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THE RAINFALL–RUNOFF RELATIONSHIP IN A SMALL CATCHMENT

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ABSTRACT

Following Gregory and Walling (1971), who described how relatively simple equipment can be used to monitor processes in a small drainage basin, experiments were started in the catchment close to the Leonard Wills Field Centre (grid reference ST 0537) in 1973. The purpose of these experiments was to provide a set of long-term data which could be easily incorporated into sixth form courses and which would illustrate the hydrological principles included in A-level syllabuses. This paper outlines the results collected over the last twelve years and provides an interpretation of both the water balance within the catchment and also the long and short term responses of the stream to rainfall.

INTRODUCTION

WATER is of vital importance to man and the study of hydrology encompasses the relationship of water, in all its forms, between the atmosphere, the land and the sea. These relationships are rarely simple. When precipitation falls over landmasses only a small proportion enters surface water bodies directly. The remainder reaches streams, rivers and lakes via the vegetation, soils and rocks of the surrounding area; processes which take a variable amount of time and during which large volumes of water may return to the atmosphere through evaporation and transpiration. The amount of water entering streams and rivers may, therefore, only be a small proportion of that which fell as precipitation.

The processes by which precipitation reaches a stream are shown diagrammatically in Fig. 1. Only a small proportion of it enters the stream directly as *channel precipitation,* the remainder falls onto the surrounding land. Of this, some will be *intercepted* by the vegetation cover, the degree of interception depending on the type of vegetation, the time of year and the intensity of the precipitation. Some of the intercepted water will evaporate but the rest will reach the ground either by flowing along stems, branches and trunks of vegetation (*stemflow*) or by falling from the leaves to the ground (*throughfall*). On reaching the ground the water moves into the soil, a process known as *infiltration* and resulting from the pull of both gravitational and capillary forces.

The rate of infiltration depends on the *hydraulic conductivity* of the soil—its *porosity* and *permeability*. The porosity describes the water-holding capacity and is controlled by the number of voids or pores present. The permeability is the extent to which the soil allows movement of water and, although influenced by porosity, it is the size of the pores, rather than their number which is important.

When investigating the movement of water in soil, it is necessary to consider three sets of forces:— molecular, capillary and gravitational. Molecular forces hold a thin volume of water to the area immediately surrounding each soil particle and they are so powerful that

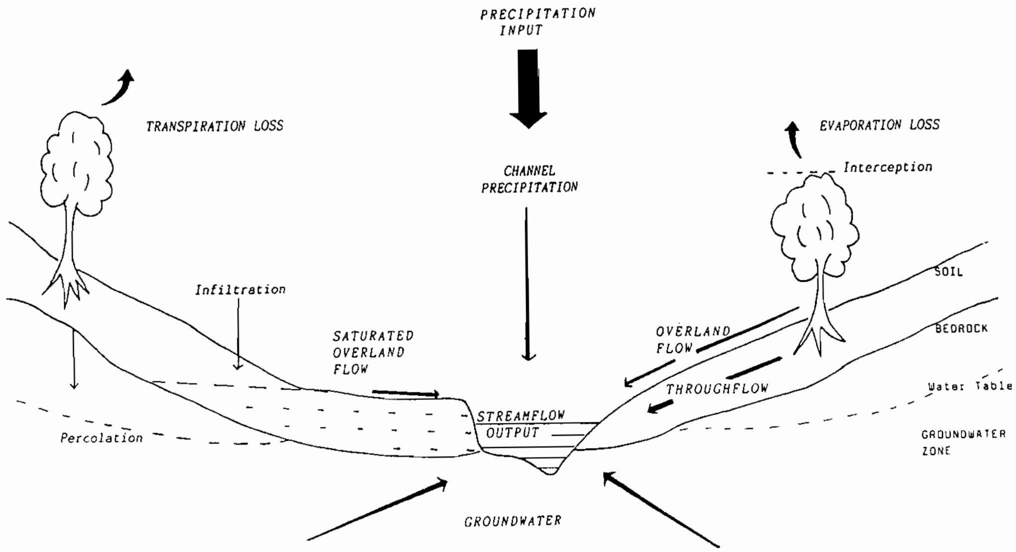


FIG. 1.
Diagrammatic form of the basin hydrological cycle.

this *hygroscopic water* can be ignored in the hydrological system. Capillary forces act slightly further away from the soil particle, exerting a suction effect which draws water towards the particle even against the force of gravity. Gravity itself operates in large pores (*macropores*), where soil particles are sufficiently far apart for capillary forces to be ineffective, causing water to move down through the soil profile.

Most soils show a decrease in permeability with depth. This results in the downslope movement of water within the surface layers, sometimes in pipes; a process known as *throughflow*. A similar effect may be caused by a decrease in permeability at the junction between soil and bedrock. The throughflow thus generated contributes to the runoff of the stream but the time taken is highly variable, depending on the prevailing moisture conditions. However, this downslope movement does result in a relatively high moisture content at the base of the slope leading to early saturation of this area in the event of heavy rainfall, partly from local rainfall and partly from throughflow arriving from upslope. Under these conditions rain may well fall at a faster rate than infiltration can take place with the result that water flows over the surface as *saturated overland flow*, resulting in an almost immediate transfer of rainwater to stream runoff. This saturated zone is normally restricted to the flat floodplain area adjacent to the stream, and to the valley head. It is often referred to as the *partial contributing area*. This is a dynamic zone, whose size varies according to the moisture conditions and it may extend to the valley sides if, for example, surface saturation of the soil is maintained by throughflow from higher up the slope. However, movement over the surface will not occur until all irregularities in the ground have filled with water (*depression storage*) and start to overflow. This idea of a partial contributing area is in contrast to Horton's (1945) suggestion that overland flow, resulting from an excess of precipitation over infiltration, commonly occurs over most of the drainage basin.

Finally, mainly in winter, some water will *percolate* from the base of the soil, into and through the underlying bedrock, the rate of percolation being controlled by the porosity and the permeability of the rock itself. The water moves slowly downwards towards the zone of permanent saturation—the *groundwater zone*. The top of the groundwater zone is often referred to as the *water table*, although it is not a horizontal surface. Groundwater provides the long term baseflow to the stream, supplying it even during dry conditions, when there is little, or no, precipitation. (Note that, under some circumstances, there may be a water table within the soil).

There are, therefore, four routes by which precipitation may reach a stream; directly as channel precipitation; over the surface, as saturated overland flow; via the soil, as throughflow; and through the bedrock, as groundwater. The time taken for each to reach the stream is different and their relative importance controls the stream's response to precipitation.

The Catchment

The drainage basin studied (Fig. 2) has a total area of 1.098 km^2 and, using the Strahler scheme of ordering, is a first order tributary of the River Doniford. The geology of the catchment is a sequence of Devonian sandstones and slates with an infill of periglacial head occupying the valley bottom. Cores taken from boreholes have shown the thickness of head to vary from a few centimetres in the valley bottom adjacent to the middle weir (B in Fig. 2), to several metres adjacent to the most downstream weir (C in Fig. 2). Whereas the head is an unconsolidated material composed of angular stone fragments in a fine matrix and as such is relatively permeable, the underlying rocks are metamorphosed and are permeable only through structural weaknesses.

The form of the valley varies considerably. Above the pond, in the area which is normally dry, it is open with valley-side slopes of less than 15°; below this the valley becomes steeply incised and the angle of the valley sides increases to between 20 and 25°. Downstream of the middle weir the valley widens out and by the level of the downstream weir it has lost its incised nature. The channel slope between the upper and lower weirs averages 5°.

The source of the stream is a spring immediately above the pond shown on the map. Its position is not fixed but varies by several metres with different moisture conditions. During very wet periods, saturated overland flow occurs in the normally dry part of the valley above the spring.

In the past, there has been considerable human interference with the stream and some channelling of water from it. The pond shown at the source of the stream was constructed at least 200 years ago. It is surrounded by trees and, during the autumn, the outflow is cleared of leaves, causing artificial peaks in water level (Fig. 19d). Water from it now feeds four cattle troughs within the catchment, with the level in each controlled by a ball valve. Downstream of the highest weir (A in Fig. 2) water is occasionally pumped off to supply a local cottage. Immediately upstream of the lowest weir the stream was channelled underground during an eighteenth century landscaping programme but a large proportion of this has now been excavated.

The present land use of the catchment (Fig. 3) is a mixture of pasture, arable, bracken and woodland. Over the last 10 years there has been an increase in the amount of arable with associated hedge removal but this does not appear to have had any significant effect on the response of the stream to rainfall.

Equipment

Precipitation Input

The aim, in the measurement of precipitation, is to collect a volume of water from a

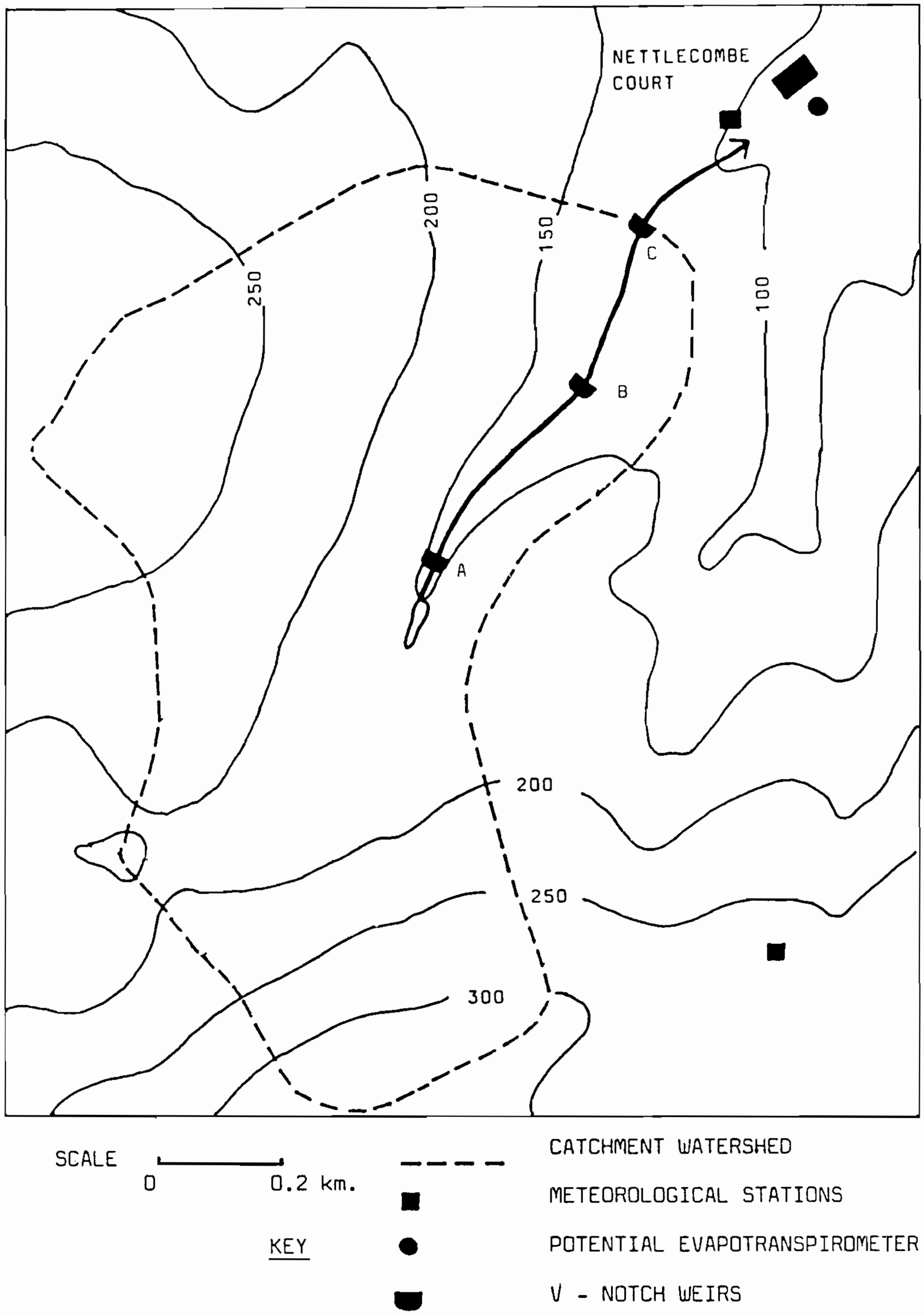


FIG. 2.
Map of the catchment, showing the position of the equipment.

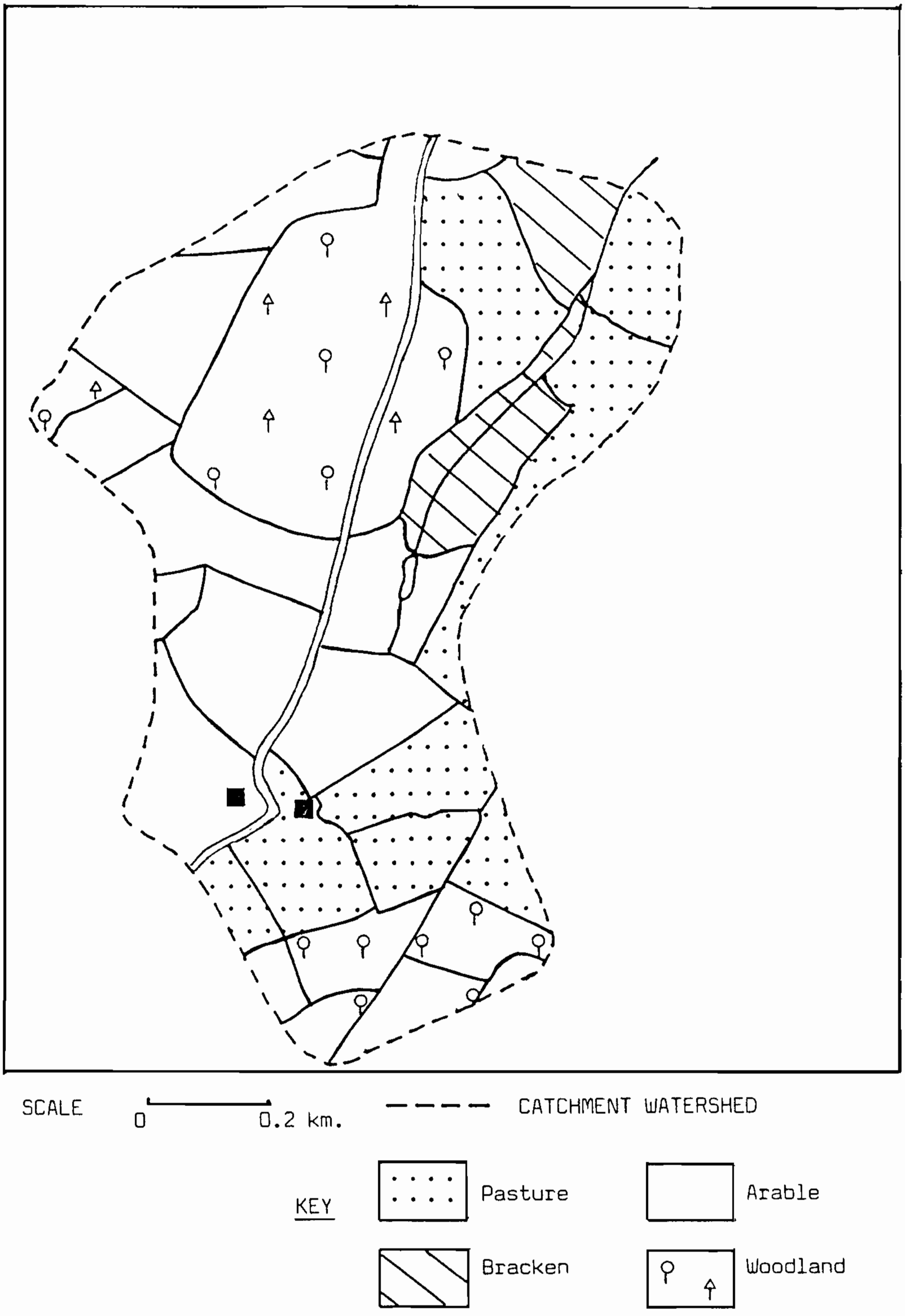


FIG. 3.
Map of the catchment, showing land use.

known area and to express it as units of depth, usually millimetres. The area over which the precipitation is collected is defined by the collecting funnel or raingauge rim and it is assumed that the depth of water collected by the gauge is the same as that falling on the surrounding area. There are a number of problems associated with this assumption. A tall raingauge acts as an obstacle to the flow of wind (and may result in falling rain being carried past the gauge—leading to an underestimate). Alternatively, if the rim is very close to the ground, insplashing may occur giving rise to an overestimate. The Meteorological Office has standardised by issuing raingauges which are 127 mm in diameter and placed so that the rim is 305 mm off the ground.

The standard meteorological raingauge is emptied and the precipitation measured and recorded every 24 hours at 0900 GMT. If the intensity and duration of the rainfall are required then it is necessary to instal an autographic raingauge. Those used in this experiment are of the self-syphoning type, fitted with daily charts on which 1 mm of rain is represented by a rise of 10 mm on the chart (Fig. 4).

The location of raingauges is also important, as the value of all data collected depends on the extent to which the sites chosen are representative of the surrounding area. Ideally, a network of gauges should cover the entire basin. The monitoring of such a network is impractical for the Field Centre and measurements are made at just two sites, one situated just below the catchment at 100 m and the other on the watershed of the River Doniford at 270 m (Fig. 2). Over the twelve year period of this study, rainfall at the higher site has been 10% greater than that at the lower but, throughout the results, averages of the two sites have been calculated. Both sites are official climatological stations, recording data for the Meteorological Office and for the Wessex Water Authority, and include both standard and autographic raingauges.

Streamflow Output

Stream *discharge* is the volume of water moving per unit of time, for small streams normally expressed as litres per second ($l\ s^{-1}$). This can be measured in a number of ways, the simplest being to multiply the cross sectional area of the channel by the "average velocity of the stream". This method is relatively straightforward and is frequently used in fieldwork but it is difficult to obtain accurate results especially on small streams. Variations of 200% have been obtained by groups of sixth form students!

For long term records it is necessary to adopt a more accurate method, made possible by the construction of a V-notch weir (Fig. 5). The weir is formed by cutting a sharp-edged triangular notch into a metal plate and installing it in a watertight manner across the stream. The angle of the notch is chosen to suit the size of the stream. In this study angles of 90° were used. The stream is channelled to flow through the notch and the height of water over the V, measurable anywhere in the weir pool, is proportional to the discharge.
For a 90° notch this relationship can be written mathematically as:

$$Q = 0{\cdot}015\, H^{2{\cdot}48}$$ (Gregory and Walling, 1971)

where

Q is the stream discharge in litres per second.

H is the head of water over the notch in centimetres.

The equation can be plotted as a line on logarithmic graph paper from which the head of water can be converted directly to discharge (Fig. 6).

Three 90° weirs were installed on the stream draining the Nettlecombe basin; the one nearest the source positioned immediately downstream of the pond (A in Fig. 2) with a drainage area of 0.505 km^2; the middle (B in Fig. 2) an area of 0.721 km^2; and the farthest

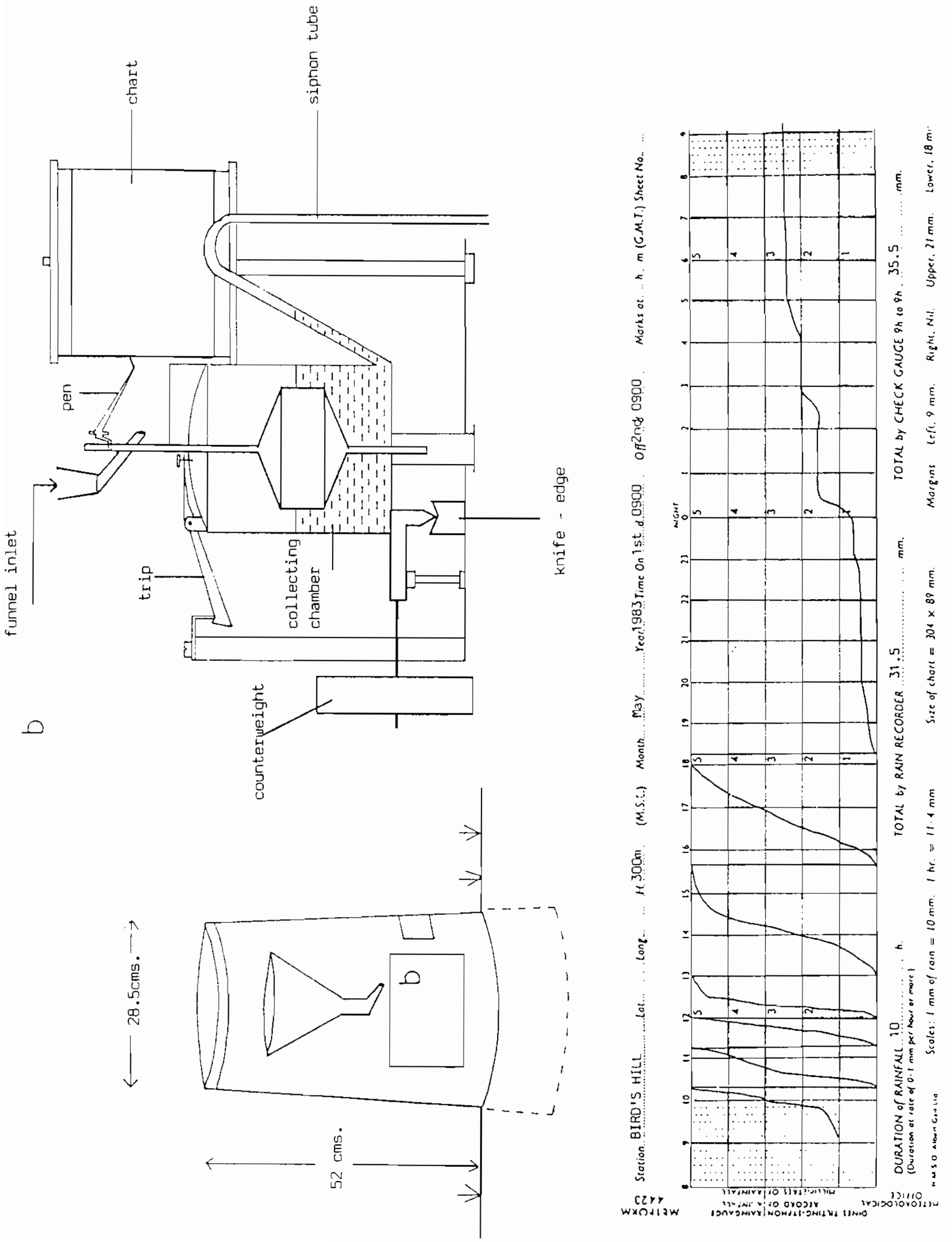


Fig. 4.
Diagram of an autographic rain gauge and chart.

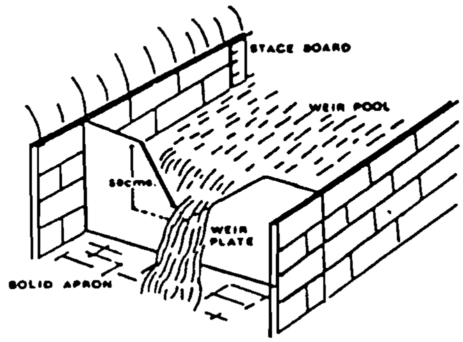


FIG. 5.
Diagram of a sharp crested V-notch weir.

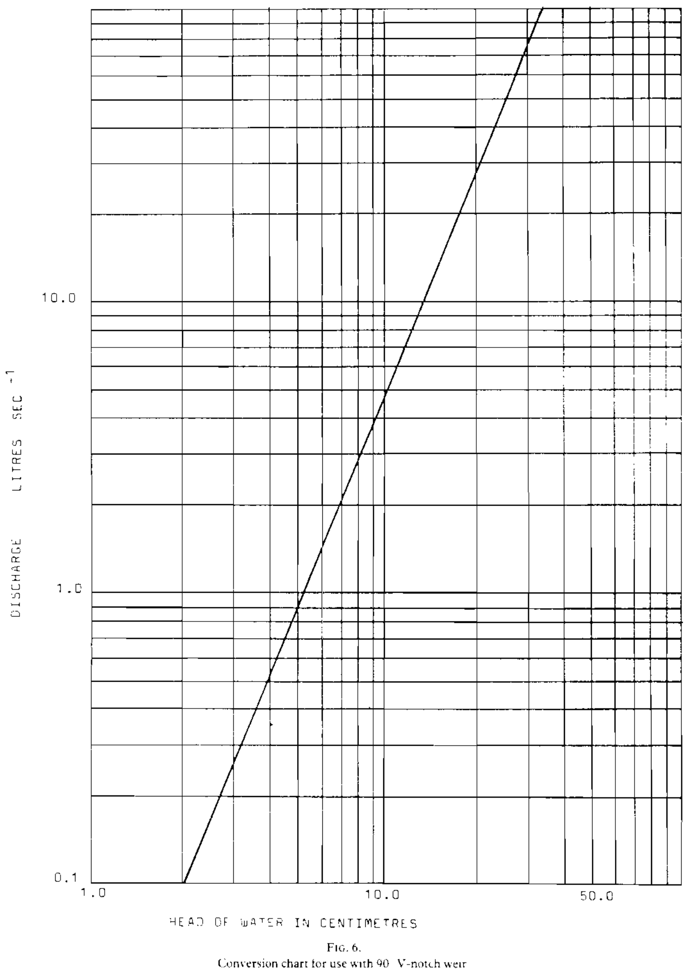


FIG. 6.
Conversion chart for use with 90° V-notch weir.

downstream (C in Fig. 2) an area of 1.098 km^2. At the start of the experiment all three weirs were read daily between 0900 and 1000 GMT with the inevitable result that several storm peaks were left unrecorded. This situation was gradually rectified with the installation of water level recorders at all three weirs. All are fitted with weekly charts but they record varying changes in level. The one at present installed on the upstream weir records changes of up to 1 foot (300 mm); that on the middle weir of up to 500 mm and that on the lower weir of up to 1 m. The ratio at the lower weir is such that small changes in level do not show.

The Water Table

During the first year of the experiment, instead of the expected increase in discharge downstream, the data showed a decrease during the summer months when the discharge was less than 5 $l s^{-1}$. To investigate this in relation to the water table, seven 30 mm diameter boreholes were drilled to varying depths across the valley adjacent to the lower weir (Fig. 20a). Each borehole was lined with perforated plastic piping and the position of the water table was measured using a conductivity probe, consisting of two electrodes between which a current flows only on contact with a water surface. Measurements were taken monthly.

Evapotranspiration Losses

For the first five years of the experiment, an attempt was made to measure the losses by evaporation and transpiration. Both parameters are notably difficult to measure accurately and are often considered together as evapotranspiration. Attempts to measure potential rather than actual rates have been most successful where potential evapotranspiration rates assume a constant supply of moisture, sufficient to meet the transpiration needs of the vegetation cover. In other words, potential rates assume that the soil is constantly at *field capacity*; where field capacity can be defined as the state of the soil after rainfall, when excess moisture has had time to drain away and the rate of downward movement has materially decreased. In reality, unless there really is constant rainfall, some of this moisture is taken up by plants in transpiration, gradually depleting the supply and creating a *soil moisture deficit (SMD)*. The SMD is calculated as the amount of moisture required to bring the soil back to field capacity and is greatest during the summer months. Potential rates of evapotranspiration therefore tend to be overestimates of actual rates, the discrepancy increasing with increasing SMD.

One method of measuring potential evapotranspiration is to record water losses from moist vegetated surfaces. To do this an evapotranspirometer was constructed, based on the design in Fig. 7. The instrument consists of three watertight oil drums, each with a single exit through a piece of plastic piping leading to a collecting vessel, housed in a fourth, central, drum. Each of the peripheral drums is filled with soil and supports vegetation similar to that of the surrounding area, in this case grass. The soil moisture content of the drums is maintained near field capacity, which, allowing for rainfall, involved adding the equivalent of 3 mm of rain each day during the winter and 6 mm during the summer. The only exceptions to this were during the very dry summer of 1976 when 12 mm had to be added, and during February 1978 when the ground was covered by 600 mm of snow for several days and no water was added artificially. Whenever possible the percolate was recycled to reduce leaching.

There are therefore only 2 exits for the water; through the pipe at the bottom which is collected in the central drum, or by evapotranspiration. Consequently, it is easy to calculate a water balance for each of the peripheral drums—the difference between the volume

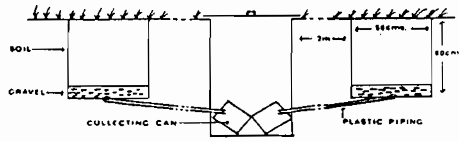


FIG. 7.
Diagram of an evapotranspirometer showing two of the three soil tanks.

of collected water and the water added gives a measure of potential evapotranspiration; potential since the drums are kept at field capacity. The readings were taken daily and an average for the three drums calculated. To express the loss as an equivalent depth of rainfall, the volume was divided by the surface area of the drum. Daily figures were used to calculate monthly totals.

The accuracy of this method depended on the success with which the drums were kept at field capacity and the extent to which they were representative of the catchment area. Although changes in soil moisture storage are important on a daily basis, their importance is negligible in the long term; a fact illustrated by the daily variation in the three drums cancelling out in the monthly totals. The three drums were sited on a flat piece of ground where there were no obstructions to precipitation. They, and the surrounding land, were all covered by grass.

When using the results in the water balance equation, the biggest drawback was the difference between potential and actual rates—especially during the summer months.

In 1978 the evapotranspirometer began to give spurious results; probably one or more of the drums had rusted through. In the same year the Meteorological Office introduced their Rainfall and Evaporation Calculation System (MORECS) in which potential and actual rates of evapotranspiration are calculated using the Penman variables for 40 km by 40 km grid squares. The method of calculation was modified slightly in May 1981. The figures calculated are for grassland, based on one soil type which is assumed to have a typical water availability. These figures, although calculated for a large area, have been used in the water balance equation since 1978. The potential rather than actual rates have been used in order to keep the methods as consistent as possible; however, it is difficult to make any direct comparison between the two as, out of the five years of data collected using the evapotranspirometer, two were the exceptionally hot and dry summers of 1975 and 1976.

RESULTS

The Water Balance

The Water Balance for any drainage basin may be summarized by

$$P = Q + ET \pm \triangle S$$

where P is precipitation (mm).
Q is stream discharge expressed as a depth of runoff (mm).
ET is loss by evapotranspiration (mm).
$\triangle S$ is changes in storage (mm).

In a "typical" year there is a balance between the input into the catchment (that is the precipitation), and the output from the catchment (that is the sum of evapotranspiration and streamflow output as runoff). Most hydrological data is calculated for the *water year* which starts on 1 October and ends on 30 September. This practice assumes that storage in the soil and rocks increases during the winter months (October to March) and is gradually depleted during the summer months (April to September) falling to a minimum towards the end of September. The Water Year is therefore designed to start and end during the period of minimum storage. (Note that the water year 1978 is the 12 month period that ended on 30 September 1978). Although water is stored separately in the soil and rock, for the purpose of this study the two are considered together.

Table 1. *Monthly averages of the water balance 1973–85*

	P	ET	Q	S
Oct	103	42	20	+41
Nov	98	21	30	+47
Dec	134	20	74	+40
Jan	123	17	83	+23
Feb	99	17	80	+2
Mar	103	33	57	+13
Apr	45	53	38	−46
May	69	80	27	−38
Jun	53	92	12	−51
Jly	44	107	5	−68
Aug	72	100	3	−31
Sep	100	69	3	+28
	1043	651	432	−40

P is precipitation input
ET is potential evapotranspiration
Q is stream flow output
S is changes is soil and groundwater storage

Tables 2 and 3 and Fig. 8 show the water balance results for the water years 1973 to 1985. The precipitation is calculated as the average of the daily recordings from the two standard raingauges. For the five years 1973–78, the potential evaporation figures are the monthly totals of the daily recordings from the evapotranspirometer; from 1978 onwards they are the calculated figures from the MORECS scheme. Stream discharge is taken from the water level recorded at 0900 GMT at middle weir for each day. The records from the middle weir were used in preference to the lower, partly because that recorder proved to be by far the most reliable and partly because of the observed loss of water from the lower reaches of the stream at low discharges.

To enable comparison between rainfall and runoff, the discharge in ($\mathrm{l\ s^{-1}}$) is converted to a depth of runoff equivalent to the rainfall data (mm) by the following equation:

$$\text{discharge (mm) of rainfall} = \frac{\text{discharge cm}^3\ \text{sec}^{-1})^{\star} \times 10 \times 86{,}400}{\text{catchment area (cm}^2)}$$

*1 litre is equal to 1000 cm^3
10 converts cm to mm.
86,400 is the number of seconds in 24 hours.

Daily results are then totalled for each month.

The first graph in Fig. 8 (Table 1) shows the monthly averages for each element of the water balance, calculated for the twelve years 1973 to 1985. Rainfall is concentrated in the

Table 2. *The water balance for each water year 1973–85*

	1973–4				1974–5			
	P	ET	Q	S	P	ET	Q	S
Oct	53·8	24.5	3.0	+26.3	72.6	22.0	51.8	−1.2
Nov	53.4	13.9	3.4	+36.1	122.5	0.9	65.5	+56.1
Dec	73.9	19.4	16.9	+37.6	105.9	25.6	62.3	+18.0
Jan	188.5	32.1	107.3	+49.1	159.6	15.1	97.7	+46.8
Feb	189.2	33.3	164.0	−8.1	29.9	8.0	68.7	−46.8
Mar	61.3	12.9	49.5	−1.1	82.4	17.3	42.3	+22.8
Apr	19.2	28.0	21.2	−30.0	51.5	33.9	27.6	−10.0
May	94.3	65.2	18.0	+11.1	20.8	63.7	17.7	−60.6
Jun	64.5	88.0	11.6	−35.1	8.3	101.2	6.8	−99.7
Jly	43.1	93.9	5.1	−55.9	53.6	87.0	4.1	−37.5
Aug	90.5	76.1	2.9	+11.5	38.4	64.9	1.8	−28.3
Sep	204.8	42.3	14.5	+148.0	151.9	79.9	1.2	+70.8
	1136.5	529.6	417.4	+189.5	897.4	519.5	447.5	−69.6

	1975–6				1976–7			
	P	ET	Q	S	P	ET	Q	S
Oct	36·0	25.0	0.2	+10.8	243.4	89.4	105.5	+48.5
Nov	73.9	26.9	0.0	+47.0	107.4	55.4	47.9	+4.1
Dec	53.9	32.5	12.7	+8.7	133.5	30.9	108.2	+5.6
Jan	51.2	24.4	8.8	+18.0	126.5	0.0	90.0	+36.5
Feb	48.6	16.8	16.7	+15.1	174.9	25.7	127.9	+21.3
Mar	82.9	18.7	21.4	+42.8	104.0	51.8	62.6	−10.4
Apr	6.8	35.8	21.7	−50.7	32.8	49.3	39.3	−55.8
May	45.0	77.5	7.7	−40.2	51.2	97.9	15.5	−62.2
Jun	13.8	108.8	3.2	−98.2	93.7	95.3	9.4	−11.0
Jly	8.7	186.5	0.9	−178.7	20.3	107.6	5.4	−92.7
Aug	53.7	210.4	0.1	−156.8	117.5	126.2	3.3	−12.0
Sep	209.1	134.5	1.1	+73.5	22.2	67.8	3.9	−49.5
	683.6	897.8	94.5	−308.7	1227.4	797.3	618.9	−188.8

	1977–8				1978–9 (calculation of ET changed to MORECS system)			
	P	ET	Q	S	P	ET	Q	S
Oct	68·0	73.0	2.6	−7.6	6.2	40	1.8	−35.6
Nov	109.0	0.0	14.7	+94.3	56.6	26	0.8	+29.8
Dec	106.9	13.7	49.4	+43.8	207.4	19	15.3	+173.1
Jan	150.2	27.2	65.9	+57.1	114.6	17	70.3	+27.3
Feb	232.5	0.0	109.6	+122.9	72.8	22	82.8	−32.0
Mar	112.0	60.9	83.6	−32.5	133.3	42	84.7	+6.6
Apr	49.6	32.5	51.2	−34.1	83.0	67	64.5	−48.5
May	58.1	84.8	36.0	−62.7	123.3	68	20.6	+34.7
Jun	33.6	64.0	11.9	−42.3	33.3	77	31.0	−74.7
Jly	119.6	126.1	7.5	−14.0	21.9	105	9.4	−92.5
Aug	89.0	124.2	7.3	−42.5	104.0	81	3.9	+19.1
Sep	18.0	83.6	4.2	−69.8	23.0	62	2.2	−41.2
	1146.5	690.0	443.9	+12.6	979.4	626	387.3	−33.9

Table 2. *Continued*

	1979–80				1980–1			
	P	ET	Q	S	P	ET	Q	S
Oct	121·6	41	1.5	+79.1	158.7	35	47.7	+76.0
Nov	71.0	27	11.0	+33.0	92.7	31	42.2	+19.5
Dec	217.3	31	89.0	+97.3	87.4	21	55.5	+10.9
Jan	74.8	12	80.8	−18.0	46.4	16	44.0	−13.6
Feb	118.2	16	100.6	+1.6	70.4	19	27.0	+24.4
Mar	132.2	33	68.6	+30.6	173.3	30	120.3	+23.0
Apr	20.8	76	63.1	−118.3	30.7	55	32.0	−56.3
May	37.4	112	10.7	−85.3	108.7	76	18.3	+14.4
Jun	123.8	95	6.7	+22.1	41.9	85	12.5	−55.6
Jly	55.7	98	4.4	−46.7	46.7	87	5.8	−46.1
Aug	59.9	89	3.1	−32.2	43.5	105	2.4	−63.9
Sep	93.8	70	1.7	+22.1	136.7	64	4.0	+68.7
	1126.5	700	441.2	−14.7	1037.1	624	411.7	+1.4

	1981–2				1982–3			
	P	ET	Q	S	P	ET	Q	S
Oct	136·4	44	17.0	+75.4	136.2	31	3.0	+102.2
Nov	76.8	19	43.6	+14.2	141.9	17	54.0	+70.9
Dec	200.3	12	106.0	+82.3	168.4	8	216.4	−56.0
Jan	67.8	16	132.9	−81.1	175.4	14	98.2	+63.2
Feb	62.9	15	44.3	+3.6	30.8	13	57.6	−39.8
Mar	164.6	33	91.4	+40.2	45.5	26	14.7	+4.8
Apr	18.7	54	21.7	−57.0	136.2	62	23.4	+50.8
May	27.6	100	10.5	−82.9	147.1	67	129.7	−49.6
Jun	76.1	85	6.1	−15.0	30.5	79	23.6	−72.1
Jly	67.8	94	3.8	−30.0	27.7	107	8.5	−87.8
Aug	71.4	68	1.6	+1.8	23.7	92	4.1	−72.4
Sep	61.1	49	0.5	+11.6	134.0	58	2.8	+73.2
	1031.5	589	479.4	−36.9	1197.4	574	636.0	−12.6

	1983–4				1984–5			
	P	ET	Q	S	P	ET	Q	S
Oct	90·9	43	6.6	+41.3	114.7	35	1.4	+78.3
Nov	52.7	16	6.0	+30.7	212.7	19	68.5	+125.2
Dec	171.9	16	85.3	+70.6	85.6	15	73.2	−2.6
Jan	253.2	19	155.0	+79.2	73.0	10	38.9	+24.1
Feb	108.0	20	109.1	−21.1	51.0	16	55.5	−20.5
Mar	61.5	27	26.1	+8.4	81.6	40	19.6	+22.0
Apr	8.6	82	21.9	−95.3	82.8	56	72.8	−46.0
May	76.7	74	14.5	−11.8	42.6	73	18.3	−48.7
Jun	34.4	115	8.2	−88.8	77.1	106	7.8	−36.7
Jly	14.8	99	3.7	−87.9	42.6	88	3.9	−49.3
Aug	41.8	76	1.4	−35.6	128.8	93	5.6	+30.2
Sep	106.5	73	1.1	+32.4	33.6	46	3.1	−15.5
	1021.0	660	438.9	−77.9	1026.1	597	368.6	+60.5

Table 3. *Monthly and yearly runoff totals converted from the discharge at the middle weir*

	1973–4	1974–5	1975–6	1976–7	1977–8	1978–9	1979–80	1980–1	1981–2	1982–3	1983–4	1984–5	Total	Mean
Oct	3.0	51.8	0.2	105.5	2.6	1.8	1.5	47.7	17.0	3.0	6.6	1.4	242.1	20.2
Nov	3.4	65.5	0.0	47.9	14.7	0.8	11.0	42.2	43.6	54.0	6.0	68.5	357.6	29.8
Dec	16.9	62.3	12.7	108.2	49.4	15.3	89.0	55.5	106.0	216.4	85.3	73.2	890.2	74.2
Jan	107.3	97.7	8.8	90.0	65.9	70.3	80.8	44.0	132.9	98.2	155.0	38.9	989.8	82.5
Feb	164.0	68.7	16.7	127.9	109.6	82.8	100.6	27.0	44.3	57.6	109.1	55.5	963.8	80.3
Mar	49.5	42.3	21.4	62.6	83.6	84.7	68.6	120.3	91.4	14.7	26.1	19.6	684.8	57.1
Apr	21.2	27.6	21.7	39.3	51.2	64.5	63.1	32.0	21.7	23.4	21.9	72.8	460.4	38.4
May	18.0	17.7	7.7	15.5	36.0	20.6	10.7	18.3	10.5	129.7	14.5	18.3	317.5	26.5
Jun	11.6	6.8	3.2	9.4	11.9	31.0	6.7	12.5	6.1	23.6	8.2	7.8	138.8	11.6
Jly	5.1	4.1	0.9	5.4	7.5	9.4	4.4	5.8	3.8	8.5	3.7	3.9	62.5	5.2
Aug	2.9	1.8	0.1	3.3	7.3	3.9	3.1	2.4	1.6	4.1	1.4	5.6	37.5	3.1
Sep	14.5	1.2	1.1	3.9	4.2	2.2	1.7	4.0	0.5	2.8	1.1	3.1	40.3	3.4
Year	417.4	447.5	94.5	618.9	443.9	387.3	441.2	411.7	479.4	636.0	438.9	368.6	5185.3	
Average	34.8	37.3	7.9	51.6	37.0	32.3	36.8	34.3	40.0	53.0	36.6	30.7	432.1	

Table 4. *Monthly and yearly rainfall totals taken as the average of the two meteorological stations*

Month	1973–4	1974–5	1975–6	1976–7	1977–8	1978–9	1979–80	1980–1	1981–2	1982–3	1983–4	1984–5	Total	Mean
Oct	53.8	72.6	36.0	243.4	68.0	6.2	121.6	158.7	136.4	136.2	90.9	114.7	1238.5	103.2
Nov	53.4	122.5	73.9	107.4	109.0	56.6	71.0	92.7	76.8	141.9	52.7	212.7	1170.6	97.6
Dec	73.9	105.9	53.9	133.5	106.9	207.4	217.3	87.4	200.3	168.4	171.9	85.6	1612.4	134.4
Jan	188.5	159.6	51.2	126.5	150.2	114.6	74.8	46.4	67.8	175.4	253.2	73.0	1481.2	123.4
Feb	189.2	29.9	48.6	174.9	232.5	72.8	118.2	70.4	62.9	30.8	108.0	51.0	1189.2	99.1
Mar	61.3	82.4	82.9	104.0	112.0	133.3	132.2	173.3	164.6	45.5	61.5	81.6	1234.6	102.9
Apr	19.2	51.5	6.8	32.8	49.6	83.0	20.8	30.7	18.7	136.2	8.6	82.8	540.7	45.1
May	94.3	20.8	45.0	51.2	58.1	123.3	37.4	108.7	27.6	147.1	76.7	42.6	832.8	69.4
Jun	64.5	8.3	13.8	93.7	33.6	33.3	123.8	41.9	76.1	30.5	34.4	77.1	631.0	52.6
Jly	43.1	53.6	8.7	20.3	119.6	21.9	55.7	46.7	67.8	27.7	14.8	42.6	522.5	43.5
Aug	90.5	38.4	53.7	117.5	89.0	104.0	59.9	43.4	71.4	23.7	41.8	128.8	862.1	71.8
Sep	204.8	151.9	209.1	22.2	18.0	23.0	93.8	136.7	61.1	134.0	106.5	33.6	1194.7	99.6
Year	1136.5	897.4	683.6	1227.4	1146.5	979.4	1126.5	1037.0	1031.5	1197.4	1021.0	1026.1	12510.3	1042.5
Average	94.7	74.8	57.0	102.3	95.5	81.6	93.9	86.4	86.0	99.8	85.1	85.5	1042.5	

winter months, with the seven months, September to March showing above average amounts. Potential evapotranspiration is greatest during the summer months, with the increase in temperatures and the growth of vegetation, and all six months from April to September show above average amounts. The amount of water leaving the basin as runoff was highest in the winter with above average quantities in the four months December to March but with very low amounts in the four months June to September. In theory, the storage component of the water balance equation should show a surplus of water entering

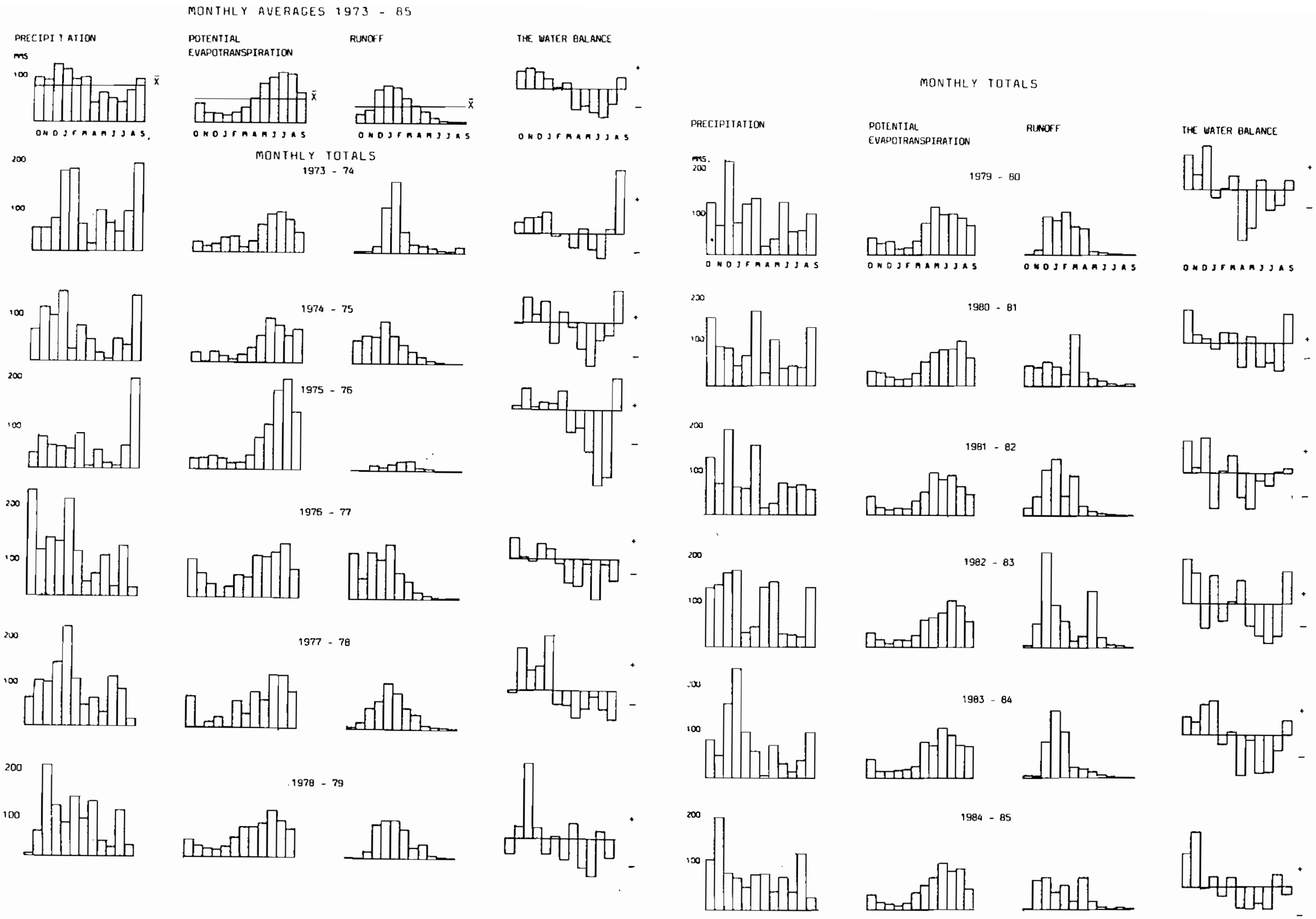


FIG. 8.

Monthly estimates of the four major components of the water balance for the average of the twelve years 1973–1985 and for each year individually. Precipitation values are an average of the amounts recorded at two meteorological stations and stream discharge is converted from the flow recorded at the middle weir at 0900 GMT each day.

storage during the winter months, October to March, and an overall removal from storage (deficit) during the summer months, April to September. The results show the expected pattern with the exception of a surprisingly large surplus in September, due to the unexpectedly high rainfall in several Septembers. This is illustrated by the facts that in two of the years during the study period, September was the wettest month and in six out of the twelve years it had over 100 mm of rain. The overall balance shows an average loss of 40 mm per year, an amount at least partially explained by the excess of potential over actual rates of evapotranspiration during the summer months. It is important to remember that it is during the summer, when the soil moisture deficit is highest, that there will be the greatest difference between potential and actual rates. Another problem concerning the reliability of the potential evapotranspiration figures arises from the change in method from measurement to calculation at the start of the 1979 water year.

Results for the individual water years (Fig. 8 and Table 2) show some interesting patterns. 1975 had a low total rainfall of 897.7 mm resulting in a significant soil and groundwater deficit at the end of the year. Under normal conditions this deficit would have been largely replenished by winter rainfall but in the winter of 1975–6 this was well below average with only 347 mm falling during October to March—compared to a twelve year average of 660.4 mm. This, combined with an unusually hot summer in 1976 with resulting high rates of evapotranspiration, reduced the depleted storage and, consequently, the stream discharge to such an extent that only 94.5 mm left the drainage basin as runoff compared to a twelve year average of 432.3 mm. Both these years show a significant total annual loss of water from storage in excess of 200 mm.

Several other water years showed slight losses and only 1974 a significant gain. These results, although demonstrating general trends, need to be viewed with an awareness of potential inaccuracies in measurement, especially with respect to the estimates of evapotranspiration.

The Rainfall–Runoff Relationship

Annual and monthly totals of rainfall and runoff are shown in Tables 3 and 4, and in Fig. 9. The average amount of rain per year was 1043.3 mm. 1976 was the driest water year with only 683.6 mm, and 1977 the wettest with 1227.4 mm. Rain fell on an average of 196 days in each year, ranging from 140 days (1976) to 216 (in 1974). On average, 63% of the rain fell during the winter months, October to March, with a range of 51% in 1974 to 72% in both 1977 and 1984. Over the twelve years, December was the wettest month and July the driest, with averages of 134 mm and 44 mm respectively. However, April was the driest month in 6 out of the 12 years, and September the most variable, having above average rain in eight years and being the wettest in two of them, but in contrast also being the driest in two other years. The wettest month in the records was January 1984 with 253.2 mm of rain and the driest was October 1978 with 6.2 mm.

Over the 12 water years, 41% of rainfall reappeared as runoff varying from 14% in 1976 to 53% in 1983. 1976 was an exceptional water year; the next lowest was 36% in 1985. Seasonally 52% of winter (October to March) rainfall reappeared as runoff compared to 23% of summer (April to September).

The period of highest flow represented by the three months December, January and February accounts for 55% of the rain falling during those months, compared to only 3% in the period of low flow, represented by the months July, August and September. During

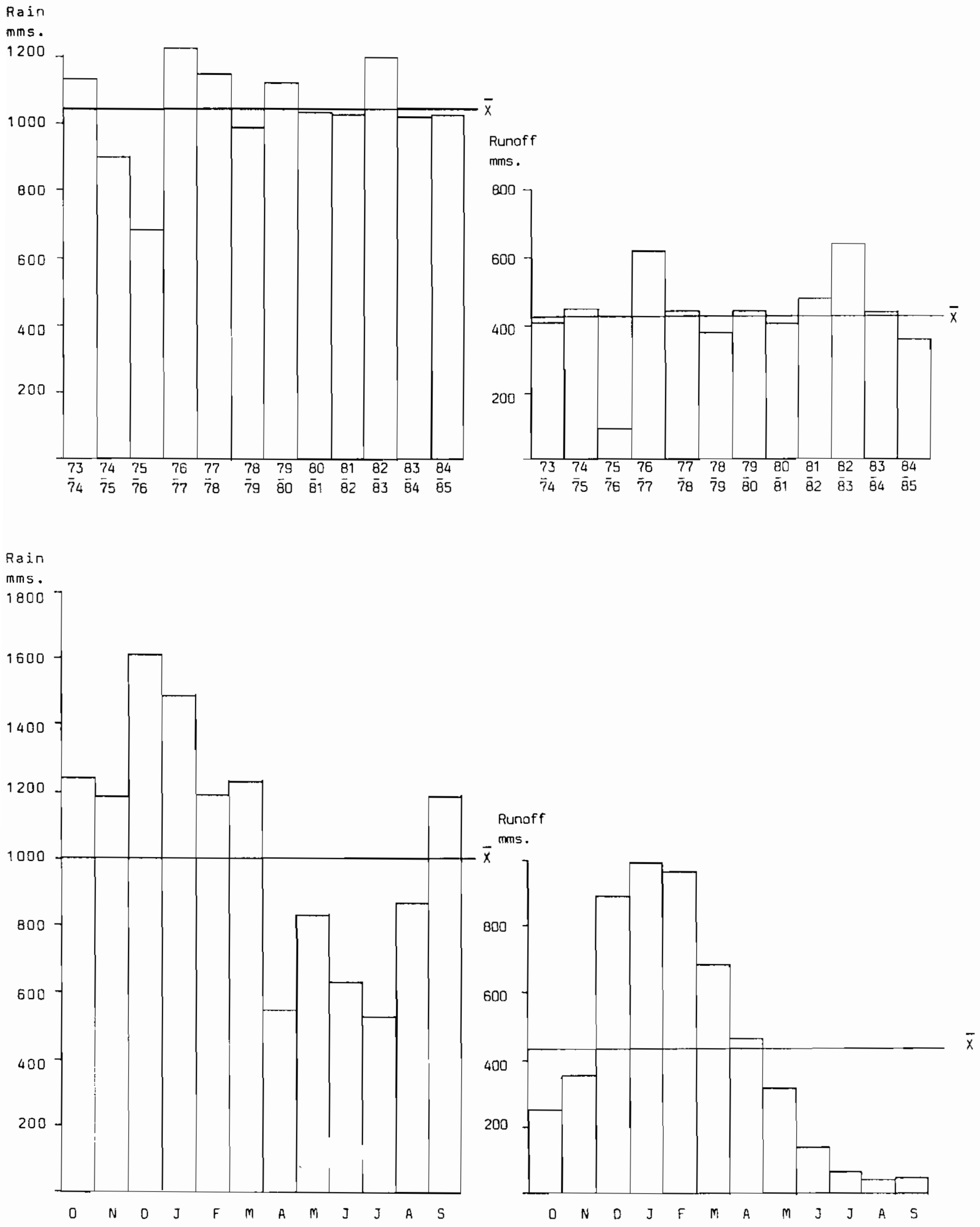
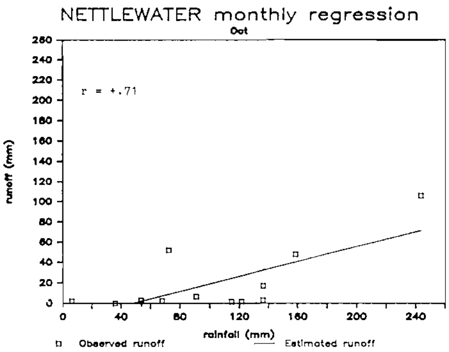
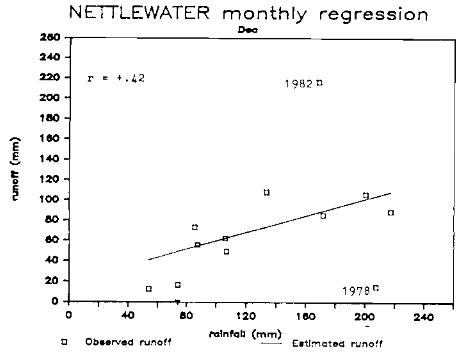


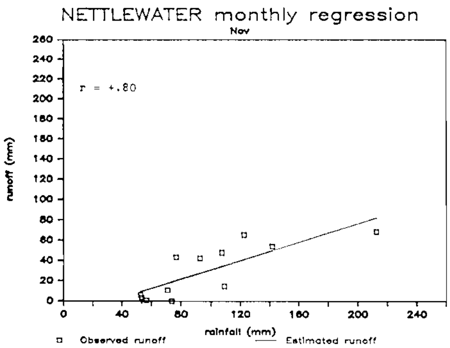
FIG. 9.
(a) Annual rainfall totals. (b) Annual runoff totals. (c) Monthly rainfall totals. (d) Monthly runoff totals.

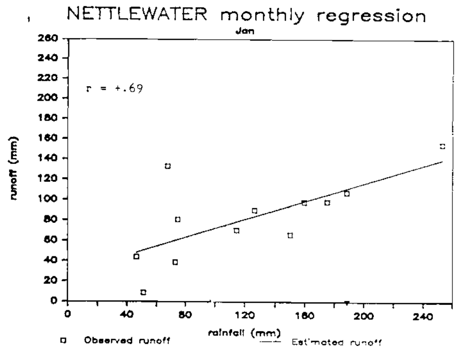
these months, water loss by evaporation and transpiration is greatest but large amounts of rainfall remain, held in the soil and rocks as storage.

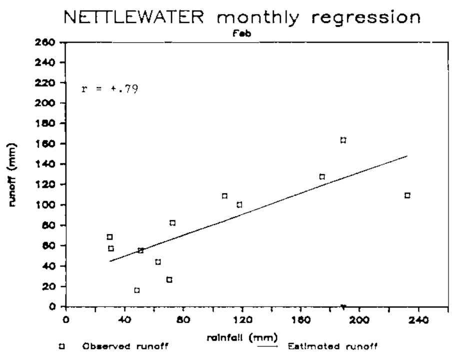
Taking the twelve years as a whole, 19% of all runoff occurs in January representing the equivalent of 989.8 mm of rain. The lowest proportion 0.7% occurs in August, representing 37.9 mm.

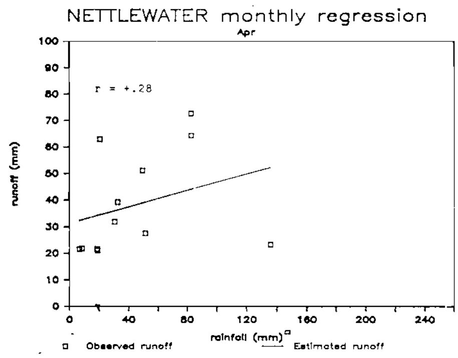

NETTLEWATER monthly regression
Oct
r = +.71
runoff (mm)
rainfall (mm)
Observed runoff
Estimated runoff

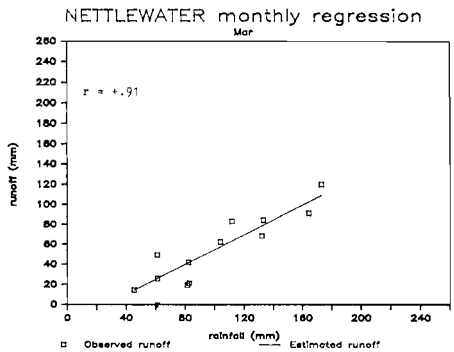

NETTLEWATER monthly regression
Dec
r = +.42
1982
1978
runoff (mm)
rainfall (mm)
Observed runoff
Estimated runoff

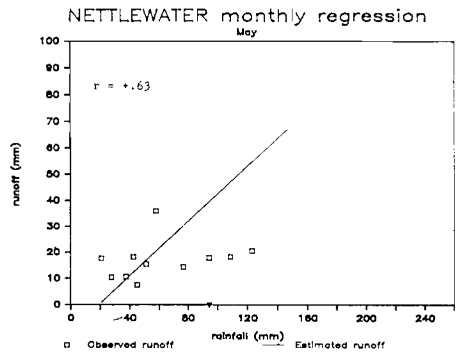

NETTLEWATER monthly regression
Nov
r = +.80
runoff (mm)
rainfall (mm)
Observed runoff
Estimated runoff


NETTLEWATER monthly regression
Jan
r = +.69
runoff (mm)
rainfall (mm)
Observed runoff
Estimated runoff


NETTLEWATER monthly regression
Feb
r = +.79
runoff (mm)
rainfall (mm)
Observed runoff
Estimated runoff


NETTLEWATER monthly regression
Apr
r = +.28
runoff (mm)
rainfall (mm)
Observed runoff
Estimated runoff


NETTLEWATER monthly regression
Mar
r = +.91
runoff (mm)
rainfall (mm)
Observed runoff
Estimated runoff


NETTLEWATER monthly regression
May
r = +.63
runoff (mm)
rainfall (mm)
Observed runoff
Estimated runoff

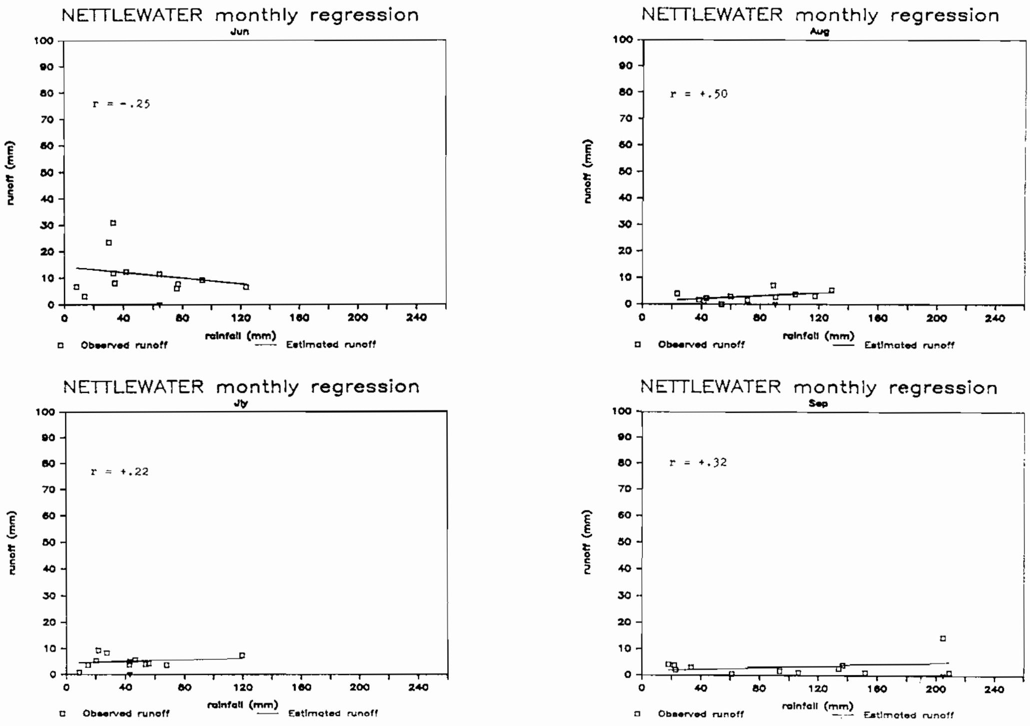


FIG. 10.

Monthly regressions between rainfall and runoff. Correlation coefficients calculated using the least squares method (Gregory, 1963). The runoff for May 1983 is greater than the scale of the graph and is omitted. The figures are 129.7 mm of runoff against 147.1 mm of rainfall.

The runoff pattern is much less evenly distributed than that of rainfall, the contrast resulting from storages within the catchment. It is, therefore, not surprising that there is no simple relationship between rainfall and runoff. Correlation coefficients (r) were calculated using the least squares method (Gregory, 1963) for monthly totals of rainfall and runoff and the results are shown graphically in Fig. 10. As expected the winter months show the most striking relationships. All winter months (except December) show correlations significant at the 0.1% level. The lack of a significant relationship in December is due to the response of the stream in 1978 and 1982. Autumn 1978 was unusually dry with a combined rainfall for October and November of only 63 mm (compared with a twelve year average of 201 mm). Storage in the soil and rock was therefore unusually low, resulting in much of the 207 mm of rain that fell in December entering storage rather than leaving the drainage basin as runoff. In contrast, the autumn of 1982 was wetter than average with 278 mm of rain in October/November and a further 168 mm falling during December. The large quantities of water already held in storage resulted in the runoff for December being greater than expected. By March, storage in the soil and rocks is at its maximum and it is then that rainfall and runoff show the closest correlation (+0.91).

During the summer months the stream does not show a significant response to rainfall. This lack of a relationship is due to two factors; the high rates of evapotranspiration during the summer months and the large amounts of water entering storage. May is the only

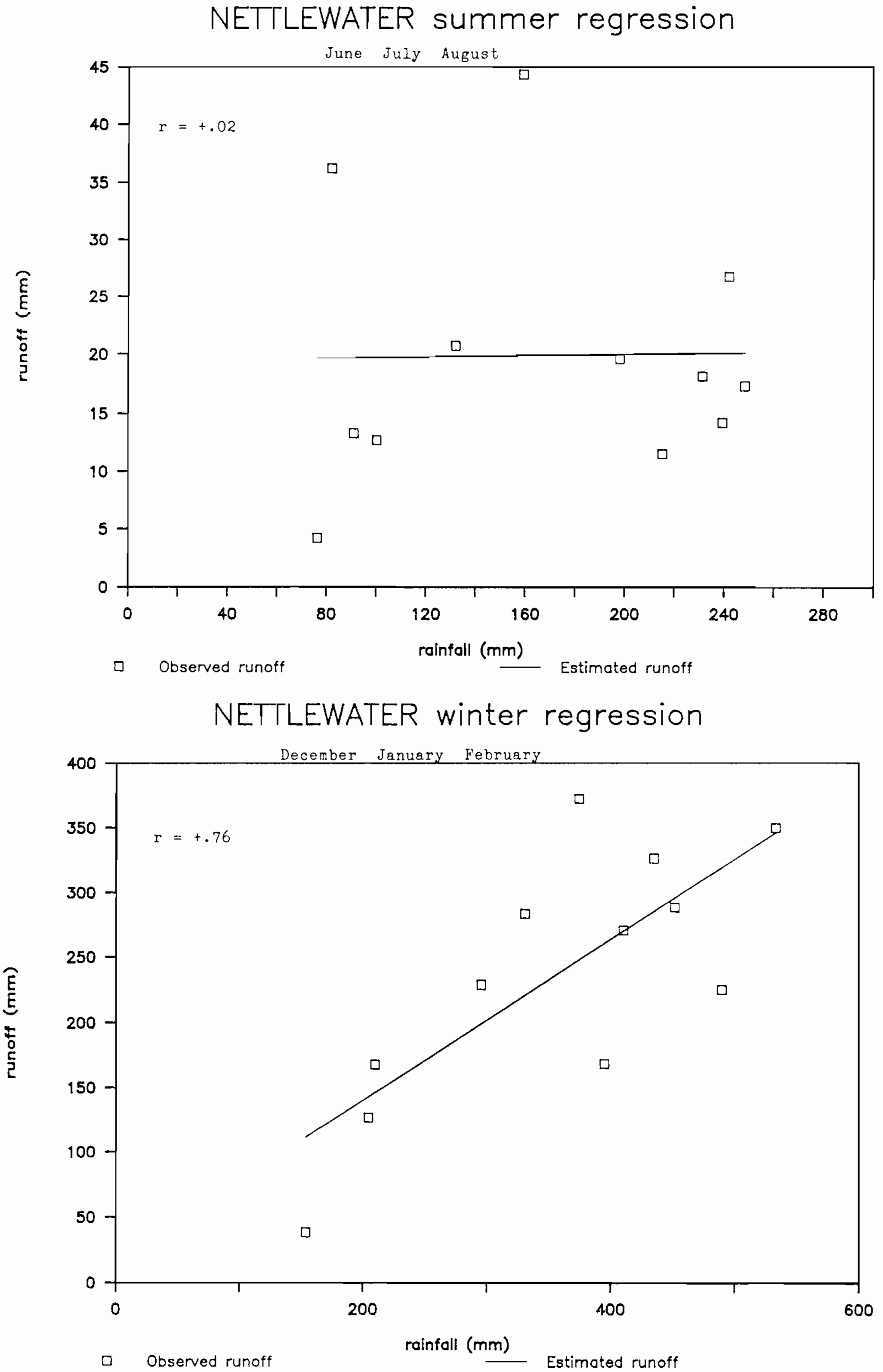


FIG. 11.

(a) Winter regression between rainfall and runoff, calculated using the sum of the totals for the months December, January and February. (b) Summer regression between rainfall and runoff, calculated using the sum of the monthly totals for June, July and August.

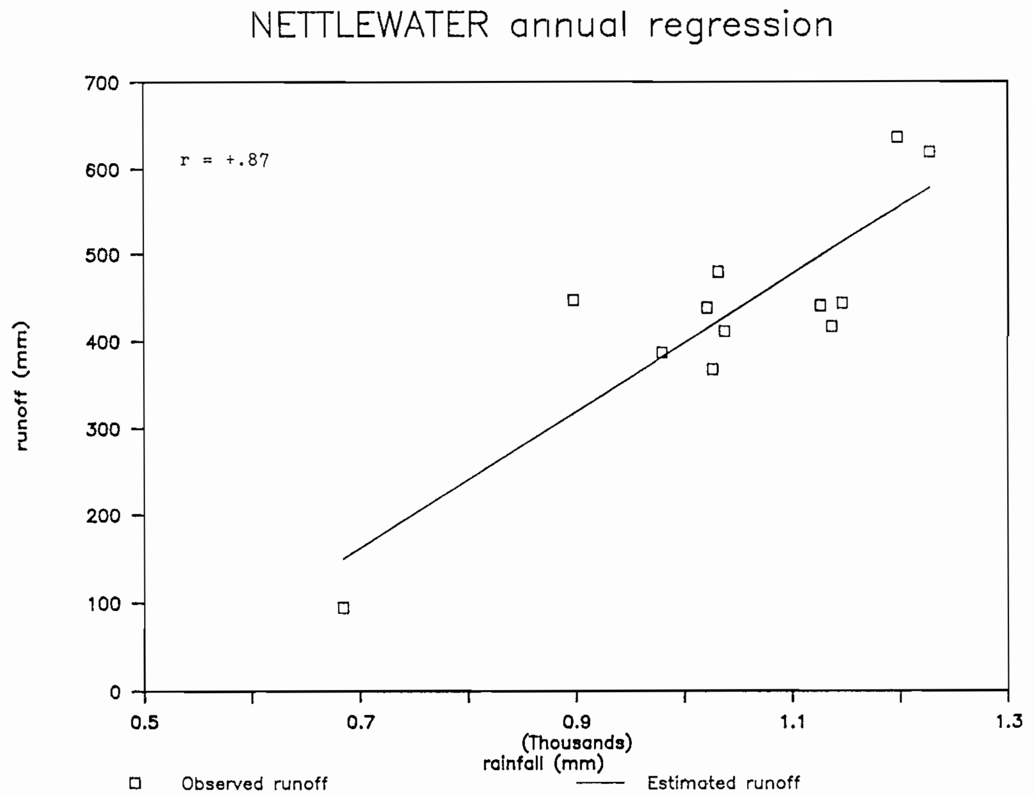


FIG. 12.
Annual regression between rainfall and runoff, calculated using the sum of the totals for all twelve months.

month to show a significant relationship (at the 0.5% level), largely resulting from the conditions in 1983. In that year April had had an unusually high total rainfall (136.2 mm, compared with a twelve year average of 45.1 mm) followed by 147.1 mm in May (compared with an average of 69.4 mm). As a result, the soil moisture deficit was considerably less than normal for the time of year, resulting in less rainfall entering the soil and rock as storage and, consequently, a high level of runoff. Seasonal aggregates of monthly totals show similar results; the totals for December, January and February give a correlation coefficient of +0.76, significant at the 0.1% level (Fig. 11a), whereas those for June, July and August give a coefficient of +0.02, showing no association between rainfall and runoff during those months (Fig. 11b). The annual totals show a correlation significant at the 0.1% level with a coefficient of +0.87 (Fig. 12).

The seasonal variations within a catchment are known as the *river regime*. Most rivers in Britain have a simple regime of one period of high runoff followed by one period of low runoff. Although it is normal practice to demonstrate a regime with a run of data at least 30 years long, the mean monthly runoff for the years 1973 to 1985 has been used to construct Fig. 13. Mean monthly rainfall is included for comparison. Runoff shows the expected pattern whereas rainfall is less regular.

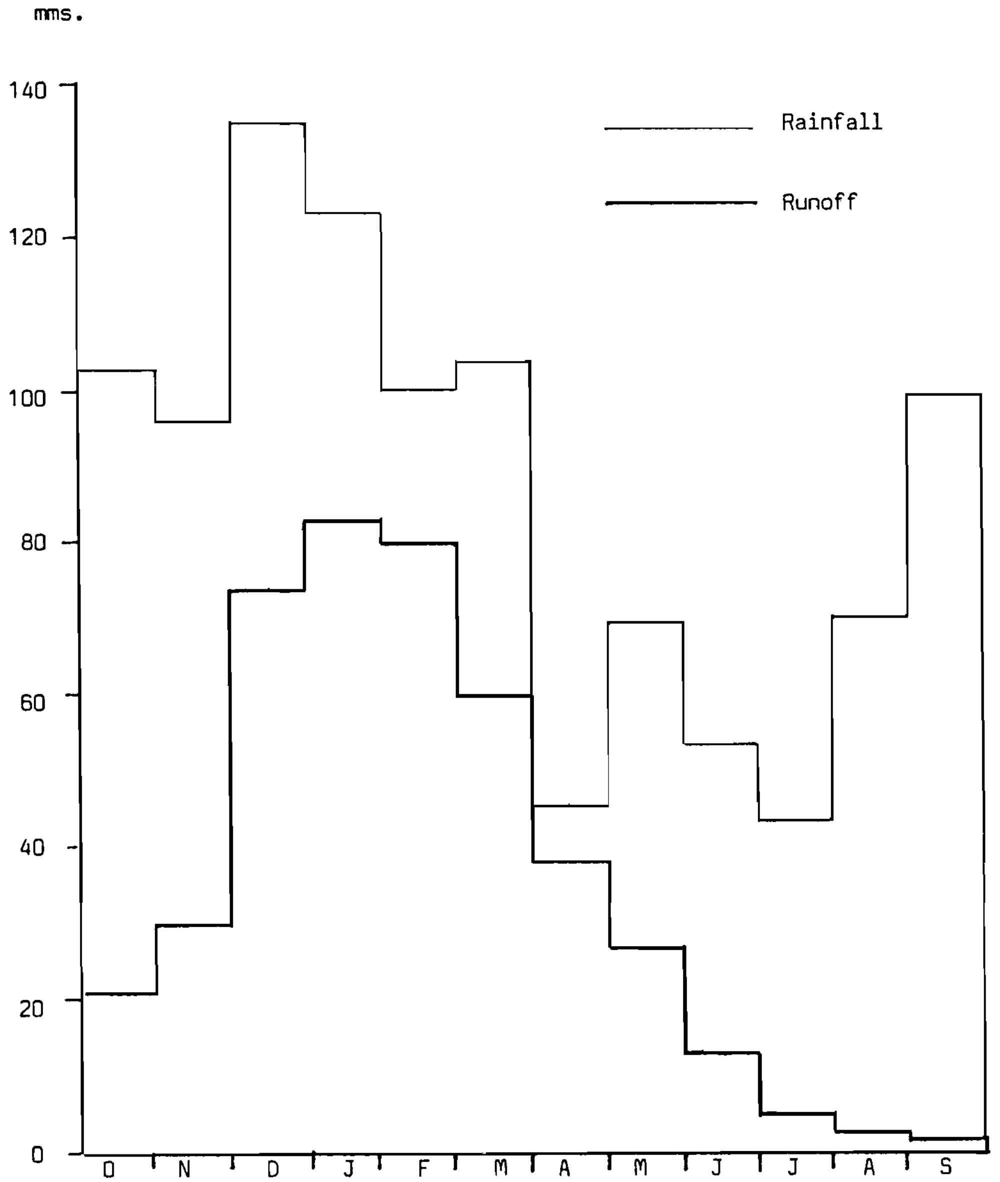


FIG. 13.
Regime diagram, showing the seasonal variation in rainfall and runoff. The monthly figures are the averages for the twelve years 1973–1985.

The variability in both rainfall and runoff is shown in Fig. 14, where the average for each month is expressed as a fraction of the average for all 144 months in the 12 years, referred to as the overall average (Table 5). A figure of 1 therefore represents a monthly average equal to the overall average. The results show a range in rainfall from 0.5 to 1.6 times the overall mean but, although there is a winter maximum and a summer minimum, the pattern is irregular. Runoff shows a much greater variability, with monthly figures falling as low as 0.09 times the overall mean in August and September and rising to 2.3 times the overall mean in January, but with a more regular distribution.

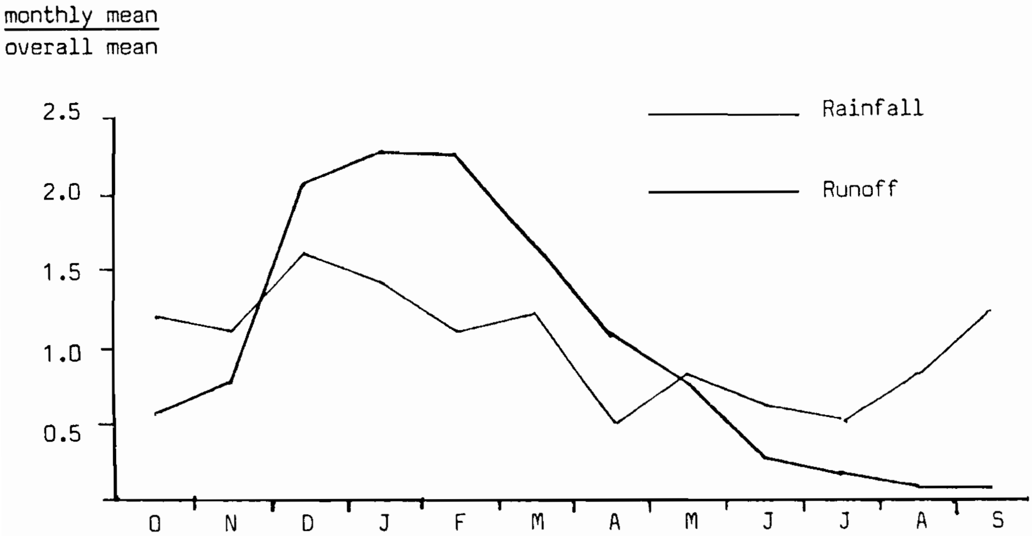


FIG. 14.
Variability in rainfall and runoff. The average for each month is expressed as a fraction of the average for all 144 months. The figures are shown in Table 5.

Table 5. *Variability in rainfall and runoff for the twelve water years1973–85*

overall mean (of 144 months) for rainfall 86.9 mm
overall mean (of 144 months) for runoff 36.0 mm

	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep
1. Mean monthly rainfall		2. Mean monthly rainfall/overall mean										
1.	103.2	97.6	134.4	123.4	99.1	102.9	45.1	69.4	52.6	43.5	71.8	99.6
2.	1.2	1.1	1.6	1.4	1.2	1.2	0.5	0.8	0.6	0.5	0.8	1.2
3. Mean monthly runoff		4. Mean monthly runoff/overall mean										
3.	20.2	29.8	74.2	82.5	80.3	57.1	38.4	26.4	11.6	5.2	3.1	3.4
4.	0.6	0.8	2.1	2.3	2.2	1.6	1.1	0.7	0.3	0.1	0.09	0.09

This same pattern is shown in Fig. 15, where monthly rainfall and runoff are expressed as cumulative percentages of the annual rainfall for each year. Whereas rainfall shows a fairly regular increase, runoff has reached it maximum level by March or April in every year except 1983, when May showed a secondary rise. In every case these graphs represent the increased loss of water by evapotranspiration and the influence of storage during the summer months. As discussed in the water balance section, the unusual pattern of runoff during the 1976 water year, reflects both the soil moisture deficit which was carried over from the previous year and the small amount of rainfall during that year itself.

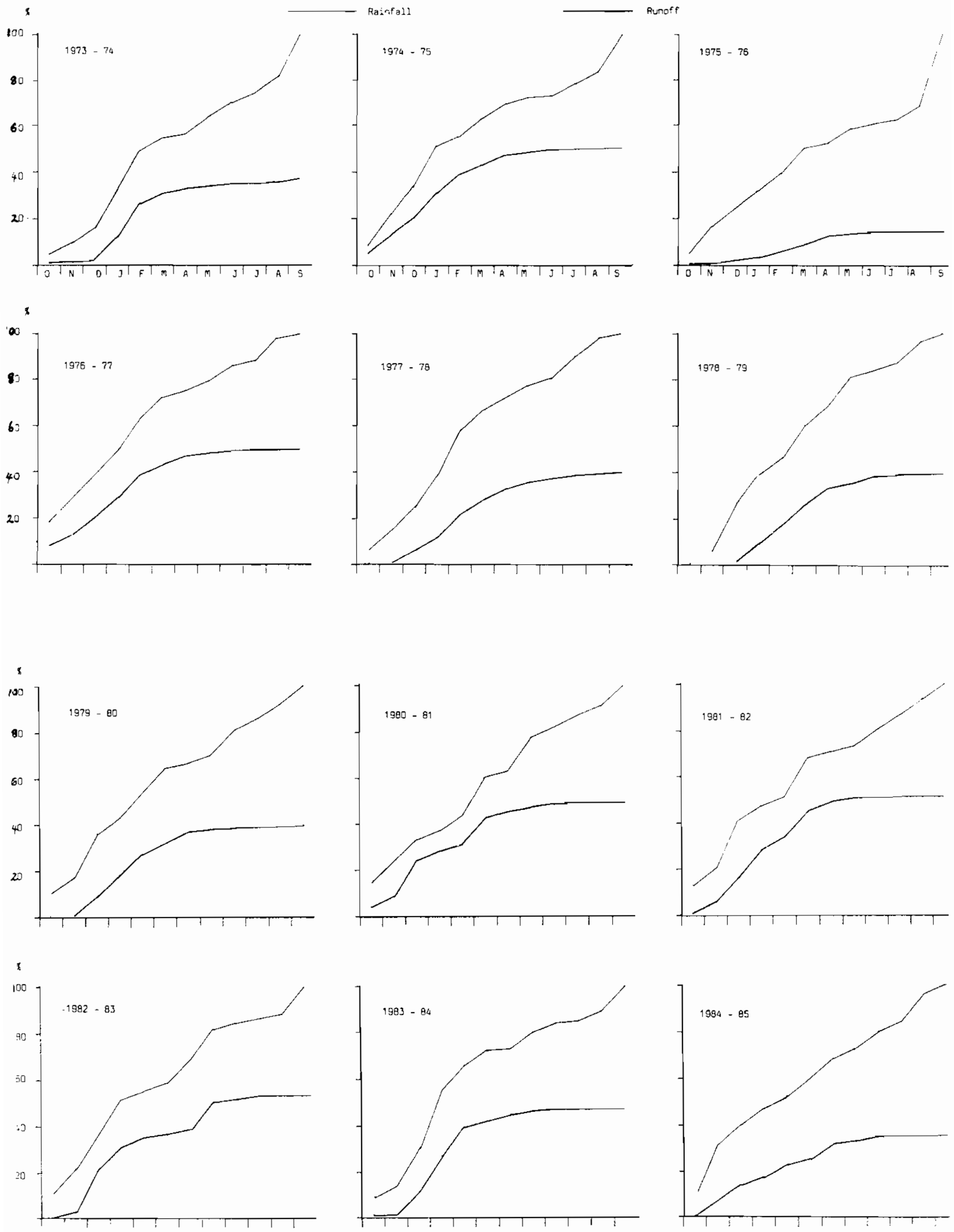


FIG. 15.
Monthly rainfall and runoff expressed as cumulative percentages of the annual rainfall for each year.

Flow Duration

So far, the discussion has dealt entirely with stream discharge converted to a runoff equivalent of rainfall, but a considerable amount of extra information is available from the discharges themselves. A *hydrograph* is a graph of runoff and the annual hydrographs for the years 1973 to 1985 are shown in Fig. 16. From them it can be seen that the stream's response to rainfall is highly variable and shows no standard pattern. However, it is possible to separate flow entering the stream directly from flow delayed through storage, by identifying the greatest flow sustained by the stream without additional rainfall. From the hydrographs this can be seen to be 16.5 ls^{-1}, reached in February and March 1980.

The variability in flows can be demonstrated by the *flow duration curve* which shows the percentage of time by which individual flows are equalled or exceeded. As this curve itself gives no indication of season its usefulness can be increased by combining it with a simple regime diagram. The flow duration diagram in Fig. 17 represents all flows from 1973 to 1985.

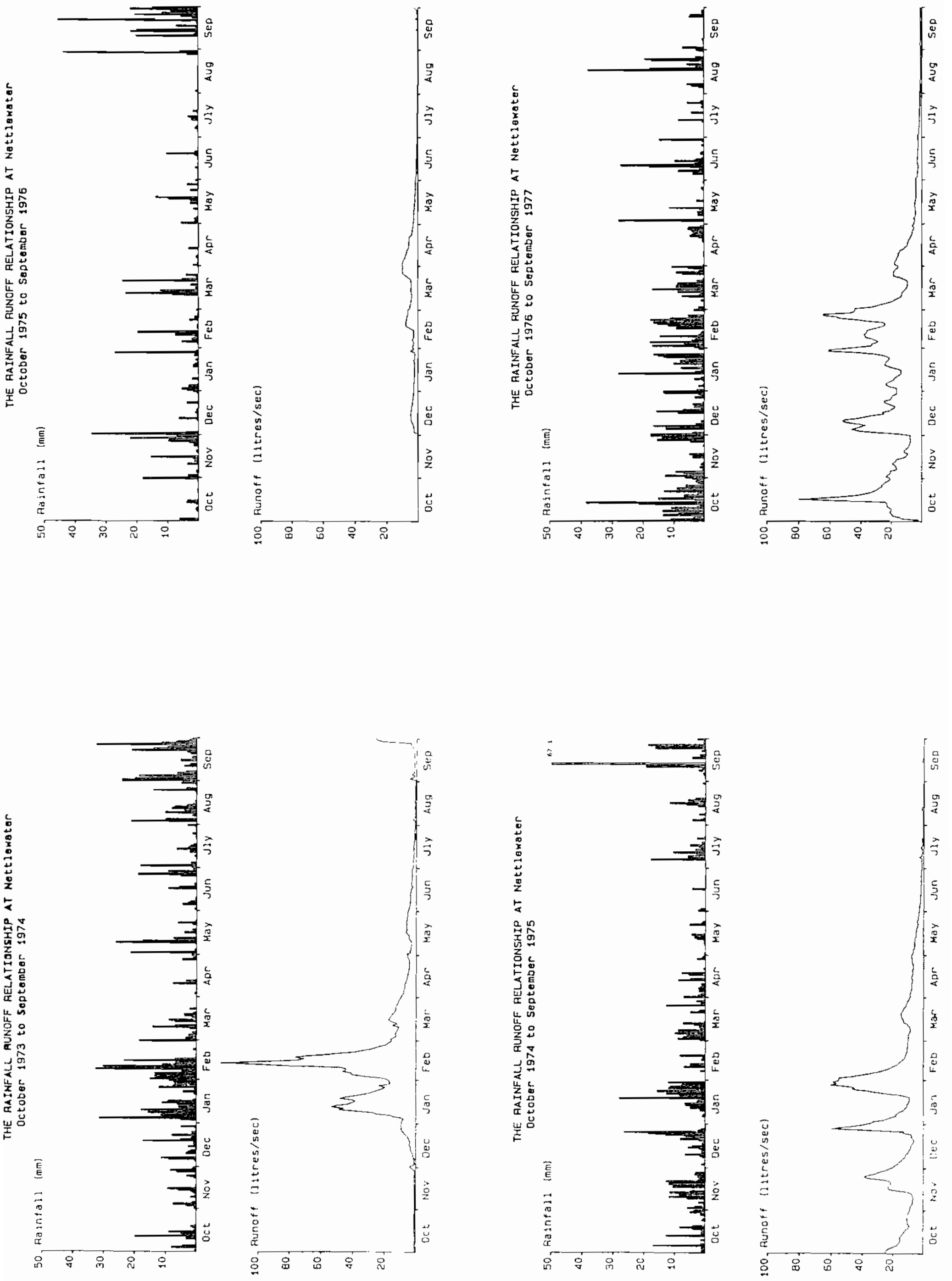
From this diagram it can be seen that the discharge of the stream varied from 124 ls^{-1}, to nil, the latter reflecting occasional periods when the stream dried out. Discharges are less than 16.5 ls^{-1} for 80% of the time, representing the period when the stream is largely maintained by sub-surface, delayed flow, contributions. The 20% of time when flows are in excess of 16·5 ls^{-1}, represents the period when baseflow is augmented by quickflow and rapid throughflow, conditions which occur mainly during the winter months and especially during December, January and February. Flows greater than 80 ls^{-1} were only recorded on 33 occasoins, representing 0.8% of the time and all associated with periods of heavy rainfall or storm events.

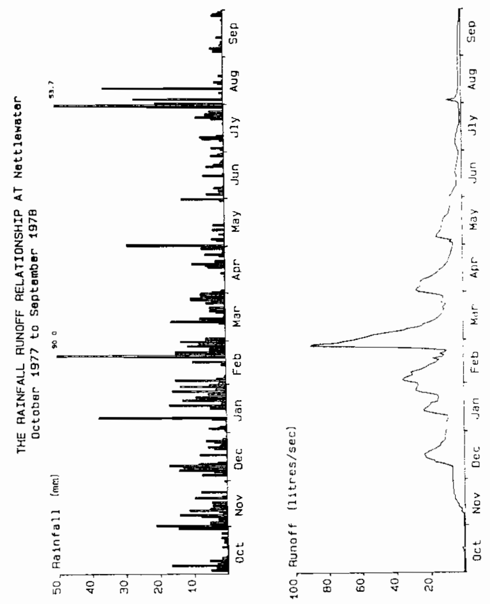
Storm Hydrographs

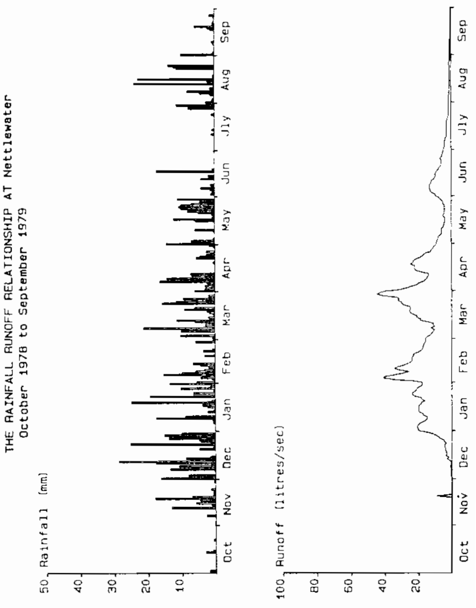
In many ways, the influence of drainage basin characteristics on runoff is best illustrated by the response of the stream to storms, the effect often being intense but short term. This response is recorded in the *storm hydrograph* and, although largely determined by the basin characteristics, it is also influenced by the spatial and temporal distribution of the rainfall.

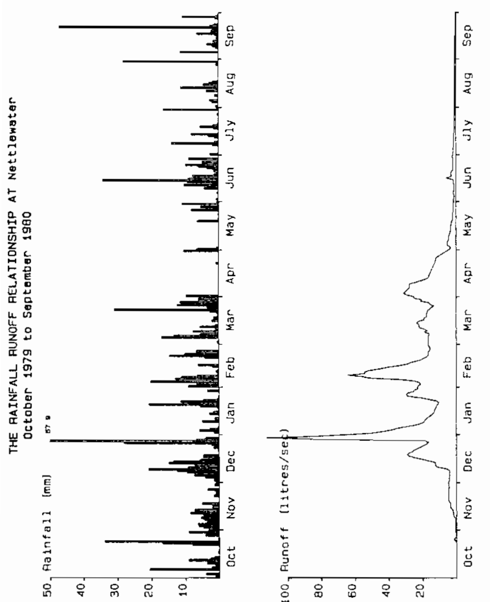
Storm hydrographs are often characterized by an almost immediate response to heavy rainfall shown by a rapid rise in discharge followed by an almost equally rapid decline. This peak is referred to as the *quickflow peak* and results from channel precipitation and overland flow, two water sources which are speedily depleted. The time interval between the most intense rainfall and the quickflow peak is referred to as the *lag time*, and the rise in discharge as the *peak rise*. Following the quickflow peak there is often a second, slower, rise in discharge resulting from throughflow, especially under high moisture conditions, and referred to as *delayed flow*. Rises in groundwater also contribute to delayed flow. On the hydrograph itself the two types of flow can be separated by drawing a line from the start of the hydrograph rise to an inflection point on the recession limb. Although this technique has been used in this study, the identification of these points is to some extent arbitrary.

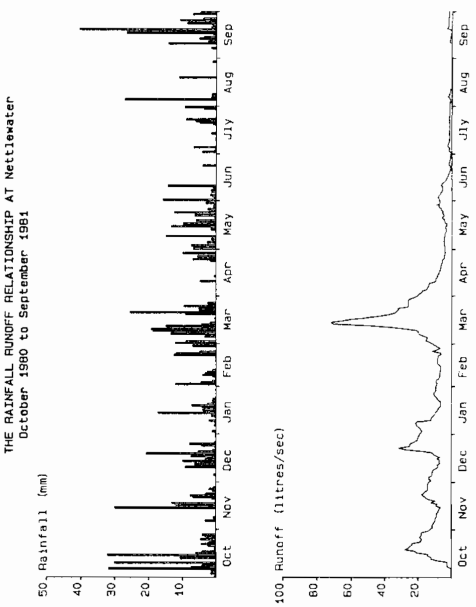
The response of the stream to storms is well illustrated by three storm hydrographs recorded during the water year 1983. The storms in question occurred in December 1982, May 1983 and September 1983. Those in December and May were very similar in both the initial amount of rain and its distribution, although the situation in December was complicated by several heavy falls subsequent to the first storm. The rainfall in September, although similar in quantity, was spread over a much longer period of time, and was,


THE RAINFALL RUNOFF RELATIONSHIP AT Nettlewater
October 1973 to September 1974
Rainfall (mm)
Runoff (litres/sec)
THE RAINFALL RUNOFF RELATIONSHIP AT Nettlewater
October 1974 to September 1975
Rainfall (mm)
Runoff (litres/sec)
THE RAINFALL RUNOFF RELATIONSHIP AT Nettlewater
October 1975 to September 1976
Rainfall (mm)
Runoff (litres/sec)
THE RAINFALL RUNOFF RELATIONSHIP AT Nettlewater
October 1976 to September 1977
Rainfall (mm)
Runoff (litres/sec)
Oct Nov Dec Jan Feb Mar Apr May Jun Jly Aug Sep


THE RAINFALL RUNOFF RELATIONSHIP AT Nettlewater
October 1977 to September 1978
Rainfall (mm)
Runoff (litres/sec)


THE RAINFALL RUNOFF RELATIONSHIP AT Nettlewater
October 1978 to September 1979
Rainfall (mm)
Runoff (litres/sec)


THE RAINFALL RUNOFF RELATIONSHIP AT Nettlewater
October 1979 to September 1980
Rainfall (mm)
Runoff (litres/sec)


THE RAINFALL RUNOFF RELATIONSHIP AT Nettlewater
October 1980 to September 1981
Rainfall (mm)
Runoff (litres/sec)

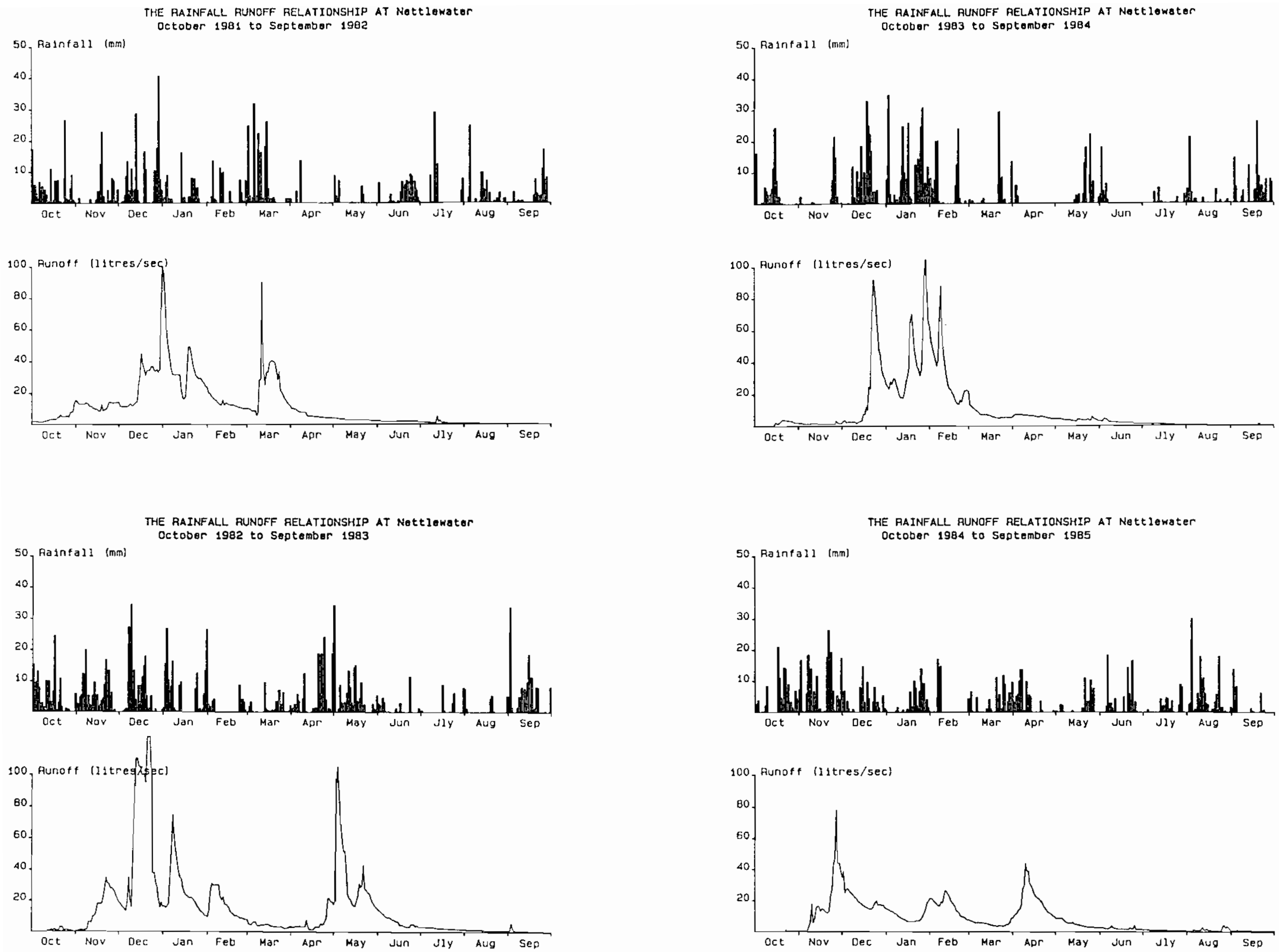


FIG. 16.

Annual hydrographs and daily rainfall for the years 1973–85. The hydrographs are based on stream discharge recorded at the middle weir at 0900 G.M.T. each day and the rainfall is the average recorded at the two meteorological stations.

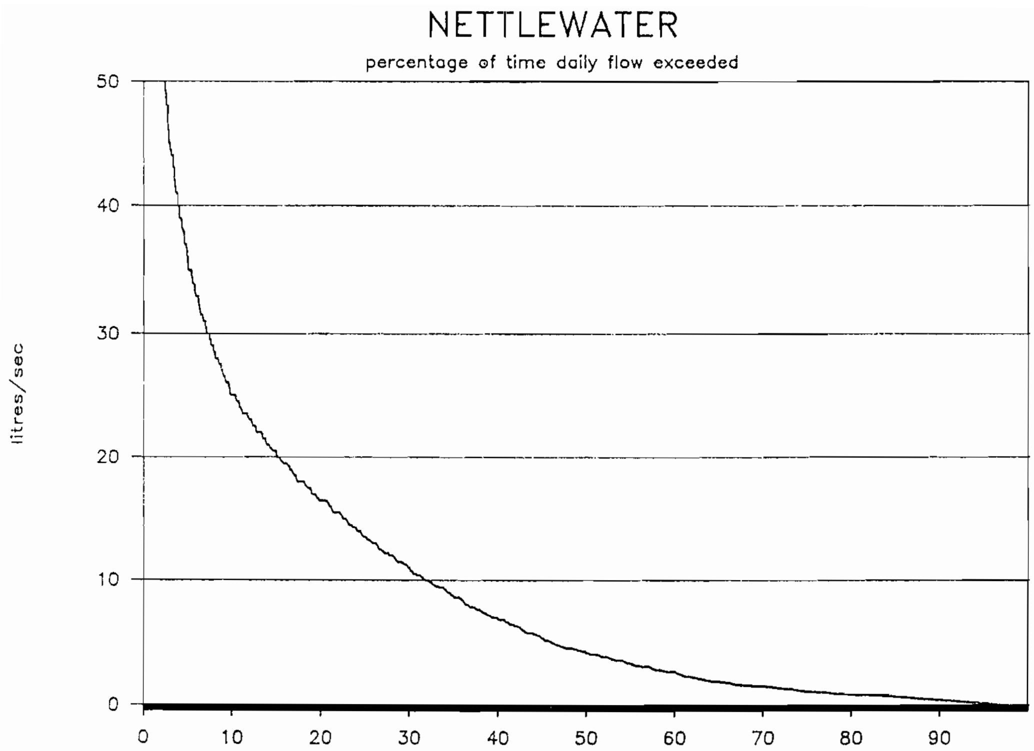


FIG. 17.
Flow duration curve for the middle weir for the twelve years 1973–85.

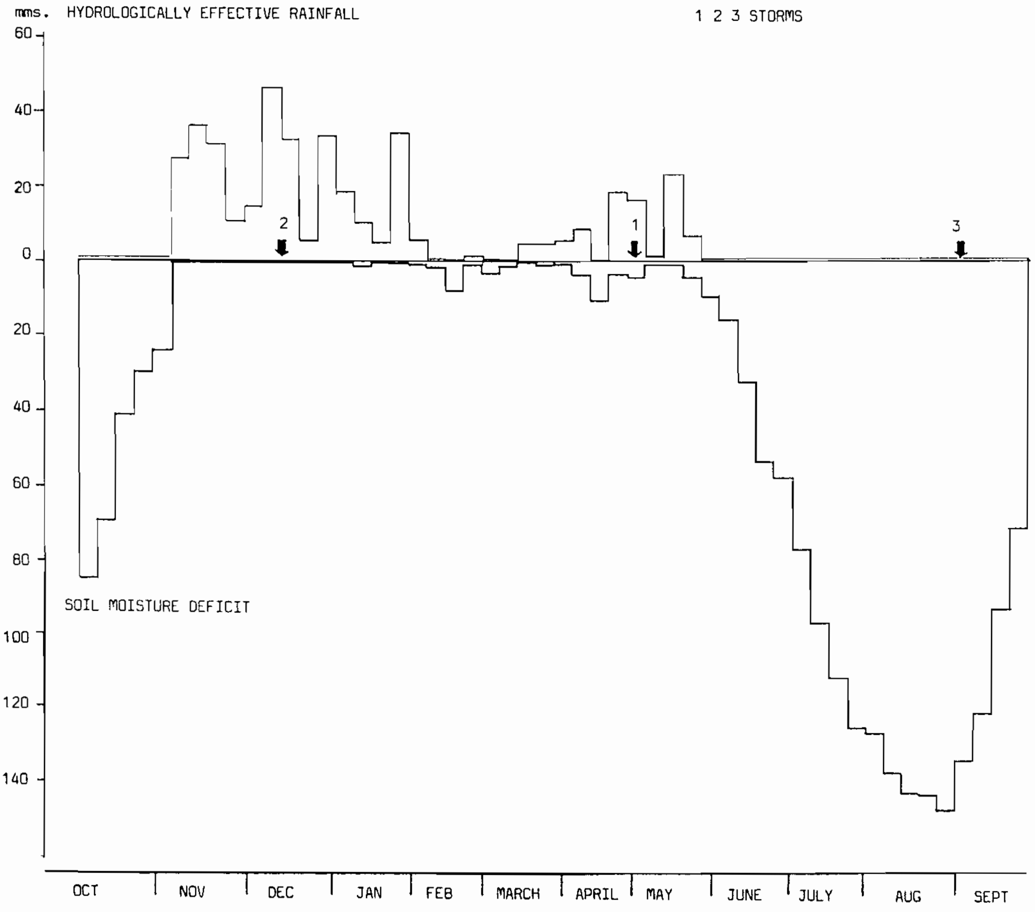


FIG. 18.
Weekly estimates of hydrologically effective rainfall and soil moisture deficit calculated by MORECS (Meteorological Office Rainfall and Evaporation Calculation System), for the water year 1982–83.

therefore, of considerably lower intensity. The relationship of these storms to antecedent moisture conditions can be seen in Fig. 18, which shows the weekly estimates of the soil moisture deficit calculated by MORECS.

Of these three storms, the May response (Fig. 19a) is most typical of a single storm event occurring under relatively high moisture conditions. This storm is therefore discussed first, before the one which occurred in the preceding December. In the 30 days prior to the storm, there had been 136 mm of rain, 81 mm of which had fallen in the preceding 10 days. On 1 May, 31.5 mm of rain fell over a period of 10 hours with a maximum intensity of 6.5 mm h^{-1}. Stream discharge at the start of the storm was 18 ls^{-1}. Within two hours of the maximum rainfall intensity, the stream had reached a quickflow peak of a 130 ls^{-1}, a peak rise of 112 ls^{-1}. This peak resulted from channel precipitation and saturated overland flow. It was one of the rare occasions when water could be seen flowing over the floodplain (Plate 1). During the quickflow peak the concentration of suspended load increased visibly (Plate 2), but unfortunately was not measured. The sources of this quickflow were rapidly depleted and within one hour the stream had fallen to 36 ls^{-1}; the amount of water which had left the drainage basin as quickflow was equivalent to 0.41 mm, or 1.3% of the rainfall. Following this fall in discharge, the stream rose gradually over the next 21 hours reaching a secondary throughflow peak 24 hours after the period of maximum rainfall intensity. Five days later the discharge of the stream levelled off at 28 $l\,s^{-1}$.

The December storm (Fig. 19b) occurred at a time of year when moisture conditions were generally higher than in the succeeding May, although the period immediately preceding the storm had been considerably drier. In the 30 days prior to the storm, 114.6 mm of rain had fallen but only 20.7 mm of this fell in the previous 10 days. 29.9 mm of rain fell, over a period of 11 hours, on 9 December with a maximum intensity of 7.3 mm h^{-1}. The intensity and duration of the rainfall was, therefore, reasonably similar to that of the May storm, although spread over a slightly longer period. However, the response of the stream was complicated by further heavy falls on succeeding days. The discharge of the stream before the storm was the same as in May, 18 ls^{-1}.

The initial response of the stream was in many ways comparable to that described for May. The lag time to the quickflow peak was two hours although the peak rise was only 66 ls^{-1}. However the quickflow peak lasted considerably longer than the one in May, six hours compared with two. During the quickflow peak, the runoff was equivalent to 0.9 mm or 3.0% of the rainfall. The differences in the response of the stream to the two storms is probably related to the rainfall in the 10 days preceding the storm and in the temporal distribution of the rain during the storm itself. There is no simple throughflow peak comparable to that which occurred in May, but the increased discharge resulting from throughflow continued for several weeks due to continued replenishment by repeated rainfall. However, it should be pointed out that the plateau nature shown by the throughflow component of this hydrograph is unusual and probably results from an equipment failure or limitation.

In contrast to both these, the storm in September occurred after a fairly prolonged dry period, when only 82 mm of rain had fallen in two months. 13.9 mm fell in the 30 days prior to the storm with 9.1 mm in the preceding 10 days. The discharge of the stream before the storm was 1.1 ls^{-1}. The storm itself was more prolonged than either of the other two, with 38 mm of rain falling over a period of 26 hours with a maximum intensity of 6 mm h^{-1}.

The stream rose to two quickflow peaks, each one half hour after a period of heavy rainfall. The first peak reached a discharge of 7.7 and the second 4.5 ls^{-1}; in both cases

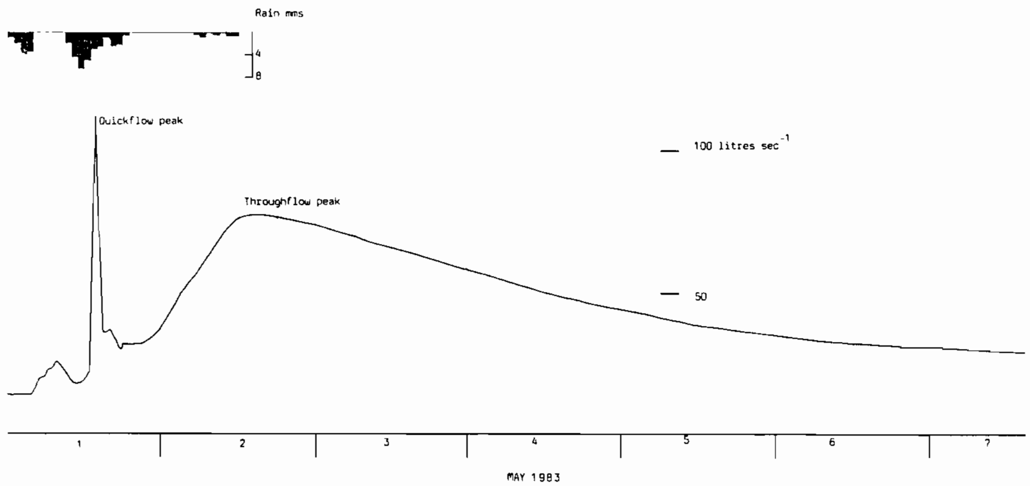


FIG 19(a).
Storm hydrograph for 1–7 May 1983. Middle weir

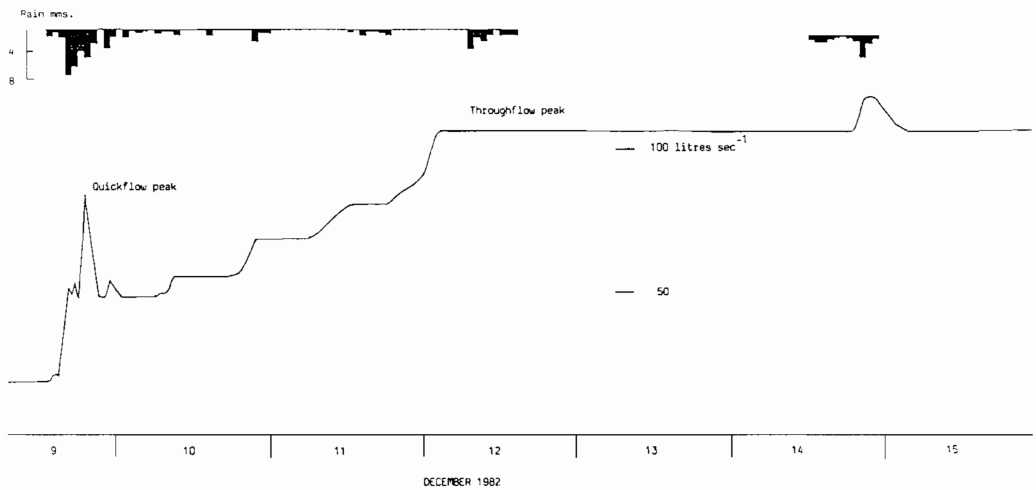


FIG. 19(b).
Storm hydrograph for 9–15 December 1983. Middle weir

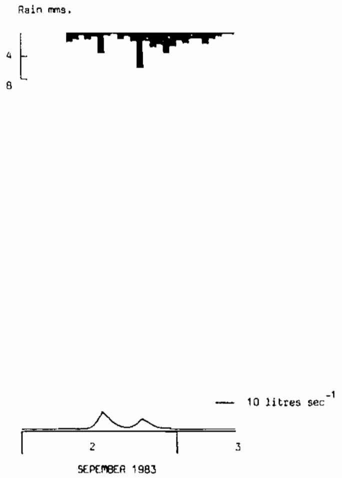


FIG. 19(c).

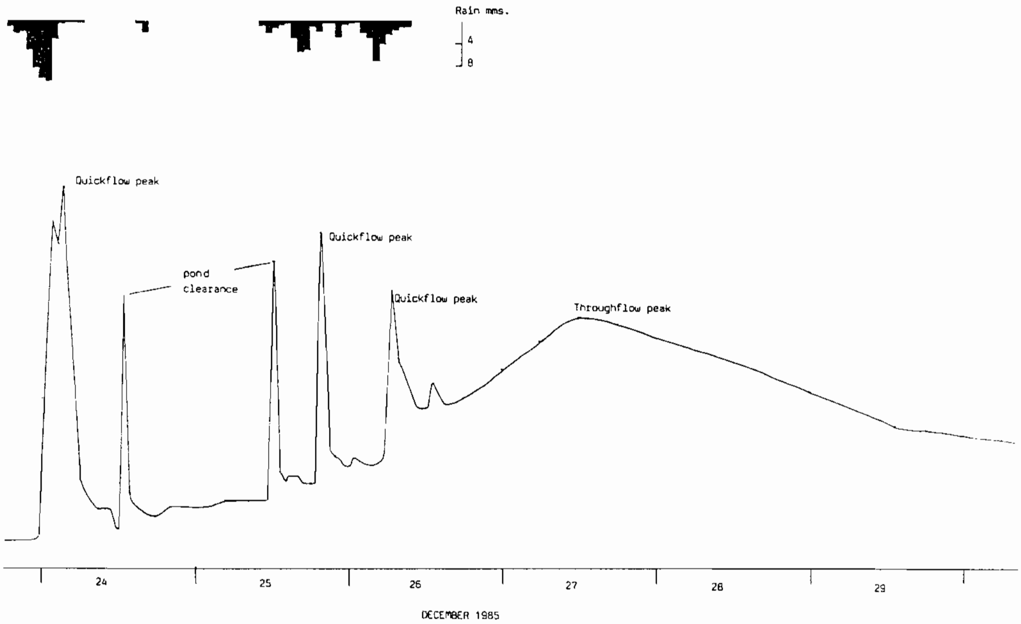


FIG. 19.

Storm hydrographs taken from the middle weir. (c) Storm hydrograph for 2–3 September 1983. (d) Storm hydrograph for 24–29 December 1985, showing the artificial peaks created by clearance of the pond outflow.

these peaks will have represented channel precipitation alone. There was no throughflow peak, a situation typical of summer storms, and again, illustrating the importance of evaporation, transpiration and storage during summer months.

From these three storms a number of assumptions can be made concerning the drainage basin. As saturated overland flow is a relatively rare occurrence in the catchment, a significant part of the quickflow peak must result from direct channel precipitation. Other small catchment studies have come to similar conclusions (Troake and Walling, 1973; Weyman, 1974). In a sample of ten storms occurring at all times of the year, an average of only 1.6% of the rainfall left the drainage basin as quickflow, varying from 0.1% in a July storm to 4.6% in a December storm. When saturated overland flow does occur it can only be from a relatively small proportion of the catchment and is probably largely restricted to the flat floodplain areas adjacent to the stream; and the hollow at the source, where moisture conditions are high and saturation occurs relatively easily.

Throughflow is probably too slow to contribute to the quickflow peak directly, although it may contribute indirectly by feeding the saturated area at the base of the slope. It does, however, provide a significant contribution to delayed flow, especially in periods of high antecedent moisture conditions, as shown by the May and December hydrographs. Although the soils throughout the catchment show poor horizon differentiation the upper part is underlain by relatively impermeable metamorphosed rock which encourages the lateral movement of throughflow at the junction between the soil and the rock, thereby increasing the throughflow contribution. In contrast the lack of a secondary peak on the September and other summer hydrographs demonstrates the importance of evapotranspiration and storage during the summer months.

The response of the stream to storm rainfall is undoubtedly complicated by the pond at its source, but to date, no attempt has been made to quantify that influence. However, it can be assumed that the pond catches and stores water that would otherwise leave the catchment as runoff, and it certainly receives the throughflow from the area of the catchment above it. Figure 20 shows the response of all three weirs to the storm of May 1983, and the flattening of the throughflow peak at the upper weir, sited immediately downstream of the pond, almost certainly reflects the influence of the pond. The outlet of the pond is cleared at irregular intervals, giving rise to artificial "quickflow" peaks as illustrated on the storm hydrograph for December 1985 (Fig. 19d).

The Groundwater Table

The position of the boreholes drilled across the valley is shown in Fig. 21a, and the results of measurements taken from them in Fig. 21b. Unexpectedly, the results show the top of the saturated zone to be inversely related to relief. At this site, the valley is wider, and the thickness of periglacial head is considerably greater than further upstream and it is possible that these two factors are combining to create a "sponge" effect. In other words, as the stream flows from the restricted valley at the middle weir to the open valley at the lower weir, the water seeps from the channel into the permeable head deposit, creating a "groundwater mound" above the regional water table and with a hydraulic gradient away from the stream. This would explain the decrease in discharge recorded between the middle and lower weirs at low flows.

There are, however, other possible explanations for this decrease in discharge. There may be structural weaknesses in the bedrock, allowing "leakage" into an adjacent catchment, or a similar process may be operating through some old drainage scheme associated

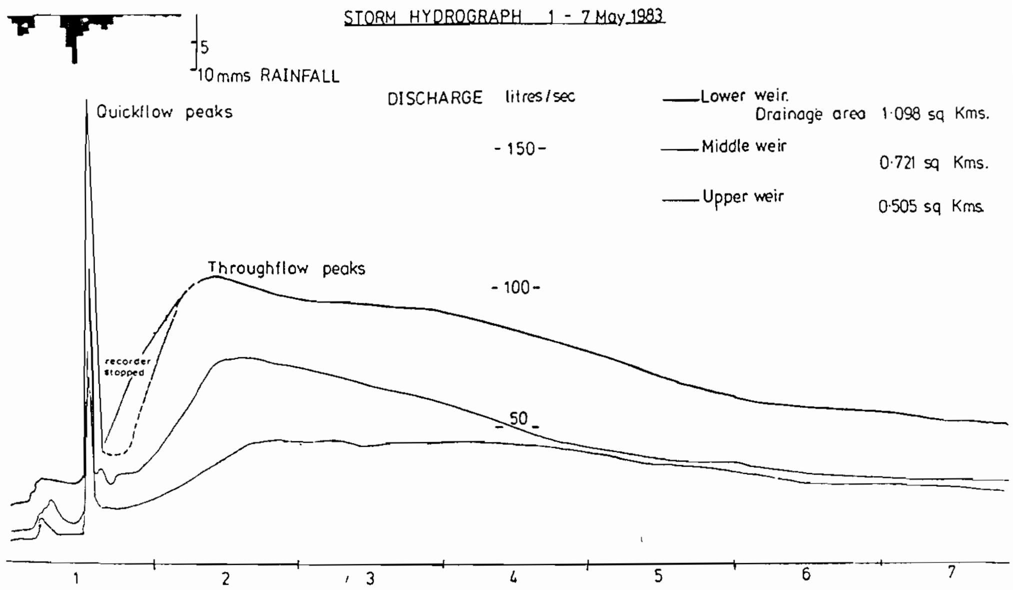


Fig. 20.
Storm hydrograph for 1–7 May 1983 showing the response of the stream at all 3 weirs.

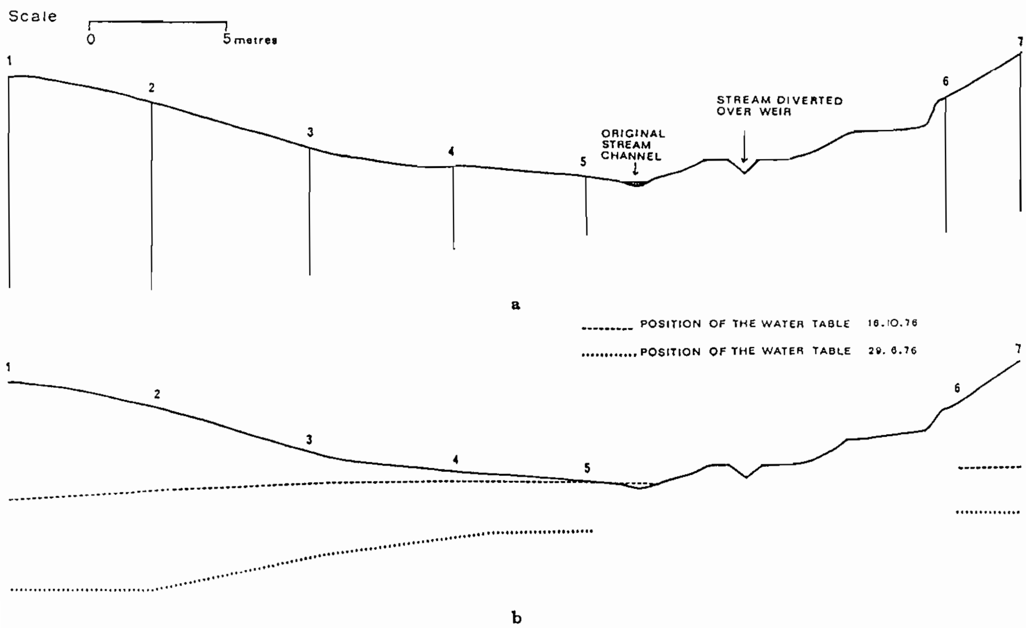


Fig. 21.
The transect line of bore holes adjacent to the lower weir. (a) Position and depth of bore holes. (b) The assumed height and shape of the water table as recorded from the bore holes in the summer and autumn of 1976.

with the landscaping of the estate. It is known that the stream was channelled underground during the eighteenth century, but as yet the full extent of the scheme has not been discovered.

Conclusion

Although some of the methods in this study are open to criticism (shedding doubt on the calculated results—especially with regard to the water balance), the experiments were designed to illustrate techniques and provide background data. In this context the results have been invaluable for both geographical and ecological freshwater courses at the Leonard Wills Field Centre. In spite of the inaccuracies associated with the water balance, interesting patterns have been identified—especially the highly seasonal variation in runoff, reflecting the importance of both evapotranspiration and storage during the summer months.

This relationship is illustrated even more clearly by the detailed information concerning rainfall and runoff shown in the annual and storm hydrographs, the flow duration curve and the seasonal totals. The storm hydrographs enable separation of quickflow and delayed flow, and show that quickflow (which results mainly from channel precipitation and saturated overland flow) accounts for a relatively small proportion of total runoff. Throughflow makes a significant contribution during periods of high moisture conditions, but during periods with a soil moisture deficit, large amounts of moisture are lost by evapotranspiration. Infiltrating rainfall is held in storage and does not generate runoff.

These trends are reflected in the monthly and seasonal comparisons between rainfall and runoff with significant correlations during the winter months and a lack of correlation during the summer months.

The results all suggest that there is considerable storage potential in the soils of the catchment, and the need for detailed information concerning both this and the movement of water within the soil has become apparent. Consequently a grid of soil moisture tensiometers accompanied by several throughflow troughs has been established during 1984 and 1985 and it is hoped that the results now being collected from these will help provide a more detailed explanation of the streams response to rainfall.

Glossary

baseflow
Water which maintains the flow of streams and rivers, even during periods of no rainfall, by the slow release from storage within the rocks. Baseflow may be loosely equated with throughflow (q.v.).

channel precipitation
Precipitation which falls directly into the stream or river channel.

depression storage
Water held in hollows on the ground surface.

delayed flow
Precipitation reaching the stream via the soil and/or rocks.

effective rainfall
The proportion of rainfall that reaches the stream or river directly, i.e. the proportion that generates quickflow.

evapotranspiration
Water loss from evaporation of water used in transpiration from the leaf surfaces of the vegetation cover, as well as that lost by direct evaporation.

field capacity
The state of the soil after rainfall, when excess moisture has had time to drain away and the rate of downward movement has materially decreased.

flow duration curve
A graph showing the percentage of time when individual flows are equalled or exceeded.

groundwater
Water held in the rock and contributing the *baseflow* of the stream or river.

hydrograph
A graph of runoff against time.

hygroscopic water
Water held in the area surrounding individual soil particles by strong molecular forces.

interception
Precipitation caught by the vegetation cover some of which is evaporated directly back to the atmosphere.

lag time
In a storm the lag time is the time interval between the most intense rainfall and the quickflow peak discharge.

peak rise
In a storm the peak rise is the rise in discharge from the start of the storm to the quickflow maximum.

permeability
The capacity of soil or rock to allow the movement of water.

porosity
The water holding capacity of soil or rock.

potential evapotranspiration
The rate at which evaporation and transpiration would take place if the soil was constantly at *field capacity*.

quickflow peak
The initial peak in stream discharge following heavy rain.

runoff
Stream or river discharge.

saturated overland flow
Surface runoff resulting from saturation of the surface layers, usually in the area immediately adjacent to the stream or river.

soil moisture deficit
The amount of water needed to return the soil to *field capacity.*

stemflow
Precipitation *intercepted* by the vegetation cover which reaches the ground by flowing down plant stems, tree trunks, etc.

throughfall
Precipitation *intercepted* by the vegetation cover which reaches the ground by dripping through the canopy.

throughflow
The downslope movement of water through soil.

Acknowledgements

The recording for this study has been undertaken by many staff at the Leonard Wills Field Centre. We would like to extend our thanks to all of them and in particular to the Warden, Dr J. H. Crothers. The recordings obtained from the boreholes were made possible thanks to the staff of the Department of Civil Engineering and Construction, Hatfield Polytechnic.

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PLATE 1(a).
Saturated overland flow and the swollen stream in the Nettlecombe valley, 1 May 1983. (Photo: D. H. Dalby)

PLATE 1(b).
The usual appearance of Nettlecombe Court, Late April 1985. (Photo: J. H. Crothers)

PLATE 2(a).
The lower V-notch weir in the Nettlecombe valley, showing the high level of suspended load on 1 May 1983.
(Photo: H. J. Howcroft)

PLATE 2(b).
The lower V-notch in late April 1985. The cover has been removed from the water level recorder. (Photo: J. H. Crothers)