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RUNOFF AND SEDIMENT OUTPUT FROM A SMALL LOWLAND CATCHMENT—THE EXAMPLE OF PRESTON MONTFORD BROOK, SHROPSHIRE

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ABSTRACT

Data are presented on annual runoff and sediment output and their seasonal variations for a small (3.15 km^2) intensively farmed lowland catchment on drift deposits in the North Shropshire Plain. Simple low cost installations for monitoring rainfall, runoff, evapotranspiration, soil moisture, groundwater, and suspended and bedload sediment are described. Of a mean annual rainfall of 678 mm about one third was discharged as runoff, though problems were encountered in producing an accurate water balance. A highly seasonal pattern of runoff with some 90% occurring between December and April was related to the winter build up of storage consequent upon reduced evapotranspiration, giving increased surface runoff, throughflow and base flow. Variations in runoff associated with eighteen storm events were analysed statistically and strong correlations were obtained with measures of antecedent conditions. Extensive overland flow events were rare, but when high intensity rainfall coincided with high soil water storage, amounts of runoff and sediment loss were dramatic. Flow duration data indicated a 90% occurrence of delayed flow contributions to the stream with rates below 50 l sec^{-1}, while quickflow events were limited to 10% of the flows but gave discharge rates up to 634 l sec^{-1}. Estimated annual suspended sediment output exceeded 100,000 kg compared to a bedload loss of up to 3,300 kg. Bedload movement was confined to high flow events between December and April, though some depletion of load during late winter storms was associated with reduced availability of load. High suspended sediment concentrations (6,756 mg l^{-1}) were associated particularly with high intensity rain falling on well saturated arable fields without complete crop cover. The usefulness of the cascading system model is discussed in relation to the temporal behaviour of the catchment and the importance of negative feedback mechanisms and thresholds in the control of runoff and sediment loss are considered.

INTRODUCTION

Since water is one of man's most basic resources, there is a need to understand the processes which control its quantity and quality, particularly as these processes may be affected by human activity. Considerable progress has been made in hydrology over recent years and field-based process studies have achieved much in helping to explain the controls of runoff, sediment and solute output from drainage basins, in addition to working towards models capable of predicting runoff events. Quantification and explanation in hydrology can be achieved with relatively simple instrumentation (Troake and Walling, 1973; Howcroft, 1977). These and other comparable studies have investigated small catchments (c. 1.0 km^2) with relatively steep slopes, a solid and reasonably watertight geology and limited influences due to arable farming. The aims of this paper are to present the results of three years' measurement of runoff and sediment from an intensively farmed lowland catch-

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ment with gentle slopes and drift deposits on the North Shropshire Plain as well as to propose explanations of the results and to relate these findings to current hydrological theory. Consideration will also be given to the field methods, instrumentation, data collection procedures and analytical methods.

Before proceeding to describe the methods and present findings, a brief review is given of some current ideas concerning the controls of runoff and sediment output from catchments (for more complete accounts reference may be made to Weyman, 1975; Weyman and Weyman, 1977; Hilton, 1979 and Thornes, 1979).

The functioning of a stream catchment may be modelled simply as an input-output system in which water enters the drainage basin as rainfall and is lost from the system either as runoff or evapotranspiration with varying periods of storage, mainly in the soil or in the rocks, between the time of input and that of output. Over long periods of time a balance is to be expected between the amount of rainfall input and the combined outputs, assuming that water is neither entering the catchment from elsewhere by groundwater seepage, nor being lost to other catchments by subsurface flows. Such balances are normally computed from October to September, the start and finish representing periods of usually minimum storage. Shorter term changes in input, output and storage may be investigated and a number of hypotheses put forward concerning the relationship between the input, output and storage variables. High levels of evapotranspiration in summer for example might be expected to result in diminished amounts of rainfall reaching the stream channel. Excesses of output over input will result in a deficit of water with a reduction in the amount of water stored in the catchment as soil moisture and groundwater. Conversely, reduced levels of evapotranspiration during the winter months should result in rainfall contributing to storage, giving short term surpluses making good the deficits of the previous summer, to be followed by an increase in runoff once stores have been recharged.

Whilst such "water balance" concepts are readily testable once values of input, output and storage have been obtained, they do not in themselves offer any explanation of the pathways which water follows between entering the catchment and being discharged as runoff or lost by evapotranspiration. Explanation and ultimately the prediction of runoff requires a more detailed model than the so-called "black box" approach of a simple input-output system.

It has been proposed that the internal processes in the catchment may be modelled as a stack or "cascade" of storage boxes, each one draining to the store below as well as having an overflow to the stream channel through which water will pass once the storage box is topped up (Fig. 1).

Of the rain falling on the catchment, a small proportion will fall directly into the stream channel (*direct channel precipitation*), though the greater part falling on the hillslopes will encounter the vegetation canopy where the first of the catchment's storage boxes comes into operation. While much of the initial rainfall will be trapped and stored on the surfaces of leaves and stems, once the vegetation is wetted, further rainfall will tend to drip through (*throughfall*) or trickle down stems, branches and tree trunks (*stemflow*). Amounts of interception will vary spatially depending upon the density and height of the vegetation canopy (c.f. grassland and woodland) while seasonal variations will also be considerable, both under the natural conditions of deciduous woodland, and under cultivation where bare fields in winter contrast with dense crop cover in summer.

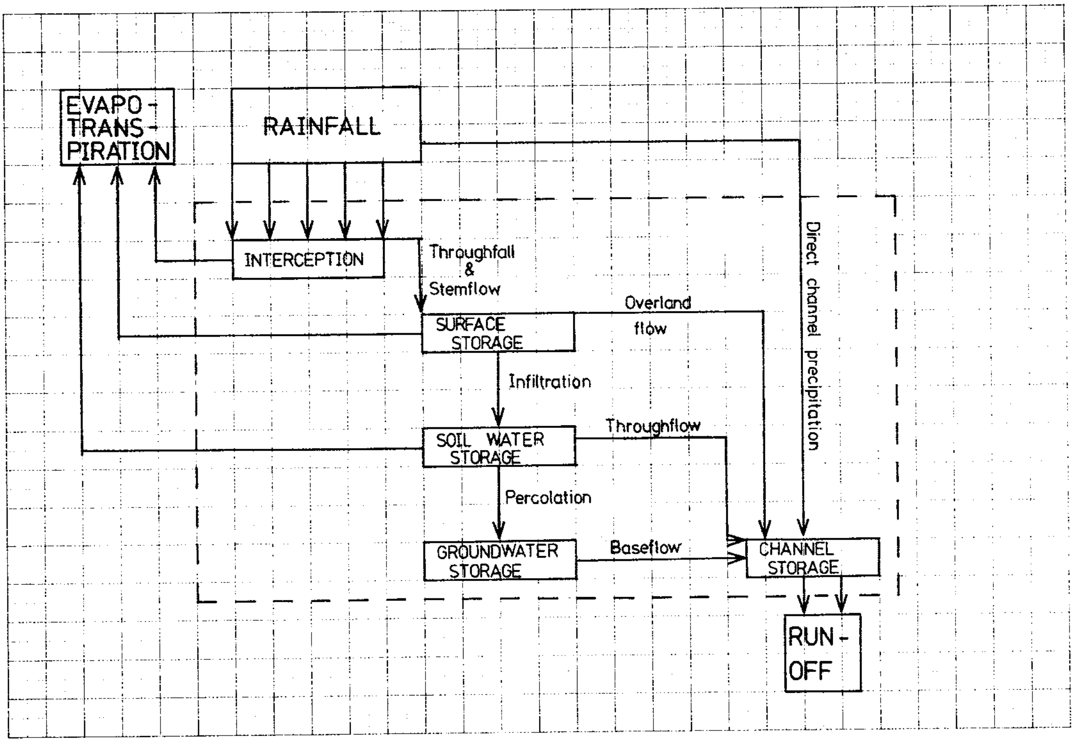


FIG. 1.
The catchment runoff system

The ground surface with its complex of depressions, large and small, constitutes the second storage box in the cascade. If water is reaching the surface more quickly than it can infiltrate into the soil, then depressions will begin to fill and then overflow into those downslope, establishing *surface runoff* or *overland flow* over certain areas, allowing large amounts of water to be transferred rapidly from hillslope to channel. It is thought that such conditions are unlikely to operate over large areas of the catchment, but will tend to be confined to previously saturated areas of soil near the base of slopes, in hollows on slopes and at the heads of streams giving rise to restricted areas of saturated overland flow, the area over which this process occurs being referred to as the *partial contributing area*. As this is likely to be a dynamic zone which will expand at times of high soil moisture (e.g. in winter or after a prolonged wet spell), the term "expanding contributary area" is sometimes used. This interpretation of the surface runoff process is somewhat in contrast to the Hortonian view which proposes that a reduction in infiltration rates occurs during rainfall until a point is reached when infiltration rate falls below the rainfall rate, giving rise to infiltration excess overland flow over much of the catchment.

Pore spaces in the soil form the next storage box in the system. Apart from controlling surface runoff generation, soil water may itself contribute to stream runoff by the process of *throughflow;* the lateral downslope movement of soil water. The size of soil pores is important in this context since water occupying the larger pore spaces (macropores) is subject to movement by gravity (*gravitational water*) and will tend to flow downslope through the soil towards the stream channel. Water occupying smaller pore spaces, however, becomes increasingly affected by the retaining capillary forces exerted by the soil matrix (*capillary water*). When soil water content rises, gravitational forces dominate leading to increased throughflow contributions to the stream, while under conditions of reduced soil moisture, capillary forces counteract the downslope pull of gravity, limiting the loss of water as throughflow.

A modified and more rapid form of throughflow has been identified which depends upon the flow of water down subsurface channels or pipes (*pipeflow*), possibly associated with tree roots or interconnected systems of animal burrows. In artificially drained catchments, tile drains constitute a type of pipeflow.

Assuming a permeable bedrock and the presence of free water in the soil, downward percolation into the groundwater store is assumed. Once again, high levels of storage will result in a strong baseflow contribution to the stream where the water table outcrops at the surface in valley bottoms.

Having identified a number of possible processes whereby water may enter stream channels and contribute to runoff, it may be possible to explain and even predict the pattern of runoff variation over time following rainfall. Since they will have a different amount of delay associated with them according to their flow velocities, in small catchments at least, the different runoff processes may result in temporally discrete peaks of flow in the stream channel with direct channel precipitation producing the most immediate response, followed by overland flow, and then with a greater delay, throughflow and groundwater contributions. It may also be possible to identify which processes are responsible for runoff at any particular time by reference to the dissolved and suspended load concentrations in the stream. Overland flow, for example, will tend to introduce water with high suspended loading, but usually low solute concentrations, while the onset of throughflow or groundwater contributions may be indicated by an increase in stream solute concentrations and a reduction in suspended sediment. Variations in suspended and solute load may be partly interpreted in terms of the relative contributions of the different types of flow to the stream. In considering the coarser load fraction (bedload), however, yield might be expected to relate more closely to energy availability in the channel, though a further consideration may also be the availability of material of a suitable size in the bed and banks of the channel.

From this theoretical approach to the catchment's dynamics, hypotheses may be generated in an attempt to explain the dependent variables of runoff and sediment yield (sufficient data are not yet available for this catchment to deal fully with solute loadings).

THE CATCHMENT

The catchment of Preston Montford brook covers an area of some 3.0 km^2 of drift-covered lowland of subdued relief in the Severn Valley, 6.0 km west of Shrewsbury, Shropshire. It is a first order stream, though culverted drainage from a pond in the south east part of the catchment may once have been a surface tributary. The total length of the brook from its source to its confluence with the River Severn is some 4.0 km, but the catchment area studied only includes the area upstream of the gauging structure adjacent to the Field Centre, giving a channel length of about 3.0 km. The stream rises at an altitude of 100 m O.D. (National Grid Reference SJ 491200) on a low boulder clay plateau at the southern end of the

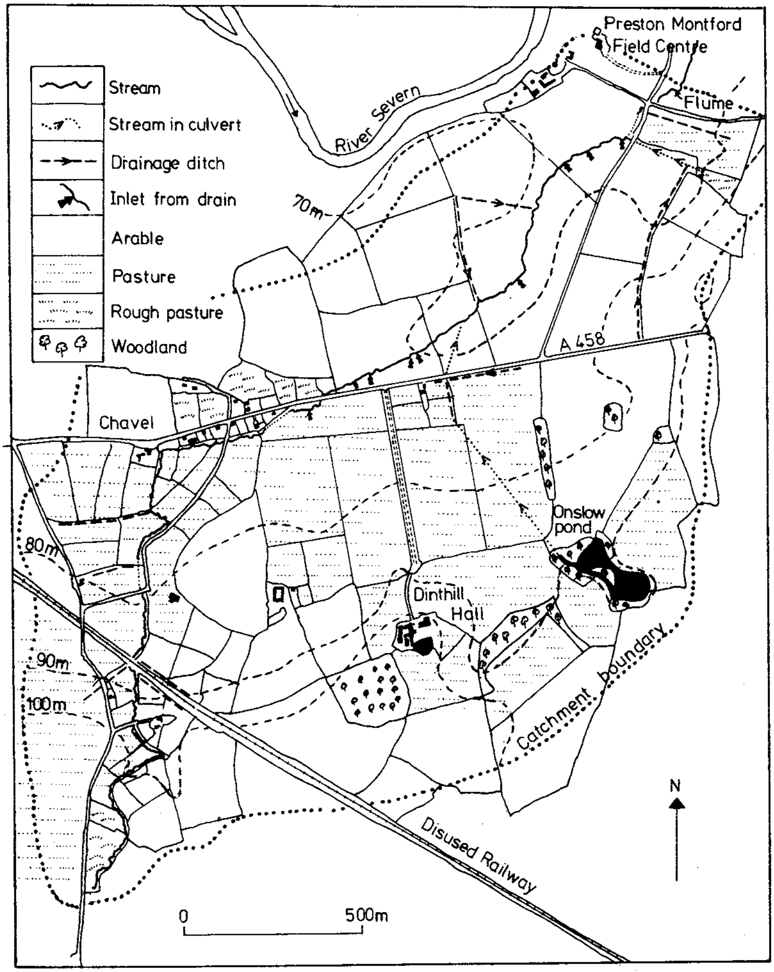


FIG. 2.
Preston Montford catchment

catchment. Within the first 1.5 km the stream descends to 70 m O.D. as it flows northwards down to the Severn Valley, continuing northeastwards at a gentler gradient across the upper Severn terrace.

While the boulder clay plateau forms the southern boundary to the catchment, the eastern watershed is a north-south trending moraine ridge which crosses the Severn Valley just east of the Field Centre. The western boundary is ill-defined, while the northern watershed follows the top of the high bank of the River Severn. The drift deposits are complex and include both glacial and fluvioglacial material. Outwash sands and gravels are widespread particularly in the lower parts of the catchment, though lake clays underlie these and outcrop in the stream bed just downstream of the flume. Sandy clay loam soils are characteristic of much of the arable land but textures become heavier to the south of the A458 where pasture is dominant. Arable land is planted with barley, sugar beet or potatoes and, apart from some winter barley, arable fields often lie bare between December and March. Tile drainage is widespread throughout the catchment, and the stream also receives inputs of water from field ditches, culverted drainage from Onslow Pond and a number of road drains.

MEASUREMENT OF WATER INPUT, OUTPUT AND STORAGE

1. Rainfall

Rainfall measurements commenced at Preston Montford in August 1976 with the establishment of a Meteorological Office climatological station. The standard "5 inch" gauge (12.7 cm diameter funnel standing 0.305 m above ground level) provides daily totals for an open site at the northern end of the catchment. In view of the subdued relief it may be reasonable to assume that the raingauge site is representative of the water input into the catchment as a whole, although this may not be so for the infrequent, high intensity convectional rainstorms. When possible during major storm events, the catch was measured at 20 minute intervals to give data on intensity and timing of rainfall. In July 1978, a tilting-siphon rainfall recorder was installed at the meteorological station. This instrument consists of a collecting chamber containing a float linked to a pen arm which records rainfall on a paper chart as a rising trace. When full, a catch is tripped which causes the collecting chamber to tilt, allowing the water to siphon out down the overflow pipe so that the trace drops vertically to the bottom of the chart. By reading off the vertical displacement of the trace during each hour or, with care, shorter periods of time, the amount, timing and intensity of rainfall is obtained.

Both raingauges suffer from the problems associated with the funnels standing above ground level. It has been suggested (Rodda, 1970) that significant amounts of rainfall may be missed due to turbulence and deflection of raindrops away from the funnel, particularly at exposed sites and under conditions of driving rain. To overcome this problem, a groundlevel gauge was constructed, similar in design to that developed by the Institute of Hydrology, and daily readings commenced at the beginning of the water year, 1979/80. The gauge is identical to the 5″ standard Meteorological Office pattern, but it is set into the ground so that the top edge of the funnel is flush with the ground surface. To prevent rain from splashing in, dripping in from the surrounding vegetation or surface runoff from spilling into the funnel, the gauge is set in the middle of a pit 1.3 m square and 0.3 m deep. The surface is covered by a plastic grid with a 5 cm square mesh supported on a light timber frame to create a surround which is aerodynamically similar to the grass sward of the site.

A comparison of the two "5 inch" gauges over the 12 months, October 1979–September 1980 (Table 2), shows that the groundlevel gauge catch is some 5% greater than that of the standard gauge.

When analysing rainfall over short periods of time, the total catch is taken from the groundlevel gauge and then partitioned throughout the period of rainfall according to the proportions obtained from the autographic recorder. Thus the rainfall for any particular time period (e.g. 1 hour) during an event is calculated as:

$$\begin{array}{l}\text{Rainfall in mm from}\\ \text{autographic trace for the hour}\end{array} \times \frac{\text{24 hr. total catch (groundlevel gauge) mm}}{\text{24 hr. total catch (autographic gauge) mm}}$$

The amount of "undercatch" associated with the autographic gauge is invariably greater than for the 5″ standard gauge owing to the larger funnel size and greater height above ground level.

Table 1. *Monthly rainfall totals 1976–1980 (mm)*

Month	1976/77	1977/78	1978/79	1979/80	Mean	Standard Deviation
October	93.3	32.5	19.9	63.8	52.4	32.9
November	39.3	77.1	42.2	52.1	35.2	19.5
December	50.0	63.6	112.6	117.9	86.0	34.3
January	78.1	51.4	40.4	55.5	56.4	15.8
February	113.7	42.4	40.0	84.4	70.1	35.5
March	33.9	35.9	75.4	87.3	58.1	27.2
April	43.9	43.2	51.6	10.0	37.2	18.5
May	49.0	63.5	86.6	26.9	56.5	25.1
June	105.5	64.6	33.5	80.0	70.9	30.1
July	8.3	70.3	7.3	36.2	30.5	29.7
August	88.7	54.8	43.8	76.4	65.9	20.3
September	31.8	38.5	26.2	68.2	41.2	18.7
TOTALS	735.5	637.8	579.5	758.7	677.9	83.9

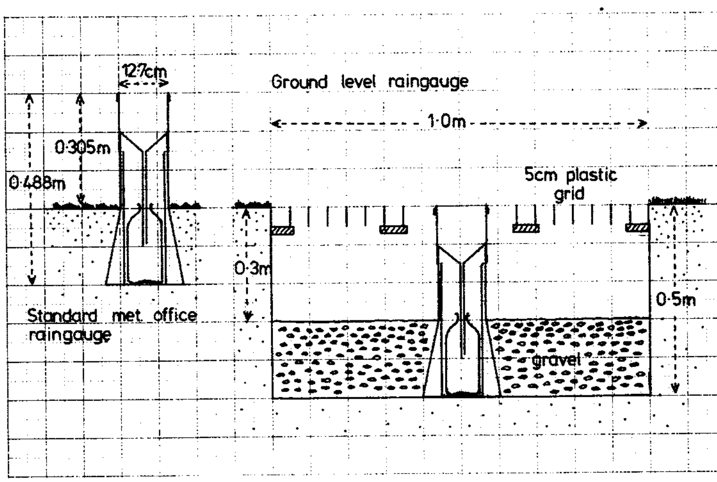


FIG. 3.
Standard and ground level raingauges

Table 2. *Comparison of standard and groundlevel raingauge catches*

	Oct	Nov	Dec	Jan	Feb	March	April	May	June	July	Aug	Sep	Total
Standard Meteorological Office Raingauge (mm)	63.8	52.1	117.9	55.5	84.4	87.3	10.0	26.9	79.9	34.1	76.4	68.2	756.5
5″ Groundlevel Gauge (mm)	65.6	54.8	122.4	58.9	88.4	90.5	10.7	28.2	83.8	36.2	82.0	72.3	793.8
Difference in Totals as % of standard gauge catch	+2.8	+5.2	+3.8	+6.1	+4.7	+3.7	+7.0	+4.8	+4.9	+6.2	+7.3	+6.0	Mean +4.9

2. Runoff

Some clarification is necessary on the units of measurement of runoff and the different ways in which it can be expressed. A distinction may be made between measurement of the rate of streamflow per unit of time, normally referred to as *discharge*, and total quantities of water output or *runoff*. Stream discharge for small catchments is most conveniently measured in *litres per second* ($l\ sec^{-1}$) though for larger channels *cubic metres per second* ("cumecs") is the more widely used unit. When dealing with total water output during storm events or over a month, the volumes of water involved become very large and it is more convenient to express them in millions of litres (*Megalitres*).

It is often necessary to make direct comparisons between quantities of runoff and the rainfall that produced it. A problem then arises in that rainfall is expressed as a depth measurement in mm while water output as streamflow is expressed as a volume. In order to compare the two quantities, they must be expressed in the same unit, either by multiplying the rainfall by the catchment area, thus converting rainfall to a volume, or, more conveniently, taking the volume of runoff and dividing by the catchment area, thereby expressing runoff as a depth measurement. The calculation for expressing runoff in mm is as follows:

$$\text{Runoff (mm)} = \frac{\text{Runoff } (\text{mm}^3)}{\text{Catchment area } (\text{mm}^2)}$$

$$= \frac{\text{Runoff (Megalitres)} \times 10^{12}}{\text{Catchment area } (\text{km}^2) \times 10^{12}}$$

At first sight this appears somewhat unwieldy but since the number of mm^3 in a Megalitre is equal to the number of mm^2 in a km^2 (i.e. 10^{12}), the calculation becomes very simple.

Runoff measurement on Preston Montford brook commenced in January 1977 with the installation of a 90° "V" notch weir (see Howcroft, 1977; Gregory and Walling, 1973, pp. 135–145) in conjunction with a home-made water level recorder constructed largely of Meccano parts. A number of problems were encountered with the "V" notch weir, culminating in its removal in March 1977. On several occasions the structure was overtopped; with the low gradient of the stream the backing up of water behind the weir was considerable, extending some 30 m upstream during high flows. This resulted in a large pressure on the plate which became distorted, while the increase in water level in the channel upstream began adversely to affect drainage in a low-lying part of the adjacent field. It was envisaged that with very high suspended sediment loadings, silting up in the pool behind the weir was likely to become a problem. As an interim measure, the water level recorder was reinstalled over a stilling well adjacent to a reasonably stable reach of the existing natural channel. A stage board was set up and for a wide range of flows discharge was measured using a Braystoke flowmeter to obtain velocity readings and cross section area from width and depth measurements. From this data, a *rating curve* was established to relate stage (i.e. stream height) to discharge (Gregory and Walling, 1973).

In August 1977 a trapezoidal concrete flume was constructed (Gregory and Walling, p. 139). This consists of a reinforced 3 metre length of channel incorporating a lateral constriction, or throat, in its middle section (Fig. 4). A stable cross section shape is thus ensured while the constriction amplifies changes in depth as discharge varies. For a low gradient stream with high sediment loadings, this structure has proved to have the following advantages over a "V" notch weir:

(i) Very little "ponding up" of water behind the structure, therefore no adverse effects upon field drainage.

(ii) Little reduction in velocity upstream of the structure (except at low flows), therefore minimal sedimentation.

(iii) Very little maintenance required apart from scrubbing the surfaces free of algal growth and occasional removal of sediment.

These advantages arise largely from a fundamental difference between a thin plate weir and a flume; when constructing the former, the base of the "V" must be a minimum height above the bed of the stream, whereas in a flume, the base of the throat is at the same level as the stream bed upstream of the structure. A stage board and water level recorder were installed and field rating was carried out by flow metering in the throat during a wide range of flows. Low flow measurements were checked using the gulp injection salt dilution method (Gregory and Walling, pp. 131–133).

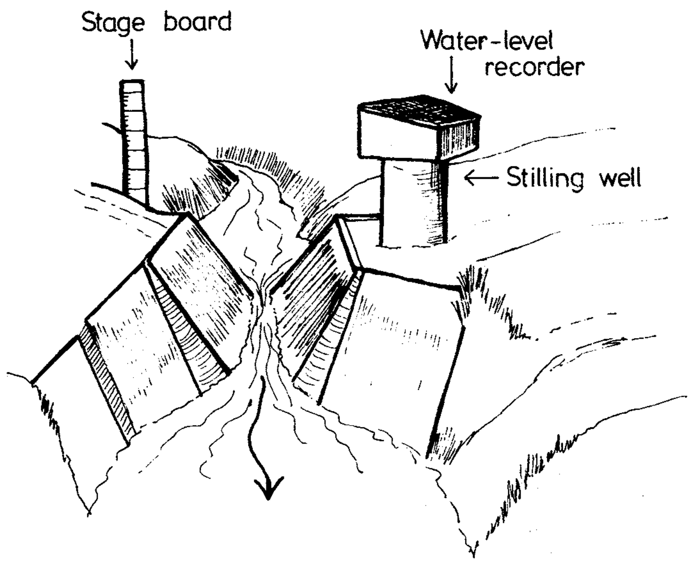


FIG. 4.
Trapezoidal flume on Preston Montford brook

The water level recorder used was initially a further development of the home made instrument, but in October 1978 a Lea recorder was borrowed from the Severn Trent Water Authority and now provides greater accuracy. Hourly readings of stage are taken from the charts and converted to discharge (l sec^{-1}) using the rating plot (Fig. 5). Daily, and hence monthly, runoff figures are then computed for the catchment.

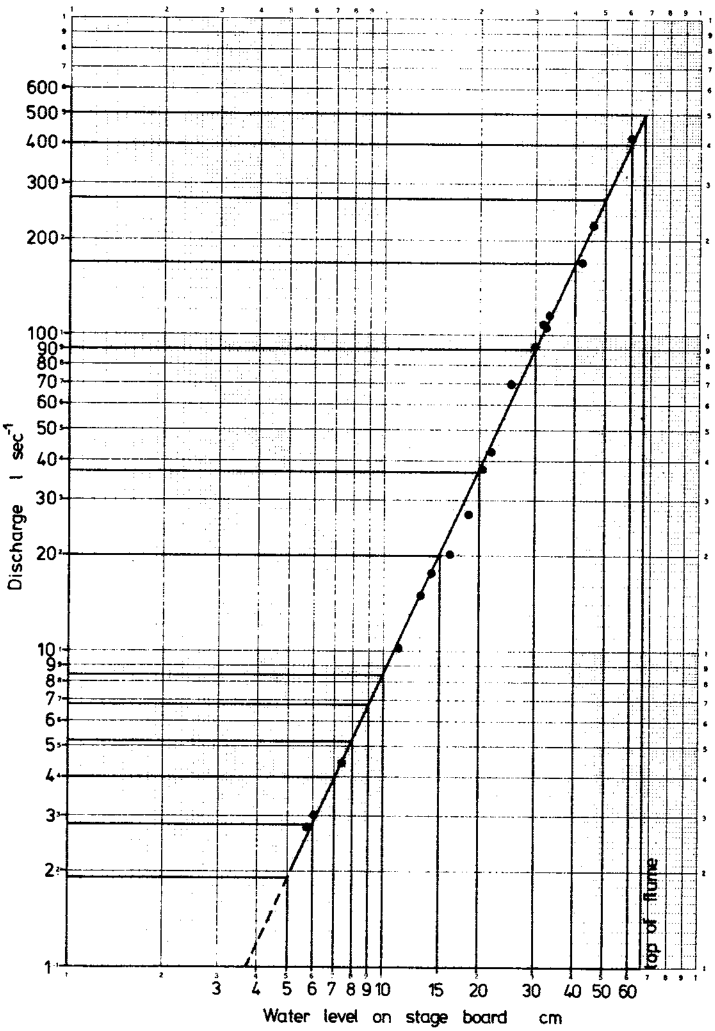


FIG. 5.
Stage/discharge rating for trapezoidal flume

From August 1977 until October 1980 most of the major storm events (here defined as isolated rainfall events in excess of 10 mm) have been recorded and the rainfall and runoff response plotted. The reaction of the stream to rainfall variations cannot be explained purely in terms of the amount and intensity of rainfall. In order to investigate variations in storm response more fully, eighteen events at different times of year were analysed. An attempt was made to distinguish *storm runoff* (or *quickflow*) from the lower *delayed flow* occurring prior to and after the event. The onset of quickflow is readily identified by a sharp upturn in the hydrograph referred to as the *rising limb*, as water reaches the channel by rapid runoff processes (Fig. 6). A *peak storm discharge* may be identified and the time lapse between the period of greatest rainfall intensity and the timing of the peak discharge can be measured (*the lag time*). In order to indicate the steepness of the rising limb, the time between the initial rise in stream level and the peak discharge was recorded (*time to* *peak*). The *height of the peak* was determined by subtracting the stream discharge before the rising limb (*antecedent discharge*) from the peak discharge and the *rate of the rise* calculated by dividing the height of the peak by the time to peak. Following the steep rise to peak discharge most hydrographs then show a gentler decline (*the recession curve* or *falling limb*) until the trace levels out once the rapid runoff processes cease. There is then a return to delayed flow conditions although the discharge level may exceed the antecedent flow. The point where storm runoff ceases is often difficult to determine objectively, but normally there is some change in hydrograph gradient.

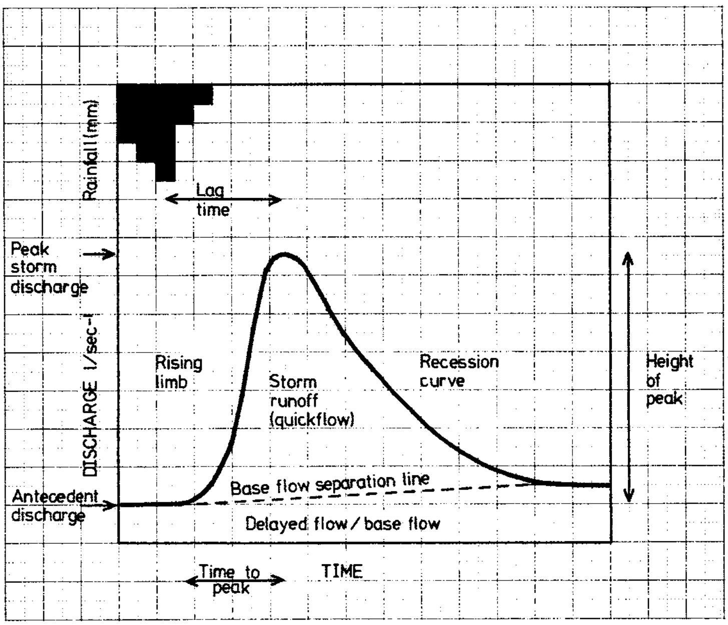


FIG. 6.
Storm hydrograph components

A separation line may now be drawn in to distinguish the quickflow from the delayed flow. If plotted on graph paper, the total storm runoff can be estimated by counting the number of a convenient size of square which lie above the separation line and below the storm peak. The volume of water represented by a single square is readily calculated by multiplying the number of litres sec^{-1} represented on the vertical axis by the number of seconds represented on the horizontal axis. Total storm runoff can now be derived by multiplying the number of litres per square by the number of squares counted. If total storm runoff is expressed in Megalitres, the conversion to mm of runoff is readily achieved and the comparison between storm runoff and total storm rainfall (runoff %) is then possible:

$$\text{Runoff \%} = \frac{\text{Storm runoff mm}}{\text{Storm rainfall mm}} \times 100$$

Rainfall events contributing to storm hydrographs were analysed in terms of *total storm rainfall, duration of rainfall, maximum hourly rainfall* and *mean rainfall intensity* (total rainfall/duration). Weyman (1975) refers to the importance of antecedent moisture conditions in controlling runoff responses. Since direct measurements of storage conditions (soil moisture and groundwater) were unavailable for the periods preceding all eighteen storms, data have been compiled on the total rainfall and total evapotranspiration for the periods 5, 10, 20 and 30 days prior to each storm event. By subtracting evapotranspiration from rainfall for the periods 10, 20 and 30 days before each storm, a surplus or deficit of water was determined which is indicative of the storage levels in the catchment prior to the storm occurring. These surplus or deficit figures are here referred to as the antecedent moisture index (AMI).

3. Evapotranspiration

Of all the hydrological variables, evapotranspiration is probably the most difficult to measure. It is a complex process involving not only the direct evaporation of water from the ground surface, but also transpired water taken up by plants from the soil.

Since April 1977 direct measurements of evapotranspiration have been attempted using a simple irrigated lysimeter system. This consists of a 26 gallon oil drum sunk into the ground with a 20 cm diameter polythene funnel bolted to the base of the drum, and the area above the funnel hollowed into a slight concavity and drilled with 5 mm holes. A length of 2 cm diameter PVC waste pipe was attached to the funnel and this leads down with a 5° gradient to a collecting vessel at the foot of an adjacent slope (Fig. 7). Gravel chippings were placed in the base of the drum to a depth of 0.3 m and the remaining 0.6 m filled with soil from the field adjacent to the lysimeter plot. Care was taken to preserve as far as possible the natural structure and horizon sequence of the soil by inverting an oil drum over the turf, hammering the rim into the ground to a depth of 0.30 m and thus extracting a core of turf and topsoil largely intact. 0.30 m of the underlying subsoil was then transferred to the oil drum and then the 30 cm turf and topsoil core was fitted into the top of the column (Fig. 7). The system was set up in triplicate thus allowing for instrumental error.

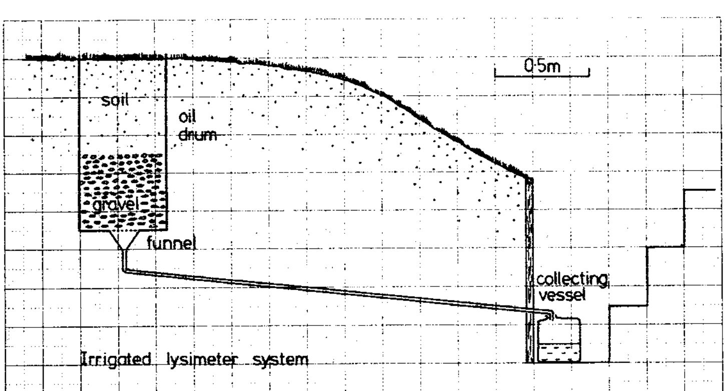


FIG. 7.
Irrigated lysimeter system

The operation of a lysimeter is based on the water balance concept that input = output ± storage. Rain falling on the grass surface of the drum forms an input into the system, while water draining through the soil/gravel column and down the outlet pipe forms one of the outputs. Evapotranspiration from the grass surface forms the other one. If rainfall input and drainage water output are measured for a given time period, the difference between the two quantities gives an estimate of evapotranspiration, provided changes in water storage within the drum can be eliminated, or at least reduced to a minimum.

The major problem in the operation of a lysimeter of this type is that of maintaining the soil moisture at a similar level from month to month. The aim is to maintain soil moisture as near as possible to field capacity. This is the state when water which occupied the large pore spaces (gravitational water) has drained out, but water in the small pore spaces (capillary water), which is held by capillary attraction, is at a maximum (see Briggs, 1977, Ch. 3). If no rain occurs on a particular day, or if rainfall is less than the water lost from the system by evapotranspiration, then soil moisture levels will fall below field capacity as capillary water is lost. Irrigation is then necessary to return the soil moisture to field capacity. Immediately after irrigation, soil moisture is likely to be in excess of field capacity, but the surplus water added will readily drain through as gravitational water and will be collected and measured. The following rule of thumb was applied to determine the amount of irrigation water added:

If no rain on preceding day, add 6 mm of water
If less than 6 mm of rain on preceding day, add a further 3 mm water
If more than 6 mm of rain on preceding day, no irrigation necessary.

Six mm of water is about the maximum daily amount of evapotranspiration likely to occur on a hot summer's day from a grass surface. To determine the volumes of irrigation water necessary, the quoted figures must be multiplied by the surface area of the drum (159 cm^2). Thus, 6 mm is obtained by adding 954 ml of water.

In order to calculate the evapotranspiration for a month, the following quantities must be determined:

(i) total rainfall input (mm) measured at the adjoining meteorological station site.
(ii) total irrigation water (mm).
(iii) total water collected—measured as a volume (ml), then converted to mm by dividing by surface area of drum.

(iii) subtracted from the sum of (i) and (ii) gives an estimate for evapotranspiration. It should be noted that these figures are likely to be estimates of *potential* rather than *actual* evapotranspiration in that no soil moisture deficit is allowed to develop in the drums. Under natural conditions, as soil moisture deficits develop, the actual rate will drop below the potential rate since the capillary forces retaining water in the soil pores begin to exceed the drawing power of plant roots.

In order to provide a check on evapotranspiration figures obtained from lysimeters, monthly estimates of potential evapotranspiration were obtained from the Meteorological Office. Since April 1978, monthly values have been published on a 40 km grid for Britain. Fig. 8 shows a plot of monthly evapotranspiration figures for Preston Montford lysimeters against the equivalent monthly figures produced by the Meteorological Office using climatological data for the North Shropshire area. Although a high correlation coefficient (significant at the 99% level) was obtained, there are major inconsistencies at certain times of year. Excessively high lysimeter

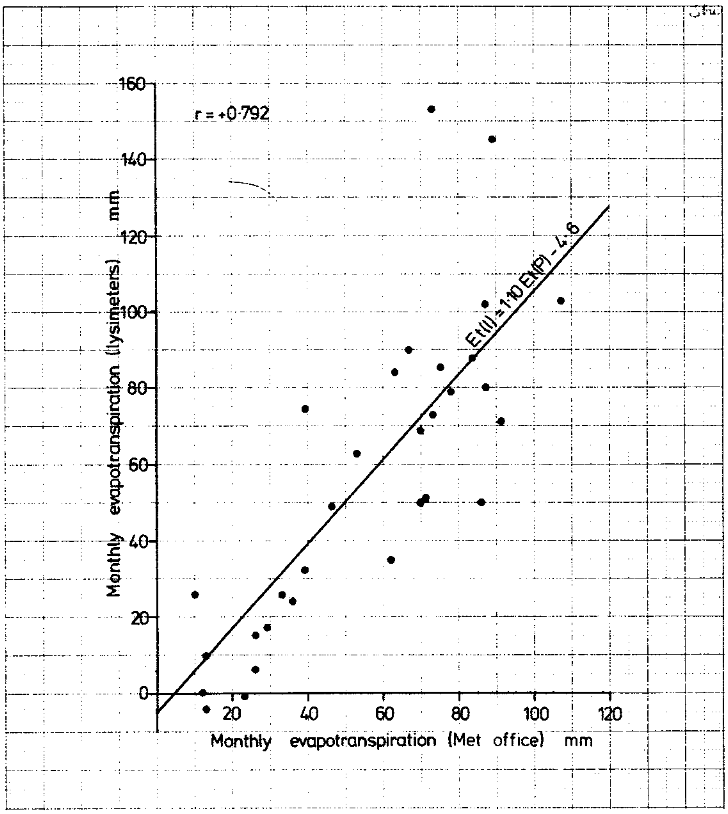


FIG. 8.
Comparison between calculated values of monthly evapotranspiration (Meteorological data) and lysimeter data

figures for some summer months suggest insufficient irrigation in the previous month and the development of a soil moisture deficit. On two occasions during the winter months, the water output from the drum exceeded the water input for the month, indicating the release of stored water from the previous month's rainfall. (On one of these occasions this was due to some of one month's snowfall persisting through to the following month, and thus producing an apparently negative estimate of evapotranspiration.)

4. Soil moisture

While the moisture content of a soil sample may be readily determined by weighing, heating to drive off moisture and reweighing, alternative non-destructive methods are required if regular frequent readings of moisture conditions in the same body of soil are required. Soil tensiometers provide a cheap and easily constructed instrument for this purpose. Burt (1978) describes a simple tensiometer system consisting of a porous pot glued firmly to one end of a length of clear acrylic tubing, and the instrument filled with de-aired water. A mercury manometer of fine bore (2 mm) flexible plastic tubing is suspended inside the tensiometer so that the longer limb protrudes through a hole in the rubber bung which seals the top of the instrument (Fig. 9). The shorter manometer limb is open to the water inside the tensiometer and is water filled above the mercury.

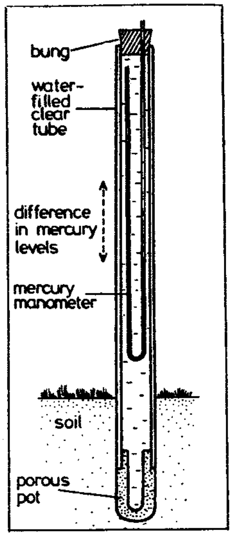


FIG. 9.
Soil tensiometer design

The tensiometer is installed by augering a hole in the soil to the required depth, inserting the tube, and back filling with slurried soil to ensure a sound contact between the porous pot and the soil. When surrounded by dry soil, capillary action draws water from the porous pot, producing a negative pressure (or *tension*) in the water column which draws up the mercury level in the shorter limb, but causing a corresponding drop in the longer manometer limb. The drier the soil becomes, the greater the tension and so the greater the difference in mercury levels. On wetting however, the state of tension in the instrument will draw water from the soil back through the walls of the porous pot, causing the mercury levels to come closer together. At saturation, soil water begins to exert a pressure, forcing water into the instrument, thereby depressing the mercury level in the shorter limb. The deeper the porous pot below the level of saturation, the greater the soil water pressure and the more the mercury level will be depressed.

Before installation, each instrument is calibrated by half immersing the porous pot in water and measuring the small differences in levels when neither tension nor pressure is being exerted (zero soil water potential). Measured differences in mercury levels are then subtracted from the difference at calibration, a positive result indicating soil water pressure, a negative result indicating soil water tension. Normally the amount of tension or pressure is expressed, not as a length of mercury column drawn up or depressed, but as the equivalent amount of water. As mercury has 13.6 times the density of water, the complete determination of soil water potential is expressed as:

$$\underset{\text{(cm)}}{\text{soil water potential}} = \underset{\text{(cm)}}{\text{difference in mercury levels at calibration}} - \underset{\text{(cm)}}{\text{measured difference in in mercury levels}} \times 13.6$$

Tensiometers were established at the meteorological station in silty clay loam soil at 10 cm and 30 cm depths in April 1979. Some problems in operation were encountered, particularly with freezing during heavy air frosts and fracturing of the mercury column at high tensions in mid-summer. A continuous record of soil water potential was therefore not possible; to provide a more complete record of soil moisture conditions over the study period, monthly soil moisture deficit figures were obtained from the Meteorological Office. These values record in mm the amount of water which would have to be added to the soil in order to return it to field capacity. A soil moisture deficit of zero therefore indicates that the soil is at field capacity, while increasingly high values indicate drying out of the soil and loss of capillary water.

5. Groundwater

In the context of the North Shropshire environment, a distinction must be made between ground water in the underlying Triassic sandstones and the more superficial perched water tables that occur locally within the drift where impermeable clays maintain saturation in the overlying coarser, permeable drift deposits. In low lying parts of the Preston Montford catchment saturated drift occurs fairly close to the surface, and in hollows, surface pools occur. To investigate seasonal variation in the perched water table, three boreholes were sunk on a gently rising slope above a seasonally water filled marshy hollow adjacent to the Field Centre. Using an extendable screw auger, three holes were sunk in line, 5 m apart, each to a depth of 3 m (Fig. 10). A 3 m length of 2 cm diameter plastic pipe with the sides drilled at 5 cm intervals with 2.0 mm holes was inserted down each hole to prevent the sides from caving in. The depth of water in the pipes was then measured at weekly intervals using a simple electrical dipper. This consisted of a length of two core cable with a jack plug attached to one end and a circuit incorporating a 9v battery and a milliammeter at the other. On touching water, the jack plug completed the circuit causing the ammeter to jump. By carefully lowering the cable down the borehole and noting the length of cable when the needle flipped, the depth of the water table below the surface was obtained. This was then converted to a height above an arbitrary datum, level with the bottom of the lowest borehole (see Fig. 10).

MEASUREMENT OF BEDLOAD SEDIMENT

Attempts to measure bedload movement in the brook began in 1978 with the installation of small trap 20 m upstream of the flume. This consisted of a strong wooden box constructed from marine ply, its length corresponding to the width of

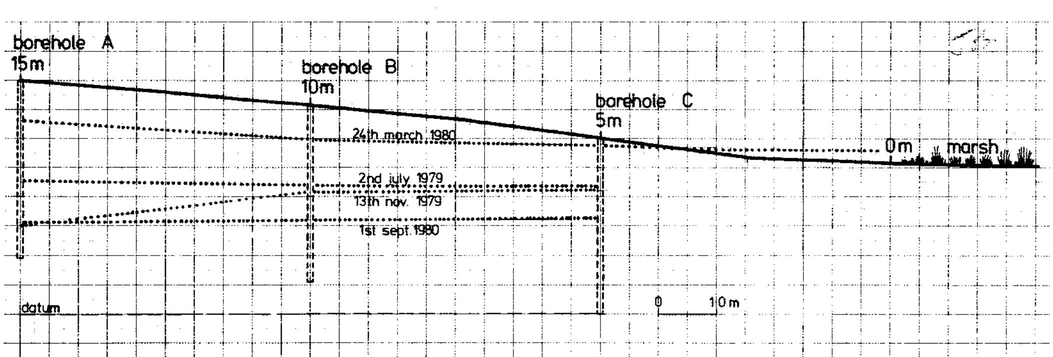


FIG. 10.
Borehole transect and water table levels

the stream (1.3 m), 0.4 m deep and 0.25 m wide, with handles built into the end panels. The bedload box was sunk into the stream bed so that the top was flush with the bed, the retaining hole being lined with concrete. A gap of 5 mm was left between the box and the concrete along the downstream side and at the ends to facilitate removal and replacement of the trap. After storm events the box was removed and the load was weighed and sieved.

It soon became evident that the trap was of insufficient capacity to hold all the load brought down by a single storm event, as even after events of medium magnitude the trap was found to be full of sediment. It was assumed that once full, further sediment was passing over the trap and being lost. Attempts to empty the trap during storms proved fruitless, owing to the difficulties of removing and replacing the trap under high flow conditions.

To overcome these problems, a larger trap was constructed in September 1979. This is a permanent structure, 1.6 m in length, 1.3 m wide and 0.6 m deep with a flat floor of paving slabs and brickwork walls (Fig. 11). A bypass channel was dug around the site, initially to divert the stream water during construction, but subsequently to allow temporary diversion of the flow during emptying. Brick slots were constructed to allow a sluice board to be placed across the channel diverting flows during emptying or at other times to block off the bypass channel. Problems were experienced in making a water-tight seal between the sluice board and the stream bed. A concrete beam was sunk into the bed of the stream to provide a flat, firm base for the sluice board to rest on, but it was still necessary to use clay, pressed in along the base and ends of the sluice board on the upstream side to maintain a complete seal. This is obviously a time-consuming and uncomfortable task, and a more efficient solution would be to install a worm driven sluice gate which could be raised and lowered.

Having diverted the flow, the water standing in the trap was pumped out with an electric pump (power being available from an adjacent cottage) and the accumulated sediment carefully shovelled onto the concrete unloading platform. The wet sediment was allowed to drain for two hours, then weighed on site (by the bucketful) with a spring balance.

From each bucket a trowelful of sediment was retained to make up a subsample which was taken back to the laboratory, weighed, placed in a drying oven at 100°C for 24 hours and then reweighed to determine its moisture content. The dry weight of the total load could then be estimated. From the dried subsample, a 2 kg fraction was taken and passed through a nest of sieves from –5.0 phi (32.0 mm) down to +3.5 phi (0.09 mm) in order to determine the particle size distribution of the sediment. Mean and standard deviation were estimated by plotting the cumulative frequency of each size class on arithmetic probability paper (see Briggs, 1977, ch 3). To distinguish the mineral content from the organic matter in the bedload samples, a teaspoonful of each sieved fraction was weighed accurately, heated at 400°C in a muffle furnace for 8 hours to burn off the organic matter, then allowed to cool in a dessicator and reweighed to give the organic content.

Measurement of Suspended and Dissolved Loads

During a number of storm events, water samples were collected immediately downstream of the bedload trap by immersing a clean 500 ml polythene bottle in the stream and allowing it to fill while moving it up and down in mid-channel to

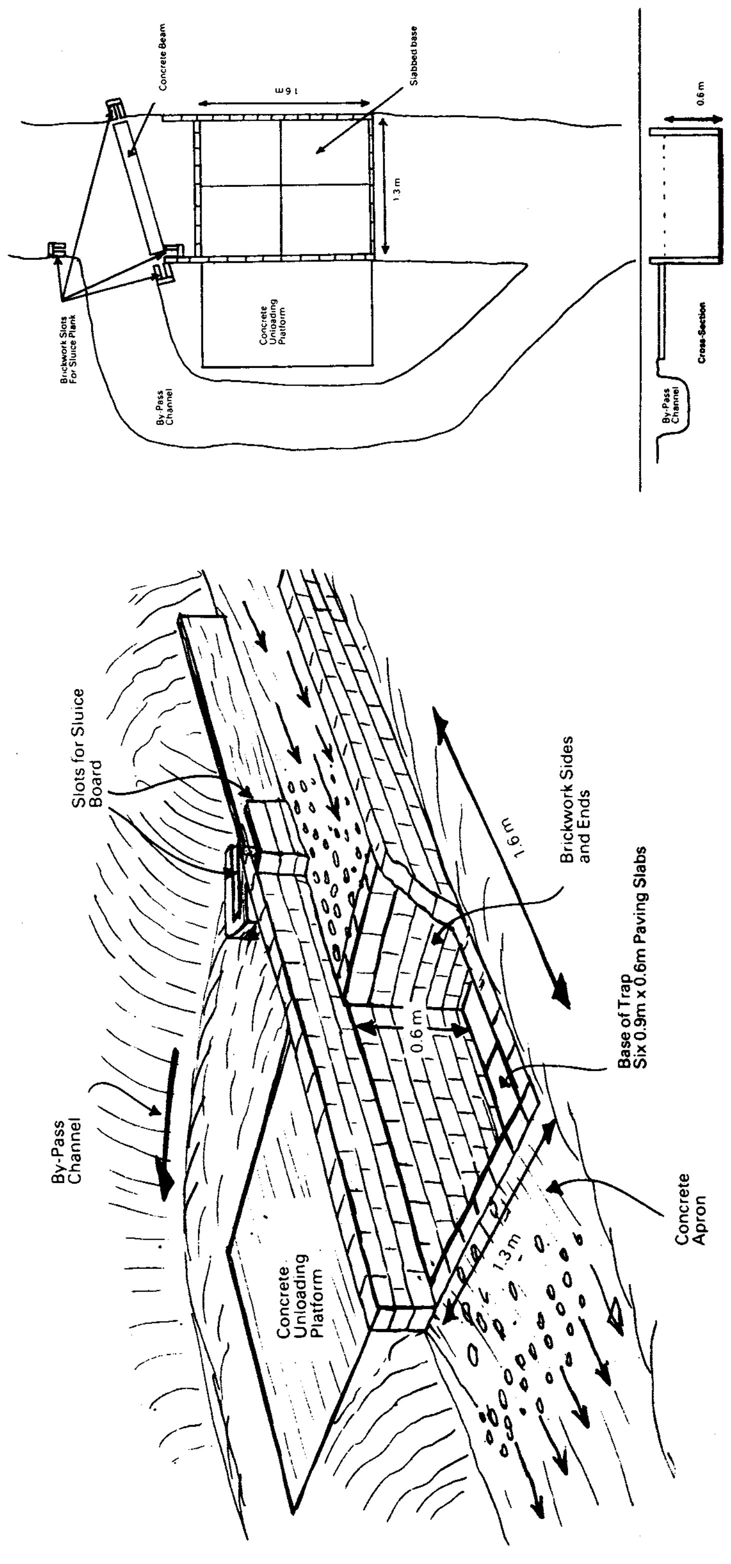


FIG. 11.
Bedload trap constructed on Preston Montford brook

integrate water from all depths. Initially, an hourly sampling interval was used, but it became evident that very rapid changes in suspended and dissolved load were taking place during the early phases of a storm event, and more frequent sampling (every 10 or 15 minutes) was therefore undertaken during rising stages. Samples were returned to the laboratory and analysed as soon as possible thereafter. The bottle was first shaken vigorously to mix up the sediment which had settled out, the precise volume measured, and then the water sample was passed through a dry fine grade fibreglass filter paper (Whatman GF/C 7.0 cm) which had previously been weighed to the nearest mg. The filter paper was placed in a Buchner funnel fitted to a Buchner flask with a vacuum pump attached to assist filtration.

With sediment concentrations in excess of about 500 mg l^{-1}, the filter paper became clogged before filtration of the sample was complete and it was necessary to continue the filtration through a second filter paper. Very heavily loaded samples required several separate filter papers. The filter paper was then carefully removed from the funnel, dried for one hour at 100°C and reweighed to determine the amount of dry sediment deposited. Suspended sediment concentration was then expressed in mg l^{-1}.

The clear filtrate now containing only the dissolved load was retained for further analysis. The conductivity was measured using a WPA conductivity meter and, since conductivity is somewhat temperature dependent, the temperature of the sample was recorded and the conductivity converted to a standardised value at 20°C. Conductivity provides a readily determined relative, rather than absolute, measure of solute concentration, since the higher the concentration of dissolved ions in the sample, the greater the electrical conductivity (conductivity, measured in mhos, is the reciprocal of resistance, measured in ohms, hence the unit of measurement, the mho). It does not, however, provide any information about which ions are present and in what concentrations. Such determinations require more complex methods, normally involving spectrophotometry, but some commonly occurring solutes can be determined by titration. Total hardness (concentration of dissolved calcium and magnesium ions), for example, can be readily determined by titration with EDTA (Douglas, 1969).

Seasonal Trends in Catchment Variables

Monthly values of the main hydrological variables for the water years 1978/9 and 1979/80 are summarised in Table 3, while Fig. 12 shows a comparative month by month plot of rainfall, evapotranspiration and runoff. Rainfall data have not been collected over a sufficiently long period of time for reliable annual or monthly means to be determined.

Over the four years from 1976/77 to 1979/80 an annual mean of 677.9 mm was recorded, but monthly means show no strong seasonal trend (Table 1). The two water years under detailed consideration, however, both show December maxima in excess of 100 mm and something of a summer or spring minimum (7.3 mm July 1979, 10.0 mm April 1980). Monthly evapotranspiration figures show a far stronger seasonal variation with rates below 20 mm during winter months rising to summer peaks in excess of 80 mm. Warm sunny conditions in May 1980 generated an extreme high of 107 mm. Monthly runoff figures show an even stronger seasonal contrast with consistently high values from December to April and late winter maxima in February or March (62.6 mm in February 1980). From May to

Table 3. *Monthly summary of hydrological variables, Preston Montford catchment*

Year	Month	Rainfall mm	Evapo-trans-piration mm	Total Runoff Megalitres	Total Runoff mm	Mean daily discharge l sec⁻¹	Maximum Instan-taneous flow l sec⁻¹	Minimum Instan-taneous flow l sec⁻¹	Soil Moisture deficit mm	Mean Ground-water level (m above datum)
1978	October	19.9	39.0	3.48	1.1	1.3	2.6	1.2	127	1.58
,,	November	42.2	29.0	5.60	1.8	2.2	10.4	1.3	109	1.78
,,	December	112.6	13.0	117.09	37.2	43.7	218.0	2.2	13	2.12
1979	January	40.4	12.0	75.95	24.1	27.9	54.9	13.4	0	2.52
,,	February	40.0	15.0	141.63	45.0	57.6	288.0	22.5	0	2.66
,,	March	75.4	40.0	167.85	53.3	63.5	320.0	22.5	2	2.68
,,	April	51.6	62.0	92.46	29.4	35.7	147.0	10.3	25	2.79
,,	May	86.6	73.0	31.52	10.0	12.2	55.0	7.0	1	2.51
,,	June	33.5	75.0	19.33	6.1	7.6	21.5	3.0	25	2.42
,,	July	7.3	89.0	5.55	1.8	2.1	2.9	1.5	98	2.05
,,	August	43.8	72.0	6.52	2.1	2.4	11.6	1.3	94	1.76
,,	September	26.2	53.0	4.45	1.4	1.7	2.8	1.3	114	1.60
1979	October	63.8	36.0	7.80	2.5	2.9	21.0	1.9	100	1.52
,,	November	52.1	26.0	11.36	3.6	4.3	10.3	2.9	61	1.93
,,	December	117.9	23.0	132.57	42.0	49.5	342.0	3.8	1	2.65
1980	January	55.5	10.0	116.13	36.9	43.3	116.0	17.0	0	2.78
,,	February	84.4	13.0	197.29	62.6	78.7	317.0	21.4	1	2.83
,,	March	87.3	33.0	176.03	55.9	65.7	516.0	28.0	0	2.76
,,	April	10.0	70.0	55.60	17.6	21.4	89.0	10.2	65	2.63
,,	May	26.9	107.0	15.85	5.0	5.9	14.7	3.9	111	1.97
,,	June	79.9	86.0	18.15	5.8	7.0	29.3	3.9	91	1.92
,,	July	34.1	91.0	10.66	3.4	4.0	14.8	2.6	124	1.85
,,	August	76.4	87.0	11.73	3.7	4.4	37.5	2.2	119	1.65
,,	September	68.2	70.0	20.80	6.6	8.0	103.0	2.4	104	1.60
Water Year '78/79	Totals/ Means	579.5	572.0	671.43	213.3	21.5	320.0	1.2	—	—
Water Year '79/80	Totals/ Means	756.5	652.0	773.97	245.6	24.6	516.0	1.9	—	—

Based on catchment area of 3.15 km²

November, monthly runoff totals are all 10.0 mm or less, both years showing minimum values in October (1.1 mm and 2.5 mm).

The water balance

Regarding the catchment initially as a simple input/output system (the "black box" approach), there should, in a typical water year, be a balance between rainfall input and the combined output of evapotranspiration and runoff, assuming storage conditions to be the same at the start of the water year as at the end. For both water years, the annual water budget shows a net deficit:

	Rainfall		Evapo-transpiration		Runoff		Deficit
1978/79	579.5	=	572.0	+	213.3	–	205.8
1979/80	756.5	=	652.0	+	245.6	–	141.1

It would appear from these results that over both water years there was an overall loss of water from storage, though reference to ground water levels and soil moisture deficit values indicate very little difference in storage between the start and finish of the water years. For the water year 1978/79, ground water levels were marginally lower in October 1979 compared with October 1978, while the soil moisture deficit was slightly less in October 1979 relative to October 1978. It seems necessary, therefore, to refer to other factors in order to account for the apparent imbalance in the annual water budget of the catchment. In view of the difficulties in measurement of hydrological variables, consideration must be given to possible

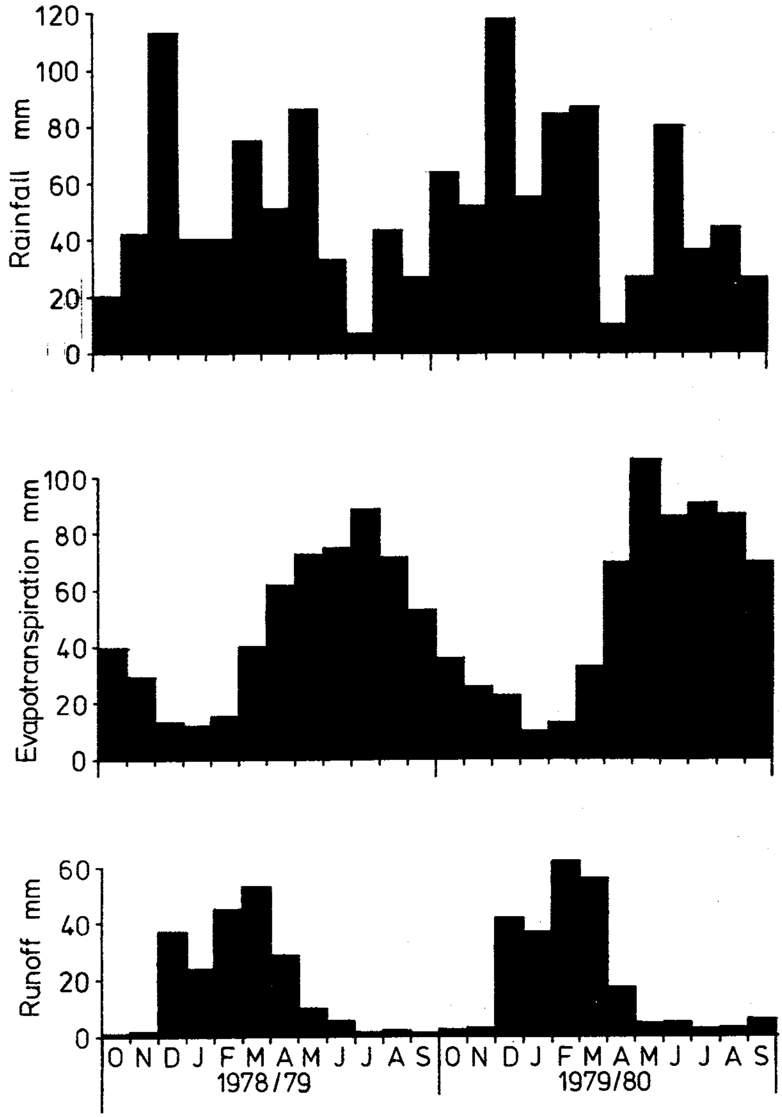


FIG. 12.
Monthly totals of rainfall, evapotranspiration and runoff, 1978/79, 1979/80.

sources of instrumental error. The rainfall totals used are derived from the standard Meteorological Office gauge. On the basis of the comparative catches of ground level and standard gauges (Table 2), the yearly totals may be assumed to be underestimates of the order of 5%. The measured ground level catch for 1979/80 was 793.8 mm, while the estimated ground level catch for 1978/79 would give a total of 608.5 mm. Despite this correction, a substantial apparent water deficit remains for both years. Since the evapotranspiration figures quoted are all potential values, these may be considered to be overestimates, particularly for summer months with high soil moisture deficits. Although no reliable figures are available for actual evapotranspiration, it is likely that the development of high soil moisture tension as the soil dried out during the summer months resulted in a reduction in transpiration rates. With respect to the runoff figures, the values for volume of runoff per month are as accurate as is possible using the present gauging structure and water level recorder. The most likely source of error arises in converting volume of runoff to mm. Reference has already been made to the problem of defining the catchment watershed in an area of low undulating relief. Overestimating the catchment area would result in an underestimate of runoff (mm) and vice versa. Finally, in an area of complex drift deposits, the subsurface catchment may not necessarily coincide with the catchment as defined by the surface topography and contours. Bands of tilting impermeable drift (such as the heavy clays which occur in the catchment) may allow water to seep across apparent topographic watersheds from one catchment to another.

While the computation and interpretation of annual water balances is beset with problems, the seasonal trends are more easily analysed. At the start of the water year the deficits of the previous summer are soon reversed as evapotranspiration falls off leaving a surplus of water to make up soil moisture deficits, and subsequently to augment depleted ground water storage. November figures illustrate this situation well with a surplus of 15.0 mm in 1978/79 and 22.5 mm in 1979/80. On both occasions a corresponding reduction in soil moisture deficit and an increase in ground water levels was observed. Throughout this late autumn/early winter period, little rainfall is getting through to the stream channel and runoff figures remain at a low level. Increasingly though, from December to March, soil moisture deficits are eliminated, ground water is topped up and more and more of the rainfall is making its way to the channel, giving maximum total runoff and peak flow conditions in February and March. With increasing air temperatures, reduced humidity and the onset of the growing season, evapotranspiration rises sharply in April, establishing the pattern of summer deficits with a corresponding reduction in stream run-off, the development of soil moisture deficits and a depletion of ground water storage.

Monthly rainfall/runoff relationships

In view of the numerous factors controlling the runoff from the catchment, it is not surprising that there is no simple correlation between monthly rainfall and runoff. Nevertheless, for all 24 months, a correlation coefficient (derived by the least squares method, see Ebdon, 1977) of +0.558 (significant at the 99% level) was obtained. If "winter" (December–April) and "summer" (May–November) months are separated, two quite distinct trends become evident (Fig. 13a). The regression line for summer months with its very gentle slope shows that higher rainfalls only produce small increases in runoff during this period. Wet summer months thus generate relatively little runoff. The wet May of 1979 with 86.6 mm of rainfall generated a mere 10.0 mm of runoff, whereas a similar rainfall in March 1980 gave rise to a spectacular 56.0 mm of runoff, including a peak flow of 516.0 l sec^{-1}. The winter regression line has a steeper slope and intersects the runoff axis at a much higher level, indicating that even if no rain were to fall during a winter month, runoff would still be substantial. Particularly during winter, certain months deviate considerably from the general trend as summarised by the regression line. In both years, February and March show strong positive residuals, that is they had more runoff relative to their rainfall than the general trend suggests, and therefore lie above the line. This period of the year has already been identified as that of greatest storage, and, therefore, likely to generate greatest amounts of runoff. Both sets of December data show negative residuals; for their high rainfalls they have rather less runoff than the general overall winter trend would indicate. This is not surprising in view of the spare storage capacity still present. Negative residuals again occur in April as rising evapotranspiration levels begin to reduce the amount of rainfall available for runoff.

Monthly runoff and other catchment variables

In view of the apparent significance of evapotranspiration and storage variables in explaining anomalies in the relationship between rainfall and runoff, fairly high levels of correlation between runoff and these other key variables in the catchment are to be expected. The tendency for high evapotranspiration to be associated with low runoff is expressed in Fig. 13b, the data showing a negative correlation of −0.614. The presence of large negative or positive residuals for certain months shows that it is not merely the amount of evapotranspiration in that month which determines runoff, but more particularly, it is the cumulative effect on storage of the surpluses or deficits of previous months. October and November, for example, despite low evapotranspiration rates, show low runoff totals, while in March evapotranspiration is already on the increase but maximum storage conditions ensure that high runoff is maintained.

The high dependence of monthly runoff on storage levels as well as on that month's evapotranspiration is well illustrated by Fig. 13c and d. Monthly soil moisture deficit and runoff show a strong negative correlation ($r = -0.820$) though the scatter of points suggests something of a curvilinear, rather than a linear, trend. Instead of there being a regular increase in runoff with decreasing soil moisture deficit, it would appear that in most cases runoff will not show any substantial increase until soil moisture deficits have been largely eliminated. Until such conditions are reached, soil water tension will remain high, thus retaining soil water rather than releasing it as throughflow.

Ground water levels and runoff show a positive correlation ($r = +0.810$) and again a curvilinear relationship is suggested. While ground water levels are in their lower range, a rising water table produces little or no corresponding increase in runoff. A critical level seems to be reached at just over 2.0 m above datum when further increases in ground water level are associated with substantial increases in runoff (Fig. 13d). This rapid increase in runoff during the winter months is also dependent upon greater contributions from ground water when water tables rise to a level at which they reach the surface in hollows and valley bottoms and thus contribute flow to the channel.

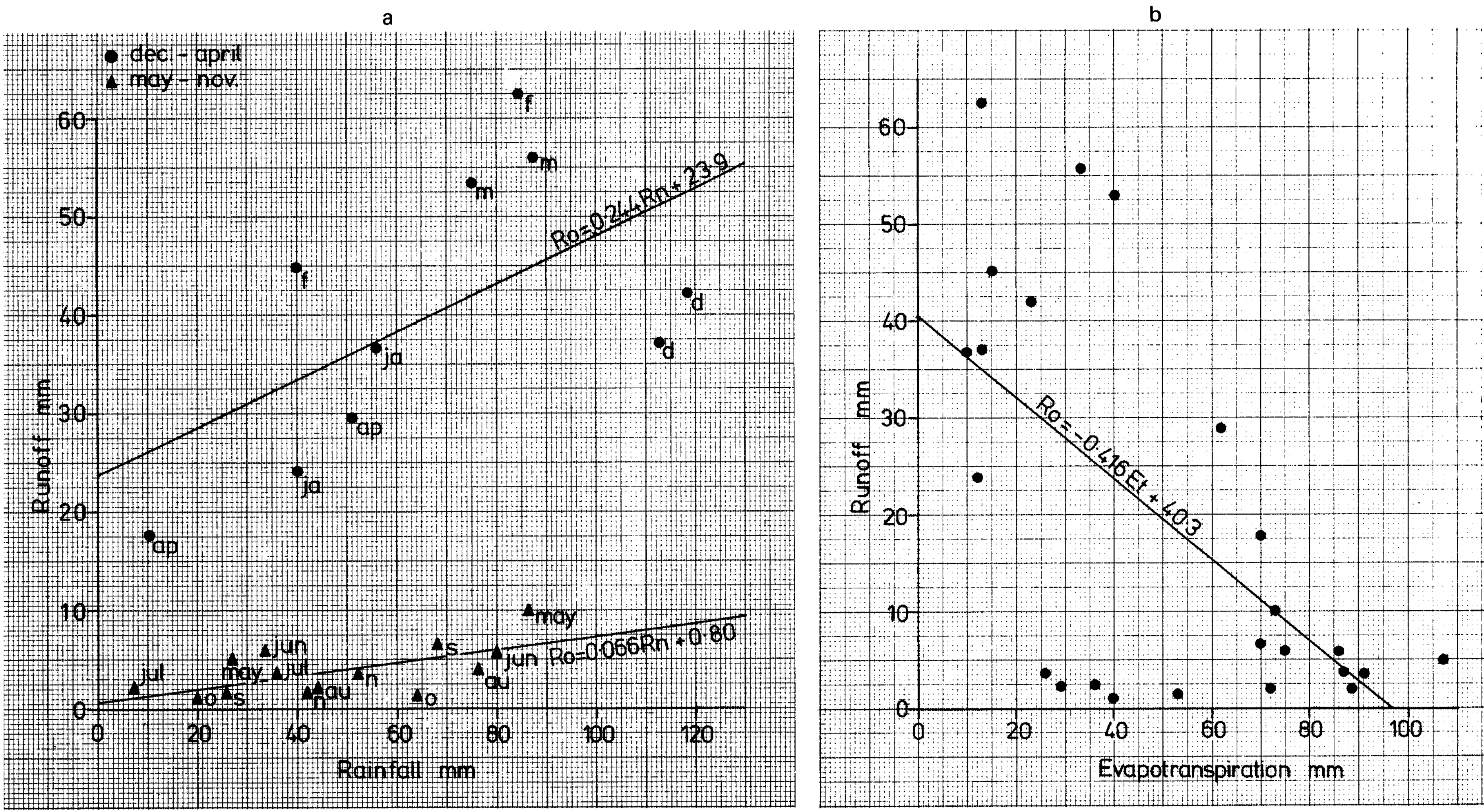
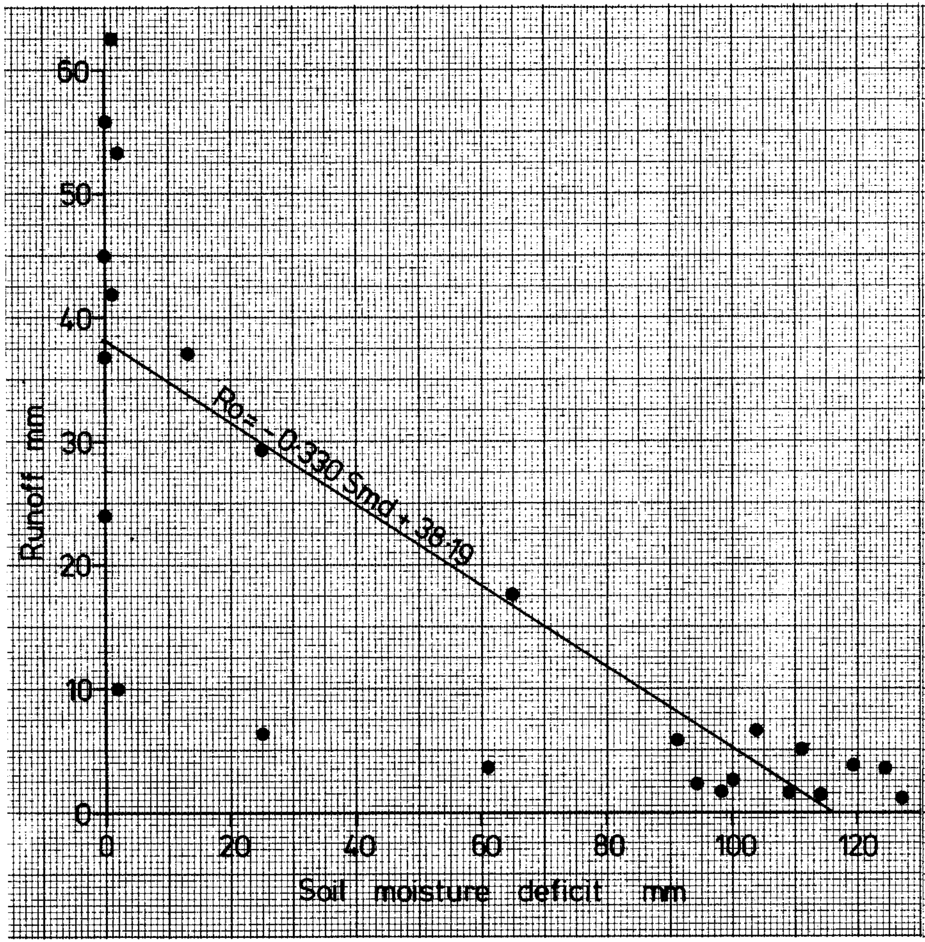


FIG. 13.
Relationships between monthly values of hydrological variables
a) rainfall and runoff
b) evapotranspiration and runoff
c) soil moisture deficit and runoff
d) groundwater level and runoff

c


Runoff mm
Ro = -0·330 Smd + 38·19
Soil moisture deficit mm
0
20
40
60
80
100
120
10
30
50

d

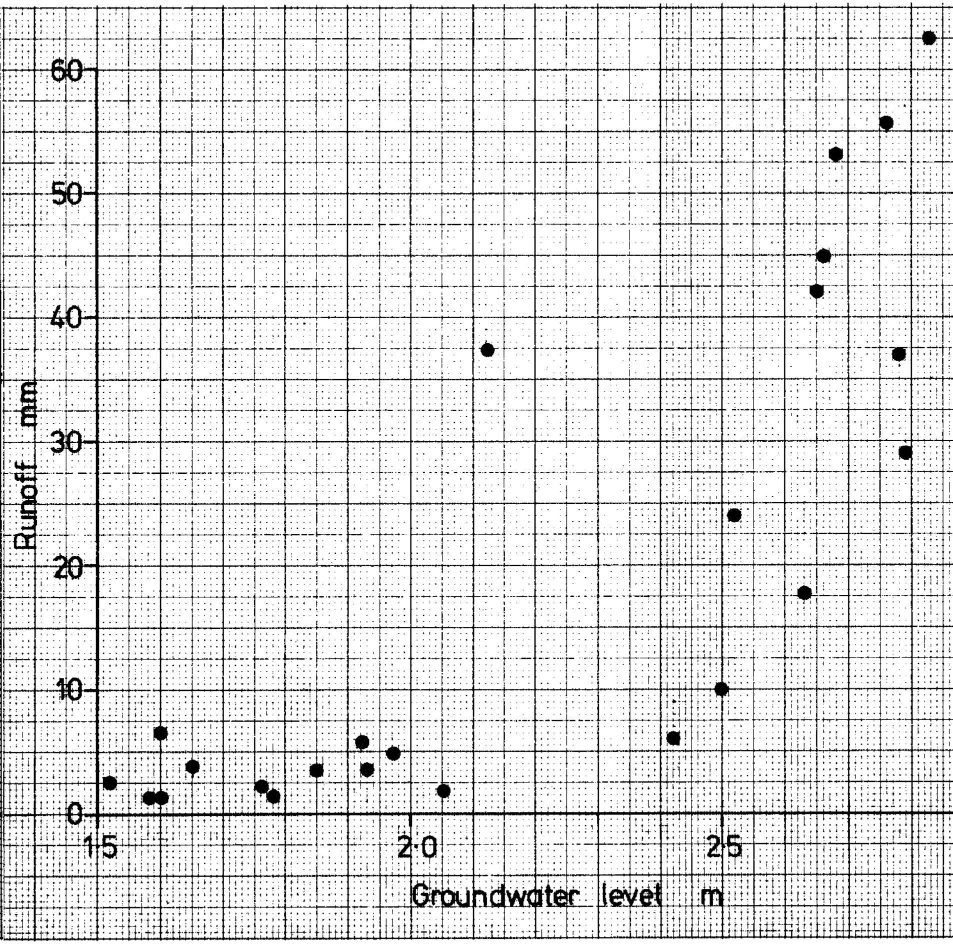

Runoff mm
Groundwater level m
1·5
2·0
2·5
0
10
20
30
40
50
60

Fig. 14 shows the pronounced seasonal nature of ground water levels over two years. Minimum levels in September/October are followed by periods of recharge, though maximum levels were not attained until March or even April, the rate of recharge depending upon the amounts of autumn and winter rainfall. These results indicate that groundwater contributions to the stream may not reach their maximum until the end of the winter period. Rising evapotranspiration in late April and May produces a sharp halt to recharge, followed by a steady depletion as base-flow drains the groundwater store through the summer months.

For the period October 1979–May 1980, Fig. 15 illustrates the relationships between rainfall, soil water potential (from tensiometer data) and stream discharge. Initially, high soil moisture tensions in October and early November results in little of the rainfall being converted to runoff, the rainfall input being taken up to make good the soil moisture deficit. It is not until late November that reduced soil moisture tensions are established, thus allowing much more of the rainfall to reach the channel from early December onwards. The maintenance of low soil moisture tensions until the end of March results in a succession of high discharge events following each significant rainfall input. The combination of a dry April and rising evapotranspirations gave sharply increasing soil water tensions (culminating in the failure of the tensiometer mercury column in late May) and a rapid drop-off in stream discharge.

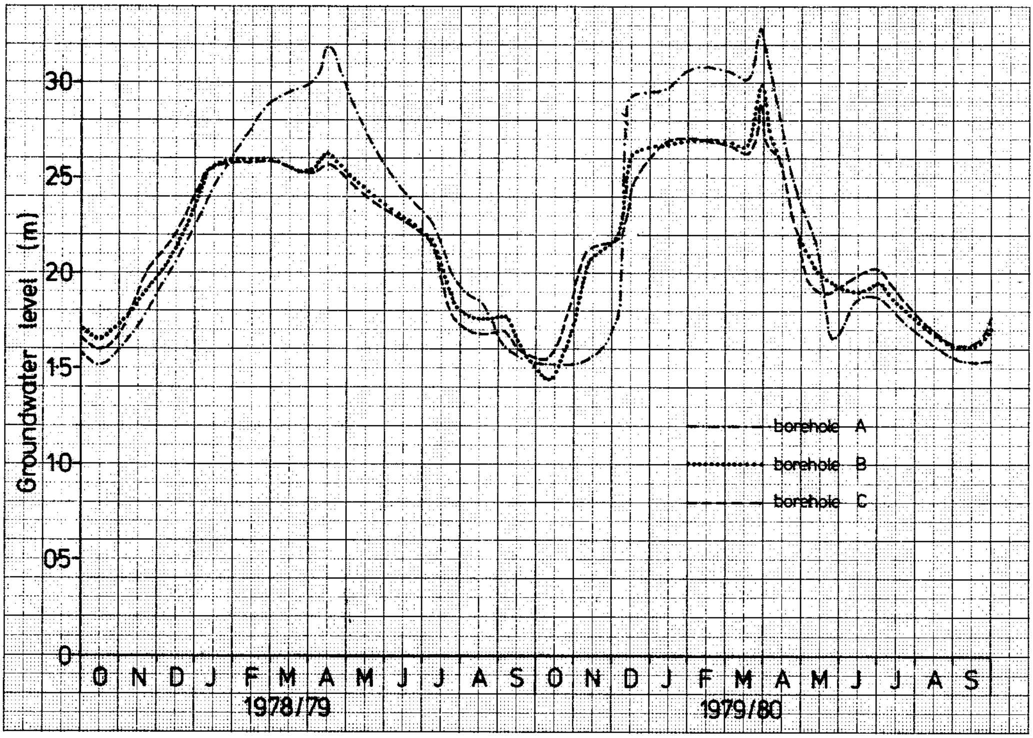


FIG. 14.
Water table variations 1978/79, 1979/80

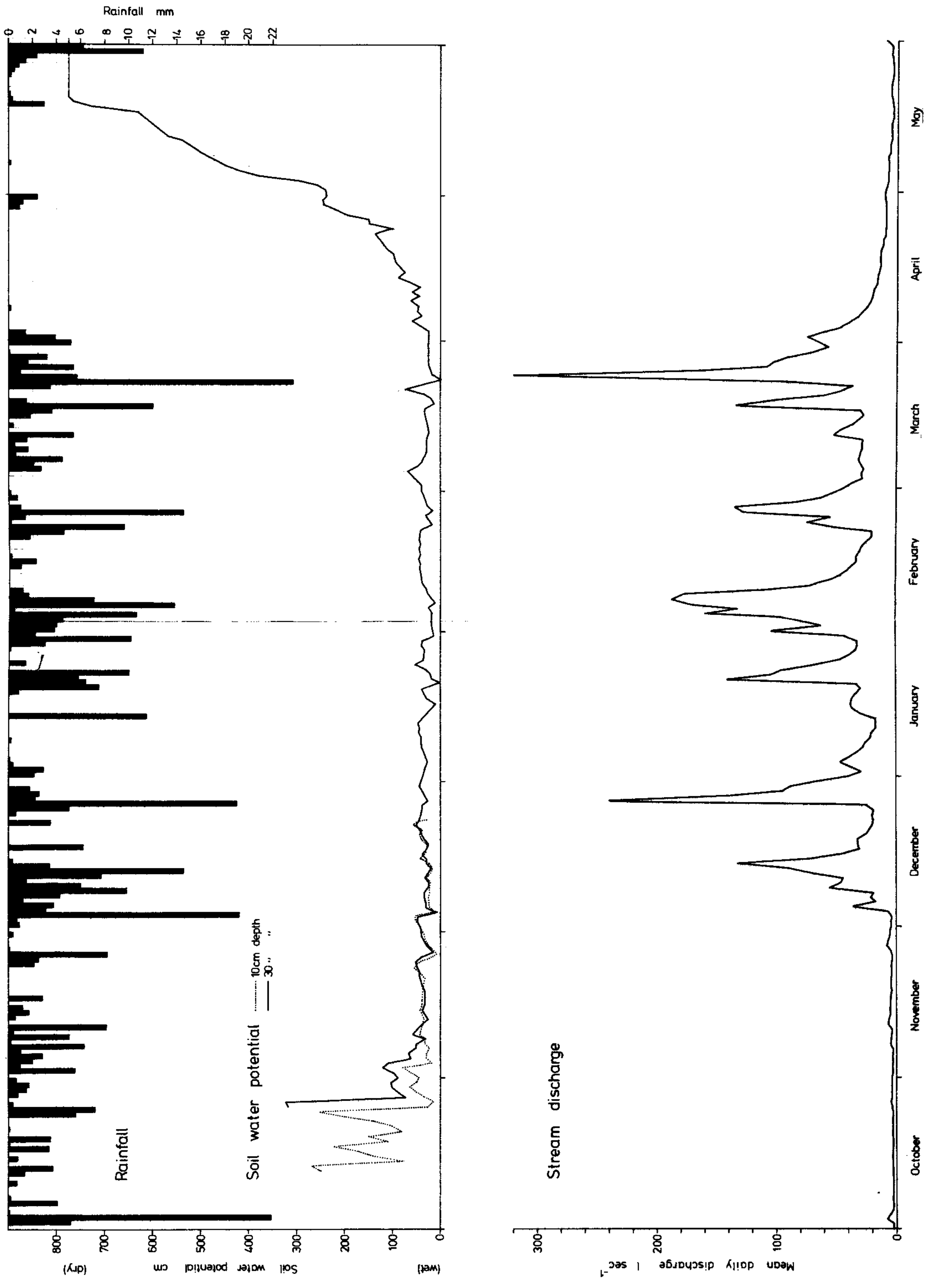


FIG. 15.
Soil water potential, runoff and rainfall, Oct. 1979–May 1980

Flow duration data

The overall frequency of occurrence of different flows may be expressed as a flow duration curve (Fig. 16) which indicates the percentage of time during the water year when a particular flow level was equalled or exceeded. While 100% of the time the flow was in excess of 1.2 l sec^{-1}, the curve drops off very sharply, so that only 50% of the flow was above 10.0 l sec^{-1} and only 10% of the flows were greater than 50 l sec^{-1}. The high discharges occupy an extremely small percentage of time through the year, flows in excess of 200 l sec^{-1} occurring for rather less than 1% of the time. The form of the flow duration curve is best explained by reference to the periods of time during which different types of runoff are occurring. Flows between storm events when there is little or no quickflow contribution to the stream fall within the range, 1.2 l sec^{-1} at times of minimum storage (September/October) to some 50 l sec^{-1} at times of high storage, this being the highest discharge which has been sustained for any length of time without rainfall. Since only 10% of the flows are in excess of this level, it may be estimated that for 90% of the time the stream is maintained at relatively low flow levels largely by delayed flow contributions. For the remaining 10% of the time, flows in excess of 50 l sec^{-1} are attained when delayed flow is augmented by contributions from quickflow processes.

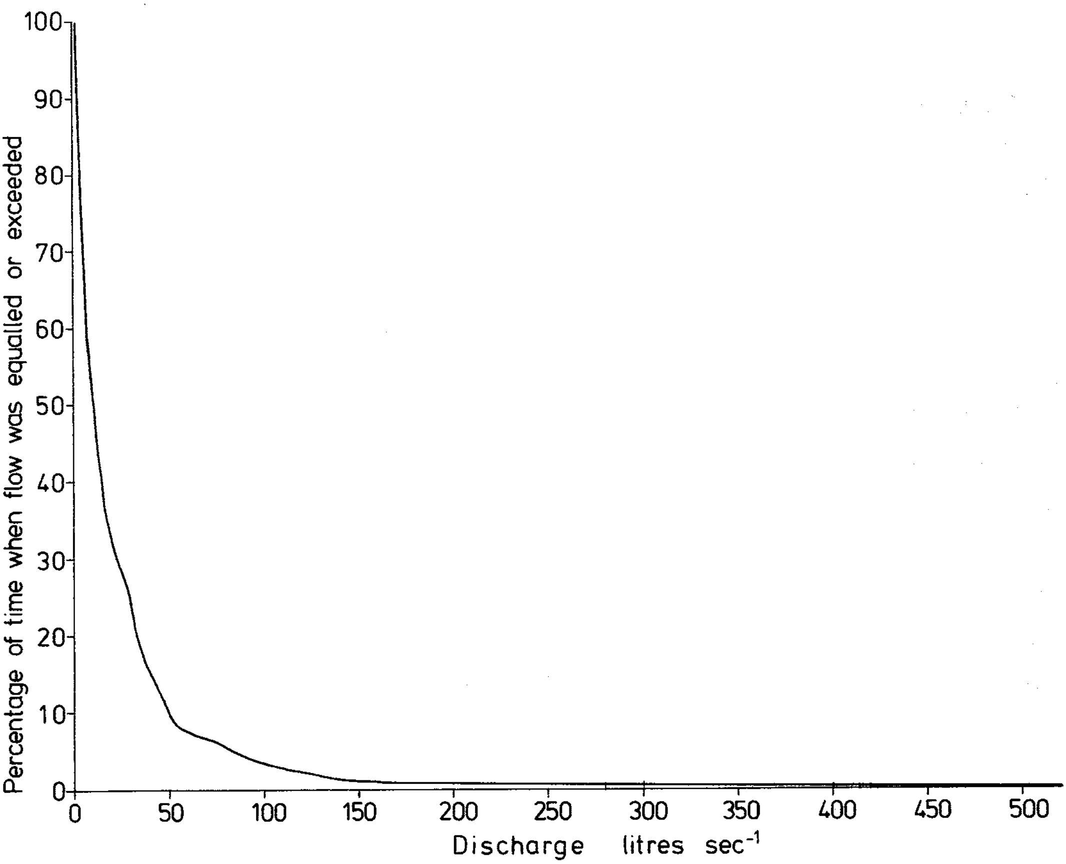


FIG. 16.
Flow duration curve, Preston Montford brook

Analysis of Storm Hydrographs

The results of the eighteen storms with their rainfall data, resulting runoff and antecedent conditions are listed in Table 4. The amount of storm runoff generated bears little direct relationship to the total storm rainfall. For example, the largest of all rainfall events (40.1 mm) occurring in August, yielded storm runoff amounting to a mere 0.47 mm, representing a runoff percentage of 1.2 and a peak discharge of 180.0 l sec^{-1}. In contrast, March rainfall half as great (20.9 mm) generated 8.57 mm of storm runoff (41% of the rainfall) with a peak discharge of 518.0 l sec^{-1}. This marked difference in runoff response cannot be explained either in terms of the amount or intensity of rainfall (the August storm was also the more intense as well as producing the greater total). The more important variables would appear to be those relating to antecedent conditions. For example, the antecedent 30 day rainfall for the August storm was 37.8 mm compared with 77.9 mm for the March event, while the 30 day evapotranspiration totals were 50.0 mm and 26.7 mm respectively. This gives a 30 day deficit of −12.2 mm prior to the August event, and a 51.2 mm surplus preceding the March event, indicating much higher storage conditions and, therefore, less capacity to absorb further rainfall. A very simple indirect indicator of storage conditions, and hence a useful variable for predicting subsequent runoff, is antecedent discharge, the stream flow rate prior to the storm. For the August event, antecedent discharge was 2.0 l sec^{-1} compared to 90 l sec^{-1} in March, suggesting very different base flow contributions, and hence ground water and soil moisture storage levels. To investigate more fully the variables which control runoff from the catchment, correlation coefficients for each variable against each in turn of the other variables were computed to give a correlation matrix (Table 5). It must be emphasised that some of the significant values are a result of the variables being interdependent, rather than because one variable is directly dependent upon another in any causal sense. In drawing conclusions from the correlation matrix, it is more useful to pick out variables which are considered to be dependent and see how closely they correlate with other independent variables which might be affecting them. Taking storm runoff in mm as a dependent variable and storm rainfall as the independent variable, a nonsignificant correlation is obtained ($r = +0.114$), showing that rainfall in itself is not a good predictor of runoff. Scanning the other coefficients in the runoff row, however, indicates that other independent variables correlate quite closely (e.g. 30 day AMI, $r = +0.624$; antecedent discharge, $r = +0.706$). These high values provide support for the hypothesis that preceding moisture conditions are of great importance to runoff control. Note that antecedent discharge is itself highly correlated with 30 day AMI, showing that while these two variables are independent with respect to storm runoff, they are not independent of each other. Antecedent discharge is dependent on 30 day AMI. Runoff percentage correlates even more closely with the antecedent moisture indices (10 day AMI, $r = +0.630$; 20 day AMI, $r = +0.708$; 30 day AMI, $r = +0.731$), demonstrating that AMI is likely to be a very useful variable in predicting runoff and that the longer time periods provide better explanations than the shorter ones.

Having established strong correlations between runoff and certain independent variables, the next possibility is that of using these relationships to predict runoff by means of regression techniques. (For the following more sophisticated methods, analysis will be based upon the storms recorded during 1979/80 as accurate instrumentation was not available for some of the earlier events.) Antecedent discharge

Table 4. *Storm hydrograph data, Preston Montford catchment*

Date of storm event	Total storm rainfall (mm)	Duration of rainfall (mm)	Maximum hourly rainfall (mm)	Mean rainfall intensity (mm hr⁻¹)	Peak storm discharge (l sec⁻¹)	Total storm discharge (Megalitres)	Total storm runoff (mm)	Runoff %	Lag time (hrs)	Time to peak (hrs)	Antecedent discharge (l sec⁻¹)	Height of peak (l sec⁻¹)	Rate of rise (l sec⁻¹ hr⁻¹)	Antecedent rainfall 5 days (mm)	Antecedent rainfall 10 days (mm)	Antecedent rainfall 20 days (mm)	Antecedent rainfall 30 days (mm)	Antecedent evapotranspiration 5 days (mm)	Antecedent evapotranspiration 10 days (mm)	Antecedent evapotranspiration 20 days (mm)	Antecedent evapotranspiration 30 days (mm)	Antecedent rainfall less antecedent evapotranspiration (mm) 10 days	20 days	30 days
21.8.77	40.1	4.5	10.9	8.9	180.0	1.49	0.47	1.2	3.5	5.0	2.0	178.0	35.6	18.7	23.2	32.9	37.8	10.2	20.3	40.6	50.0	2.9	−7.7	−12.2
1.5.78	17.2	11.0	2.6	1.6	200.0	9.72	3.08	17.9	7.0	10.5	10.0	190.0	18.1	7.2	18.1	35.1	37.4	8.6	16.3	31.6	46.9	1.8	3.5	−9.5
4.5.78	29.1	7.0	13.3	4.2	634.0	26.30	8.35	28.7	1.5	1.5	22.0	612.0	408.0	26.1	34.4	46.0	54.4	10.6	18.3	33.6	48.9	16.1	12.4	5.5
15.6.78	26.4	26.0	4.5	1.0	34.0	1.20	0.39	1.5	5.0	17.0	3.0	31.0	1.8	0.8	11.2	21.3	21.3	14.0	28.0	54.3	79.5	−16.8	−33.0	−58.2
23.11.78	11.7	8.0	3.2	1.5	10.5	0.38	0.12	1.0	6.0	6.5	2.0	8.5	1.3	7.5	18.6	24.6	30.1	4.9	9.7	19.4	31.4	8.9	5.2	−1.3
28.3.79	16.6	24.0	1.7	0.7	245.0	4.40	1.39	8.4	4.0	15.0	55.0	190.0	12.7	16.9	18.4	57.6	73.9	6.5	12.9	25.8	36.5	5.5	31.8	37.4
26.5.79	14.0	10.0	2.9	1.4	54.5	2.19	0.70	5.0	3.0	7.0	8.5	46.0	6.6	20.8	29.6	42.2	61.7	11.8	23.5	47.0	69.1	6.1	−4.8	−7.4
23.6.79	12.5	9.0	3.5	1.4	21.0	0.50	0.16	1.3	3.5	4.0	4.0	17.0	4.3	0.8	8.3	19.5	61.4	12.5	25.0	50.0	73.9	−16.7	−30.5	−12.5
3.10.79	22.3	13.0	15.7	1.7	22.5	0.50	0.16	0.7	5.0	6.0	6.0	16.5	2.8	6.0	13.2	23.2	25.0	7.6	16.5	34.2	51.9	−3.3	−11	−26.9
4.12.79	25.4	14.0	3.8	1.8	61.0	3.10	0.98	3.87	10.0	12.0	4.0	57.0	4.8	2.3	13.9	20.3	47.7	4.0	8.3	17.6	25.7	5.6	2.7	22.0
13.12.79	14.3	9.0	2.5	1.6	220.0	12.38	3.93	27.5	5.0	7.0	52.0	168.0	24.0	30.2	58.9	74.4	80.8	3.7	7.4	15.8	24.5	51.5	58.6	56.3
26.12.79	26.1	26.0	3.8	1.0	342.0	16.06	5.1	19.5	7.5	9.0	20.0	322.0	35.8	4.5	6.6	63.2	96.9	3.7	7.4	14.8	22.9	−0.8	48.4	74.0
25.2.80	15.1	29.0	1.6	0.5	214.0	15.5	4.9	32.6	9.0	9.0	54.0	160.0	17.8	18.4	20.9	47.0	87.8	2.1	4.2	8.4	12.0	16.7	38.6	75.8
23.3.80	20.9	9.0	6.5	2.3	518.0	27.0	8.57	41.0	6.0	9.0	90.0	428.0	47.6	16.0	22.4	49.7	77.9	5.3	10.6	21.2	26.7	11.8	28.5	51.2
7.8.80	13.0	5.0	4.3	2.6	20.4	0.58	0.18	1.4	6.0	9.5	3.8	16.6	1.8	10.1	19.9	29.8	39.9	14.1	28.6	57.9	87.2	−8.7	−28.1	−47.3
29.8.80	23.5	9.0	8.8	2.6	37.5	1.51	0.48	2.0	4.5	10.0	2.0	35.5	3.6	0.4	1.5	23.6	55.2	14.1	28.1	56.2	84.5	−26.6	−32.6	−29.3
20.9.80	36.6	6.0	15.4	6.1	103.0	5.50	1.75	4.8	7.0	5.0	2.6	100.4	20.1	7.6	13.4	20.5	46.2	11.7	23.3	47.1	75.2	−9.9	−26.6	−29.0
6.10.80	16.0	11.0	3.6	1.5	25.0	2.52	0.80	5.0	6.0	10.0	4.0	21.0	2.1	4.4	18.4	63.8	71.7	5.9	17.6	40.9	64.2	0.8	22.9	7.5

Height of peak ($l\ sec^{-1}$)	Rate of rise ($l\ sec^{-1}\ hr^{-1}$)	Antecedent rainfall 5 days (mm)	Antecedent rainfall 10 days (mm)	Antecedent rainfall 20 days (mm)	Antecedent rainfall 30 days (mm)	Antecedent evapotranspiration 5 days (mm)	Antecedent evapotranspiration 10 days (mm)	Antecedent evapotranspiration 20 days (mm)	Antecedent evapotranspiration 30 days (mm)	AMI 10 days (mm)	AMI 20 days (mm)	AMI 30 days (mm)
0.311	0.296	−0.029	−0.180	−0.237	−0.234	0.212	0.171	0.146	0.096	−0.214	−0.223	−0.187
0.027	−0.180	−0.142	−0.254	0.250	0.357	−0.385	−0.386	−0.400	−0.390	−0.008	0.370	0.432
0.209	0.397	−0.001	−0.120	−0.371	−0.404	0.344	0.317	0.303	0.295	−0.238	−0.390	−0.402
0.198	0.269	0.205	0.075	−0.244	−0.306	0.338	0.306	0.295	0.231	−0.088	−0.309	−0.309
0.993	0.745	0.574	0.358	0.502	0.454	−0.287	−0.393	−0.444	−0.485	0.451	0.547	0.544
0.917	0.639	0.516	0.365	0.518	0.542	−0.394	−0.483	−0.524	−0.539	0.499	0.601	0.625
0.918	0.639	0.515	0.365	0.517	0.542	−0.393	−0.482	−0.523	−0.538	0.499	0.600	0.624
0.773	0.428	0.579	0.473	0.606	0.627	−0.522	−0.594	−0.624	−0.636	0.631	0.709	0.731
−0.222	−0.457	−0.366	−0.263	−0.053	0.154	−0.546	−0.507	−0.477	−0.427	0.042	0.228	0.342
−0.289	−0.481	−0.355	−0.289	0.052	0.016	−0.065	−0.016	−0.006	−0.005	−0.207	0.034	−0.006
0.554	0.119	0.555	0.414	0.606	0.639	−0.532	−0.575	−0.589	−0.617	0.579	0.690	0.726
1.000	0.797	0.541	0.327	0.455	0.397	−0.231	−0.340	−0.394	−0.434	0.403	0.491	0.481
	1.000	0.470	0.328	0.171	0.055	0.079	−0.023	−0.077	−0.093	0.255	0.145	0.087
		1.000	0.879	0.585	0.347	−0.252	−0.329	−0.355	−0.396	0.810	0.549	0.431
			1.000	0.575	0.226	−0.281	−0.320	−0.326	−0.337	0.895	0.527	0.328
				1.000	0.776	−0.563	−0.531	−0.505	−0.501	0.678	0.882	0.732
					1.000	−0.537	−0.526	−0.506	−0.494	0.415	0.748	0.853
						1.000	0.980	0.954	0.933	−0.670	−0.862	−0.858
							1.000	0.994	0.980	−0.708	−0.865	−0.881
								1.000	0.990	−0.710	−0.852	−0.876
									1.000	−0.712	−0.844	−0.875
										1.000	0.799	0.658
											1.000	0.923
												1.000

Table 5. *Correlation matrix for storm hydrograph variables*

	Total storm rainfall (mm)	Duration of rainfall (hrs)	Maximum hourly rainfall (mm)	Mean rainfall intensity (mm hr^{-1})	Peak storm discharge (l sec^{-1})	Total storm discharge (Megalitres)	Total storm runoff (mm)	Runoff %	Lag time (hrs)	Time to peak (hrs)	Antecedent discharge (l sec^{-1})
Total storm rainfall (mm)	1.000	−0.110	0.705	0.785	0.254	0.114	0.114	−0.081	−0.054	−0.170	−0.211
Duration of rainfall (hr)		1.000	−0.416	−0.578	0.065	0.118	0.118	0.201	0.366	0.616	0.287
Maximum hourly rainfall (mm)			1.000	0.662	0.155	0.066	0.066	−0.134	−0.289	−0.522	0.250
Mean rainfall intensity (mm hr^{-1})				1.000	0.145	−0.017	−0.017	−0.157	−0.271	−0.491	0.252
Peak storm discharge (l sec^{-1})					1.000	0.940	0.940	0.825	−0.192	−0.246	0.649
Total storm discharge (Megalitres)						1.000	1.000	0.942	0.016	−0.230	0.710
Total storm runoff (mm)							1.000	0.942	0.015	−0.230	0.709
Runoff %								1.000	0.131	−0.125	0.832
Lag time (hrs)									1.000	0.373	0.075
Time to peak (hrs)										1.000	0.128
Antecedent discharge (l sec^{-1})											1.000
Height of peak (l sec^{-1})											
Rate of rise (l sec^{-1} hr^{-1})											
Antecedent rainfall 5 days (mm)											
Antecedent rainfall 10 days (mm)											
Antecedent rainfall 20 days (mm)											
Antecedent rainfall 30 days (mm)											
Antecedent evapotranspiration 5 days (mm)											
Antecedent evapotranspiration 10 days (mm)											
Antecedent evapotranspiration 20 days (mm)											
Antecedent evapotranspiration 30 days (mm)											
AMI 10 days (mm)											
AMI 20 days (mm)											
AMI 30 days (mm)											

for these nine storm events during the water year 1979/80 correlates very closely with storm runoff ($r = +0.916$). The relationship between storm runoff (Qs) and antecedent discharge (Qa) may be summarised by the regression equation:

$$Qs = 0.861 + Qa\ 0.0815$$

By substituting known values of Qa for each storm, predicted values for Qs can be derived and compared to the observed values (Table 6). In most cases this linear regression model provides a reasonably good prediction of runoff (predicted values correlate very closely with observed, $r = +0.916$). To assess the differences between predicted and observed runoff values, the residuals may be calculated by subtracting predicted from observed, giving a positive residual where observed is more than predicted. These residuals may be considered to represent the amount of variation in runoff which cannot be explained in terms of Qa, and it is therefore useful to turn to Table 4 to see if the pattern of variation in the residuals corresponds with any other independent variable. There would appear to be some relationship between the size and sign of the residuals and total storm rainfall, in that storm runoff values with positive residuals are generally associated with higher rainfall totals and storms with negative residuals with the lower rainfalls. It seems then that rainfall might be a useful independent variable to explain that part of the variation in storm runoff not accounted for by Qa.

Table 6. *Predicted values of total storm runoff*

Date of Storm	Antecedent Discharge l sec^{-1} (Qa)	Total storm Runoff mm (Qs)	Predicted Runoff mm Qs (i)	Residuals mm (i)	Storm Rainfall mm Q (Rns)	Predicted Runoff mm Qs (ii)	Residuals mm (ii)
4.12.79	4.0	0.98	1.18	−0.20	25.4	1.52	−0.54
13.12.79	52.0	3.93	5.09	−1.16	14.3	4.19	−0.26
26.12.79	20.0	5.10	2.49	+2.61	26.1	2.97	+2.13
25. 2.80	54.0	4.90	5.26	−0.36	15.1	5.10	−0.20
23. 3.80	90.0	8.57	8.16	+0.41	20.9	7.25	+1.32
7. 8.80	3.8	0.18	1.17	−0.99	13.0	0.61	−0.43
29. 8.80	2.0	0.48	1.02	−0.54	23.5	1.20	−0.72
20. 9.80	2.6	1.75	1.07	+0.68	36.6	2.21	−0.46
6.10.80	4.0	0.80	1.18	−0.38	16.0	0.85	−0.05

(i) predictions and residuals with Qa as the independent variable in the regression
(ii) predictions and residuals with Qa and Rns as the independent variables in the regression

The bivariate regression so far applied can only attempt explanation of a dependent variable with respect to a single independent variable. If we wish to attempt an explanation with respect to more than one independent variable, a multiple regression model of the following general form can be applied (King, 1969):

$$y = a + x_1b_1 + x_2b_2 + \ldots x_nb_n$$

where y = the dependent variable
a = the regression constant
x_i = the i th independent variable
b_i = the partial regression coefficient for variable i

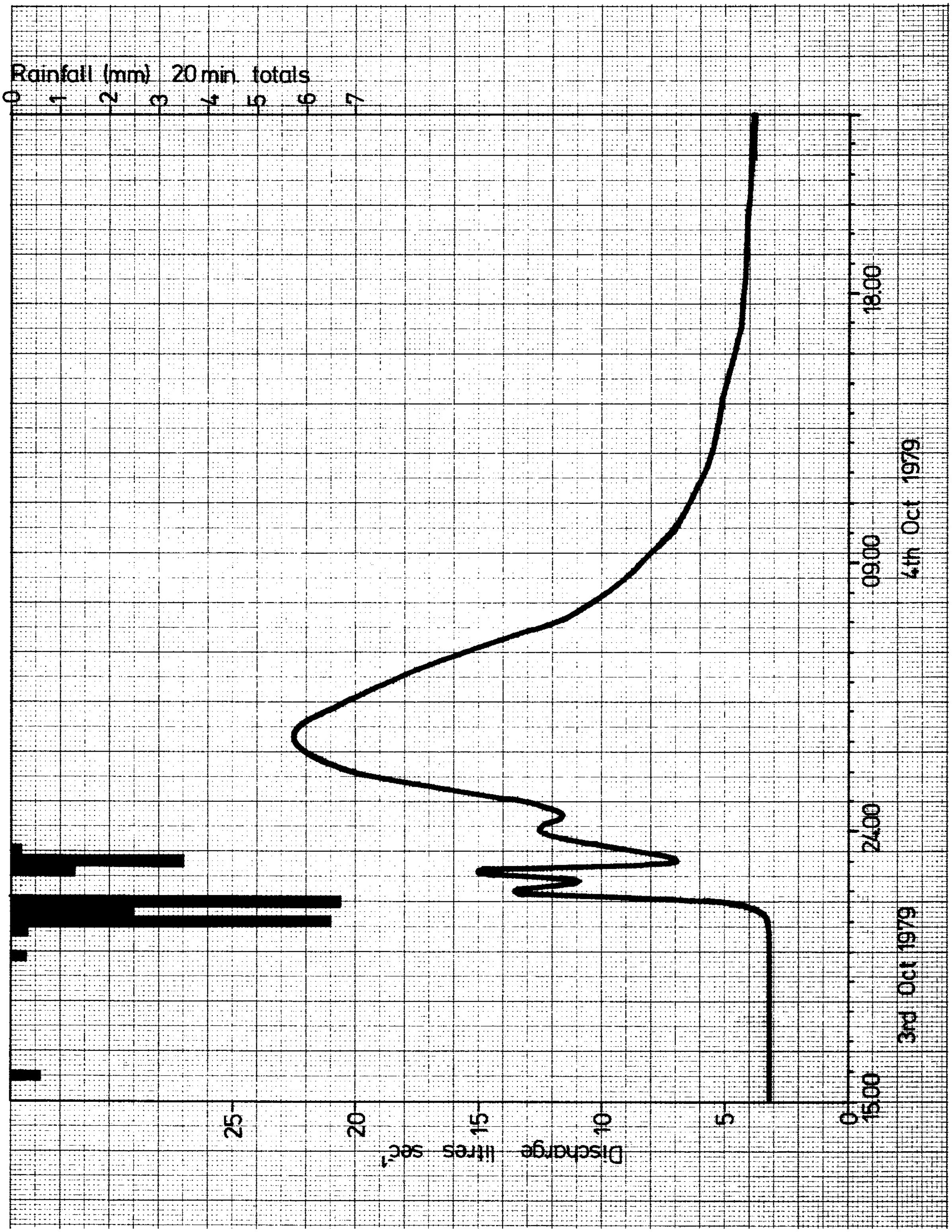


FIG. 17.
Storm hydrograph, 3rd–4th October 1979

30 days (mm)
).187
).432
).402
).309
).544
).625
).624
).731
).342
).006
).726
).481
).087
).431
).328
).732
).853
).858
).881
).876
).875
).658
).923
.000

A multiple regression on the same nine storm events with Qs as the dependent variable (y), Qa as the first independent variable (x_1), and storm rainfall (Rns) as the second independent variable (x_2), gives the following regression equation:

$$Qs = -0.659 + Qa\ 0.087 + Rns\ 0.072$$

To assess the level of correlation between Qs and Qa and Rns together, a multiple correlation coefficient (r) may be calculated giving a value of +0.95. By squaring this value, the coefficient of determination may be obtained which expresses as a percentage the amount of variation in Qs which is explained by Qa and Rns together. This gives a value of 90.9, indicating that a very high proportion of storm runoff variation can be accounted for by reference to Qa and Rns. The multiple regression model can be used in the same way as the bivariate equation to predict values of Qs, in this case for given values of Rns as well as Qa. These predicted values are listed in Table 6 and it will be noted that, by building rainfall into the regression, the residuals have in six cases been reduced relative to those where Qa was the sole independent variable. In only three cases does Qa alone give a better prediction of Qs. The level of correlation between predicted and observed values of Qs also increases to +0.95 with the multiple regression model.

While statistical analysis can indicate which variables are most useful in predicting storm runoff, it cannot explain directly the mechanism of the runoff process in terms of how much water is being contributed from each of the possible source areas. To achieve a more complete explanation of storm hydrographs, the amount and timing of runoff from the different contributary areas need to be determined. The following sources of storm runoff in the catchment have been identified:

(a) Direct channel precipitation (about 0.0004 km^2)
(b) Pond at Onslow Hall (0.0124 km^2)
(c) Road surfaces (maximum 0.016 km^2)
(d) Surface runoff from fields (variable area)
(e) Throughflow assisted by tile drains from farmland.

Major problems were encountered in attempting to monitor flow from each of these sources, not least being denial of access to south-eastern parts of the catchment. In the case of piped inputs (road drains, tile drains and culverts) to the channel, direct measurement (by timed collection or flowmetering) is possible at low stream discharge, but as the stream rises, the points of outflow become submerged below stream level. Where measurement was possible, road drains were found to have a lag time of 0.5–1.0 hours and yielding water with high suspended sediment and low conductivity, while tile drains showed lag times of 5.0–6.0 hours yielding water low in sediment but with higher conductivity (i.e. higher solute concentrations). Figs. 17 and 18 show results for two high intensity rainfall events when it was possible to relate different peaks on the hydrograph to specific sources of runoff. During the October event (Fig. 17) the three sharp shortlived runoff peaks, occurring about one hour after three separate bursts of rainfall, yielded a total runoff of 32,400 litres. Only 9,000 litres of this can be accounted for by direct channel precipitation, indicating that other rapid runoff sources must also have been making contributions to the stream. Relatively high suspended sediment loadings (166 mg l^{-1}) and the short lag favour road drainage and, possibly, some surface runoff from fields as likely sources.

The peak discharge for this event had a lag time of 5 hours which, together with the lower sediment loadings (65 mg l^{-1}) and higher conductivity, is consistent with a throughflow origin via tile drains.

Similar trends are evident for the storm event of 4–6 May 1978 (Fig. 18), though the stream response was very much more dramatic, reflecting not only a higher total rainfall (29.1 mm compared with 22.3 mm), but greater intensity (10 mm in 20 minutes compared with 6.7 mm) and higher storage levels in the catchment prior to the event (30 AMI = +5.5 compared with −36.9 for the October event). While the distinction between an initial rapid runoff peak (lag time 1.5 hours) and a secondary peak with a 5 hour lag is similar to the October storm, in this case there was only a single initial runoff peak (reflecting the isolated, rather than three-peaked, rainfall input). This exceeded the secondary peak, reaching a record discharge of 634 l sec^{-1} and contributing in total 1,250,000 litres of water. Assuming 100% runoff from the channel surface, roads and the pond, a total yield of only 600,000 litres is obtained. To account for the additional rapid runoff (650,000 l), an area of at least 0.0325 km^2 must have been contributing surface runoff to the stream. Observation the following morning confirmed the occurrence of widespread surface runoff from the lower slopes of arable fields adjoining the channel and incipient rill and sheet wash had occurred. The exceptional suspended load concentration (6,756 mg l^{-1}) at 20.10 hrs on 4 May, as the stream was approaching its initial peak, adds further support to the substantial contribution of surface runoff from fields. The secondary storm peak, with its longer lag, much reduced suspended sediment (58 mg l^{-1}) and rising conductivity, indicates a predominant contribution from tile drains rather than surface sources.

Both the storm events described here in detail are responses to high intensity rainfall events. They show initial direct runoff peaks some 1.0–1.5 hours after the most intense rainfall, to be followed by a later peak after a lag of some 5 hours. Most of the other storm events recorded (in the main, responses to less intense falls of rain) do not show a distinct direct runoff peak, or else it is only a very minor component of the hydrograph. Widespread surface runoff is not, therefore, a frequent event in the catchment, but when conditions coincide to give rise to the process, the results are dramatic in terms of both stream runoff and sediment yield.

SUSPENDED SEDIMENT LOAD

Fig. 19 shows no simple relationship between suspended sediment concentration and discharge, although the results of 148 water samples taken at intervals through a number of storm events at different times of the year show a significant positive correlation ($r = +0.45$). The scatter of samples taken between June and November (normally a period of low soil moisture values, low runoff and reduced storage) shows a distinctly different pattern from samples taken during higher runoff events from December to May. Summer and autumn samples tend to have high suspended load concentrations relative to discharge. This may be associated with high intensity convectional rainfall combined with dry surface soil conditions, resulting in heavy concentrations of fine particles being washed into the stream, even though the total amounts of runoff may not be substantial. On 21 August 1977 a concentration of 1,036 mg l^{-1} was recorded immediately after 8 mm of rain fell on previously very dry ground in half an hour, although the discharge only rose to 12 l sec^{-1}. During the period, December to May, despite much higher discharges, suspended sediment

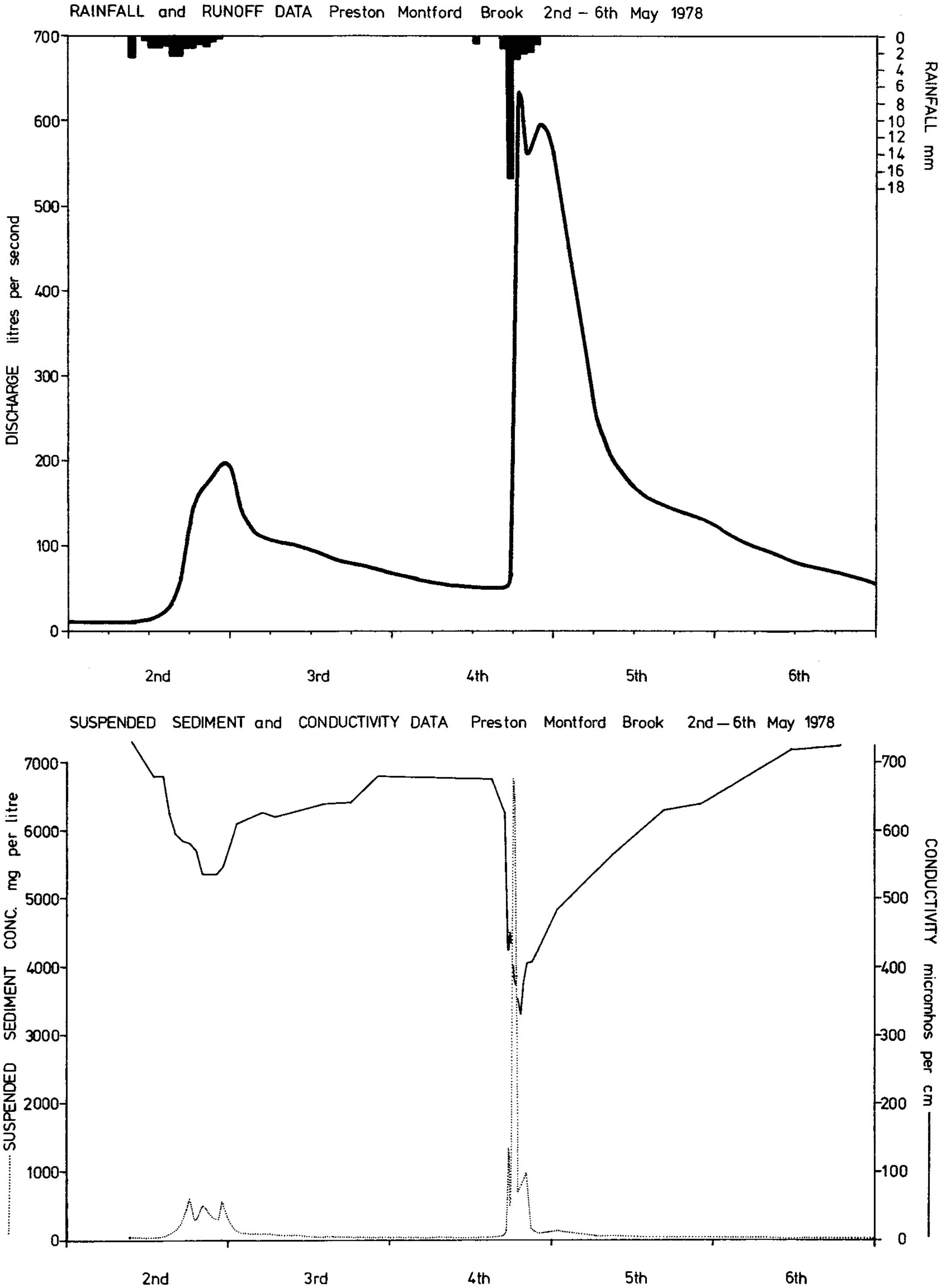


FIG. 18.
Storm hydrograph, rainfall, suspended sediment and conductivity 2nd–6th May 1978

concentrations were often not much greater than summer values, except for the few very high runoff events. During extremely high discharges, as, for example, on 4 May 1978, very heavy suspended loadings were recorded, reaching a peak of 6,756 mg l^{-1}. Over a period of $2\frac{1}{2}$ hours an estimated 14,416 kg of suspended sediment was removed from the catchment.

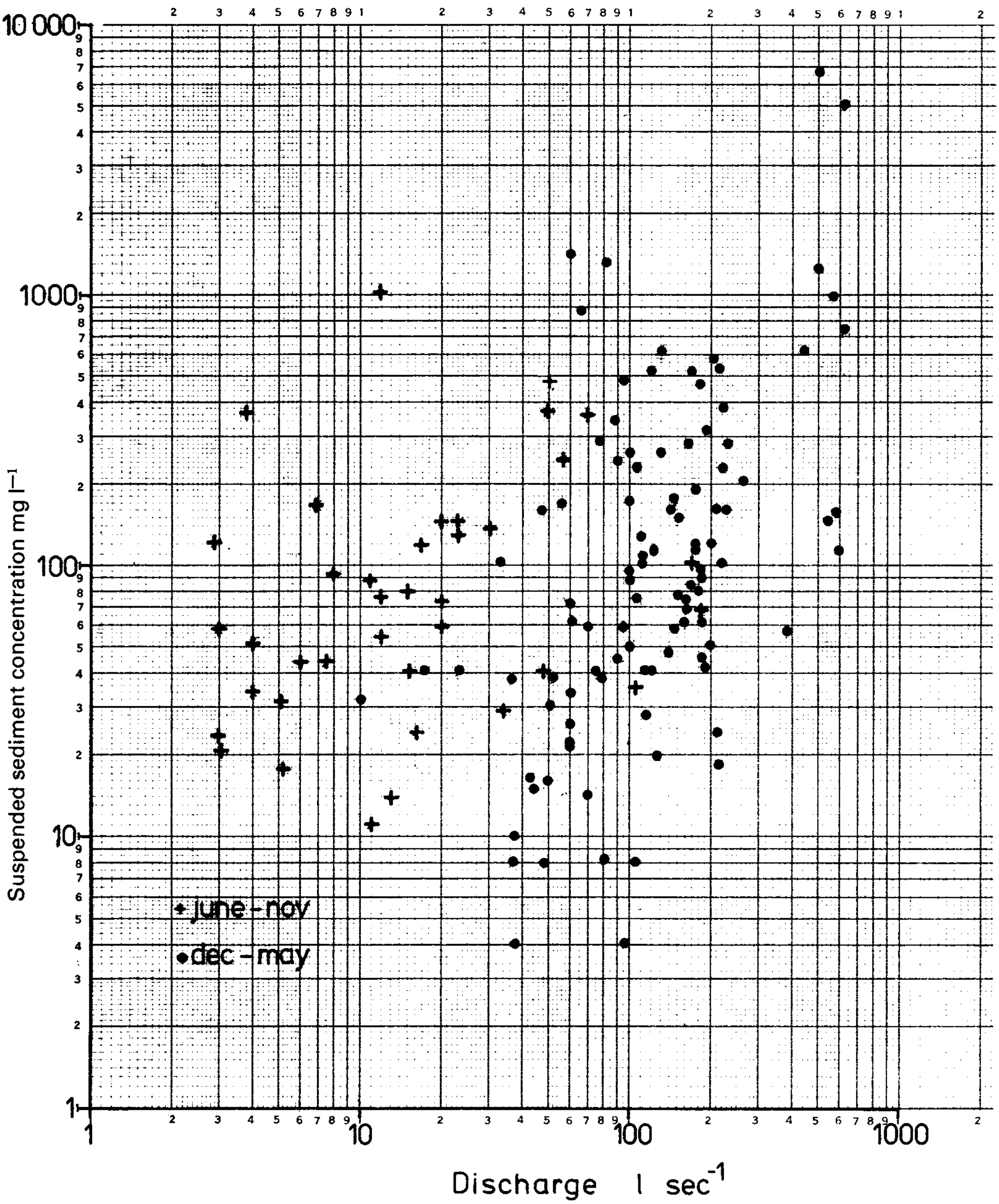


FIG. 19.
Scattergraph of suspended sediment and discharge

In addition to seasonal variations in the suspended sediment/discharge relationship, examination of individual storm events indicates that sediment loadings are by no means wholly dependent upon discharge. Invariably, the plot of suspended sediment concentration shows a steep rise during the rising limb of the hydrograph, a peak concentration which precedes peak flow, and a sharp decrease in sediment thereafter (Fig. 20). If two water samples were taken at the same discharge, one as

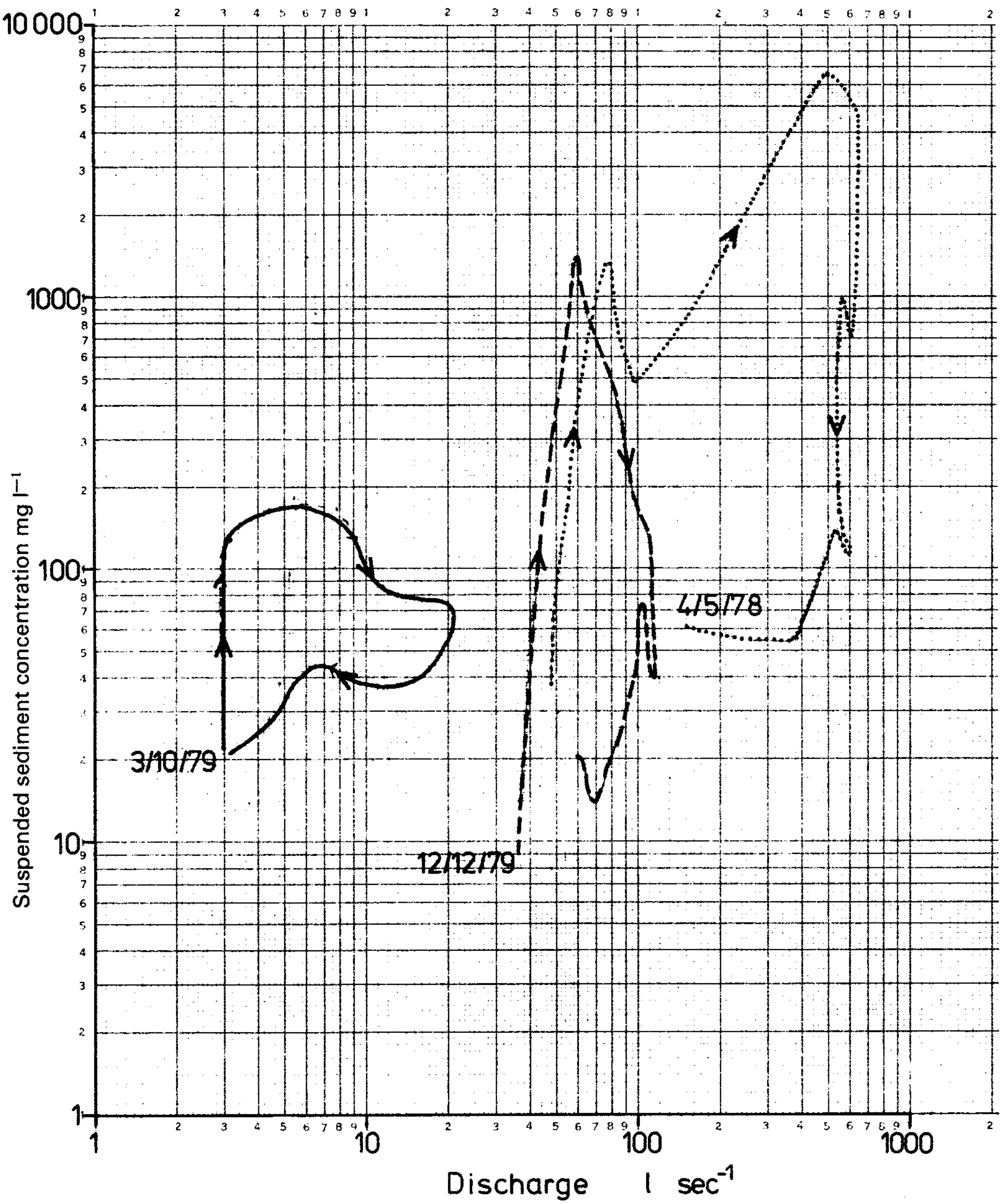


FIG. 20.
Variations in suspended sediment and discharge over three storms

the stream was rising, the other during recession, the former will contain a much higher suspended load concentration. This has interesting implications with respect to determining which flow events are the most important in terms of total suspended sediment removal. For most storm events, the greatest amounts of sediment removal occur at intermediate flows during the rising limb. At peak discharge, the greater volume of water does not usually compensate for the reduced sediment concentration, so that total sediment loss has begun to decline once the storm peak is reached. It would be useful to be able to determine a total suspended sediment loss for a full year, but this would necessitate at least daily sampling supplemented by hourly sampling during storm events, and even more frequently during rising stages. With the data available, a very approximate estimate of total annual suspended sediment loss has been attempted by combining flow duration data with a suspended sediment concentration/discharge regression. For the three years from October 1977 to September 1980, the mean number of days per year falling into each 5 litre flow class was determined. For each flow class, an estimate of suspended sediment concentration was made using the regression:

$$\text{suspended sediment concentration (mg l}^{-1}) = 2.5 \times \text{discharge} - 23.9$$

For each discharge class, the total sediment output was determined as follows:

$$\begin{array}{l} \text{Total annual} \\ \text{sediment yield} \\ \text{for flow class} \\ \text{(kg)} \end{array} = \frac{\begin{array}{l}\text{Concentration} \\ \text{of suspended} \\ \text{sediment for} \\ \text{flow class} \\ (\text{mg l}^{-1})\end{array} \times \begin{array}{l}\text{Discharge} \\ (\text{l sec}^{-1}) \\ \text{at mid-point} \\ \text{of flowclass}\end{array} \times \begin{array}{l}\text{Mean no. of} \\ \text{days when flow} \\ \text{class occurred}\end{array} \times \text{secs day}^{-1}}{10^6}$$

By summing the estimates for all flow classes, a total annual sediment yield of 103,470 kg (103.5 tonnes) was obtained. If averaged out over the catchment, this represents an annual loss of 32.8 tonnes km^{-2}. These estimates must, however, be treated with some caution owing to the crude method of calculation.

BEDLOAD DATA

After completion of the bedload trap in October 1979, there was no measureable sediment accumulation until early December. Thereafter, the trap was emptied at intervals of between one and three weeks (when possible, after every major storm event) until April 1980 when sediment movement again became negligible. Sediment yield varied from a maximum load of 1,111 kg during the period 17 December 1979 to 3 January 1980 when a peak storm discharge of 342 l sec^{-1} was reached, to a minimum of 15 kg from 12 to 20 February 1980 when no significant storm events occurred (Fig. 21, Table 7). It is noticeable that high flow events occurring after long periods of low flow yield far more bedload than events of similar (or even greater) magnitude taking place later in the winter. This suggests that bedload yield is not purely a function of stream energy, but also depends upon the availability of sediment; after a succession of high flows, available sources of sediment are depleted such that subsequent storms yield less bedload than expected in relation to their energy. A peak flow of 220 l sec^{-1} in December 1979 yielded 612 kg of load, whereas a peak of 210 l sec^{-1} in late February 1980 yielded only 220 kg.

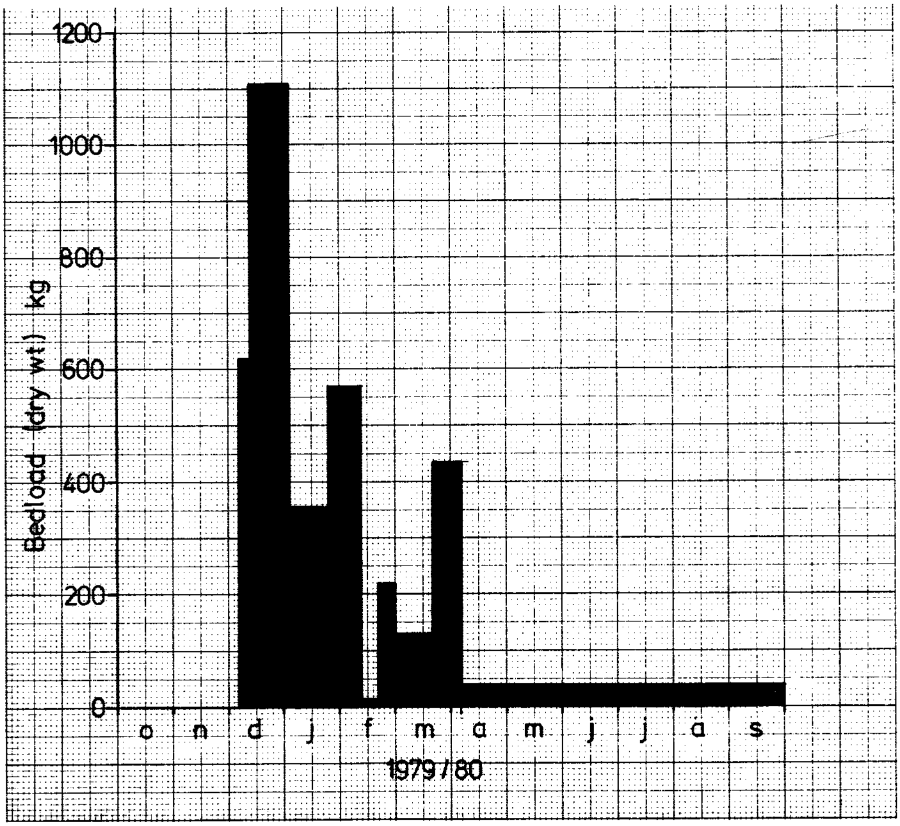


FIG. 21.
Bedload yield, October 1979–September 1980

Table 7. *Bedload and peak discharge data*

Date From	Date To	Total Bedload Dry Weight kg	Peak Discharge l sec^{-1}
7/12/79	17/12/79	612	220
17/12/79	3/ 1/80	1111	342
3/ 1/80	25/ 1/80	356	174
25/ 1/80	12/ 2/80	572	317
12/ 2/80	20/ 2/80	15	42
20/ 2/80	29/ 2/80	220	210
29/ 2/80	21/ 3/80	128	185
21/ 3/80	6/ 4/80	437	516
6/ 4/80	18/11/80	42.5	103
18/11/80	15/ 1/81	123	236
15/ 1/81	23/ 2/81	169	201
23/ 2/81	31/ 3/81	715	350
31/ 3/81	7/ 9/81	359	195
7/ 9/81	21/ 9/81	11	42
21/ 9/81	1/10/81	14	41
1/10/81	9/11/81	63	100
9/11/81	7/ 1/82	1438	450
7/ 1/82	25/ 1/82	340	302

From April to November 1980, 42.5 kg of sediment accumulated, but much of this was very fine suspended sediment and organic debris which had settled out in the low energy conditions of the trap. Table 7 includes bedload data collected up to January 1982 with corresponding peak flow rates for each collection period. While total flow shows no significant correlation with bedload yield, peak flow data give a significant correlation at the 99% level ($r = +0.754$). In Fig. 22, the bedload/peak flow (Qpk) regression line intersects the peak flow axis at 53.4 l sec^{-1}, indicating that bedload movement becomes negligible at lower flows.

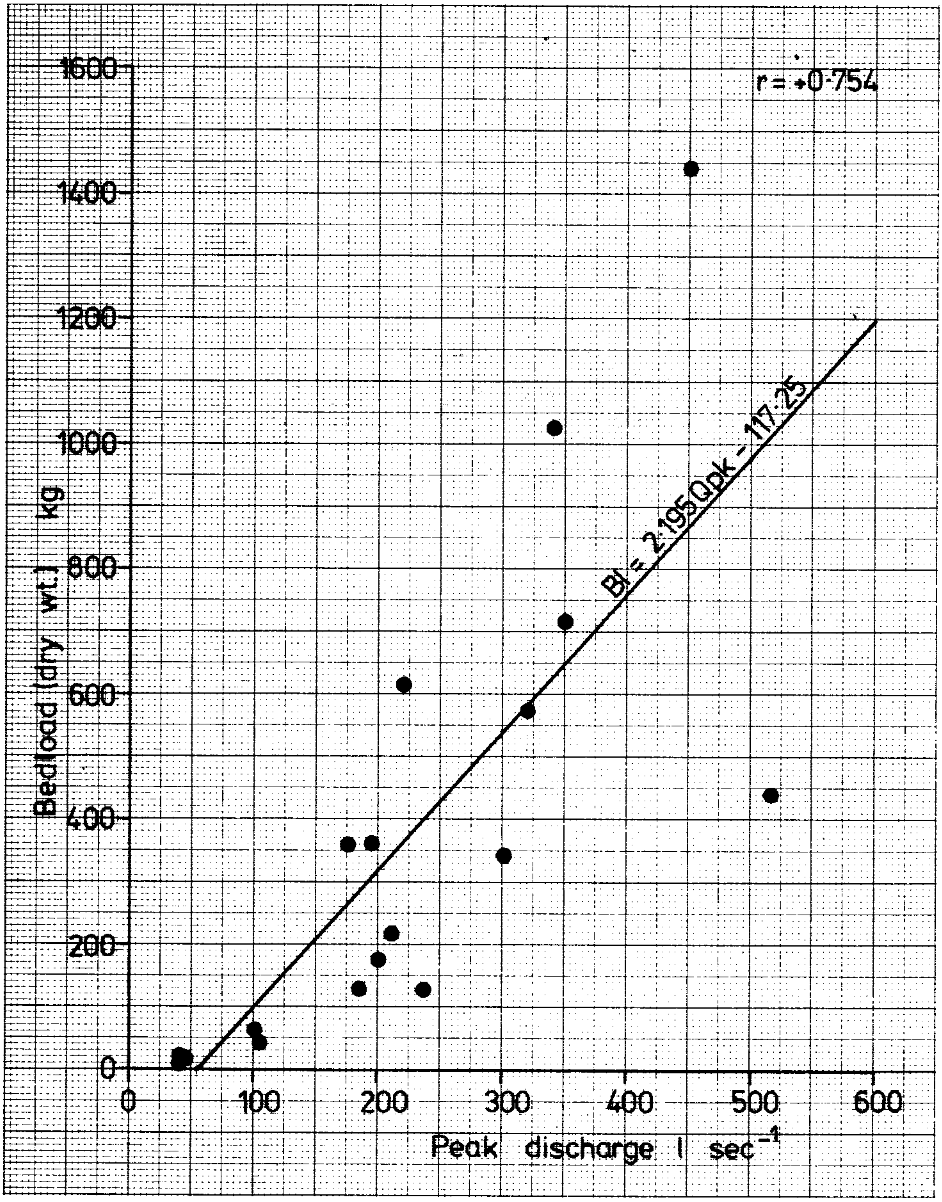


FIG. 22.
Bedload and peak discharge relationship

For the two years during which bedload has been measured, annual total yields of 3493.5 kg (1979–80) and 1391 kg (1980–81) were obtained. When compared with an estimated annual suspended load in the order of 100,000 kg, bedload losses are a very minor component of the total sediment output from the catchment.

The particle size distribution of a typical bedload sample is illustrated in Fig. 23. The small particle size of much of the material is at once evident with a mean size of 1.28 phi and some 75% by weight of the material in the coarse/medium sand range. A standard deviation of 0.85 denotes a moderate degree or sorting. The largest material trapped was in the –5 phi class, but this tended to consist of low density fragments of cinder from an old rubbish tip upstream. Although the bed of the stream is made up largely of coarse gravel and cobbles derived from the glacial drift, this appears to show little movement under the flows which occurred during the period of study. Eroding stream banks and the coarser fraction of soil washed into the channel by surface runoff appeared to be the more likely sources for much of the bedload. Organic matter accounts for about 2% of the total bedload weight although this is unevenly distributed among the size classes. Up to 20% of the material of gravel size consisted of small twigs, while the material in the finest size

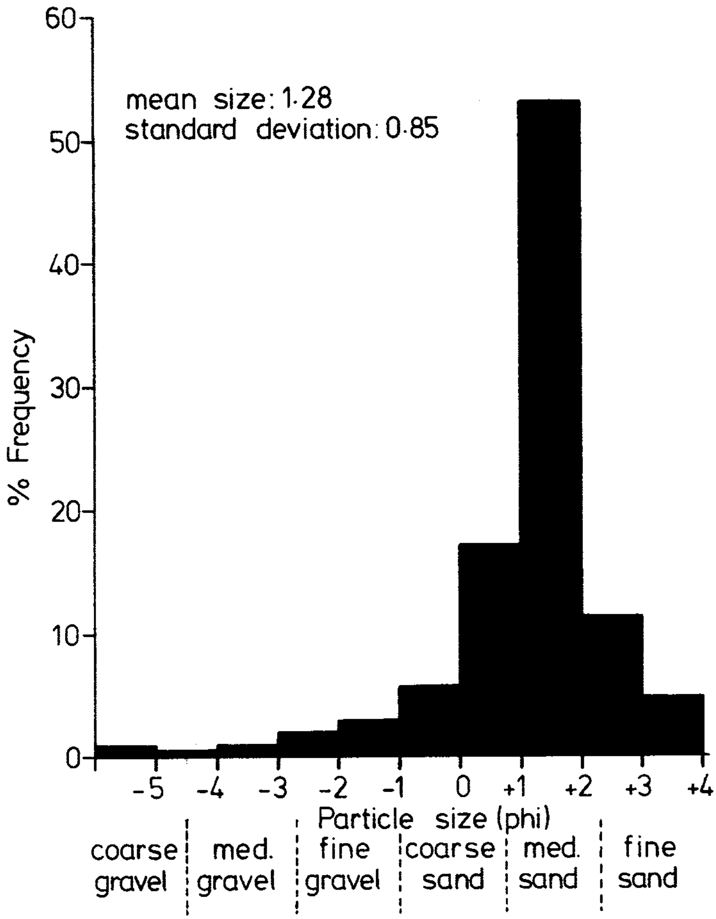


Fig. 23.
Bedload particle size distribution

class was 6% organic. The modal size class (+1.0 – +2.0 phi) contained less than 1% organic matter.

CONCLUSION

Of an annual rainfall input of 680 mm (mean 1976/7–1979/80), some thirty-five percent is lost from the catchment as runoff. The computation of precise annual water balances has been confounded by the problem of defining the catchment area in a lowland environment of complex drift deposits. Of the mean annual runoff of 235 mm (1977/8–1979/80), 88% occurs between December and the following April, partly in response to slightly greater winter rainfall, but more particularly to the effects of the highly seasonal variation in evapotranspiration. High rates between May and September result in less of the monthly rainfall totals reaching the stream, while the depletion of stored soil moisture and ground water reduce the contributions to the stream from throughflow and baseflow.

The range of stream discharges varies from 1.2 l sec^{-1} to 500+ l sec^{-1}, but, owing to the predominance of delayed flow rather than quickflow, for 90% of the time flows are less than 50 l sec^{-1}. The magnitude of storm runoff events was found to show no simple correlation with rainfall amounts or intensities. In explaining variations in storm runoff, antecedent conditions were found to provide the strongest correlations, with runoff positively correlated with antecedent rainfall, negatively correlated with antecedent evapotranspiration and strongly positively correlated with 30 day AMI. In attempting to derive predictive models of storm runoff, antecedent discharge (itself a function of pre-existing storage levels within the catchment) was found to provide reasonable predictions, though these were improved when storm rainfall was also included in the model, rainfall helping to explain the variation in runoff not accounted for by antecedent discharge alone.

The shapes of storm hydrographs were found to vary both seasonally and in response to rainfall intensity. High intensity events produced one or more initial storm peaks with lag times of about one hour after the main rainfall input, followed by a second peak with a lag time of at least five hours, probably related to throughflow processes assisted by the presence of tile drains. When catchment storage is high, the initial storm peak is the dominant feature of the hydrograph under high intensity rainfall, reflecting not only the rapid runoff from direct channel precipitation, roads and possibly a pond, but also variable amounts of surface runoff from low lying areas of saturated soil. Lower intensity rainfall events show only a single delayed peak suggesting that for most storms surface runoff is not of great significance.

Suspended sediment concentrations show no simple relationship with discharge. Relatively low runoff summer storms may generate high suspended sediment concentrations when intense rainfall coincides with a dry soil surface. The highest loadings, however, are associated with the very infrequent events where large amounts of overland flow from arable fields occur and on one occasion some 14 tonnes of suspended sediment were removed in 2½ hours. During storms, suspended sediment loadings are invariably highest in the early phases of the event when surface runoff is occurring, and by the time peak discharge is reached concentrations have already declined quite sharply. Although the total annual suspended sediment output (103.5 tonnes, 32.8 tonnes km^{-2}) is very much of an estimate, it is in keeping with measurements from other catchments in cool temperate environments (Gregory and Walling, 1973).

While much of the suspended sediment removal is associated with the intermediate flows occurring before peak discharge is reached, the amount of bedload sediment removal correlates more closely with peak flow events. This is to be expected in view of the higher energy necessary to move the larger sediment particles, but initial findings suggest that bedload yield is not simply a function of energy levels in the stream. The supply of material is an equally important factor and the availability of sediment seems to be higher during the early winter months, giving peak yields in December and January. In February and March equally high, or even higher, flows yield much less sediment. The total annual yield of 3.3 tonnes (1979–1980) is only 3.2% of the estimated suspended sediment loss.

Although no complete analysis of dissolved load concentrations and losses is attempted here, high conductivity levels indicate high solute concentrations. Baseflow conductivities in the order of 700 micromhos indicate a total dissolved solids concentration within the range 385–525 mg l^{-1} (Gregory and Walling, 1973, pp. 170–172) which would suggest a total solute loss two to three times as great as the total suspended sediment and bedload yield. High solute loadings are to be expected in a catchment with large areas of arable land receiving annual treatments of inorganic fertiliser.

The model of the catchment as a "cascading system" (Fig. 1) whereby water is passed from one store to the next and overflow from any store will generate runoff is a useful one which fits many of the observed features of runoff and its generation through time and space. The strong seasonal trends in runoff, for example, are well explained by the depleting effects of evapotranspiration on the storage "boxes", thus reducing the likelihood of overflow into the runoff box. Two further points of detail might be added to this model of the catchment and its dynamics. Firstly, let us consider the extent to which the catchment functions as an equilibrium system. Over the complete water year a state of balance is theoretically achieved with input equal to output and storage equal at start and finish (in practice however the measurements are fraught with difficulties). Over shorter periods of time, a temporary imbalance is evident giving rise to summer deficits and winter surpluses which result in changes in storage. Variations in storage may be considered to be held within upper and lower limits by the operation of what may be interpreted as negative feedback mechanisms. When water is lost from storage during summer, processes are set into operation which will tend to reduce further water loss. Remaining soil moisture is held at increasingly high tensions which reduce the rate at which it can be lost to the stream by throughflow and ultimately the ability of transpiring plants to take it up. Similarly, a reduced water table will slow up rates of ground water loss to the stream—if water tables fall far below the bed of the stream, baseflow input will cease and ground water may even gain at the expense of the stream. At times of high storage, the catchment reacts in such ways that as much as possible of any further water input is discharged as runoff. High soil water values will maintain the supremacy of gravitational over capillary forces in the soil, thus ensuring a high level of throughflow. On saturated soils the high amounts of runoff will be ensured by the operation of overland flow. The area of the catchment with saturated soils will expand as storage increases, giving an increasingly large area of land with both surface runoff and high throughflow output ("the partial contributing area", Weyman, 1975). At the same time, high ground water levels will maintain powerful baseflow contributions to the stream.

The sediment system may also be treated as an equilibrium one characterised by negative feedback mechanisms. With regard to bedload, high rates of movement take place only when material of a calibre corresponding with the competence of the stream is available. Late winter storms have been shown to remove relatively low amounts of sediment compared to storms earlier in the winter. Sediment output is thus being reduced at times of greatest energy availability by lack of sediment availability. Furthermore, when field surfaces are at their least consolidated (during conditions of low soil moisture in summer), runoff generally, and surface runoff in particular, is restricted, thereby reducing the amount of energy available for sediment removal. Thus, in winter when energy levels are at their highest, sediment losses, though considerable, are ultimately controlled by lack of sediment availability, while in summer when the supply is plentiful, energy is at a premium due to low runoff conditions. Similar feedback effects may be seen to operate within a storm when a combination of high energy availability during the rising limb of the hydrograph, together with an availability of load, produces high suspended sediment concentrations. At peak flow, when energy is at its highest, availability has dropped leading to lower concentrations.

A second feature of the dynamics of the catchment system is the significance of thresholds: the effect of one variable on another may not be felt until a critical point is reached. Bedload movement provides a simple example in that sustained periods of moderate flow will produce a high total flow for that period, but the critical velocities necessary for removing the particle sizes of the available bedload may not be reached. Little or no bedload movement appears to occur until flows of at least 50 l sec^{-1} are attained, and these levels usually only occur during storm events. In the case of throughflow contributions to the stream, gravitational water will only be available when soil moisture levels rise above field capacity. Provided a soil moisture deficit is present, retaining capillary forces will remain stronger than gravitational forces. Similarly, where perched water tables occur in the drift deposits of the catchment, a rise in ground water level will not be reflected in increased baseflow contribution until it reaches or approaches the surface in the valley bottoms or hollows, thus demonstrating a further threshold factor. If threshold levels are involved in relationships between catchment variables, curvilinear rather than linear models may be more appropriate in analysing those relationships.

Although some progress has been made towards quantifying processes in the catchment, in several respects it remains very much a "black box". The problems of measuring directly the relative contributions of surface runoff (from different types of surface), throughflow and ground water remain to be solved. While high solute loadings are indicated, more intensive sampling is required to derive a total annual dissolved load output for the catchment, to investigate seasonal variations and to determine concentrations in different types of flow, as well as identifying the sources of solutes particularly in relation to fertiliser application. Only when these questions can be answered can truly predictive models be constructed and the effects of human activity on runoff, sediment and solute yields be objectively assessed.

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