was discussed and the requirement of two or more years of data for the optimization of parameters was partly explained by the significant effects of these parameters upon simulated flows It was also mentioned that the process of optimization was dependent upon the objectives for which the model was to be used

In chapter ten, the simulated results for each year were compared with the corresponding recorded flows These results were the total yearly runoff, the mean monthly flow, the standard deviation of monthly flow and the correlation coefficient between recorded monthly and simulated monthly flow The hydrographs of simulated and recorded mean daily flow and monthly flow were also compared

The results obtained revealed that during the five year period, the recorded mean monthly value exceeded the simulated value by some 2 per cent The monthly correlation coefficient over the five year period was 0.94, showing that the simulated flow was highly correlated with recorded flow

Considering the individual years, the simulated monthly flows were quite close to the recorded flows during the water years 1969, 1970, 1971 and 1972 The correlation coefficients between the simulated monthly and the recorded flows range from 0 99 in 1972 to 0 90 in 1969 For the water year 1973, however, there was a wide discrepancy between the simulated and recorded flows The monthly correlation coefficient was 0 57 Several explanations were put forward as possible sources of error in the simulation of flow during this year The fact that the 1973 water year had been an exceptional year (lowest total runoff during the 17 years of study), was probably a very important factor During this year more than 72 per cent of the precipitation occurred during the summer and most of the moisture had been used for evapotranspiration

The yearly baseflow calculated by the model varied from 2 62 inches (66.5 mm) in 1973 to 7 39 inches (187 7 mm) in 1970 or 11 64 to 26 03 per cent of the yearly precipitation As a percentage of the total recorded flow, the groundwater contribution was 58 13 per cent The average summer baseflow during the five year period was about 70 per cent of the total recorded flow, while during the winter it was about 61 per cent

The values of actual evapotranspiration were used to draw the diagrams of the water budget of the Browney basin for each of the five years 1969 to 1973 Differences between the yearly values of actual and potential evapotranspiration during the five water years ranged from 94 mm (1970) to 53 mm (1972).

Finally the effect of input potential evapotranspiration data upon the simulated runoff was studied For this purpose the Penman EO_2 values were replaced by the Penman Et values The model was then run for each of the five water years The results obtained showed that yearly simulated runoff with the Penman Et values were more than that with the Penman EO_2 values during each water year. The greater yearly simulated runoff with the Penman Et value as compared with the Penman EO_2 data was the result of lower Et values during the summer season In fact the monthly simulated runoff during the winter months was almost the same for both evapotranspiration input data This can be explained by the fact that the Penman Et and EO_2 values were virtually the same during the winter season The effect of lower input Et upon daily runoff was shown by plotting the hydrographs of simulated daily runoff during August 1971 using the Penman Et and EO_2 values It was shown that the peak runoff on 14th August increased sharply as a result of using the Penman Et value.

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APPENDICES

APPENDIX I

UPPER CARBONIFEROUS COAL SEAMS

	Name	Thickness	Workable extent
	Hebburn Fell	4 ft	
	Usworth	Up to 5 ft	Restricted to coast
	Ryhope Five-Quarter	2½ - 3 ft	Restricted to Eastern areas
	Ryhope Little	3 ft	Little worked, due to erosion
Middle Coal Measures	High Main	1 - 8 ft	Mid-east Durham - dirty
	Five Quarter	1 - 6 ft	East of Durham City
	Main	2½ - 7 ft	One of the principal seams
	Maudlin	Up to 6 ft	Worked in North Durham
	Durham Low Main	2 - 6 ft	Important over most of Durham
	Brass Thill	2½ - 3 ft	Predominant in Central Durham
	Hutton	Up to 7½ ft	One of thickest seams
	Ruler	Up to 2 ft	Worked in north west
	Harvey	1 - 5 ft	Good widespread seam
	Tilley	3 - 5 ft	Worked in the west
Lower Coal Measures	Busty	1 - 5 ft	Bottom Busty worked in west
	Three-Quarter	Up to 3 ft	Impersistent Improves offshore
	Brockwell	2 - 6 ft	Very important in south-west
	Victoria	2 ft	Important in the north-west
	Marshal Green	1 - 2 ft	Open casted in west Durham
	Ganister Clay	Up to 1½ ft	Thin and of no economic value
	Gubeon	Up to 1½ ft	Persistent but uneconomic.

After Smith, 1972, modified from Geology of Durham

APPENDIX II

LIST OF SYMBOLS

Chapter One

p^f is log of soil moisture tension when expressed in terms of cm of water.

k_p is compactness coefficient (shape index).

p is perimeter

A is area.

A_b is area in Miller's circularity ratio.

A_c is area of a circle with the same perimeter as that of the basin

E is elevation.

a is area between successive contours

e is mean elevation of the catchment.

Σ is summation.

L_g is length of overland flow

D is drainage density.

S_c is channel slope

S_y is average ground slope

L is total length of streams in the basin.

Chapter Two

p is yearly precipitation

h is height in ft

a is ordinates intercept

b is coefficient of regression.

Tr is return period (recurrence interval)

n is number of years of record

Pr is probability

m is the rank of the annual rainfall according to its magnitude

Chapter Three

E is evaporation rate.

K_v is eddy transfer coefficient of water vapour in cm^2/sec

ρ is density of air in gm/cm^3

q is specific humidity of the air

Z is elevation above surface.

H is vertical flux density of sensible heat

K_h is eddy transfer coefficient of sensible heat

C_p is specific heat capacity of the air

T is temperature

τ is vertical flux density of momentum

K_m is eddy transfer coefficient of momentum

U is wind velocity

E is ratio of the densities of water vapour and dry air

e is vapour pressure

P is atmospheric pressure

K is Von Karman constant

R_n is net radiation

L is latent heat of evaporation

E is the energy used for evaporation

S is the energy used for heating the soil

β is Bowen ratio

e_s is saturation vapour pressure at evaporating surface

e_z is the mean actual vapour pressure at the height Z

e_z^o is saturation vapour pressure at height Z

T_s is temperature at soil surface

T_z is temperature at height Z.

R_1 is radiation flux

r is reflection coefficient

R_B is long wave radiation.

R_A is theoretical maximum radiation if there was no atmosphere

n/N is ratio of actual to maximum possible hours of sunshine

σ is Stefan Boltzmann constant.

$P.E^*$ is unadjusted potential evapotranspiration by Thornthwaite formula

P.E is adjusted Thornthwaite potential evapotranspiration

R is runoff.

Et is evapotranspiration.

ΔS is change in soil moisture content.

ΔG is change in the groundwater content

I is irrigation.

D is drainage

<u>Chapters Four and Five</u>

Penman Et is Penman potential evapotranspiration

Penman EO_1 is Penman potential evaporation with the wind speed function of $f(U_2) = 0\ 35\ (1 + U_{2}/100)$

Penman EO_2 is Penman potential evaporation with the wind speed function of $f(U_2) = 0\ 35\ (0\ 5 + U_{2}/100)$

Chapter Six

K_1 is ratio of mean areal rainfall over the catchment to that of raingauge at Durham Observatory.

P_r is probability

Chapter Seven

Q is peak flow.

A is the drainage area

C is a coefficient as a function of land use or topography

n is a constant in empirical formula for estimations of peak rate of runoff

I is rainfall intensity.

W is average infiltration rate

P is precipitation

Q is total amount of runoff from the storm

s is the summation of interception and depression storage

API is antecedent precipitation index

P_n is the precipitation n days before the storm.

A_n is a constant

I_o is the initial value of API index

I is the index values n days later

K is recession constant.

t_p is time lag in hours.

C_t is a coefficient for estimation of time lag of a basin

L is length of main streams from the outlet to the upstream divide

L_{ca} is length in miles along the main stream from the gauging station to a point opposite (nearest) to the centroid.

Q_p Unit hydrograph peak discharge

C_p is a coefficient for estimations of peak discharge.

B is the time base of unit graph

A is the drainage area

d is deficiency in soil moisture.

Chapters Eight, Nine and Ten

K_1 is ratio of average segment rainfall to average gauge rainfall

EXPM is interception storage parameter

UZSN is nominal upper zone storage

LZSN is nominal lower zone storage.

CB is infiltration index

CC is interflow parameter

K3 is actual evapotranspiration loss index

E_p is potential evapotranspiration

r is evapotranspiration opportunity

K24L is index to inactive groundwater recharge.

K24EL is fraction of basin with shallow groundwater

SS is overland flow slope

NN is Manning's n for overland flows

IRC is daily interflow recession rate.

KK24 is daily groundwater discharge recession rate.

KV is parameter for variation of groundwater recession rates

UZS is initial upper zone storage

LZS is initial lower zone storage.

SGW is initial groundwater storage

GWS is initial groundwater slope

RES is initial surface detention storage

SRGX is initial interflow detention storage

APPENDIX III

1 - Daily data of potential evapotranspiration measured by evapotranspirometers at Durham Observatory

2 - Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

3 - Daily data of actual evapotranspiration measured by lysimeters at Durham Observatory

APPENDIX III 1

Daily data of potential evapotranspiration measured by evapotranspirometers at Durham Observatory

Date	Water added cc	Water collected (average) cc	Diff (added-collected) cc	Diff (added-collected) mm	Rainfall mm	Et mm
1/5/73	1500	407	1093	4 46	-	4.46
2/5/73	1500	713	787	3.21	-	3.21
3/5/73	1500	887	613	2.50	3.1	5.60
4/5/73	1500	2078	-578	-2 36	12.1	9 74
5/5/73	1500	3289	-1789	-7.30	-	-7 30
6/5/73	1500	900	600	2 45	0 1	2.55
7/5/73	1500	843	657	2.68	0.8	3.68
8/5/73	1500	885	615	2 51	-	2.51
9/5/73	1500	895	605	2.47	3.4	2.87
10/5/73	1500	1115	385	1.57	1.1	1 67
11/5/73	1500	586	914	3.73	-	3.73
12/5/73	1500	887	613	2 50	0 1	2 60
13/5/73	1500	929	571	2 33	-	2.33
14/5/73	1500	877	623	2.54	-	2 54
15/5/73	1500	855	645	2 63	-	2 63
16/5/73	1500	897	603	2.46	-	2.46
17/5/73	1500	826	674	2.75	-	2.75
18/5/73	1500	659	841	3.43	-	3.43
19/5/73	1500	760	740	3.02	7.8	10 82
20/5/73	1500	1118	382	1.56	5.6	7.16
21/5/73	1500	3446	-1446	-7.94	7.8	-0.14
22/5/73	1500	3029	-1529	-6.24	1.4	-4.84
23/5/73	1500	2059	-559	-2 28	1 3	-0.98
24/5/73	1500	1586	-86	-0.35	2.5	2.15
25/5/73	1500	1730	-230	-0.94	0 1	-0.84
26/5/73	1500	1377	123	0 50	-	0.50
27/5/73	1500	831	669	2 73	-	2.73
28/5/73	1500	686	814	3 32	-	3.32
29/5/73	1500	816	684	2.79	6 7	9.49
30/5/73	1500	1252	248	1 01	0.3	1.31
31/5/73	1500	-1078	-578	-2 36	1.6	-0 76

Daily data of potential evapotranspiration measured by evapotranspirometers at Durham Observatory

Date	Water added cc	Water collected (average) cc	Diff (added-collected) cc	mm	Rainfall mm	Et mm
1/6/73	1500	1404	96	0.39	-	0 39
2/6/73	1500	878	422	1.72	11.2	12.92
3/6/73	1500	372	1128	-4.60	0.3	-4.30
4/6/73	1500	1878	-378	-1.54	-	-1.54
5/6/73	1500	897	603	2.46	-	2 46
6/6/73	1500	760	740	3.02	-	3.02
7/6/73	1500	559	941	3.84	-	3.84
8/6/73	1500	664	836	3.41	-	3.41
9/6/73	1500	586	914	3.73	-	3.73
10/6/73	1500	654	846	3.45	-	3.45
11/6/73	1500	429	1071	4 37	-	4.37
12/6/73	1500	679	821	3.35	0 7	4 05
13/6/73	1500	444	1056	4.31	-	4.31
14/6/73	1500	390	1110	4 53	-	4 53
15/6/73	1500	476	1024	4.18	-	4.18
16/6/73	1500	630	870	3 55	1 8	5.35
17/6/73	1500	838	662	2.70	0 2	2.90
18/6/73	1500	900	600	2.45	-	2.45
19/6/73	1500	635	865	3.53	1 1	4 63
20/6/73	1500	1274	226	0.92	11 2	12.12
21/6/73	1500	3446	-1946	-7.94	0.1	-7.84
22/6/73	1500	1284	216	0.88	-	0.88
23/6/73	1500	836	664	2.71	5.6	8 31
24/6/73	1500	1552	-52	-0.21	1.5	-1.29
25/6/73	1500	1887	-387	-1.58	0.1	-1 48
26/6/73	1500	1527	-27	-0.11	0.4	-0.29
27/6/73	1500	1174	326	1.33	-	1.33
28/6/73	1500	1260	240	0.98	-	0.98
29/6/73	1500	1022	478	1.95	-	1.95
30/6/73	1500	821	679	2 77	-	2.77

Daily data of potential evapotranspiration measured by evapotranspirometers at Durham Observatory

Date	Water added cc	Water collected (average) cc	Diff (added-collected) cc	Diff (added-collected) mm	Rainfall mm	Et mm
1/7/73	1500	762	738	3.01	-	3.01
2/7/73	1500	551	949	3.87	-	3.87
3/7/73	1500	500	1000	4.08	-	4.08
4/7/73	1500	480	1020	4 16	-	4.16
5/7/73	1500	588	912	3.72	-	3.72
6/7/73	1500	711	789	3 22	12 5	15.72
7/7/73	1500	2924	-1424	-5.81	1 4	-4.41
8/7/73	1500	2150	-650	-2.65	-	-2.65
9/7/73	1500	1147	353	1 44	-	1 44
10/7/73	1500	872	628	2.56	1.4	3 96
11/7/73	1500	755	645	3.04	0.1	3.14
12/7/73	1500	1257	243	0.99	3.9	4.89
13/7/73	1500	1179	321	1 31	6.3	7.61
14/7/73	1500	2314	-814	-3.32	6.3	2.98
15/7/73	1500	3220	-1720	-7.02	0.7	-6.32
16/7/73	1500	1833	-333	-1.36	33.7	32.34
17/7/73	1500				1.8	
18/7/73	1500			Flooded	3 5	
19/7/73	1500				2.1	
20/7/73	1500	863	637	2.60	0.5	3.10
21/7/73	1500	973	527	2.15	4.1	6.25
22/7/73	1500	1252	248	1.01	-	1.01
23/7/73	1500	811	689	2.81	-	2.81
24/7/73	1500	760	740	3.02	-	3.02
25/7/73	1500	868	632	2.58	-	2.58
26/7/73	1500	816	684	2.79	-	2.79
27/7/73	1500	615	885	3.61	-	3.61
28/7/73	1500	728	772	3.15	-	3.15
29/7/73	1500	735	765	3.12	-	3 12
30/7/73	1500	836	664	2 71	-	2.71
31/7/73	1500	517	983	4.01	-	4.01

Daily data of potential evapotranspiration measured by evapotranspirometers at Durham Observatory

Date	Water added cc	Water collected (average) cc	Diff(added-collected) cc	Diff(added-collected) mm	Rainfall mm	Et mm
1/8/73	1500	640	860	3.51	1.0	4.51
2/8/73	1500	863	637	2.60	-	2.60
3/8/73	1500	735	765	3.12	0.5	3.62
4/8/73	1500	1002	498	2.03	1.4	3.43
5/8/73	1500	640	860	3.51	25.2	28.71
6/8/73	1500	1787	-287	-1.17	2.9	1.73
7/8/73	1500	2782	-1282	-5.23	1.0	-4.23
8/8/73	1500	2174	-674	-2.75	1.6	-1.15
9/8/73	1500	1069	431	1.76	-	1.76
10/8/73	1500	1370	130	0.53	2.9	3.43
11/8/73	1500	1238	262	1.07	-	1.07
12/8/73	1500	1672	-172	-0.70	-	-0.70
13/8/73	1500	1056	444	1.81	-	1.81
14/8/73	1500	973	527	2.15	-	2.15
15/8/73	1500	811	689	2.81	-	2.81
16/8/73	1500	730	770	3.14	-	3.14
17/8/73	1500	716	784	3.20	-	3.20
18/8/73	1500	559	941	3.84	1.0	4.84
19/8/73	1500	620	880	3.59	22.2	25.79
20/8/73	1500	3995	-2495	-10.18	3.2	-6.98
21/8/73	1500	3508	-2008	-8.19	-	-8.19
22/8/73	1500	2382	-882	-3.60	-	-3.60
23/8/73	1500	1118	382	1.56	-	1.56
24/8/73	1500	897	603	2.46	-	2.46
25/8/73	1500	1022	478	1.95	-	1.95
26/8/73	1500	1184	316	1.29	-	1.29
27/8/73	1500	1157	343	1.40	-	1.40
28/8/73	1500	1108	392	1.60	-	1.60
29/8/73	1500	-	-	-	2.5	2.50
30/8/73	1500	1319	181	-0.74	0.7	-0.04
31/8/73	1500	1179	321	1.31	0.1	1.41

Daily data of potential evapotranspiration measured by evapotranspirometers at Durham Observatory

Date	Water added cc	Water collected (average) cc	Diff(added-collected) cc	Diff(added-collected) mm	Rainfall mm	Et mm
1/9/73	1500	1152	348	1.42	-	1.42
2/9/73	1500	973	527	2.15	-	2.15
3/9/73	1500	961	539	2.20	1 0	3.20
4/9/73	1500	1260	240	0.98	-	0 98
5/9/73	1500	1203	297	1.21	-	1 21
6/9/73	1500	1047	453	1.85	-	1 85
7/9/73	1500	983	517	2.11	-	2.11
8/9/73	1500	-	-	-	-	-
9/9/73	1500	-	-	-	1.2	1.20
10/9/73	1500	966	534	2.18	-	2.18
11/9/73	1500	422	1078	4.40	-	4.40
12/9/73	1500	912	588	2 40	0.2	2.60
13/9/73	1500	993	507	2.07	0.5	2.57
14/9/73	1500	1213	287	1.17	-	1 17
15/9/73	1500	1434	66	0.27	2.1	2 37
16/9/73	1500	1265	235	0.96	0.1	1.06
17/9/73	1500	1537	-37	-0.15	2.0	1.85
18/9/73	1500	1686	-186	-0.76	11 4	10.64
19/9/73	1500	2198	-698	-2.85	0.3	-2 53
20/9/73	1500	2505	-1005	-4.10	-	-4.10
21/9/73	1500	2145	-145	-0 59	-	-0.59
22/9/73	1500	1223	277	1.13	16.3	17.43
23/9/73	1500	3002	-1502	-6.13	-	-6.13
24/9/73	1500	2576	-1076	-4.39	-	-4.39
25/9/73	1500	1841	-341	-1.39	1 6	-0 21
26/9/73	1500	1490	10	0.04	0.5	0 54
27/9/73	1500	1557	-257	-1.05	5.9	4.85
28/9/73	1500	1804	-304	-1.24	0.1	-1.14
29/9/73	1500	1547	-47	-0.19	5.5	5.31
30/9/73	1500	1983	-483	-1.97	0.1	-1 87

APPENDIX III 2.

Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

Date	Water added	Water collected (average)	Diff(added-collected)		Rainfall	Et
	cc	cc	cc	mm	mm	mm
14/7/73	1500	2358	-858	-3.50	12.0	8.50
15/7/73	1500	2750	-1250	-5.10	7.8	2.70
16/7/73	1500	2750	-1250	-5 10	2.0	-3 10
17/7/73	1500	3976	-2476	-10.10	22.2	12.10
18/7/73	1500	2848	-1348	-5.50	1.0	-4.50
19/7/73	1500	1402	98	0.40	2 8	3.20
20/7/73	1500	1647	-147	-0 60	0 5	-0.10
21/7/73	1500	789	711	2 90	-	2.90
22/7/73	1500	-	-	-	7.3	7 20
23/7/73	1500	-	-	-	0.2	0.20
24/7/73	1500	1132	368	1.50	-	-1.50
25/7/73	1500	691	809	3 30	-	3.30
26/7/73	1500	667	833	3.40	-	3.40
27/7/73	1500	716	784	3.20	-	3.20
28/7/73	1500	618	882	3.60	-	3.60
29/7/73	1500	-	-	-	-	-
30/7/73	1500	1353	147	0.60	3.25	3 85
31/7/73	1500	544	956	3.90	-	3.90

Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

Date	Water added	Water collected (average)	Diff(added-collected)		Rainfall	Et
	cc	cc	cc	mm	mm	mm
1/8/73	1500	838	662	2.7	-	2.70
2/8/73	1500	642	858	3.5	-	3.50
3/8/73	1500	1353	147	0 6	2.20	2.80
4/8/73	1500	-	-		2.00	2.00
5/8/73	1500	-	-	-	1.50	1.50
6/8/73	1500	4392	-2892	-11.8	35.00	23.20
7/8/73	1500	4122	-2622	-10 7	2.00	-8.70
8/8/73	1500	1426	74	0.3	1 80	2.10
9/8/73	1500	838	662	2 7	0 80	3 50
10/8/73	1500	1083	417	1.7	0.50	2.20
11/8/73	1500	-	-	-	13.50	13.50
12/8/73	1500	-	-	-	-	-
13/8/73	1500	2775	-1275	-5 2	-	-5 20
14/8/73	1500	642	858	3.5	-	3.50
15/8/73	1500	544	956	3 9	-	3.90
16/8/73	1500	301	1299	5.3	-	5.30
17/8/73	1500	79	1421	5.8	-	5.80
18/8/73	1500	-	-	-	-	-
19/8/73	1500	-	-	-	1.80	1.80
20/8/73	1500	3412	-1912	-7.8	24.00	16.20
21/8/73	1500	3927	-2427	-9.9	4.00	-5.90
22/8/73	1500	4932	-3432	-1.4	-	-1.40
23/8/73	1500	789	711	2 9	-	2.90
24/8/73	1500	887	613	2.5	-	2.50
25/8/73	1500	-	-	-	-	-
26/8/73	1500	-	-	-	-	-
27/8/73	1500	-	-	-	-	-
28/8/73	1500	887	613	2.5	-	2.50
29/8/73	1500	642	858	3.5	-	3.50
30/8/73	1500	1108	392	1 6	6.25	7 85
31/8/73	1500	1770	-270	-1.1	2.25	1.15

Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

Date	Water added cc	Water collected (average) cc	Diff.(added-collected) cc	mm	Rainfall mm	Et mm
1/9/73	1500	1157	343	1.40	1.5	2.90
2/9/73	1500	-	-	-	-	-
3/9/73	1500	1010	490	2 0	-	2 00
4/9/73	1500	544	956	3 9	0 5	4 40
5/9/73	1500	936	564	2 3	-	2.30
6/9/73	1500	593	907	3.7	-	3.70
7/9/73	1500	838	662	2.7	-	2.70
8/9/73	1500	-	-	-	-	-
9/9/73	1500	-	-	-	-	-
10/9/73	1500	1083	417	1.7	2 25	3.95
11/9/73	1500	961	539	2.2	-	2.20
12/9/73	1500	1059	441	1.8	-	1 80
13/9/73	1500	1083	417	1 7	-	1.70
14/9/73	1500	1255	245	1.0	0 80	1.80
15/9/73	1500	-	-	-	-	-
16/9/73	1500	-	-	-	0 80	0.80
17/9/73	1500	-	-	-	-	-
18/9/73	1500	1377	123	0.5	3.80	4.30
19/9/73	1500	3780	-2280	-9 3	14 80	5.50
20/9/73	1500	2480	-980	-4.0	0.5	-3.50
21/9/73	1500	1010	490	2.0	-	2.00
22/9/73	1500	1181	319	1 3	1.25	2.55
23/9/73	1500	-	-	-	8.25	8.25
24/9/73	1500	3436	-1936	-7.9	1.00	-6.90
25/9/73	1500	1206	294	1 2	-	1.20
26/9/73	1500	1304	196	0.8	1 25	2 03
27/9/73	1500	1255	245	1.0	-	1.00
28/9/73	1500	2995	-1495	-6 1	9.80	3.70
29/9/73	1500	-	-	-	1 25	1.25
30/9/73	1500	-	-	-	8.25	8.25

Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

Date	Water added	Water collected (average)	Diff (added-collected)		Rainfall	Et
	cc	cc	cc	mm	mm	mm
1/10/73	1500	2995	-1495	-6.1	-	-6.10
2/10/73	1500	814	686	2 8	-	2.80
3/10/73	1500	1230	270	1.1	-	1.10
4/10/73	1500	1255	245	0.1	1.25	1 40
5/10/73	1500	1868	-368	-0.15	-	-0.15
6/10/73	1500	-	-	-	-	0
7/10/73	1500	-	-	-	0 50	0.50
8/10/73	1500	-	-	-	0 25	0.25
9/10/73	1500	1581	-81	-0.33	4 60	4 27
10/10/73	1500	2186	-686	-2 80	4.10	1 28
11/10/73	1500	2961	-1461	-5.96	6.40	0.44
12/10/73	1500	1431	69	0 28	-	0.28
13/10/73	1500	-	-	-	-	-
14/10/73	1500	-	-	-	-	-
15/10/73	1500	2797	-1297	-5 29	8 40	3 11
16/10/73	1500	1990	-490	-2.00	-	-2.00
17/10/73	1500	1990	-490	-2.00	4.10	+2 10
18/10/73	1500	1650	-150	-0 61	-	-0.61
19/10/73	1500	2385	-885	-3.61	6.80	3.19
20/10/73	1500	-	-	-	7.60	7.60
21/10/73	1500	-	-	-	1.80	1 80
22/10/73	1500	3559	-2059	-8.40	2 80	-5.60
23/10/73	1500	2039	-539	-2 20	0.50	-1.70
24/10/73	1500	544	956	-0.39	0.25	0.64
25/10/73	1500	1309	191	0 78	-	0 78
26/10/73	1500	1309	191	0.78	-	0.78
27/10/73	1500	-	-	-	-	-
28/10/73	1500	-	-	-	-	-
29/10/73	1500	1495	5	0.02	2.80	2.82
30/10/73	1500	1853	-353	-1 44	1	-1.44
31/10/73	1500	1404	96	0.39	-	0 39

Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

Date	Water added	Water collected (average)	Diff (added-collected)		Rainfall	Et
	cc	cc	cc	mm	mm	mm
1/11/73	1500	1309	191	0 78	1 00	1 78
2/11/73	1500	1650	-150	-0 61	-	-0 61
3/11/73	1500	-	-	-	-	-
4/11/73	1500	-	-	-	0 50	0 50
5/11/73	1500	1760	-260	-1 06	1 30	0 24
6/11/73	1500	1370	130	0 53	-	0.53
7/11/73	1500	1287	213	0.87	-	0 87
8/11/73	1500	1321	179	0.73	-	0 73
9/11/73	1500	1404	96	0 39	3 60	3.99
10/11/73	1500	-	-	-	10 70	10 70
11/11/73	1500	-	-		0 70	0 70
12/11/73	1500	4147	-2647	-10 80	2 30	-8 50
13/11/73	1500	2140	-640	-2.61	2 50	-0 11
14/11/73	1500	2064	-564	-0 23	0 50	0.27
15/11/73	1500	1309	191	0.78	0 25	1 03
16/11/73	1500	1341	159	0.65	0 50	1 15
17/11/73	1500	-	-	-	-	-
18/11/73	1500	-	-	-	-	-
19/11/73	1500	1458	42	0.17	1 00	1 17
20/11/73	1500	1429	71	0 29	-	0 29
21/11/73	1500	1431	69	0.28	-	0 28
22/11/73	1500	1458	42	0.17	-	0.17
23/11/73	1500	1429	71	0 29	-	0 29
24/11/73	1500				.25	
25/11/73	1500				.25	
26/11/73	1500				-	
27/11/73	1500	No measurement			-	
28/11/73	1500				-	
29/11/73	1500				4.06	
30/11/73	1500				-	

Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

Date	Water added	Water collected (average)	Diff (added-collected)		Rainfall	Et
	cc	cc	cc	mm	mm	mm
1/12/73	1500	-	-	-	0 25	0.25
2/12/73	1500	-	-	-	0 76	0.76
3/12/73	1500	1377	123	0 50	1 01	1 51
4/12/73	1500	1287	213	0.87	-	0.87
5/12/73	1500	1341	159	0 65	-	0 65
6/12/73	1500	1652	-152	-0.62	1 27	0.65
7/12/73	1500	1375	125	0.51	-	0 51
8/12/73	1500				2.79	
9/12/73					-	
10/12/73					-	
11/12/73		No measurements			0.76	
12/12/73					-	
13/12/73					5 1	
14/12/73					-	
15/12/73	1500	-	-	-	-	-
16/12/73	1500	-	-	-	2 0	2.0
17/12/73	1500	1718	-218	-0.89	-	-0.89
18/12/73	1500	1294	206	0 84	-	0 84
19/12/73	1500	1348	152	0 62	-	0 62
20/12/73	1500	2277	-777	-3.17	6 6	3 4
21/12/73	1500	1716	-216	-8 80	9 1	0 3
22/12/73	1500	1745	-245	-1 0	1 8	0 8
23/12/73	1500	-	-	-	4.6	4.6
24/12/73	1500	2456	-956	-3.90	0 25	-3.6
25/12/73	1500	0	0	0	-	0
26/12/73	1500	-	-	-	0 25	0
27/12/73	1500	1458	42	0.17	-	0
28/12/73	1500	1409	-91	-0.37	0 76	0
29/12/73	1500	-	-	-	0 50	0
30/12/73	1500	-	-	-	1 52	1
31/12/73	1500	1882	-382	-1.56	-	-1

Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

Date	Water added cc	Water collected (average) cc	Diff (added-collected) cc	Diff (added-collected) mm	Rainfall mm	Et mm
1/1/74	1500	-	-	-	-	-
2/1/74	1500	-	-	-	-	-
3/1/74	1500	1341	159	0 65	-	0 65
4/1/74	1500	1279	221	0 90	1 52	2.42
5/1/74	1500	-	-	-	8 89	8 89
6/1/74	1500	-	-	-	2 54	2 54
7/1/74	1500	4392	-2892	-11 80	3.30	-8 50
8/1/74	1500	2581	-1081	-4.41	8 12	3 71
9/1/74	1500	3561	-2061	-8 41	12 40	3 99
10/1/74	1500	4100	-2600	-10 61	-	-10.61
11/1/74	1500	2211	-711	-2 90	8.60	5.70
12/1/74	1500	-	-	-	4 31	4.31
13/1/74	1500	-	-	-	5 84	5.84
14/1/74	1500	4147	-2647	-10 80	2 54	-8 26
15/1/74	1500	3927	-2427	-9 90	7 11	-2 79
16/1/74	1500	3559	-2059	-8 40	8 89	0.49
17/1/74	1500	2309	-809	-3.30	6 60	3 30
18/1/74	1500	765	735	-3 00	0 50	-2 50
19/1/74	1500	-	-	-	.25	0 25
20/1/74	1500	-	-	-	-	-
21/1/74	1500	-	-	-	-	-
22/1/74	1500	1392	108	0.44	-	0 44
23/1/74	1500	1409	91	0 37	0 25	0 62
24/1/74	1500	1404	96	0.39	2 28	2 67
25/1/74	1500	1770	-270	-1.10	-	-1 1
26/1/74	1500	-	-	-	-	-
27/1/74	1500	-	-	-	5 84	5 84
28/1/74	1500	3951	-2451	-10.0	15 24	5 24
29/1/74	1500	3669	-2169	-8 85	5 33	-3.52
30/1/74	1500	4147	-2647	-10 80	11 43	1 63
31/1/74	1500	3030	-1530	-6 24	5 33	-0 91

Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

Date	Water added	Water collected (average)	Diff (added-collected)		Rainfall	Et
	cc	cc	cc	mm	mm	mm
1/2/74	1500	765	735	-3.00	0.25	-2.75
2/2/74	1500	-	-	-	10.66	10.66
3/2/74	1500	2755	-1255	-5.12	5.58	-0.46
4/2/74	1500	3956	-2456	-10.02	-	-10.02
5/2/74	1500	1458	42	17	0.25	0.42
6/2/74	1500	1936	-436	-1.78	8.38	6.60
7/2/74	1500	2770	-1220	-4.98	1.52	-3.46
8/2/74	1500	2122	-622	-2.54	-	-2.54
9/2/74	1500	-	-	-	6.60	6.60
10/2/74	1500	-	-	-	3.60	3.60
11/2/74	1500	3794	-2294	-9.36	2.50	-6.86
12/2/74	1500	2319	-819	-3.34	5.10	1.76
13/2/74	1500	2086	-586	-2.39	1.80	-0.59
14/2/74	1500	1451	49	0.20	-	0.20
15/2/74	1500	1787	-287	-1.17	5.10	3.93
16/2/74	1500	-	-	-	0.50	0.50
17/2/74	1500	-	-	-	6.60	6.60
18/2/74	1500	3794	-2294	-9.36	1.80	-7.56
19/2/74	1500	1723	-223	-0.91	1.00	0.09
20/2/74	1500	1740	-240	-0.98	0.80	-0.18
21/2/74	1500	1596	-96	-0.39	1.50	1.11
22/2/74	1500	1422	78	0.32	2.00	2.32
23/2/74	1500	-	-	-	-	-
24/2/74	1500	-	-	-	-	-
25/2/74	1500	1505	-5	-0.02	-	-0.20
26/2/74	1500	1314	186	.76	-	0.76
27/2/74	1500	1213	287	1.17	-	1.17
28/2/74	1500	1404	96	39	-	0.39

Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

Date	Water added	Water collected (average)	Diff (added-collected)		Rainfall	Et
	cc	cc	cc	mm	mm	mm
1/3/74	1500	1458	42	0 17	7 4	7 57
2/3/74	1500	-	-	-	0 3	0 30
3/3/74	1500	-	-	-		-
4/3/74	1500	2074	-574	-2 34	9 4	7 06
5/3/74	1500	2652	-1152	-4 70	6 8	2 10
6/3/74	1500	2838	-1338	-5 46	-	-5.46
7/3/74	1500	1887	-387	-1 58	2 5	0.92
8/3/74	1500	1549	-49	-0 20	-	0 02
9/3/74	1500	-	-	-	-	-
10/3/74	1500	-	-	-	1 0	1 00
11/3/74	1500	1596	-96	-0 39	1 0	0 61
12/3/74	1500	1642	-142	-0 58	8 1	7 52
13/3/74	1500	2863	-1363	-5.56	4.3	-1 26
14/3/74	1500	2887	-1387	-5.66	6 4	0 74
15/3/74	1500	1740	-240	-0 98	1 8	0 82
16/3/74	1500	-	-	-	3 3	3 30
17/3/74	1500	-	-	-	0 8	0 80
18/3/74	1500	2647	-1147	-4 68	6 1	1 42
19/3/74	1500	1642	-142	-0 58	0.5	-0 08
20/3/74	1500	1549	-49	-0 20	2.5	2 30
21/3/74	1500	2530	-1030	-4.20	6 4	2 20
22/3/74	1500	1549	-49	-0 20	-	-0 20
23/3/74	1500	-	-	-	-	-
24/3/74	1500	-	-	-	-	-
25/3/74	1500	1451	49	0.20	1 8	2 0
26/3/74	1500	1451	49	0 20	1 3	1 50
27/3/74	1500	1404	96	0 39	-	0 39
28/3/74	1500	1159	341	1 39	-	1 39
29/3/74	1500	1083	417	1 70	-	1.70
30/3/74	1500	-	-	-	-	-
31/3/74	1500	-	-	-	-	-

Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

Date	Water added	Water collected (average)	Diff (added-collected)		Rainfall	Et
	cc	cc	cc	mm	mm	mm
1/4/74	1500	1260	240	0.98	-	0 98
2/4/74	1500	686	814	3.32	-	3.32
3/4/74	1500	1132	368	1 50	-	1 50
4/4/74	1500	1071	429	1 75	-	1 75
5/4/74	1500	1118	382	1.56	-	1 56
6/4/74	1500	1181	319	1 30	-	1 30
7/4/74	1500	1181	319	1 30	-	1 30
8/4/74	1500	1218	282	1 15	-	1 15
9/4/74	1500	1248	252	1 03	-	1 03
10/4/74	1500	980	520	2 12	-	2 12
11/4/74	1500	860	640	2 61	9 4	12 01
12/4/74	1500	1382	-882	-3 60	1 3	-2 30
13/4/74	1500	-	-	-	-	-
14/4/74	1500	-	-	-	-	-
15/4/74	1500	-	-	-	-	-
16/4/74	1500	2694	-1194	-4.87	-	-4.87
17/4/74	1500	689	811	3 31	0 5	3.81
18/4/74	1500	1108	392	1 60	-	1 60
19/4/74	1500	1081	419	1 71	-	1 71
20/4/74	1500	1196	304	1 24	-	1 24
21/4/74	1500	1132	368	1 50	-	1 50
22/4/74	1500	1076	424	1 73	-	1 73
23/4/74	1500	1071	429	1 75	-	1.75
24/4/74	1500	926	574	2.34	-	2.34
25/4/74	1500	1020	480	1 96	-	1 96
26/4/74	1500	1113	387	1 58	-	1 58
27/4/74	1500	1081	419	1 71	-	1 71
28/4/74	1500	1676	-176	-0 72	7 5	6 78
29/4/74	1500	4081	-2581	-10 53	12 5	1 97
30/4/74	1500	1672	-172	-0 70	-	0 70

Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

Date	Water added	Water collected (average)	Diff (added-collected)		Rainfall	Et
	cc	cc	cc	mm	mm	mm
1/5/74	1500	1118	382	1.56	-	1 56
2/5/74	1500	1213	287	1 17	-	1 17
3/5/74	1500	1118	382	1 56	2 50	4 06
4/5/74	1500	-	-	-	1.00	1 00
5/5/74	1500	640	860	3.51	0.25	3 76
6/5/74	1500	1093	407	1 66	2 50	4 16
7/5/74	1500	1333	167	0.68	-	0 68
8/5/74	1500	1022	478	1 95	-	1 95
9/5/74	1500	806	694	2.83	-	2 83
10/5/74	1500	1309	191	0 78	10.00	10 78
11/5/74	1500	-	-	-	2 30	2 30
12/5/74	1500	1902	-402	-1 64	-	-1.64
13/5/74	1500	902	598	2.44	2 50	4 94
14/5/74	1500	1333	167	0 68	0 70	1 38
15/5/74	1500	711	789	3 22	-	3 22
16/5/74	1500	711	789	3 22	-	3 22
17/5/74	1500	811	689	2 81	-	2 81
18/5/74	1500	946	554	2 26	-	2.26
19/5/74	1500	1152	348	1 42	3 80	5 22
20/5/74	1500	1213	237	1 17	-	1.17
21/5/74	1500	787	713	2 91	-	2 91
22/5/74	1500	902	598	2 44	0 30	2.74
23/5/74	1500	877	623	2.54	0 30	0 84
24/5/74	1500	3699	-2199	-8 97	19.60	10 63
25/5/74	1500	1625	-125	-0 51	0 30	0 21
26/5/74	1500	998	502	2.05	-	2 05
27/5/74	1500	644	856	3.49	-	3 49
28/5/74	1500	500	1000	4.08	-	4 08
29/5/74	1500	740	760	3 10	-	3 10
30/5/74	1500	569	931	3 80	-	3.80
31/5/74	1500	564	936	3.82	-	3.82

Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

Date	Water added cc	Water collected (average) cc	Diff (added-collected) cc	Diff (added-collected) mm	Rainfall mm	Et mm
1/6/74	1500	711	789	3 22	-	3.22
2/6/74	1500	529	971	3 96	0 5	4 46
3/6/74	1500	640	860	3 51	0.3	3 81
4/6/74	1500	789	711	2 90	-	2 90
5/6/74	1500	691	809	3 30	-	3 30
6/6/74	1500	1328	172	0 70	6 6	7 30
7/6/74	1500	1647	-147	-0 60	1 3	0 70
8/6/74	1500	1819	-319	-1 30	1 3	0
9/6/74	1500	-	-	-	9 1	9 10
10/6/74	1500	2505	-1005	-4 10	7 9	3 80
11/6/74	1500	1328	172	0 70	-	0 70
12/6/74	1500	887	613	2 50	-	2 50
13/6/74	1500	765	735	3.00	-	3 00
14/6/74	1500	716	784	3 20	-	3 20
15/6/74	1500	544	956	3 90	-	3.90
16/6/74	1500	176	1324	5 40	0.3	5 70
17/6/74	1500	1059	441	1 80	7 4	9 20
18/6/74	1500	1598	-98	-0 40	0 3	-0 10
19/6/74	1500	593	907	3 70	-	3.70
20/6/74	1500	470	1030	4.20	-	4.20
21/6/74	1500	372	1128	4.60	-	4 60
22/6/74	1500	470	1030	4.20	-	4 20
23/6/74	1500	765	735	3.00	-	3 00
24/6/74	1500	765	735	3 00	-	3.00
25/6/74	1500	1010	490	2 00	-	2 00
26/6/74	1500	1010	490	2.00	0 80	2 80
27/6/74	1500	1402	98	0 40	1.00	1 40
28/6/74	1500	985	515	2.10	-	2 10
29/6/74	1500	887	613	2 50	-	2 50
30/6/74	1500	789	711	2 90	-	2 90

Daily data of potential evapotranspiration measured by evapotranspirometers at Honey Hill

Date	Water added	Water collected (average)	Diff (added-collected)		Rainfall	Et
	cc	cc	cc	mm	mm	mm
1/7/74	1500	1282	218	0 89	7.62	8.51
2/7/74	1500	2049	-549	-2.24	-	-2.24
3/7/74	1500	3645	-2145	-8.75	22.1	13 35
4/7/74	1500	2456	-956	-3.90	2 0	-1.90
5/7/74	1500	2792	-1292	-5 27	14 0	8 73
6/7/74	1500	3647	-2147	-8 76	8 1	-0.66
7/7/74	1500	-	-	-	0 3	0.30
8/7/74	1500	2407	-907	-3.70	-	-3.70
9/7/74	1500	419	1081	4 41	-	4 41
10/7/74	1500	593	907	3.70	-	3 70
11/7/74	1500	1164	336	1 37	4 63	6.20
12/7/74	1500	973	527	2 15	1 02	3.17
13/7/74	1500	542	958	3.91	0.30	4.21

APPENDIX III 3

Daily data of actual evapotranspiration measured by lysimeters at Durham Observatory

Date	lysimeter One			lysimeter Two			Average Depth mm	Rainfall mm
	Reading	Diff in readings	Depth mm	Reading	Diff in readings	Depth mm		
15/5/73	10.1	-0 1	-0 7	6 4	-0 4	-2.7	-1.7	-
16/5/73	10 0	-0 2	-1 3	6 0	-0 2	-1 3	-1 3	-
17/5/73	9.8	-0 3	-2 0	5 8	-0 2	-1 3	-1 6	-
18/5/73	9.5	-0 1	-0.7	5.6	-0.1	-0 7	-0 7	-
19/5/73	9.4	+1 1	+7.3	5 5	+1 5	+10 0	+8 6	-
20/5/73	10.5	+0 7	+4 7	7 0	+0 6	+4 0	+4 4	7 8
21/5/73	11.2	+1 0	+6 7	7 6	+0 6	+4 0	+5 4	5 6
22/5/73	12 2	+0.2	+1 3	8 2	+0 3	+2.0	+1 6	7.8
23/5/73	12.4	+0 2	+1 3	8 5	+0 3	+2 0	+1 6	1 4
24/5/73	12 6	+0 1	+0.7	8 8	+0 1	+0 7	+0 7	1.3
25/5/73	12 7	-0 2	-1 3	8 9	-0 1	-0.6	-1	2 5
26/5/73	12.5	-0 3	-2 0	8 8	-0 3	-2 0	-2	0 1
27/5/73	12.2	-0 4	-2 7	8 5	-0 6	-4 0	-3.4	-
28/5/73	11 8	-0 2	-1 3	7.9	-0 2	-1 3	-1 3	-
29/5/73	11.6	+0 6	+4 0	7 7	+0 5	+3.4	+3.7	-
30/5/73	12.2	-0.1	-0 7	8 2	0	+0	-0 4	6 7
31/5/73	12.1	-0.1	-0.7	8 2	-0 2	-1.3	-1 0	0 3
1/6/73	12.0			8 0				1 6

Daily data of actual evapotranspiration measured by lysimeters at Durham Observatory

Date	Lysimeter One Reading	Lysimeter One Diff in readings	Lysimeter One Depth mm	Lysimeter Two Reading	Lysimeter Two Diff in readings	Lysimeter Two Depth mm	Average Depth mm	Rainfall m
1/6/73	12 0			8.0				1 6
2/6/73	-	+1 4	+9.3	-	+1 1	+7.3	+8.3	-
3/6/73	13 4	-0.1	-0 7	9 1	-0 3	-2	-1.4	11 2
4/6/73	13 3	-0 3	-2	8 8	-0 5	-3 4	-2 7	0.3
5/6/73	13.0	-0 6	4	8 3	-0 7	-4 7	-4 4	-
6/6/73	12 4	-0 3	-2	7.6	-0.3	-2	-2.0	-
7/6/73	12 1	-0 6	-4	7 3	no measurement		-4	-
8/6/73	11 5	-0 6	-4	-			-4	-
9/6/73	10 9	-0 5	-3 4	13 5	-0 4	-2 7	-3	-
10/6/73	10 4			13 1				-
11/6/73	-	-0 9	-6	-	-0 6	-4	-5	-
12/6/73	9 5	-0 7	-4 7	12 5	-0 5	-3 4	-4	-
13/6/73	8 8	-0 5	-3 7	12 0	-0 2	-1 3	-2 5	0.7
14/6/73	8 3	-0 5	-3 7	11 8	-0 4	-2 7	-3 2	-
15/6/73	7.8			11 4				-
16/6/73	-	-1 0	-6.7	-	-0 8	-5.4	-6	-
17/6/73	6 8			10.6				1 8
18/6/73	-	-1 0	-6 7	-	-0 6	-4	-5 4	0.2
19/6/73	5 8	+1 1	+7.3	10 0	+1 4	+9 3	+8.3	-
20/6/73	6.9	+0 5	+3.4	11.4	+0.5	+3 4	+3 4	9 1
21/6/73	7.4	-0.5	-3.4	11 9	-0 4	-2 7	-3.0	4.2
22/6/73	6.9	-0 7	-4 7	11 5	-0 8	-5.4	-5.0	0.1
23/6/73	6.2	+0.6	+4	10 7	+0 6	+4	+4 0	-
24/6/73	6 8	+0.3	+2	11 3	+0 3	+2	+2 0	5 6
25/6/73	7 1	-0 3	-2	11.6	-0 1	-0 7	-1 3	1 5
26/6/73	6.8			11 5				0 1
27/6/73	-	-0 8	-5 4	-	-0 7	-4 7	-5 0	0.4
28/6/73	6.0	-0 6	-4 0	10 8	-0 6	-4	-4.0	-
29/6/73	5 4	-0 3	-2 0	10 2	-0 7	-4 7	-3 4	-
30/6/73	5 1			9 5				-

Daily data of actual evapotranspiration measured by lysimeters at Durham Observatory

Date	Lysimeter One Reading	Lysimeter One Diff in readings	Lysimeter One Depth mm	Lysimeter Two Reading	Lysimeter Two Diff in readings	Lysimeter Two Depth mm	Average Depth mm	Rainfall mm
30/6/73	5 1			9 5		-11 7	-10.2	-
1/7/73	-	-1.3	-8 7		-1 7			-
2/7/73	3 8			7.8				-
		-0 4	-2.7		-0 6	-4	-3 4	
3/7/73	3 4			7 2				-
		-0 4	-2.7		-0 8	-5 4	-4 0	
4/7/73	3 0			6 4				-
		-0 6	-4 0		-0 8	-5.4	-4 7	
5/7/73	2 4			5.6				12 5
		-0 2	-1 3		-0 5	-3 4	-2 4	
6/7/73	2.2			5.1				1 4
7/7/73	-	+1 7	+11 7		+1.5	+10 0	+10 8	-
8/7/73	3.9			6.6				-
9/7/73	-	-1.0	-6 7	-	-1 0	-6 7	-6 7	1 4
10/7/73	2.9			5 6				0 1
		-0 2	-1 3		-0 3	-2	-1 6	
11/7/73	2 7			5.3				3 9
		-0 2	-1.3		-0 1	-0 7	-1 0	
12/7/73	2 5			5 2				6.3
13/7/73	-	+1 0	+6 7		+0 7	+4 7	+5.7	6 3
14/7/73	3 5			5 9				0.7
		+1 3	+8.7		+1 3	+8 7	+8 7	
15/7/73	4 8			7 2				33 0
		-0 7	-4 7		-0 5	-3 4	-4 0	
16/7/73	4.1			6 7				1 8
		+4.9	+3 3		+4 5	+30 0	+31.5	
17/7/73	9 0			11 2				3.4
		-0.1	-0 6		+0.1	+0 7	0	
18/7/73	8.9			11 3				2.0
		+0 3	+2		+0 3	+2	+2	
19/7/73	9 2			11 6				0 5
		-0.3	-2		0	0	-1	
20/7/73	8 9			11 6				4.1
		-0 2	-1 3		-0 1	-0 7	-1	
21/7/73	8.7			11 5				-
		-0 1	-0 7		+0 2	+1 4	+0 4	
22/7/73	8 6			11 7				-
		-0 1	-0 7		+0 2	+1 4	+0 4	
23/7/73	8.5			11 9				-
		+0.1	+0 7		-0 1	-0 7	0	
24/7/73	8 6			11 8				-
		-0 4	-2 6		-0 8	-5 4	-4	
25/7/73	8 2			11 0				-
		-0.7	-4 6		-0.5	-3 4	-4	
26/7/73	7.5			10 5				-
		-0 1	-0 7		-0.1	-0 7	-0.7	
27/7/73	7.4			10 4				-
		+0 2	+1 3		-0 1	-0 7	+0.4	
28/7/73	7.6			10 3				-
		-0.5	-3 4		-0 4	-2 6	-3 0	
29/7/73	7.1			9 9				-
		-0.5	-3 4		-0 3	-2	-2 7	
30/7/73	6 6			9 6				-
		-0 8	-5 4		-0 6	-4	-4 7	
31/7/73	5.8			9 0				-

Daily data of actual evapotranspiration measured by lysimeters at Durham Observatory

Date	lysimeter One: Reading	lysimeter One: Diff in readings	lysimeter One: Depth mm	lsyimeter Two: Reading	lsyimeter Two: Diff in readings	lsyimeter Two: Depth mm	Average Depth mm	Rainfall mm
31/7/73	5 8			9 0				1 0
		-0 3	-2.0		-0.8	-5 4	-3 7	
1/8/73	5.5			8 2				-
		-1 0	-6 7		-0 5	-3.4	-5 0	
2/8/73	4 5			7 7				0.5
		-0 6	-4 0		-0 4	-2 7	-3 4	
3/8/73	3 9			7.3				1 4
		-0 5	-3 4		-0 3	-2 0	-2 7	
4/8/73	3 4			7 0				25 2
		-0 5	-3 4		-0.6	-4 0	-3 7	
5/8/73	2 9			6 4				2 9
		+3	+20.0		+3 8	+25 3	+22.7	
6/8/73	5 9			10 2				1 0
		-0.6	-4.0		-0 4	-2 7	-3.4	
7/8/73	5 3			9 8				1 6
		-0 8	-5.4		-0 5	-3.4	-4 4	
8/8/73	4 5			9 3				-
		-0 6	-4.0		-0 6	-4 0	-4 0	
9/8/73	3 9			8.7				2 9
		-0 4	-2 7		-0 4	-2.7	-2 7	
10/8/73	3 5			8 3				-
		-0 5	-3.4		-0.7	-4 7	-4 0	
11/8/73	3			7 6				-
		-0 5	-3 4		-0 6	-4 0	-3.7	
12/8/73	2 5			7 0				-
		-0 2	-1 4		-0 8	-5.4	-3.4	
13/8/73	2 3			6.2				-
		-0 8	-5 4		-0 7	-4 7	-5 0	
14/8/73	1 5			5 5				-
		-0 3	-2 0		-0 9	-6 0	-4.0	
15/8/73	1 2			4 6				-
		-0.6	-4 0		-0 5	-3 4	-3 7	
16/8/73	0 6			4.1				-
		-0 8	-5 4		-0 6	-4.0	-4 7	
17/8/73	-0 2			3 5				-
		-0.5	-3 4		-0 5	-3 4	-3 4	
18/8/73	7.8			3.0				1 0
		-0 3	-2 0		-0 7	-4.7	-3.4	
19/8/73	7 5			2 3				22.2
		+3.7	+24.7		+3 4	+22 7	+23.7	
20/8/73	11 2			5.7				3 2
		+0.5	+3.4		+0 8	+5.4	+4 4	
21/8/73	11 7			6 5				-
		-0.3	-2		-0 7	-4.7	-3 4	
22/8/73	11.4			5.8				-
		-0 1	-0 7		-0 4	-2 7	-1 6	
23/8/73	11.3			5 4				-
		-0.8	-5 4		-0 7	-4.7	-5 0	
24/8/73	10.5			4 7				-
		-0.6	-4.0		-0 3	-2.0	-3 0	
25/8/73	9 9			4 4				-
		-0.5	-3.4		-0 3	-2 0	-2 7	
26/8/73	9 4			4 1				-
		-0 2	-1.4		-0 3	-2 0	-1 7	
27/8/73	9.2			3 8				-
		-0 1	-0 7		-0 2	-1 4	-1 1	
28/8/73	9 1			3 6				-
29/8/73	-							2 5
		-0 4	-2 7		-0.4	-2 7	-2 7	
30/8/73	-							0 7
31/8/73	8 7			3 2				0.1

Daily data of actual evapotranspiration measured by lysimeters at Durham Observatory

Date	lysimeter One			lysimeter Two			Average Depth mm	Rainfall mm
	Reading	Diff in readings	Depth mm	Reading	Diff in re-adings	Depth mm		
31/8/73	8 7	-0 4	2 7	3.2	0	0	-1 4	-
1/9/73	8 3	-0 3	-2	3.2	-0 9	-6	-4	-
2/9/73	8 0	-0 1	-0 7	2 9	-0 1	-0.[illegible]	-0 7	1 0
3/9/73	7.9	0	0	2.8	-0 4	+2 7	+1 4	-
4/9/73	7 9	-0.5	-3.4	2.4	-0 5	-3.4	-3 4	-
5/9/73	7 4	0	-0	1 9	0	0	0	-
6/9/73	7 4	-0 5	-3.4	1 9	-0 5	-3.4	-3 4	-
7/9/73	6.9			1.4				-
8/9/73	-	-0 6	-4 0	-	-1 0	-6 7	-5 3	-
9/9/73	-			-				1 2
10/9/73	6.3	0	0	0.4	-0 2	-1 4	-0 6	-
11/9/73	6 3			0 2				-
12/9/73	-	-0 2	-1.3	-	0	0	-0 6	0.2
13/9/73	6 1			0 2				0 5
14/9/73	-	0	0	-	+0 3	+2 0	+1 0	-
15/9/73	6 1	0.1	-0 7	0 5	+0 5	+3 4	+1 4	2 1
16/9/73	6 0	0	0	1 0	-0 1	-0 7	-0 4	0 1
17/9/73	6 0			0 9				2.0
18/9/73	-	1 0	+6.7		+0.3	+2 0	+4.4	11 4
19/9/73	5.0			1 2				0.3
20/9/73	-	-0 2	-1 3	-	-0 2	+1 3	-1 4	-
21/9/73	4.8	-0 2	-1 3	1 0	-0 1	-0 7	-1 0	-
22/9/73	4.6	+2.3	+15.3	0.9	+2 5	16 5	+1	16.3
23/9/73	6 9	0	0	3.4	-0 4	-2 7	-1 4	-
24/9/73	6 9	-0 1	-0 7	3 0	-0 2	-1 4	-1 0	1 6
25/9/73	6 8			2.8				0 5
26/9/73	-	-0 1	-0.7	-	-0 5	-3.4	-2 0	5 9
27/9/73	6 7	+0.8	+5.8	2.3	+1 9	+12 7	+9 0	0 1
28/9/73	7 5	-0 4	-2 7	4 2	-0 7	-4 7	-3 7	5 5
29/9/73	7 1	+0 1	+0.7	3 5	0	0	+0 4	0 1
30/9/73	7 2	-0 3	-2 0	3 5	-1 1	-7 3	-4.6	-
31/9/73	6 9			2.4				-

APPENDIX IV

1 - To calculate Thornthwaite Potential Et

2 - To calculate total, mean, standard deviation, minimum and maximum for each variable in a set (or subset of observations)

3 - To calculate correlation coefficient and regression equation intercept and coefficient

4 - To plot hydrographs of simulated and recorded mean daily flows

5 - To plot hydrographs of simulated and recorded mean daily flows and calculate correlation coefficient between simulated and recorded flows

APPENDIX IV 1

```
      CALCULATE THORNTHWAITE POTENTIAL ET
      DIMENSION T(12),ETP(12)
      P  )
      READ(5,100) (T(K),K=1,12)
  100 FORMAT (12F6.1)
      K=1
  200 F=P+(T(K)/5)**1.514
      K=K+1
      IF(K.LE.12) GO TO 200
      AP=6.75*(1.0E-7)*(P**3)-7.71*(1.0E-5)*(P**2)+1.792*(1.0E-2)*P
     *+0.49239
      K=1
  300 ETP(K)=1.6*(10*T(K)/P)**AP
      K=K+1
      IF(K.LE.12) GO TO 300
      WRITE (6,400) (ETP(K),K=1,12)
  400 FORMAT (12F5.2)
      STOP
      END
END OF FILE
```

APPENDIX IV 2

```
C
      CALCULATE TOTAL,MEAN,ST. DEV.,MIN.,MAX.,FOR EACH VARIABLE
      IN A SET (OR SUBSET OF OBSERVATIONS)
      REAL A(10,12),TOTAL(12),AVER(12),SD(12),VMIN(12),VMAX(12),S(12)/
     A12*1.0/
      READ(5,1)((A(I,J),J=1,12),I=1,10)
    1 FORMAT(12F6.1)
      CALL TALLY(A,S,TOTAL,AVER,SD,VMIN,VMAX,10,12)
      WRITE(6,4)
    4 FORMAT(//1X,'TOTAL',6X,'AVER',7X,'SD',7X,'MIN',7X,'MAX'//)
      DO 10 J=1,12
      WRITE(6,3) TOTAL(J),AVER(J),SD(J),VMIN(J),VMAX(J)
   10 CONTINUE
    3 FORMAT(1X,F9.2,4F10.2)
      STOP
      END
END OF FILE
```

APPENDIX IV 3

```
C
      CALCULATE CORRELATION COEFFICIENT AND REGRESSION EQUATION
      INTERCEPT AND COEFFICIENT
      REAL XPUND(12),YRAIN(12)
      REAL*4 R(4),STD(2),DATA(24),A(4),B(4),D(12),S(4),XBAR(2),SKEW(2)
     *,CURT(2),CODE(2)
      READ 1,(YRAIN(I),I=1,12)
      READ 1,(XPUND(I),I=1,12)
    1 FORMAT(12F6.1)
      DO 10 L=1,12
      DATA(L)=XPUND(L)
   10 DATA(L+12)=YRAIN(L)
      CALL MISP(12,2,DATA,CODE,XBAR,STD,SKEW,CURT,R,N,A,B,S,IER)
      PRINT 11,(XBAR(I),STD(I),I=1,2), R(2),A(2),B(2),S(2)
   11 FORMAT(1H,T15,'MEAN',T40,'STD.DEV.',//1H ,'XPUND',T10,F10.4,
     *T35,F10.4/1H ,'YRAIN',T10,F10.4,T35,F10.4//1H ,'CORRELATION',T20,
     *F6.4//1H ,'INTERCEPT',T18,F6.2,'COEFFICIENT',T50,F7.6/1H,'STANDAR
     *D ERROR',T40,F7.4)
      STOP
      END
END OF FILE
```

APPENDIX IV 4

```
C
      PLOT HYDROGRAPHS OF SIMULATED AND RECORDED (A) DAILY FLOWS
      REAL XDAY(366),YSIM(366),YACT(366),YDIF(366),HEADER(8)
      READ1,HEADER,NDAYS
      READ2,(YACT(I),I=1,NDAYS)
      READ2,(YSIM(I),I=1,NDAYS)
    1 FORMAT(8A4/I3)
    2 FORMAT(12F6.0)
      DO3 I=1,NDAYS
    3 XDAY(I)=I
      CALL PSYMB(2.0,9.0,.40,2,HEADER,0.0,32)
      CALL PSCALE(10.0,1.0,XMIN,DX,XDAY(1),366,1)
      CALL PSCALE( 8.0,1.0,YMIN,DY,YACT(1),366,1,YSIM(1),366,1)
      CALL PLTOFS(XMIN,DX,YMIN,DY,1.0,1.0)
      CALL PAXIS(1.0,1.0,'OCT 1-SEP 30 1972',-17,10.0,0.0,XMIN,DX,1.0)
      CALL PAXIS(1.0,1.0,'DISCHARGE (F3 ',14,8.0,90.0,YMIN,DY,1.0)
      CALL PLINE(XDAY,YACT(I),366,1,0,0,1)
      CALL PLNSYM(0,04)
      CALL PLINE(XDAY(I),YSIM(I),366,1,3,11,1)
      CALL PLTEND
      STOP
      END
END OF FILE
```

APPENDIX IV.5

```
C
      PLOT HYDROGRAPHS OF SIMULATED AND RECORDED MEAN DAILY FLOWS
      AND CALCULATE CORRELATION COEFFICIENT BETWEEN SIMULATED AND
      RECORDED FLOWS
      REAL XDAY(366),YSIM(366),YACT(366),YDIF(366),H1(8),H2(8),H3(8)
      REAL*4 R(4),STD(2),DATA(732),RX(4),B(2),D(2),T(2),XBAR(2)
      READ1,H1,H2,H3,NDAYS
      READ2,(YACT(I),I=1,NDAYS)
      READ2,(YSIM(I),I=1,NDAYS)
    1 FORMAT(8A4/8A4/8A4/I3)
    2 FORMAT(12F6.0)
      DO3 I=1,NDAYS
    3 XDAY(I)=I
      DO 10 L=1,NDAYS
      DATA(L)=YACT(L)
   10 DATA(L+NDAYS) YSIM(L)
      CALL PSYMB(2.0,9.0,-0.2,H1,0.0,32)
      CALL PSYMB(2.0,8.5,-0.2,H2,0.0,32)
      CALL PSYMB(2.0,8.0,-0.2,H3,0.0,32)
      CALL PSCALE(10.0,1.0,XMIN,DX,XDAY(I),366,1)
      CALL PSCALE( 8.0,1.0,YMIN,DY,YACT(I),366,YSIM(I),366,1)
      CALL PLTOFS(XMIN,DX,YMIN,DY,1.0,1.0)
      CALL PAXIS(1.0,1.0,'OCT 1-SEP 30',-12,10.0,0.0,XMIN,DX,1.0)
      CALL PAXIS(1.0,1.0,'DISCHARGE CFS ',14,8.0,90.0,YMIN,DY,1.0)
      CALL PLINE(XDAY(I),YACT(I),366,1,0,0,1)
      CALL PLTSYM(0.04)
      CALL PLINE(XDAY(I),YSIM(I),366,1,3,11,1)
      CALL PLTEND
      CALL CORRE(NDAYS,2,1,DATA,XBAR,STD,RX,R,B,D,T)
   11 FORMAT(1H ,T15,'MEAN',T40,'STD.DEV.',//1H ,'ACTUAL',T10,F10.4,
     *1H ,'CORRELATION',T20,F6.4)
      STOP
      END
C
END OF FILE
```

APPENDIX V

Table A: Monthly values of simulated flow in ft^3/sec (m^3/sec) for three different CB values for 1972 (a year of average flow)

Month / CB	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL inches (mm)
0.6	512 (14 5)	1135 (32.3)	298 (8.5)	4430 (125.8)	2747 (78 0)	1405 (39 9)	933 (26.5)	1145 (32.5)	1273 (36.2)	978 (27 8)	337 (9 6)	225 (6.4)	8.22 (208.8)
0.1	391 (11 1)	965 (27 4)	341 (9 7)	5551 (157.6)	3827 (108 7)	1755 (49.8)	991 (28.1)	737 (20.9)	916 (26.0)	1289 (36.6)	447 (12.7)	349 (9 9)	9 36 (237.7)
0 01	326 (9.3)	1556 (44.2)	343 (9 7)	7773 (220.8)	4048 (115 0)	1846 (52.4)	915 (26.0)	238 (6.8)	327 (9 3)	1174 (33.3)	127 (3.6)	116 (3.3)	10 02 (254.5)

Table B Monthly values of the simulated flow in ft^3/sec (m^3/sec) for two different CB values for 1973 (a year of low flow)

CB \ Month	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL inches (mm)
0.05	307 (8 7)	597 (17.0)	863 (24.5)	740 (21 0)	491 (13.9)	341 (9.7)	448 (1? 7)	475 (13.5)	396 (11.2)	546 (15.5)	562 (16.0)	413 (11 7)	3.30 (83.8)
0 1C	312 (8 9)	627 (17 8)	819 (23.2)	680 (19.3)	531 (15.1)	382 (10 8)	531 (15.1)	604 (17.1)	511 (14.5)	633 (18 0)	688 (19.5)	507 (14 4)	3.64 (92 4)

Table C Monthly values of simulated flow in ft^3/sec (m^3/sec) for two values of LZSN

Month / LZSN	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL inches (mm)
7.5	318 (9.0)	602 (17.1)	612 (17 4)	3787 (107.6)	4344 (123 4)	2613 (74.2)	1652 (46.9)	1043 (29.6)	1018 (28.9)	832 (23.6)	622 (17.7)	454 (12 9)	9.55 (242.6)
5.5	324 (9.2)	689 (19 6)	660 (18.7)	5212 (148.0)	4810 (136.6)	2848 (80 9)	1804 (51 2)	1021 (29.0)	971 (27 6)	830 (23 6)	633 (18.0)	450 (12 8)	10 80 (274.3)

Table D Monthly values of simulated flow in ft^3/sec (m^3/sec) for two values of UZSN

Month / UZSN	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL inches (mm)
0 3	363 (10.3)	774 (22 0)	282 (8 0)	5184 (147.2)	3779 (107.3)	1716 (48 7)	971 (27 6)	851 (24 2)	1042 (29.6)	1670 (47.4)	451 (12 8)	371 (10.5)	4 31 (109.5)
1 0	352 (10 0)	709 (20 1)	249 (7.1)	3296 (93 6)	3193 (90 7)	1753 (49.8)	996 (28.3)	874 (24.8)	1015 (28 8)	769 (21.8)	393 (11 2)	484 (13 7)	7.51 (190 8)

<u>Table E</u> Monthly values of simulated flow in ft^3/sec (m^3/sec) for two values of CC with a CB of 0 6

Month / CC	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL inches (mm)
0 5	773 (22 0)	1299 (36 9)	324 (9.2)	5321 (151 1)	3165 (89.9)	1619 (46 0)	1154 (32.8)	1446 (41 1)	1639 (46 5)	1357 (38 5)	451 (12 8)	364 (10.3)	10 09 (256 3)
3.0	773 (22.0)	1299 (36.9)	324 (9 2)	5321 (151.1)	3185 (90 4)	1630 (46 3)	1155 (32 8)	1443 (41 0)	1635 (46 4)	1358 (38 6)	449 (12 8)	363 (10 3)	10 10 (256.5)

Table F Monthly values of simulated flow in ft^3/sec (m^3/sec) for two values of CC, with a CB of 0.1

CC \ Month	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL inches (mm)
0.5	783 (22.2)	910 (25 8)	860 (24.4)	4164 (118.2)	4288 (121.8)	2662 (75.6)	1584 (45 0)	1084 (30.8)	1036 (29 4)	829 (23.5)	615 (17 5)	478 (13.6)	10 29 (261.4)
3.0	790 (22 4)	445 (12.6)	877 (24 9)	4746 (134.8)	4576 (130.0)	2621 (74.4)	1657 (47.0)	1015 (28.8)	1003 (28 5)	902 (25 6)	654 (18.6)	467 (13 3)	10 83 (275 1)

Table G· Monthly values of simulated flow in ft^3/sec (m^3/sec) for two values of K3 for 1972 (a year of average flow)

K3 \ Month	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL inches (mm)
0 20	783 (22 2)	910 (25 8)	860 (24.4)	4164 (118.2)	4288 (121 8)	2662 (75.6)	1584 (45.0)	1084 (30.8)	1036 (29.4)	829 (23.5)	615 (17.5)	478 (13.6)	10 29 (261.4)
0 25	783 (22 2)	910 (25.8)	860 (24.4)	4164 (118 2)	4288 (121 8)	2662 (75.6)	1584 (45.0)	1084 (30 8)	1036 (29 4)	828 (23 5)	615 (17 5)	477 (13.5)	10 29 (261.4)

Table H: Monthly values of simulated flow in ft^3/sec (m^3/sec) for two values of K3 for 1973 (a year of low flow)

K3 \ Month	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL inches (mm)
0 20	222 (6.30)	531 (15 1)	680 (19 3)	565 (16.0)	462 (13 1)	340 (9 7)	511 (14.5)	545 (15 5)	408 (11 6)	556 (15.8)	632 (17 9)	441 (12.5)	3 14 (79 8)
0 25	222 (6.30)	531 (15 1)	679 (19 3)	564 (16.0)	461 (13.1)	339 (9 6)	511 (14 5)	542 (15.4)	402 (11 4)	539 (15.3)	604 (17 2)	419 (11.9)	3.10 (78.7)

Table I Monthly values of simulated flow in ft^3/sec (m^3/sec) for two values of K24L

K24L \ Month	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL inches (mm)
0.0	318 (9.0)	602 (17.1)	612 (17 4)	3787 (107.6)	4344 (123 4)	2613 (74 2)	1652 (46 9)	1043 (29 6)	1018 (28 9)	832 (23 6)	622 (17 7)	454 (12 9)	9 55 (242 6)
0.25	306 (8.7)	523 (14 8)	479 (13 6)	3093 (87 8)	3594 (102 1)	2170 (61.6)	1395 (39 6)	877 (24 9)	841 (23 9)	701 (19 9)	499 (14.2)	375 (10 6)	7.92 (201 2)

Table J· Monthly values of simulated flow in ft^3/sec (m^3/sec) for two values of K24EL

Month / K24EL	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL inches (mm)
0 1	227 (6 4)	540 (15 3)	684 (19 4)	566 (16 1)	462 (13.1)	342 (9 7)	529 (15 0)	594 (16 9)	482 (13 7)	619 (17 6)	681 (19 3)	491 (13.9)	3 32 (84 3)
0 2	222 (6 3)	531 (15 1)	679 (19 3)	564 (16 0)	461 (13 1)	339 (9 6)	511 (14 5)	542 (15.4)	402 (11 4)	539 (15.3)	604 (17 2)	419 (11.9)	3.10 (78 7)

<u>Table K</u> Monthly values of simulated flow in ft^3/sec (m^3/sec) for two values of UZS

Month / UZS	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL inches (mm)
C.00	502 (14 2)	1699 (48 2)	497 (14.1)	6751 (191 7)	4171 (118 4)	1891 (53 7)	1069 (30 4)	866 (24.6)	1062 (30 2)	1826 (51 9)	492 (14 0)	435 (12 4)	11 34 (288 0)
1 00	1094 (31.1)	2144 (60 9)	531 (15 1)	6968 (197.9)	4228 (120.1)	1919 (54 5)	1093 (31 0)	866 (24 6)	1063 (30 2)	1851 (52 6)	499 (14 2)	449 (12 8)	12 11 (307.6)

Table L· Monthly values of simulated flow in ft^3/sec (m^3/sec) for two values of LZS

LZS \ Month	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL inches (mm)
5.00	363 (10 3)	774 (22 0)	282 (8 0)	5184 (147.2)	3779 (107.3)	1716 (48 7)	971 (27.8)	851 (24 2)	1042 (29 6)	1670 (47 4)	451 (12 8)	371 (10 5)	9 31 (236.5)
8 00	502 (14 3)	1699 (48 2)	497 (14.1)	6751 (191.7)	4171 (118.4)	1891 (53.7)	1069 (30.4)	866 (24 6)	1062 (30 2)	1826 (51 9)	492 (14.0)	355 (10 1)	11 34 (288 0)

Table M Monthly values of simulated flow in ft^3/sec (m^3/sec) for two IRC values

Month / IRC	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL inches (mm)
0 5	265 (7.5)	673 (19 1)	1768 (50 2)	2884 (81.9)	1823 (51 8)	2764 (78.5)	1584 (45.0)	1102 (31 3)	908 (25 8)	649 (18 4)	3033 (86 1)	921 (26.2)	9 80 (248.9)
0 7	265 (7 5)	670 (19 0)	1688 (47.9)	2919 (82.9)	1869 (53 1)	2757 (78 3)	1581 (44.9)	1115 (31.7)	908 (25.8)	649 (18 4)	3032 (86.1)	922 (26 2)	9 80 (248 9)

<u>Table N</u>· Monthly values of simulated flow in ft^3/sec (m^3/sec) for two KK24 values

KK24 \ Month	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL
0 90	801 (22 8)	1297 (36.8)	2877 (81 7)	3283 (93 2)	1407 (40 0)	3223 (91.5)	1076 (30 6)	691 (19 6)	1005 (28 5)	373 (10 6)	4057 (115.2)	240 (6 8)	10.80 (274 3)
0 99	265 (7 5)	670 (19.0)	1688 (47.9)	2919 (82.9)	1869 (53.1)	2757 (,8 3)	1581 (44.9)	1115 (31.7)	908 (25 8)	649 (18 4)	3032 (86 1)	922 (62 2)	9.80 (248 9)

Table 0 Monthly values of simulated flow in ft^3/sec (m^3/sec) for two KV values

KV \ Month	O	N	D	J	F	M	A	M	J	J	A	S	TOTAL
0 0	525 (14 9)	662 (18 8)	595 (16 9)	3479 (98.8)	3156 (89 6)	2150 (61 1)	1575 (44.7)	1256 (35 7)	1154 (32.8)	1251 (35 5)	792 (22 5)	622 (17 7)	9 18 (233.2)
1.0	629 (17.9)	763 (21 7)	696 (19 8)	4196 (119.2)	4172 (118 5)	2409 (68 4)	1357 (38 5)	946 (26.9)	948 (26.9)	1027 (29 2)	624 (17.7)	458 (13 0)	9 72 (246 9)