ABSTRACT

Elizabeth Ann Sutton
The Hydrologic Response of the Eaton River Basin, Quebec
Department of Geography
Master of Science

The hydrologic response, defined as the amount of stormflow from a basin divided by the amount of precipitation producing this flow, was examined for the Eaton River in south-eastern Quebec, for 38 major individual storms as well as for yearly snow-free season totals. These response values were related statistically to independent hydrologic and meteorologic variables representing storm size and intensity, antecedent moisture conditions, and temperature and evapotranspiration conditions. Antecedent moisture conditions were found to be the most important determinants of response. The results were interpreted in terms of various models of runoff production. Little support was found for Horton's classic model of overland flow; the variable source area model offers a more probable mechanism for runoff production on the Eaton basin.

THE HYDROLOGIC RESPONSE OF THE EATON RIVER BASIN QUEBEC

THE HYDROLOGIC RESPONSE OF THE EATON RIVER BASIN, QUEBEC

by

Elizabeth Ann Sutton

A dissertation submitted to the Faculty of Graduate Studies, McGill University, in partial fulfillment of the requirements for the degree of Master of Science.

ACKNOWLEDGEMENTS

This research was carried out under the supervision of Dr. M. A. Carson and was partially supported by a grant (290-60) from Ministère des Richesses Naturelles du Québec, Québec. Most of the diagrams were drawn by Mr. Abel Sen, Cartographer, Department of Geography, McGill University as part of a forthcoming paper.

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

SYMBOLS USED IN THE TEXT

a	area of the perennial channel network
A	area of the catchment
AP	antecedent precipitation index
D_d	perennial drainage density
e	Naperian base
f_0	infiltration capacity at the beginning of a storm
f_c	ultimate infiltration capacity attained at the end of long storms when the soil profile is completely saturated
f_p	infiltration capacity at an elapsed time t from the start of a storm
i	hydraulic gradient
k	empirical constant in formula for infiltration capacity, that depends on the soil type
k	empirical constant in formula for antecedent precipitation index, varying from 0.82-0.99
k	in the Darcy equation, the hydraulic conductivity of the soil
P	storm precipitation in inches
P_{ea}, P_b	rainfall in an individual storm at East Angus and as an average for the basin, respectively
P_{ea}, P_l	total precipitation in snow-free season at East Angus and at Lennoxville

P_n preciptation on the day n days prior to the beginning of a storm

q_s discharge of through-bank seepage per unit length of channel

r correlation coefficient

$r_{h(b)}, r_{h(h)}$ hydrologic response for a single flood event according to the Barnes and the Hewlett methods of hydrograph separation, respectively

$R_{h(b)}, R_{h(h)}$ average hydrologic response for the snow-free season, according to the Barnes and the Hewlett hydrograph separation techniques, respectively

S amount of storm flow in inches

s_b, s_h storm runoff in an individual flood event according to the Barnes and the Hewlett hydrograph separation techniques, respectively

S_b, S_h total storm runoff in snow-free season according to the Barnes and the Hewlett methods of hydrograph separation, respectively

t unit of time

v velocity

CHAPTER I

INTRODUCTION

There are few streams which flow with a steady discharge. Most channels carry increased quantities of water during or immediately after a rainstorm; the magnitude of this extra flow, even allowing for the size of the rainstorm, varies markedly from storm to storm. Notwithstanding the progress in hydrology in recent decades, very little is known about the exact mechanism by which storm rainfall over a drainage basin actually produces a peak in the stream hydrograph.

The problem can be approached in one of two general ways, termed by Amorocho and Hart (1964) as the parametric and the physical approaches. In the physical approach, attention is focussed directly on the processes contributing to storm runoff; attempts are made to monitor the path of every unit of water that is shed from the drainage basin as it moves toward the stream network. The attainment of a workable physical model is the ultimate goal in hydrologic, as in other scientific, research; once a workable model has been constructed, it should be reasonably simple to predict a basin's response to future storms. The physical approach to the problem of the hydrologic behaviour of even one drainage basin, is , however, immensely costly in terms of resources. It is beyond the scope of a M.Sc. research topic; indeed, in the case of the Eaton basin, this forms the long-term goal of the Ministère des Richesses Naturelles du Québec.

An alternative to the physical approach is the parametric method. In this type of study, little attention is paid directly to measurement of the movement of water within drainage basins. The aim is, rather, to develop a predictive equation for basin behaviour in terms of a number of independent variables. The choice of these variables is, of course, to a large extent based on an intuitive assessment of the hydrologic processes at work, but direct measurement of these processes is not necessary. It might seem that this type of approach has little value outside mere prediction of hydrologic events, and that it has little to offer in attempting to explain <u>how</u> water gets into the stream network during floods. This is an incorrect impression. The relative importance of different variables in the predicting equation must be some indication of the relative importance of different processes in the production of storm runoff. Ultimately, as Carson (1969) pointed out, the ideas suggested by a parametric study should be tested via a physical approach; nevertheless the parametric method offers a very rapid means of assessing the relative importance of various different processes in the early stages of research in a discipline.

Many hydrologists have examined runoff-rainfall relationships using a parametric approach. A recent article by Lee and Bray

(1970), for instance, presents equations that may be used for the prediction of storm runoff for five basins in New Brunswick. Others have preferred to focus attention on the ratio of storm runoff to storm rainfall, sometimes referred to as hydrologic response, and attempted to predict this. Hewlett (1967) mapped the annual hydrologic response throughout Georgia; more recently, Woodruff and Hewlett (1970) have analysed variations in response among different river basins over a much larger part of the eastern United States. Few studies, however, have concentrated on a single catchment and examined the variation in the response of different storms. The results of such a study are reported here.

Over the years, several different theories of storm runoff production have been developed. With different bases in physical reasoning, each model depends upon slightly different basin conditions to produce the requisite hydrograph peak. The classic Horton model of overland flow (Horton, 1945), for instance, depends closely on the intensity of storm rainfall in relation to the infiltration capacity of the basin soil mantle. The more recently developed variable source area model (Hewlett and Hibbert, 1967), on the other hand, is more dependent on ground water table levels in the vicinity of the channel network prior to the storm. Since these various ideas clearly affected the choice of variables that were employed in this study, some review of the existing literature on the theory of runoff production is necessary.

CHAPTER II

MODELS OF RUNOFF PRODUCTION

The infiltration theory of overland flow (Horton, 1945) is the most well-known of all models of runoff production. It still pervades most of the standard hydrologic textbooks, although, for some time, it has been challenged by many hydrologists. Simply stated, it says that runoff is produced when the infiltration capacity of the soil is exceeded by the precipitation intensity. Small dams are initially formed by the surface litter, and when their capacity is exceeded the stored water runs together, either as an intricate rill network or as a sheet of water, and moves across the surface of the basin (Figure 2.1), eventually reaching the channel.

There are a number of conditions which influence the occurrence or absence of overland flow. Most important, according to Horton, is the presence of a vegetal cover. A litter layer on the soil surface will break the impact of the raindrops and the drops from the surrounding vegetation; this leaves the soil surface itself undisturbed and encourages entry of the water into the soil. Without this cover the surface of the soil may become compacted, and pores closed, through the impact of the raindrops, with the result that the water is prevented from entering and so is forced to flow

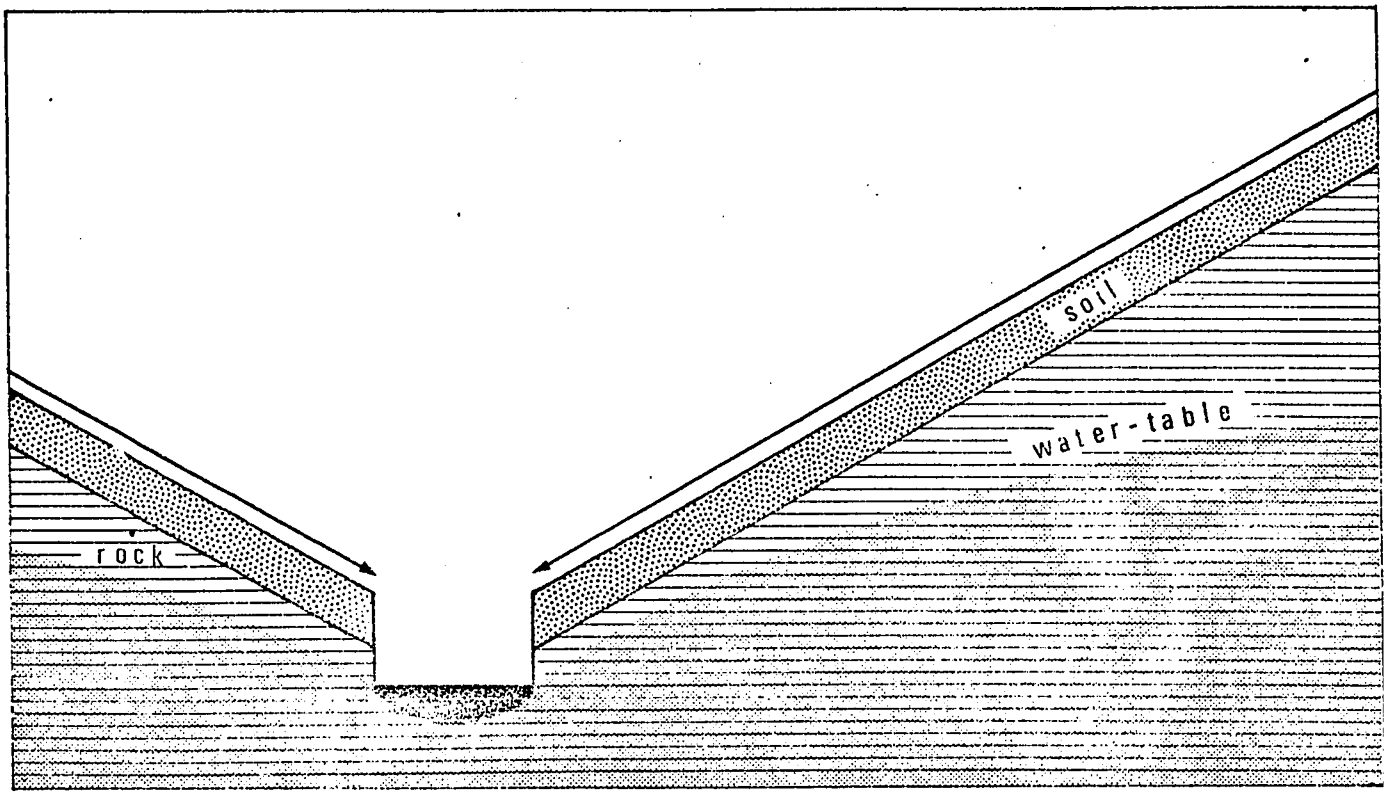


FIGURE 2.1

Horton Overland Flow

over the ground.

Vegetation is also important through its influence on the soil structure. The presence of roots, and of canals left by decayed roots, increases the permeability of the soil and gives it a less dense structure. The effect is a net increase in the rate that the water can infiltrate and then percolate through the soil mantle. The presence of a vegetation cover also may result in deeper, better formed soils, that are in turn capable of holding more water before capacity is exceeded.

The nature of the soil itself may have a great influence on whether overland flow will occur. A sand dune, for example, may have no stabilizing vegetal cover at all, and yet will exhibit virtually no overland flow. because of the extremely high infiltration capacity. In general, it is believed that the infiltration capacity of soils decreases as the content of clay-size particles increases. The infiltration capacity of a soil body is not, however, a time-invariant characteristic. It is affected considerably by the amount of moisture in the soil. Wet soils, for instance, will exert less capillary suction on the incoming moisture than drier soils; in addition, as demonstrated so clearly by Schumm and Lusby (1963), the development of dessication cracks in clay soils may drastically increase infiltration capacities during periods of drought.

Horton theorized that at the beginning of a storm there would generally be little or no overland flow. With the progress of the storm, however, the infiltration capacity of the soil decreases.

Horton found that, assuming a large enough supply rate, the change in infiltration capacity with time during a storm could be expressed in the following form:

$$f_p = f_c + (f_0 - f_c)\, e^{-kt} \qquad (2.1)$$

where f_p is the infiltration capacity at an elapsed time t from the start of the storm;

f_c is the ultimate infiltration capacity attained at the end of long storms when the soil profile is completely saturated;

f_0 is the infiltration capacity at the start of the storm;

e is the Napierian base; and

k is an empirical constant that depends on soil type.

Eventually the infiltration capacity may become lower than the rainfall intensity. At this point although some moisture continues to infiltrate the soil, water now begins to accumulate on the ground surface in small depressions. Once the requirements of surface detention are fulfilled, overland flow begins. Implicit in the Horton model is that <u>all</u> parts of the drainage basin supply runoff once overland flow begins. If one accepts the initial assumptions of the Horton model, this is a reasonable corollary; rainfall intensity and soil infiltration capacity are likely to be relatively uniform throughout the basin, at least for small basins, so that as soon as the infiltration capacity of the soil has been reduced below the rainfall intensity at one point in the basin, this condition should quickly develop at other points too.

Any point in the catchment not only produces overland flow but also receives flow from all points above it, and, as a result, there should be an increase in the rate of discharge proportional to the total length of the path of flow. The depth of flow might therefore be expected to increase, although this increase may be small because the velocity of flow may increase downslope; also, as pointed out by Hack and Goodlett (1960), concave curvature of contours should cause an increase in the depth of flow due to convergence of water.

Many ideas contained in the Horton model have recently come under challenge. Among these are the following:

a. equal contribution to stormflow from all portions of the basin;

b. a sharp delineation between water which infiltrates the soil to become groundwater and eventually baseflow, and water which fails to infiltrate and so proceeds over the surface of the basin to become storm runoff;

c. the role of ground water as the sole contributor to and determinant of baseflow;

d. rainfall intensity exceeding infiltration capacity in a significant number of storms;

e. the actual existence of the phenomenon of overland flow in well-vegetated areas.

With the expansion of research in humid forested areas, it has become increasingly evident that the Horton model simply does not apply. Very special circumstances must exist to create a system of overland flow. These will be found mainly in semi-arid areas, where the infiltration capacity of the soil is likely to be lower, and rainfall intensities are often higher: but under a forest cover the concept is no longer accepted, and several inter-related theories have been developed to take its place. Basic to all of them is the concept of sub-surface seepage, stressed long ago by Hursh (1936).

Expressed in its simplest form, this is the concept that rain water almost universally infiltrates the soil, and eventually appears in the flood water through flow beneath the ground surface, but above another, less permeable layer of soil. Basically, it is a submerged version of the overland flow model. The mechanism is as follows: in soils under humid forested conditions, the A, and sometimes the upper part of the B, horizon is generally extremely permeable, both naturally through the structure and texture of the soil, and artificially, through the work of roots, channels left by decayed roots, animal burrows, and other forms of biotic activity. In contrast, lower layers are likely to be less permeable due to higher clay content (through illuviation), increased compaction, and less mechanical reworking. The top soil layer is covered by a thin litter layer that serves in a dual capacity of cushioning the underlying soil from the impact of raindrops and permitting rapid infiltration of the water. Above this is the canopy, which also intercepts the

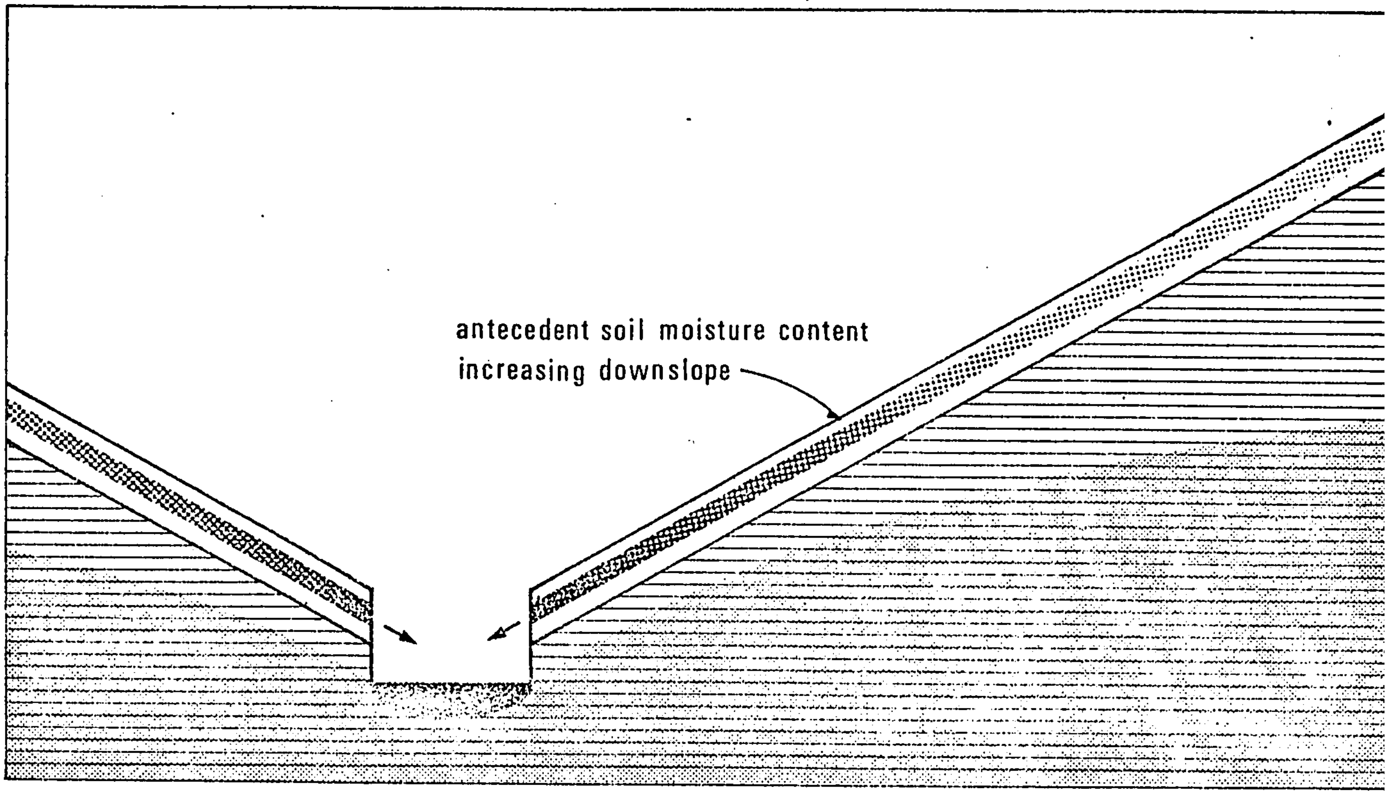


FIGURE 2.2

Throughflow

precipitation and slows it in its path to the surface. The moisture infiltrates the soil, and moves vertically downward until it encounters a soil layer that is less permeable than the top one. Some moisture continues into the lower layers, but as this occurs at a much slower rate, there is a tendency for a saturated condition to develop above the interface, and surplus moisture may begin movement laterally downslope. This process has been variously termed interflow, subsurface stormflow (Hursh, 1936), and, more recently, throughflow (Kirkby and Chorley, 1967).

The throughflow model does not totally preclude the existence of overland flow in a catchment. The shingle-like nature of hardwood litter contributes to surface detention, and to small amounts of overland flow. The thickness of the soil mantle through which the moisture is moving will vary throughout the surface of the slope, and moisture entering a shallow zone from a deeper one may be forced to the surface. The porosity and water-holding capacity of the soil may also vary from place to place irrespective of depth. Curvature of contours across the slope may cause convergence of the moisture, resulting in too great a concentration for the capacity of the soil. In this model there is, therefore, no sharp delineation between surface and sub-surface moisture, and a unit of water may very well belong to both categories at different times in the course of its movement downslope. These ideas are thus rather similar to the findings of Amerman (1965), working in an agricultural watershed in Ohio, who found that overland flow was distributed in a seemingly random fashion on ridges, valley sides and valley floors.

Perhaps the most well-known results of experimentation into the presence and importance of throughflow are those of Whipkey (1965), working in east-central Ohio, with a plot supporting a 60-year-old mixed oak stand on a $15\frac{1}{2}°$ slope. The soil, a sandy loam, was found to be extremely permeable through the first two layers, to a depth of 90 cm. Below this depth, soil was more compact and contained more silt and clay particles. At the foot of the slope a pit was dug exposing the soil layers, and a trough system was devised to collect seepage from each layer. Artificial rain was applied and seepage was found to amount to 3-16 percent of total moisture applied over the initial 24-hour period; unfortunately Whipkey gave no indication of the fate of the remaining 84-97 percent of the precipitation. Sixty-four percent of all seepage came from the top two layers, with most of this total from the layer directly above the flow-impeding interface. This is because, although moisture enters the soil from the top down, saturated conditions build from the bottom (of the throughflow zone) up. Saturated flow in this case, therefore, began in the second layer; often the top layer never attained saturated conditions, and so flow in that horizon remained low. This need not always be the case; if initial moisture conditions are dry, the wetting front (Bodman and Colman, 1943) itself may act as a barrier to moisture advancement, and a saturated layer can develop above it, resulting in flow beginning initially from the top layer. This was reported by Hansen (1955), working with irrigated soils, and substantiated by Whipkey. Figure 2.3 shows outflow hydrographs from Whipkey's tests; it will be noted that in the case of the saturated layers of soil (layers 2 and 3) these closely resemble storm hydrographs of stream discharge. Hydraulic head readings in Whipkey's study indicated the presence of a saturated zone of soil that was thickest and hence nearer the surface at the downslope face. The depth of the saturated "mound" remained almost constant as flow peaked and receded, although the depth of the saturated zone upslope decreased with time.

Work by Betson, Marius and Joyce (1968) in western North Carolina shows similar results. Using piezometers they monitored soil moisture fluctuations in the A horizon of a clay loam soil in an agricultural area, where the watershed slope is about 22°. It was thought that, in this case, the interface between the A and B horizons provided the moisture boundary. They found, as did Whipkey, that fluctuations were very sensitive to both antecedent soil moisture and storm intensity; piezometric responses to moisture were greater under shallow soils, and response disappeared altogether when a shallow stretch occurred immediately upslope of a thicker one. This leads to the conclusion that the deeper soil layer absorbed the moisture, which would then flow under unsaturated rather than saturated conditions. Thus a variable concentration of water developed throughout the

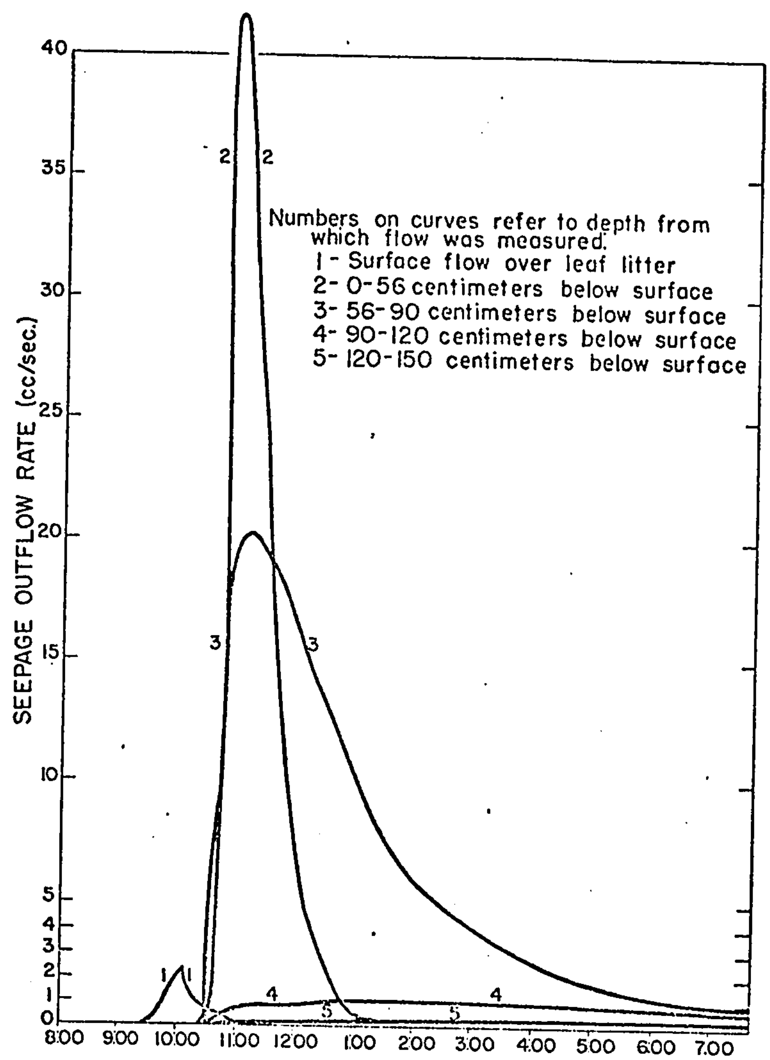


FIGURE 2.3

Outflow Hydrographs from Subsurface Seepage

(After Whipkey, 1965)

watershed, a fact that agrees well with the variable distribution of overland flow noted previously. They concluded that the size of the area that effectively contributes to stormflow varies with the amount of available storage in the throughflow horizon. Betson, Marius and Joyce's work indicated much more rapid response to precipitation in downslope areas than in upslope areas, and suggested that moisture concentration was probably greater there.

The concept of throughflow is basic to much of present-day thinking on the subject of runoff production. On its own, however, it leaves unanswered a number of fundamental questions, the most important being whether water can be supplied to the stream rapidly enough to contribute to stormflow. Kirkby (1969) cites throughflow rates of only 20-30 cm per hour, compared to average overland flow rates of 200-300 m per hour. A slope of 300 meters in length could easily drain during the time of a typical flood provided overland flow was the prevailing mechanism; on the other hand, 1000 hours would be required under the simple throughflow mechanism, and, under such circumstances, most floods would have subsided long before drainage of the hillsides had been completed. It should be remembered that throughflow rates vary greatly (many orders of magnitude) with soil type. Kirkby's values are based on Whipkey's (1965) data for the saturated hydraulic conductivity of a sandy loam. This figure (11.3 inches per hour) agrees closely with a figure of 13.5 inches per hour for similar material reported by Dunne and Black (1970a). Throughflow, as expressed in the model discussed above, does not appear to solve the problem of runoff production.

Work by Hewlett (1961; Hewlett and Hibbert, 1967) offered a solution to this contradiction with the variable source area model. The argument is that storm runoff does not originate in proportional amounts from all parts of the basin, but is almost entirely derived from the valley floors and the lower parts of the valley slopes. In the period between rainstorms, moisture slowly moves downward laterally through the soil mantle on the hillsides, feeding baseflow in the stream channels, depleting the upslope areas and producing a relative concentration of moisture at the base of the slope. Rates of flow from a trough of soil, constructed to resemble as nearly as possible, natural conditions, demonstrated outflow levels sufficiently large to maintain known baseflow levels in the region, whereas the existence of ground water had not been proved.

This concentration of moisture at the base of the slope creates a small zone of near-saturated conditions near the stream channel. More important in this context, however, is the fact that this water is immediately available to be dispatched into the stream in the event of further precipitation. In a crude analogy, one might view this water as trapped between an advancing wetting-front above and a more impermeable soil below, and, in effect, the water is squeezed out into the channel. In terms of the actual mechanics of the situation, it is probably more useful to view the process as a relaxation of soil moisture tension in the throughflow horizon. If the antecedent moisture content of the soil is quite high, as it is at the base of the slope, then a small input of additional moisture will produce an appreciable relaxation of the tension in soil pores. In the extreme case the moisture system may be completely converted to a freely-draining water source and saturated subsurface flow will take place.

As the idea has been described above, it is still difficult to appreciate why the volume of subsurface discharge should increase <u>sufficiently</u> to account for stormflow volumes. There will be some increase in the <u>velocity</u> of water seeping through channel banks, into the channels, but this increase is, nevertheless, rather small. There will also, presumably, be some increase in the <u>thickness</u> of the subsurface flow, but again it seems doubtful, as put forward above, that this is large enough to produce much stormflow.

Actually the above description is rather a narrow representation of the variable source area model, and indeed, although Hewlett and Hibbert (1967) tend to support this Whipkey-style subsurface seepage, it is not the major element in this model. Most important is the proximity of the true ground water table to the ground surface in the vicinity of the valley bottom. During rainstorms it may be expected, and indeed has been shown, that the water table may rise very rapidly in these near-channel areas. The mechanism of this sudden rise of the water table is explained as follows. Above the water table is a zone in which, although the soil moisture is in tension, the pores are very close to saturation, and, indeed, in the lower part are fully saturated. Only a small amount of additional water is necessary to drastically reduce the soil moisture tension and transform this"capillary fringe" into freely-draining water. As a result there is a sudden increase in the thickness of the ground-water zone entering the channel. The principle, in fact, is clearly identical to the mechanism of accelerated sub-surface seepage in the soil mantle, described above. In the case of this second situation, in the valley bottoms rather than on the main hillsides, the "jump" in the water table is liable to produce a much thicker zone of extra subsurface flow than on the hillsides. Actually, it is probably very difficult to separate these two zones of water in the vicinity of the valley bottoms. According to Hewlett and Hibbert (1967) (Figure 2.4), on the main part of the hillside slope subsurface soil moisture (as defined previously) will be perched at some distance above the true ground water table; moving downslope toward the valley bottom, the two moisture zones will tend to merge, and, of course, the extra subsurface seepage in the soil mantle during the storm will aid downward percolating rainfall in the transformation of the capillary fringe above the true water table by the stream.

There is an additional contribution from this swelling of the ground water system near the channel network apart from the extra thickness of saturated flow. This is the inevitable tendency

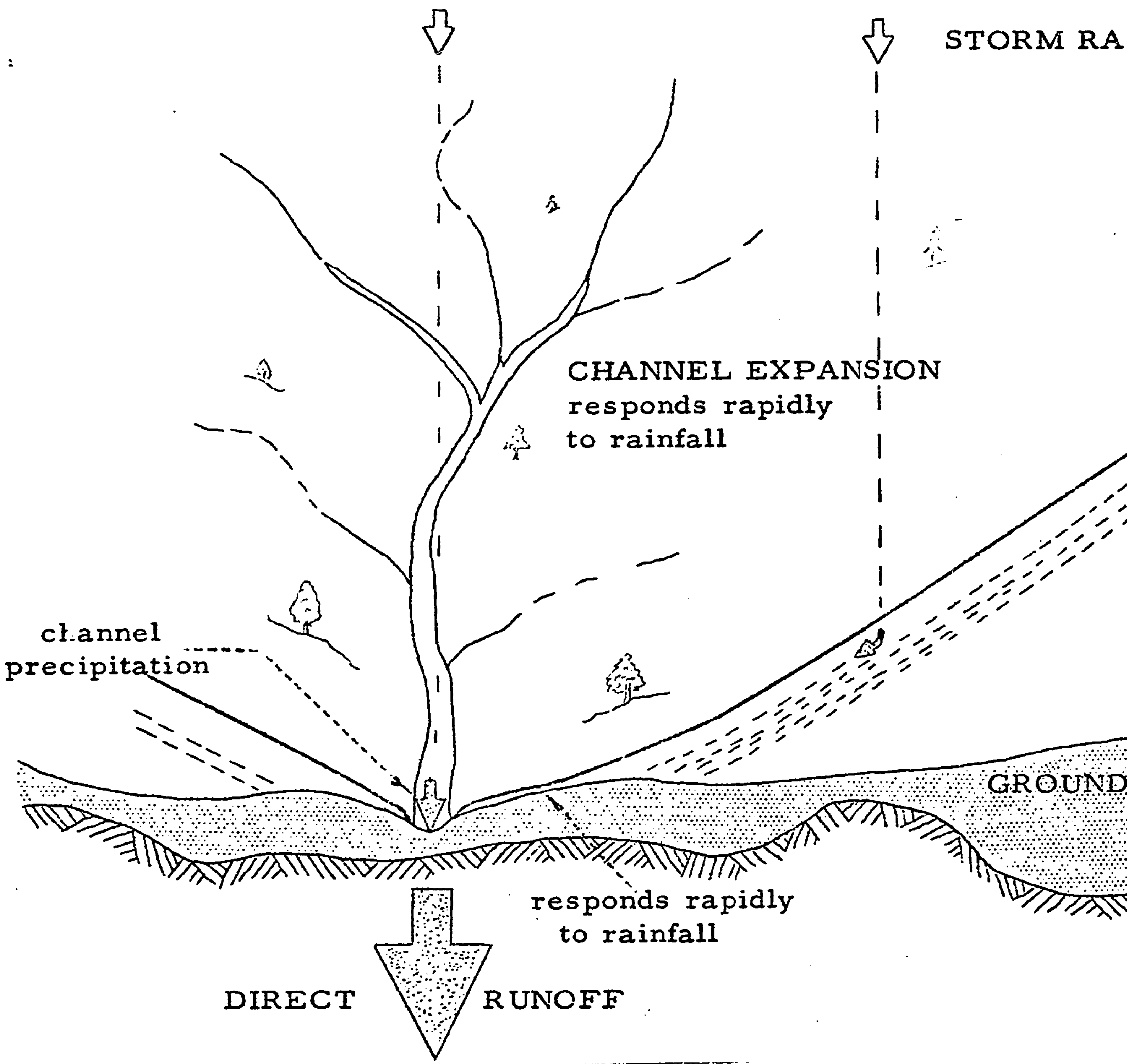


FIGURE 2.4

Variable Source Are

(After Hewlett and Hibb

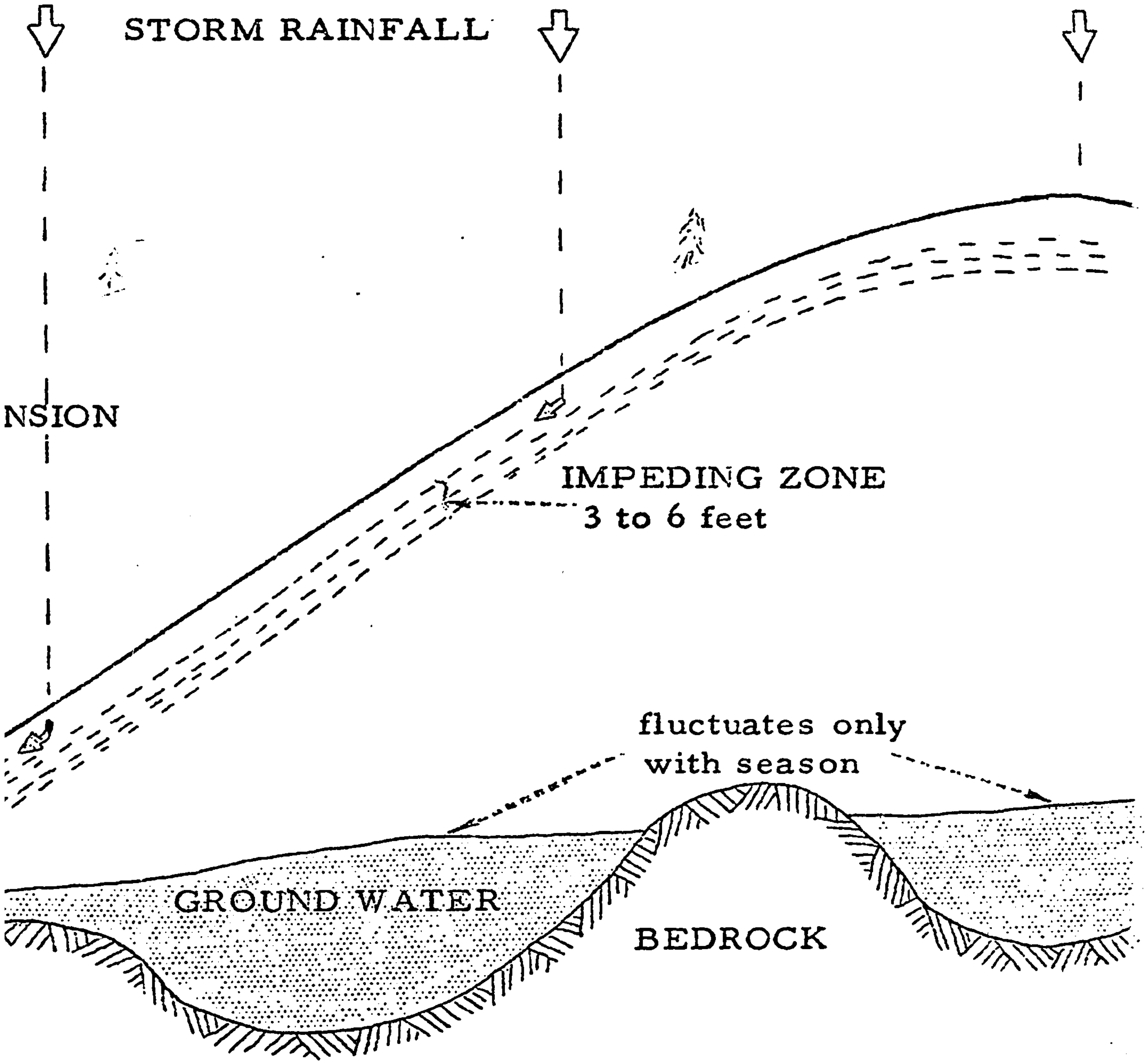


FIGURE 2.4

able Source Area Model

ewlett and Hibbert, 1967)

for the channel network, banded by an enlarged ground water system, to expand headwards within the basin. A threefold expansion of the drainage density of a tiny (60 acre) southern Piedmont watershed has, for instance, been reported by Tischendorf (1969) during a 4.14 inch summer storm. A map showing the subsequent shrinkage of the drainage network to the perennial system, after the storm, is provided in Figure 2.5. The effect of this expansion in the drainage network is twofold: firstly it results in a greater area of direct channel interception of rainfall; and, secondly, it extends the amount of "channel bank" through which subsurface seepage may take place. In the southern Appalachians it appears that the latter is the most important effect; in the case of the storm referred to above, even at the peak of the storm only 0.75 percent of the catchment area was occupied by the channel network and on its own, therefore, direct channel interception of rainfall could only account for a hydrologic response of about 0.0075. With values of response up to 0.30 for major storms, the implication is that the bulk of the storm runoff was through bank subsurface seepage.

The concept of the variable source area contribution to stormflow is perhaps best summarized by this statement of conditions in The Southeastern Piedmont by Hewlett (1969):

> "... an impeding soil zone from three to six feet in depth permits a relatively constant rate of water flow into deeper soil but turns most of the water from a large rain laterally and downward to the toe of the slope. The impeding zone is the lower part of the B horizon which resists water movement because it is enriched in colloidal particles eluviated over thousands

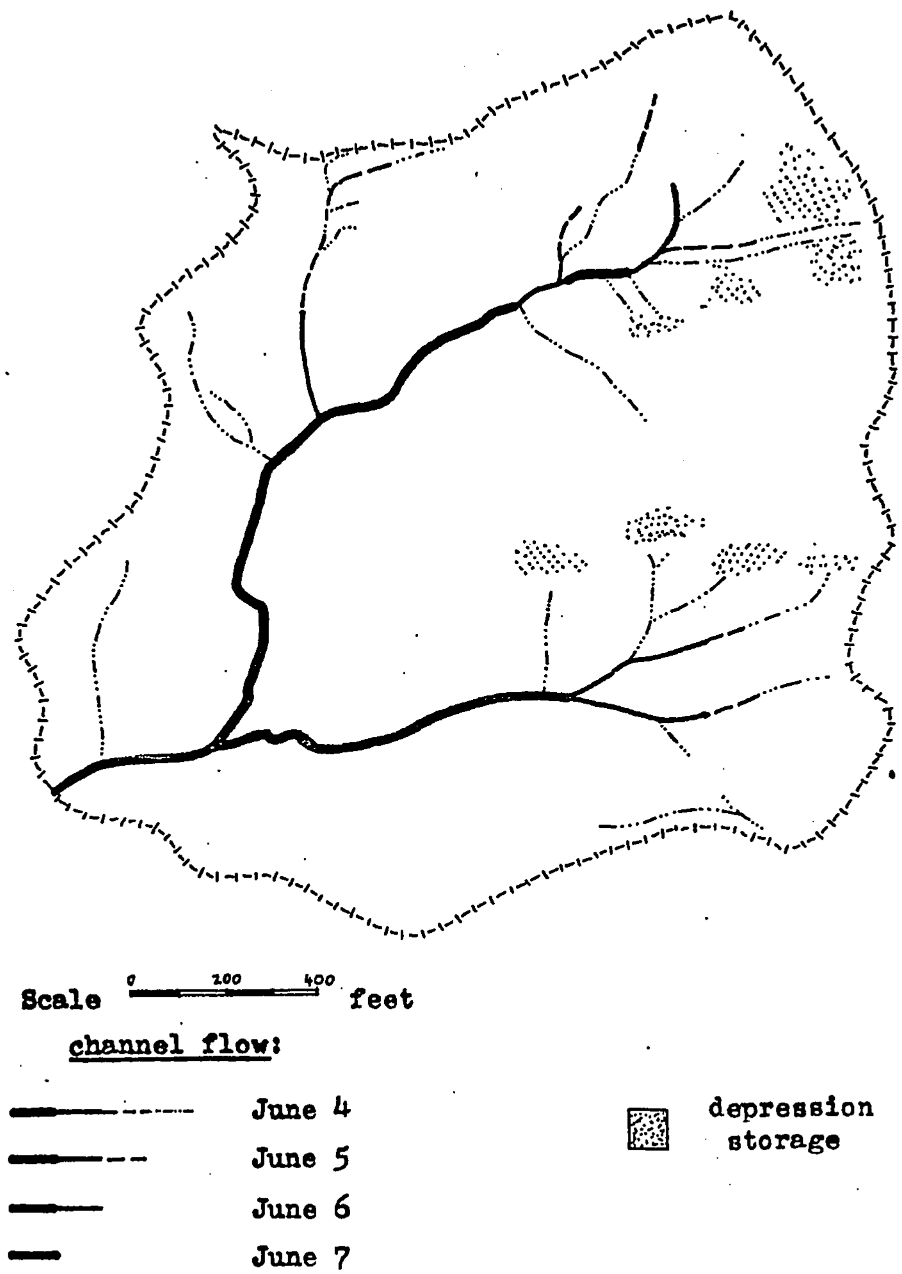


FIGURE 2.5

Headward Expansion of the Drainage Net

(After Tischendorf, 1969)

> of years from the upper soil layers. Thus two fairly distinct systems supply streamflow; part of the baseflow comes at an almost constant rate from the deep ground water aquifer, while the remainder of the baseflow and all the stormflow comes from the upper zone of the soil mantle and the open channel system. The two source systems join near the expanding and shrinking channel system, where an ephemeral rise in the ground water table helps produce the storm hydrograph and also sustains many days of baseflow. During and following exceptional rain storms, this ephemeral ground water body extends rapidly upstream along intermittent channels, forming a diffused, perched ground water system that reaches maximum extent after rainfall ceases, and then retreats to the perennial channel within a few days."

The implications of this idea are many. First, it is important to note that much of the water actually contributing to storm runoff is not derived from the rainstorm, but rather is water that has existed in the soil for weeks or months, and presumably has had time to come into chemical equilibrium with the soil in which it is contained. It might be possible, therefore, to check the validity of the model in other areas through an examination of the change in the dissolved solids concentration of the water during the course of a flood. Often there is an appreciable dilution of the dissolved solids in storm runoff and, unless this is the result of direct channel interception of rainfall, it is difficult to explain it in terms of the variable source area model.

Secondly, the water for storm runoff is not contributed equally from all parts of the basin. Upper reaches will be depleted of their moisture between storms, not only from downslope seepage but also from transpiration and direct evaporation, and seepage into lower soil depths to contribute to water table levels. Before saturated flow will occur, these losses must be replaced. These ideas are thus a distinct departure from those of Horton (1945), who, as stated previously, believed that stormflow could be supplied from the entire catchment.

Thirdly, the moisture does not have to travel long distances to reach the stream channel. The problem of timing is no longer a crucial factor. With a high moisture concentration in the lower few percent of the slope, drainage from the slopes can easily be completed in the length of time that storm basin drainage is in fact completed on the basis of hydrograph records. This removes one of the main obstacles in invoking subsurface seepage in the production of storm runoff. There still remains, however, the problem of whether, even with an expanded stream network, subsurface seepage can provide <u>a sufficiently large volume</u> of water to explain storm runoff, a slightly different question to whether or not it can supply moisture quickly enough. One solution to this particular problem has been presented by Dunne and Black (1970a), and will be discussed shortly.

Concurrent with the development of the variable source area model, other workers(Betson, 1964; Ragan, 1968; Dunne and Black, 1970a and b)have also emphasized that the bulk of storm runoff comes from the valley bottoms, and they have used the term "partial area" model. Strictly speaking, the ideas of Hewlett and Hibbert (1967) should also be considered as a partial area model; the importance attached by these two workers to the variable size of the drainage net is, however, justification for retaining the term "variable source area model" as a description of their specific ideas, while using the term "partial area" for the more general concept.

The results of a study on a 114 acre watershed in Vermont, presented by Ragan (1968), reinforce, to some extent, the ideas put forward by Hewlett and Hibbert (1967). Ragan examined the valley of a 619 foot long second order stream, with very careful instrumentation. The site was chosen because of the uniform channel geometry of the stream, the regular geomorphology of the watershed and the uniform soil conditions. The surface material was a uniform sand, about 80 feet thick, deposited during the retreat of a post glacial lake over a horizontal silt layer. An extensive and well-defined water table exists. Ragan used various techniques to monitor the water budget of the site:

a. weirs were installed at both ends of the stream length;
b. wells were established in the stream bank;
c. access tubes were used to allow nuclear probes for soil moisture determination;
d. weirs were installed to measure lateral inflow from seeps on the valley floor into the stream;

e. a seismograph was used in determining the location and configuration of the water table;

f. a pit structure, similar to Whipkey's (1965), was constructed to check for overland flow and subsurface seepage on the main valley side slopes; and

g. precipitation and interception losses were recorded.

Altogether Ragan studies eighteen storms, varying in size between 0.2 inches and 1.32 inches; from these he came to the following conclusions.

Firstly, with one exception, the pit structure on the main hillside slope indicated neither overland flow nor subsurface seepage. This is perhaps not surprising in view of the almost isotropic soil conditions in the watershed. The exception relates to the very thin (1.5 in thick) litter layer of organic soil and needles that overlies the relatively more impermeable body of the main soil mass. Occasionally during these storms the "infiltration capacity" of the surface matter was exceeded and saturated flow occurred in the shallow humic horizon. The magnitude and duration of this flow were not, however, large enough to account for any significant amount of storm runoff.

Secondly, response of the ungauged lateral inflow to rainfall was, in contrast, believed to be very rapid. Well observations near the stream channel indicated the formation of a ridge in the ground water table parallel to the stream (Figure 2.6) and this

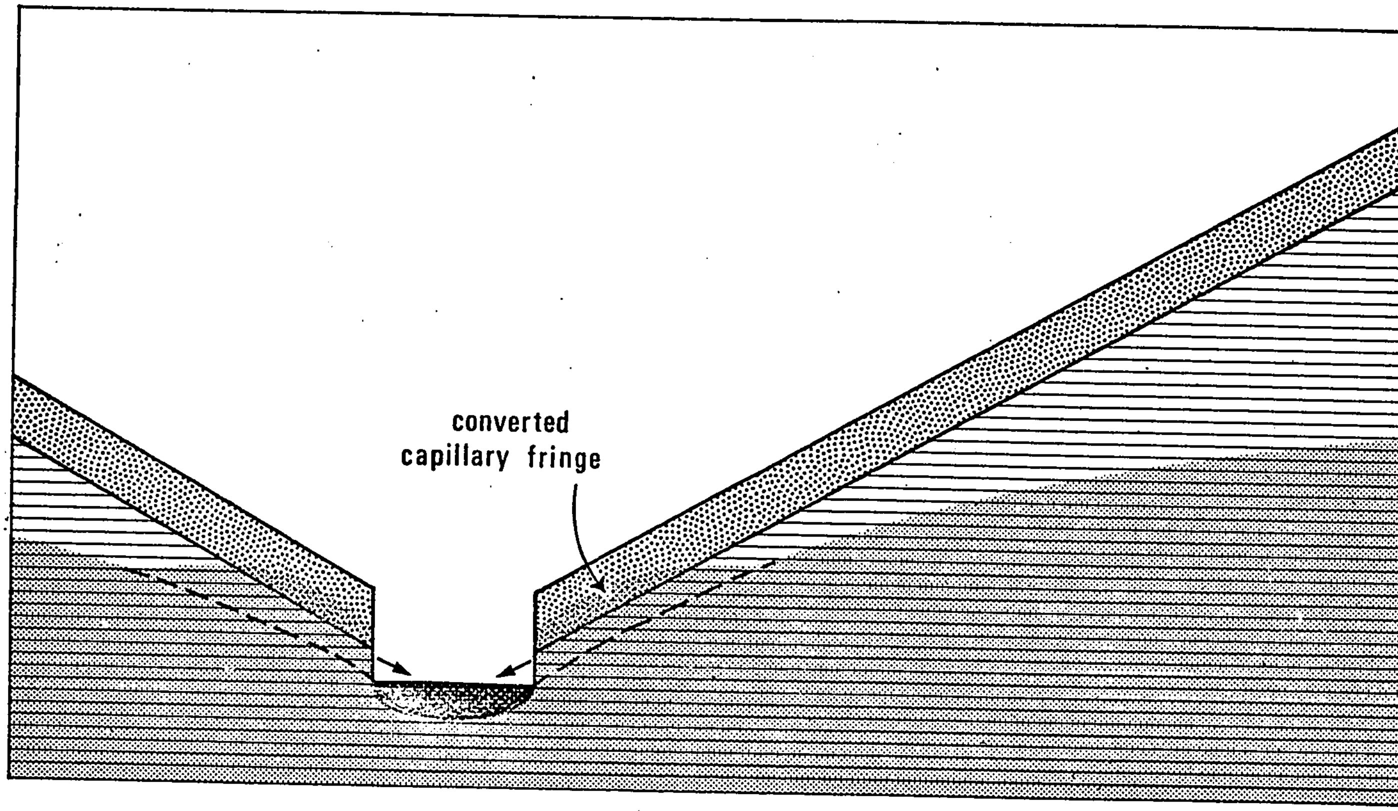


FIGURE 2.6

Direct Throughbank Seepage

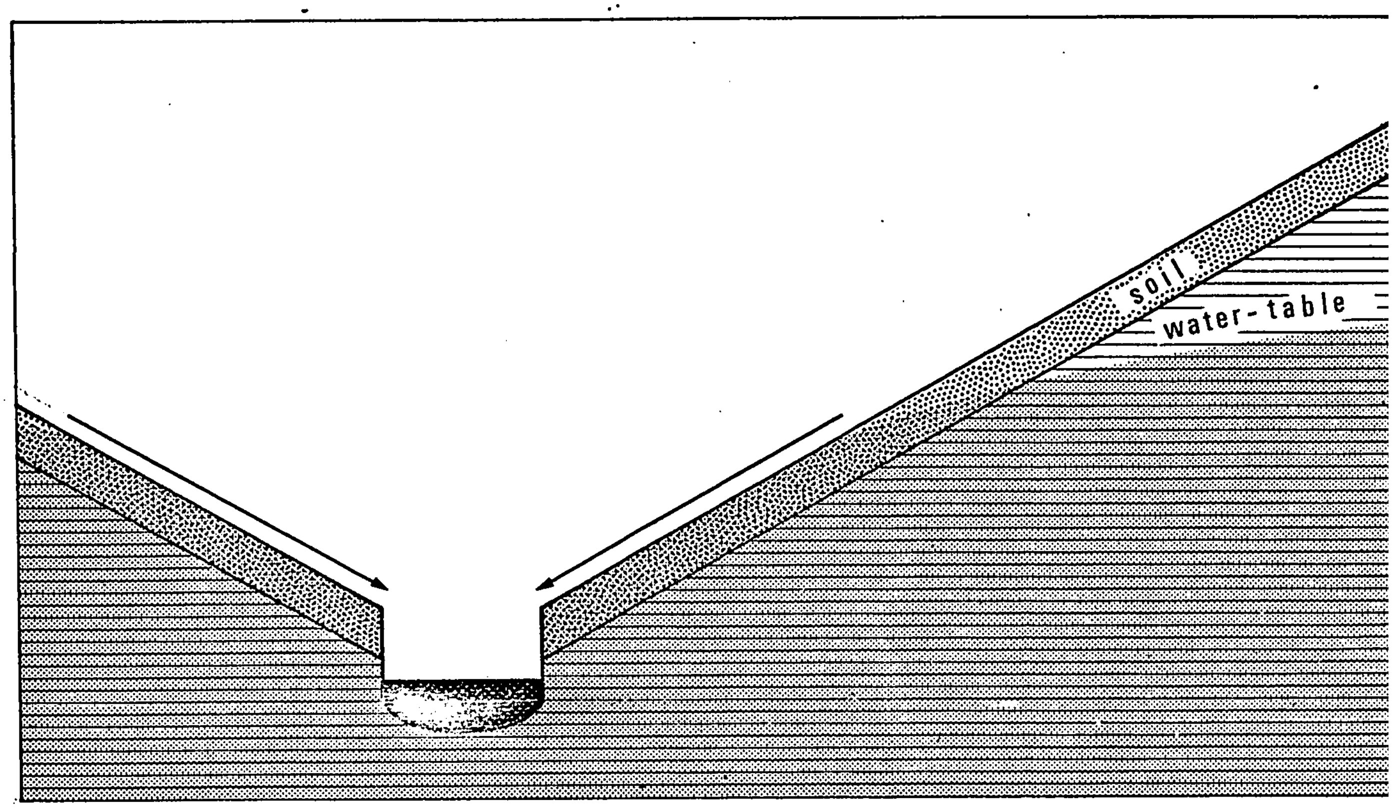


FIGURE 2.7

Lateral Expansion of the Surface Water System sudden increase in ground water levels appeared to play a major part in feeding the stream. The mechanism of this rapid rise in the water table level has already been discussed in connection with the variable source area model. Ragan did not measure this conjectured lateral inflow directly; he, in fact, estimated it as a residual after subtracting contributions from all other sources. He did find, however, that depth variations recorded in a well by the stream agreed closely with calculations of the lateral inflow using this method. In general,it was believed that this lateral inflow, essentially the same as the translatory flow conceived by Hewlett and Hibbert (1967), accounted for 43-50 percent of storm runoff.

Finally, the bulk of the remaining stormflow, and, indeed, the largest single contributor to the increased runoff, was derived from "seeps". These are areas where water emerges through the ground surface, and flows overland (and over the water table, which is coincident with the ground surface in these particular locations). These seeps are not temporary features, although their contributions to the channel system increase considerably during storm conditions. Part of this extra flow is from an expansion of the area of subsurface seepage and part is from direct interception of rainfall by those areas where the water table is at the surface.

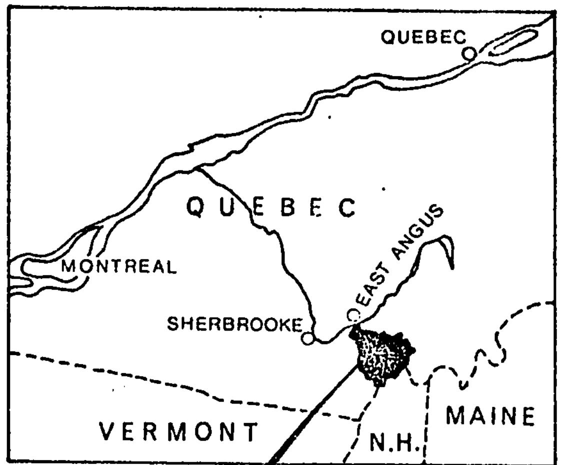
This third mechanism described by Ragan was also emphasized by Dunne and Black (1970a,b) working in a nearby catchment in Vermont. These two workers, however, tend to emphasize the dynamic nature of the situation rather more than Ragan. They not only claim that there is an appreciable rise in water table levels near the channel system, but also that in places where the water table is already just below the ground prior to the storm, the ground water system may rise to the ground surface over fairly large areas flanking the sides of the channel. This results in an acceleration of the input of water into the stream system in two ways. Firstly, direct interception of rainfall is no longer confined to the true channel network but is expanded to include these areas where the water table had risen to the surface. And, secondly, subsurface seepage is no longer confined to a narrow depth of channel bank (Figure 2.6) but is expanded to include those parts of the flood plain adjacent to the channels. Although the velocity of subsurface seepage may change only slightly, the expanded area through which it gets to the ground surface (Figure 2.7) results in a sharp increase in contributions to the channel system. Dunne and Black (1970a), from actual measurements, estimate that the ratio of the extra subsurface contribution to the direct interception of rainfall is about three to one.

One should note that Dunne and Black refer to this runoff as "overland flow" although, clearly, it is rather different to the mechanism of overland flow conceived by Horton (1945). Actually, in many ways, it is extremely similar to the variable source area model of Hewlett and Hibbert (1967) described earlier. Hewlett and Hibbert emphasized the headward extension of the channel net during storms; Dunne and Black are emphasizing the lateral expansion of the surface water system. They do disagree with Hewlett and Hibbert on one point; they argue that Whipkey-style subsurface seepage in the soil mantle on the main hillside is in-significant as a contributor to the expanded surface water system in the valley floors. They believe that the input of downward percolating water in the near-channel areas is sufficient to produce rapid response of the water table, and that downslope contributions from the main hillside are negligible. In this respect, their results concur with those of Ragan. Hewlett and Hibbert, on the other hand, have demonstrated that in the southern Appalachian environment this is not the case. To some extent, this difference is a rather minor one, and the ideas of Dunne and Black could justifiably be grouped in the general variable source area model. The idea that there is appreciable lateral, as well as headward, expansion of the channel system, makes possible quite large increases in the area of "channel bank" where subsurface seepage emerges at the surface, and may be a solution to the problem of generating sufficient storm runoff by seepage, as noted earlier.

CHAPTER III

THE EATON RIVER BASIN

The basin studied during this research is that of the Eaton River, in southeastern Quebec, situated on the northwestern flank of the Canadian Appalachians. The basin (Figure 3.1) is drained by three main rivers: the Clifton River, the North River, and the Eaton itself. The basin is bounded to the northeast by the Salmon River basin, and to the southwest by the Ascot River basin; all three basins drain to the northwest and are tributaries of the Rivière St. François. The southeastern divide of the Eaton basin is part of the main divide separating Canada and the United States. The headwaters of the Eaton, along this ridge, are at about 2000 feet; by the time it enters the St. François at East Angus, about 27 miles downstream, the Eaton has fallen 600 feet. In terms of relief, the basin is typical of the Canadian Appalachians (Cartier and LeClerc, 1964); the land is rolling, becoming increasingly rough to the southeast. The drainage density of the basin is typical of this environment and, based on maps at 1:50,000, approximates 1.24 mi/mi^2.



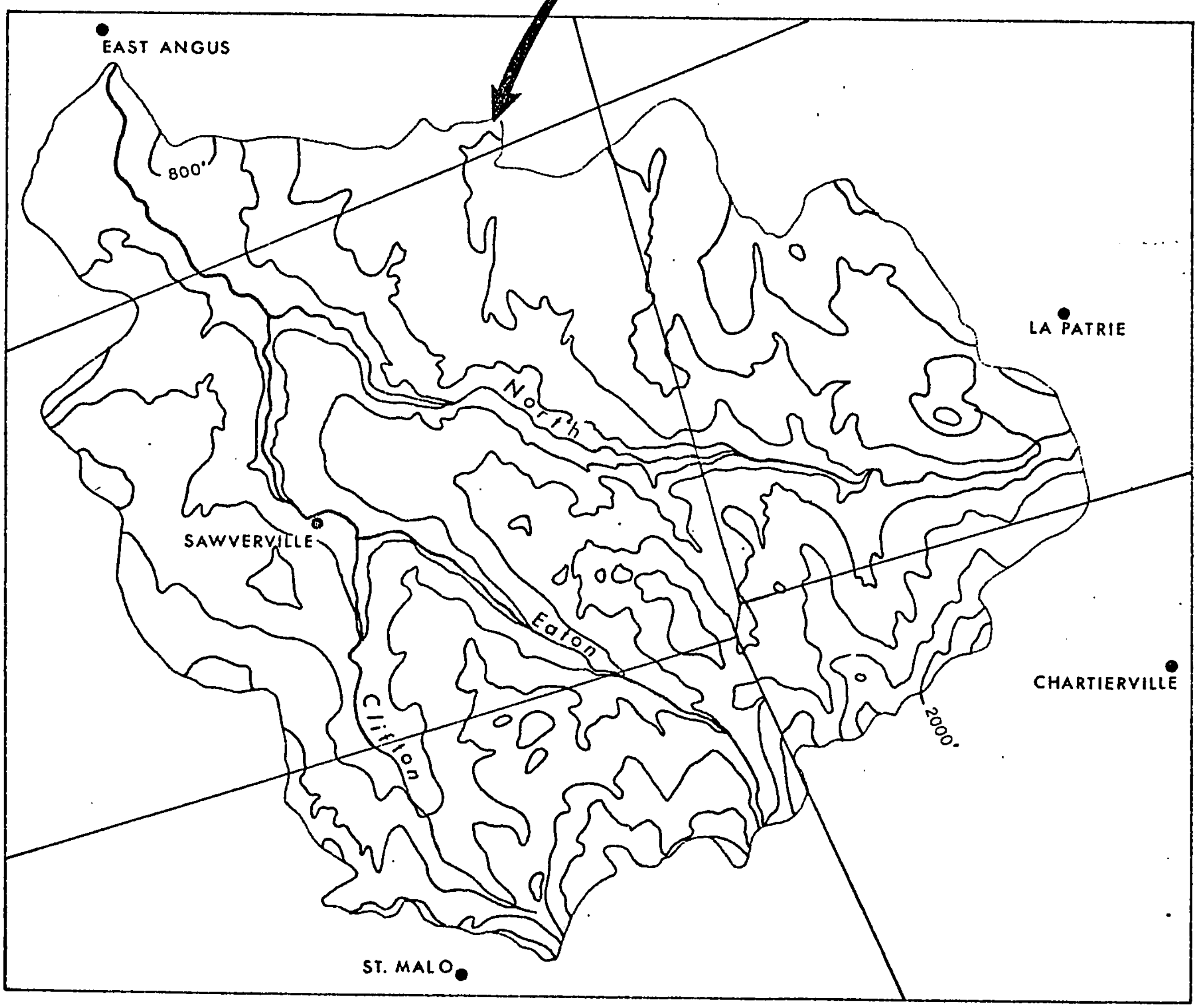


FIGURE 3.1

The Eaton River Basin

Climate

The area in which the Eaton River lies is characterized by just over 40 inches of precipitation yearly, divided almost evenly throughout the year but with minor peaks in both summer and winter. Part of this monotony of the march of precipitation is due to the insignificance of summer convective instability, although cyclonic activity continues throughout the summer months. The climograph (Figure 3.2) indicates a stronger summer peak than winter peak. This is due to the greater orographic effect of hilly terrain during the summer; the resulting instability supplements rainfall from the passage of cyclonic systems. The general lack of convective activity in this region may be accounted for by the northern location, the cool Atlantic waters and indented coastline, and possibly by the marine influence in summer of the Great Lakes (Trewartha, 1961), although it is rather doubtful if this latter influence extends to the Eastern Townships of Quebec. Cyclonic storm tracks originating almost anywhere in interior North America pass over this region, giving abundant precipitation throughout the year.

From records taken during the period 1962-1966 at Sawyerville North, a station located approximately in the centre of the basin both geographically and altitudinally, as reported by Bowker (1968), precipitation averaged 40.51 inches, with a range

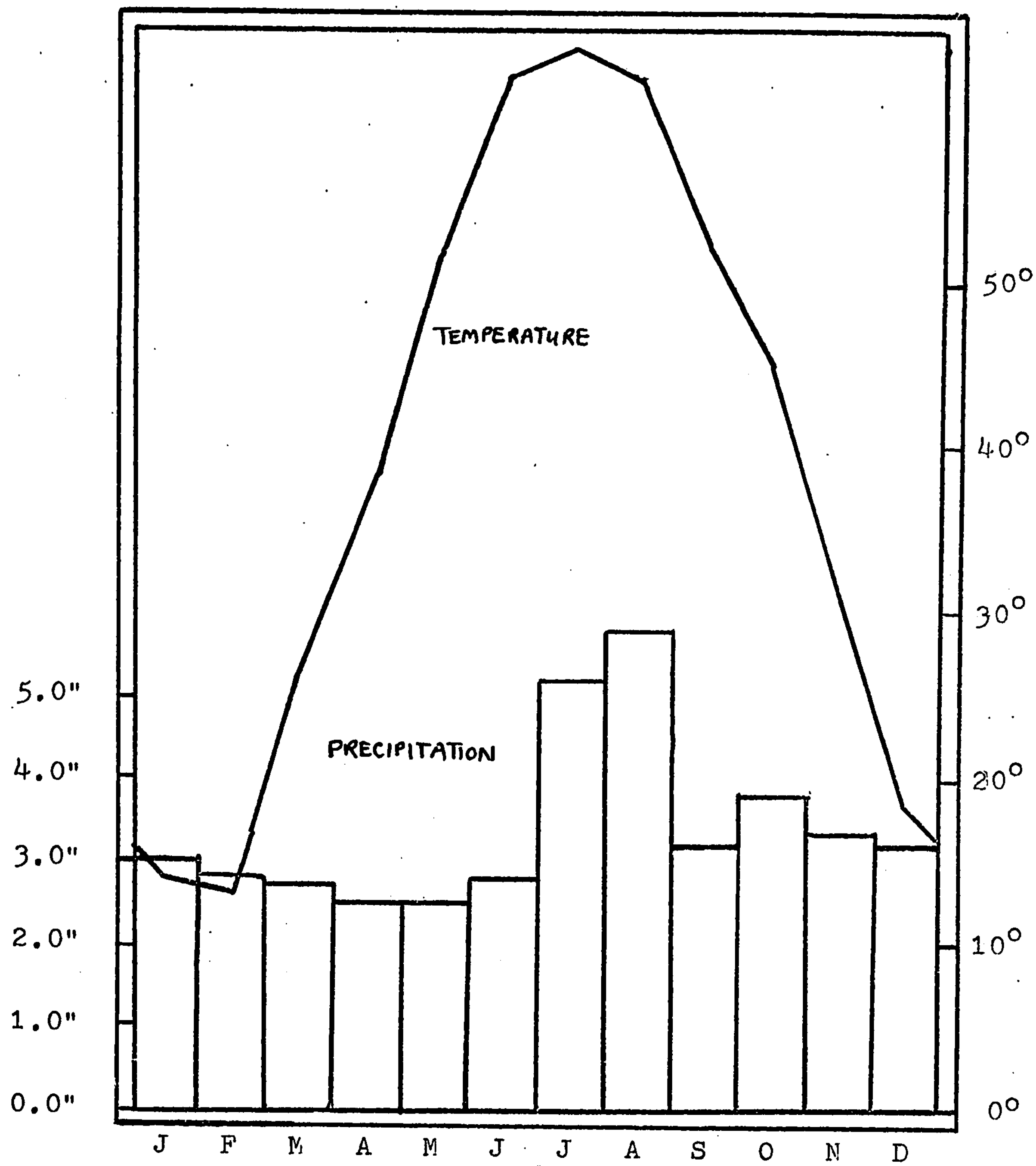


FIGURE 3.2

Climograph for Sawyerville-North, 1962-1966 of about 5 inches. Precipitation maxima were in July and August, which each received over 5 inches, and minima were in April and May, each with about 2.5 inches. The number of rainy days varied from 140-170, with a mean of 157. An average of seven heavy rainstorms (0.80 inches or greater) occurred each year. Snowfall amounts varied during this period from 75 inches to 150 inches, with a mean of 120. The average snow-free period was 170 days.

Mean monthly temperatures reached a low of +13°F in February, and a high of +64°F in July. The frost-free season, defined as the period when the temperature never drops below 32.5°F, averaged 122 days, with a range of 15-20 days on either side of this mark. On the average, 177 days had temperatures below 32°F, and July and August are the only months in which such days have never been recorded. Growing degree-days from May to October averaged 2744.

Geology and Glacial History

The surficial geology of the area in which the Eaton River is situated has been described for the Geological Survey of Canada by McDonald (1969), and most of the following account is based upon this study.

The basin is underlain by slates, limestones, and greywackes of the lower Palaeozoic St. Francis Group. Strike is consistently northeastward and the rivers generally flow across strike.

The area was repeatedly glaciated; McDonald (1967a) has found evidence for at least three glacial phases in southeastern Quebec. Glacial drift, which was derived from calcareous bedrock terrain, reaches depths of 200 feet in places in valley bottoms, thinning toward the interfluves. Drift depths, however, are very irregular, and in places the river flows on bedrock.

The older tills, as well as the surface till, are gray, compact, calcareous and stony throughout. Thinly stratified lake sediments separate the till layers. These are sand and silt-clay sediments deposited in front of advancing and retreating ice fronts. The till mantle covers most of the basin, although on the interfluves bedrock is quite near the surface and determines the local relief and drainage patterns. The thickest exposed section of surface till, measuring 75 feet, occurs in the valley of the North River. The till is oxidized and leached of carbonate near the surface. It is characteristically compact with well-striated and rounded pebbles throughout. Boulders as large as two feet in diameter are common, occupying up to five percent of the till volume. Grain size distribution of the surface till is shown in Table 3.1. The soil mantle derived from this till is generally less compact, due to eluviation and mechanical loosening.

Ice positions are marked by till ridges with intervening melt-water channels. These are locally as high as 100 feet and as broad as half a mile.

	Till Sample Number 1	2	3
Gravel (%)	1	4	0
Sand (%)	24	46	0
Silt (%)	45	32	20
Clay (%)	30	18	80
Median Particle Size (mm.)	0.02	0.06	0.0016

Till number 3 occurs in the upper Eaton valley; till samples 1 and 2 are more representative of the basin as a whole

Table 3.1

Grain size characteristics of surface till in the Eaton basin

(from McDonald, 1969)

Modern alluvium is present in all the river valleys but is extensive only along the major rivers. It contains all gradations of particle size from silt to boulder-gravel. Terraces underlain by alluvium are numerous. Modern alluvial plains are locally as much as a half mile wide along the North and lower Eaton Rivers. Changes in the width of these plains is often abrupt.

Land Use

Only about 25 percent of the land area of the catchment is cleared, the remainder being a mixed deciduous and coniferous forest, including red spruce, balsam fir, sugar maple, beech, and white and yellow birch (Braun, 1964). The leached and cobbled nature of the soil and the short growing season (120-135 days) make the area unsuitable for agriculture generally, and farms are small. The cleared land is almost exclusively in the valleys, and is mainly pasture land.

The basin has several small towns, including Cookshire, Sawyerville, West Ditton, Island Brook, St. Isidore d'Auckland, and the largest, East Angus. A number of hard and loose surface all-weather roads connect these towns; there are no major highways. Settlement is mainly in the north-west; the more rugged southeastern quarter of the basin is virtually uninhabited and no paved roads go to it.

Canadian IHD Program

Since 1964, the Eaton basin has been an official research basin in the Canadian program for the International hydrological Decade. As one of three such basins in Quebec, research within it is co-ordinated by the Service d'Hydrologie, Ministère des Richesses

Naturelles du Québec. The overall research objective of work in the basin is described (Secretariat, Canadian National Committee, 1967, p. 141) as:

> "Etudes des elements du bilan hydrique d'un bassin representatif d'une region naturelle. Preciser l'hydrologie regionale et favoriser une utilisation plus rationnelle de cet element pour les besoins de consommation."

As noted earlier, much of this work takes the form of a physical study of various elements of the hydrologic cycle. The extensive hydrologic records of the basin, together with more recent installation of meteorologic equipment, do, however, provide the opportunity for a parametric study of certain aspects of the basin's hydrology. In the light of the particular problem outlined in Chapter I, the Eaton basin clearly formed an ideal field area.

CHAPTER IV

ANALYSIS OF DATA

In an attempt to evaluate the relative importance of some of the ideas discussed in the previous chapter, several simple analytical techniques were used on data from the Eaton basin. These data, supplied on punch cards by the Ministère des Richesses Naturelles du Québec, fall into two distinct sets. The first consisted of mean daily discharge figures, in cubic feet per second (cfs), for East Angus, for the period 1933-1964; the second comprised actual hourly discharge figures (cfs) for East Angus for each of 38 storms that occurred in the Eaton basin during the snow-free seasons of the years 1953-1966. These two sets are henceforth referred to as the yearly data and the individual storm data respectively. Actually, the yearly data were modified to include only the period between the end of the spring melt and the first snowfall in each year; this was done using records supplied by the Ministère des Richesses Naturelles du Québec and the Canadian Department of Transport. The individual storm data do not encompass every summer storm that occurred in the period 1953-1966, but they do include the majority of them. The variability in the size of these storms, and in the size of the resultant floods, is quite large; storm runoff values for these events varied between 0.08 in. and 1.98 in. (actual values depend on the particular definition of "storm runoff" employed, an issue discussed below), and so include both medium and large flood events. This study is thus biased toward significant flood events rather than the response of the basin in all types of storms that are experienced.

The Data

In the analysis of these data, two parameters were used as dependent variables: storm runoff and hydrologic response. Parameters intended as predictors of these two variables were chosen to represent storm size, storm intensity, antecedent moisture conditions in the basin and possible losses of moisture due to evapotranspiration between the end of the storm and the end of the flood event. The methods employed in determining values of these parameters for individual storms, or for the snow-free seasons, are described below. This is necessary so that the limitations in the data are fully appreciated from the outset.

Storm Runoff

Several methods of separating storm runoff from baseflow have been developed. Although each has its own basis in physical reasoning, they are all, to varying degrees, essentially empirical procedures containing definite arbitrary aspects. Summaries of these techniques are available in standard hydrology texts such as those by Linsley, Kohler and Paulhus (1958), Chow (1964), and Bruce and Clark (1966). In this study, two methods were used. Comparison between results based on the two methods affords some indication of the degree to which subjectivity in the definition of "storm runoff" affects the conclusions that emerge from the data; in actual fact, it appears that, in this particular type of study, subjectivity in the separation of hydrographs into stormflow and baseflow is not important, provided consistency is maintained throughout the study.

The first method, and the only one applied <u>directly</u> to both sets of discharge data, is that used by Hewlett and Hibbert (1967) in their study of small basins in the southeastern United States. It is illustrated in Figure 4.1. Storm flow is separated from base flow by a straight line, rising at the rate of 0.05 cubic feet per second per square mile of the basin per hour (csm/hr) from the beginning of the storm until the separation line intersects the recession limb of the hydrograph. Since the constant for the slope of the separation line was developed for basins in the southern Appalachians there is absolutely no reason why it should be applicable to the Eaton basin. Moreover, in the study referred to by Hewlett and Hibbert (1967), the method was

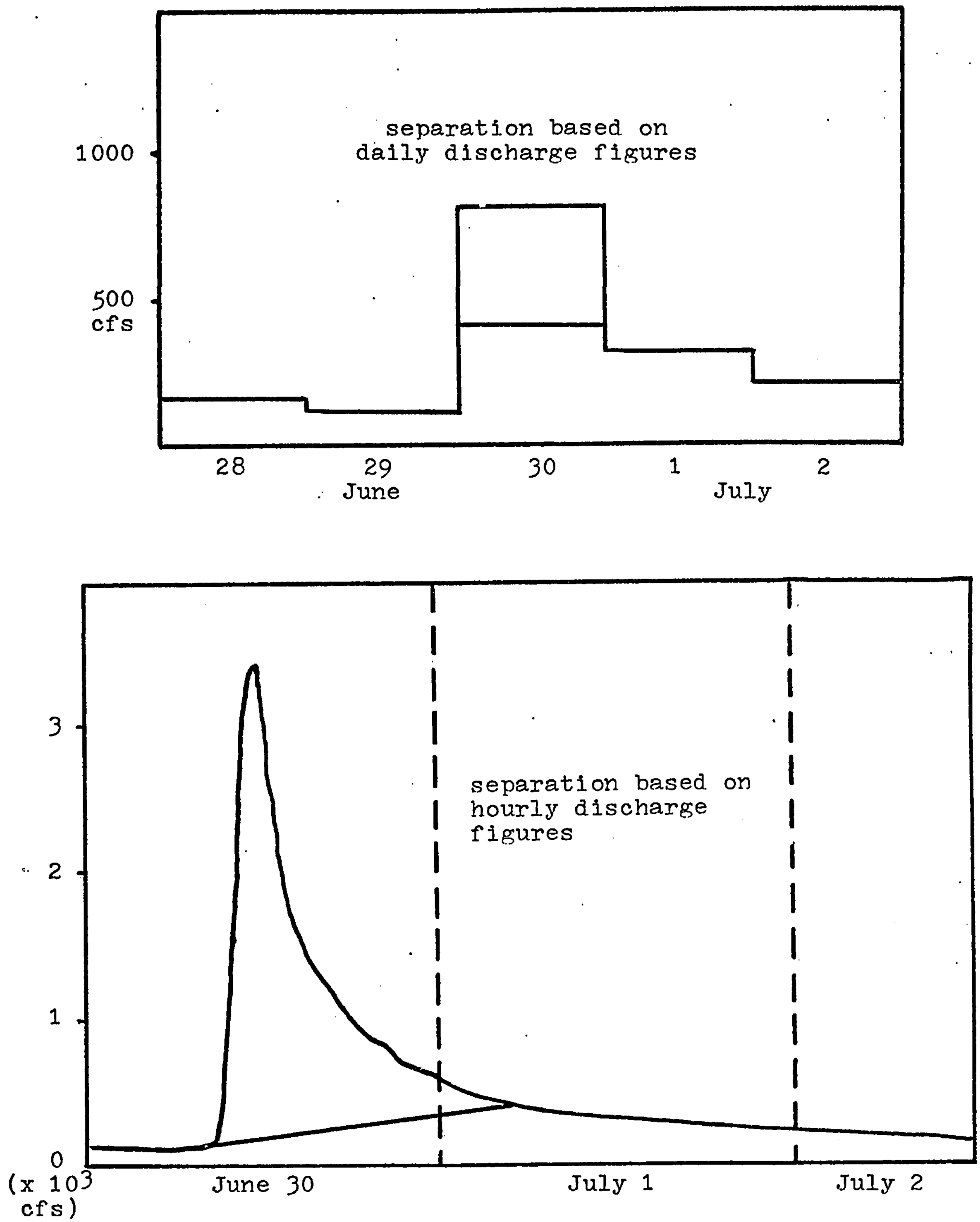


FIGURE 4.1

Hewlett Separation Procedure for the Storm of

June 30 - July 1, 1961

regarded as unsuitable for basins greater than about 200 square miles; the Eaton basin at East Angus is 248 square miles. The constant was nevertheless retained in this study. Firstly, it enables direct comparison between the Eaton basin and the basins examined by Hewlett (1967) in the southern Appalachians. Secondly, the constant is unlikely to affect significantly the correlation of storm runoff with other variables; absolute values of storm runoff are affected by the constant, but the variance of these values, and , more particularly, the covariance with other parameters, is not likely to be changed appreciably. This, indeed, is borne out in the analysis; runoff values using the Hewlett method are consistently smaller than those obtained by the second method, but the results of the correlation analysis are essentially the same.

The second separation technique used was the one developed by Barnes (1939, 1940). In this method, runoff is plotted on a logarithmic scale against time. The baseflow recession limb, which approximates to a straight line, is extended back to the time of the peak; this is then connected by a straight line to the point at the beginning of the storm. This was applied directly only to the individual storm data. Comparison of the runoff values obtained by the two separation methods for the individual storm data was quite good. The values are connected by the equation

$$s_b = s_h + 0.10'' \qquad (4.1)$$

where s_b is storm runoff by the Barnes method, and

s_h is storm runoff by the Hewlett method; the correlation coefficient (r) for this sample relationship is +0.94.

The computation of storm runoff for the yearly data, using the Hewlett method, uses the same separation line, but the constant is now expressed in the form 1.2 csm per day, rather than 0.05 csm per hour. Inevitably, the determination of storm runoff for an individual flood event using daily discharge data is much less accurate than using hourly data. Moreover, since the total (snow-free season) yearly storm runoff figure is simply the sum of storm runoff in the separate flood events of a given year, any error could be compounded in the derivation of the total yearly storm runoff. Fortunately, it was possible to check the accuracy of the Hewlett method using daily data against the Hewlett method using hourly data, with reference to 36 of the 38 individual storms for which hourly, and also daily, discharge figures were available. The two values for any storm differed in every case with the values based on daily data consistently underestimating the true (based on hourly data) values. The least-squares equation linking the two values was:

$$s_{h(h)} = 0.05'' + 1.2[s_{h(d)}] \qquad (4.2)$$

where $s_{h(h)}$ is storm runoff for a given flood using hourly data, and

$s_{h(d)}$ is storm runoff for a given flood using daily data; the difference attributable to the constant (0.05 inches) is far from insignificant physically and computation of yearly storm runoff values based on $s_{h(d)}$ data could be appreciably in error in any year with a large number of storms. All storm runoff values for the yearly data, using the Hewlett method, were therefore "corrected", using Equation 4.2.

Determination of storm runoff values, using the Barnes method, for the yearly data, was undertaken as follows: the total storm runoff in any year is given by

$$S_b = \Sigma s_b \tag{4.3}$$

summing for the number of storms in the year. Substituting for s_b (from Equation 4.1) into Equation 4.3, the following expression is obtained:

$$S_b = \Sigma(s_h) + \Sigma(0.10") \tag{4.4}$$

again summing for the number of storms in the year. In Equation 4.4, s_h is, in fact, $s_{h(h)}$ derived from daily data by Equation 4.2

Rainfall

Precipitation records for the Eaton basin are not, unfortunately, either as complete, as detailed, or as comprehensive as discharge records. Only one station within the basin, at Sawyerville, has continuous daily precipitation records for the period 1953-1966. Four other stations, at East Angus, La Patrie, St. Malo and Chartierville, fortunately surround the basin, and none is further than four miles from it.

The Thiessen polygon method (Thiessen, 1911) of estimating total rainfall on a basin was used for determining rainfall in the study of individual storm data. Although this method is merely a weighted mean, and is not as sophisticated as some methods currently in use, for a number of reasons it was felt to be adequate: the recording stations are spaced fairly evenly around the basin; the scarcity of stations did not justify use of a more subjective method, such as isohyets; and examination of precipitation data for the 38 individual summer storms made it clear that all five stations came under the same synoptic influences, and were therefore suited for use in combination. Although amounts may vary widely from station to station in a single storm,in no case did a station fail to receive rain when the others showed measurable precipitation. The days upon which rain fell also show a marked correspondance from station to station.

Table 4.1 shows the area covered by each polygon. Visual inspection of the Thiessen map (Figure 3.1) makes it clear that while quantitatively the distribution seems uneven, this is caused mainly by Sawyerville's central location and not by any pronounced unevenness in spacing of the stations.

Table 4.1

Station	Area: percent of basin	square miles
Sawyerville	48.5	121.52
La Patrie	21.2	52.08
St. Malo	15.6	39.68
East Angus	8.4	19.84
Chartierville	6.3	14.88

None of these stations is a Class I Department of Transport station, and it is unfortunate that precipitation data for three of the five stations are not complete. Data for St. Malo were missing for five storms, East Angus was missing records for four storms, and two storms lacked data from Chartierville. Two further storms lacked data for more than one station. Fortunately, records for Sawyerville and La Patrie, the stations showing most prominently in the Thiessen diagram, were complete. Altogether, one-third of the storms examined had missing data. Proportions of basin area allotted to each of the remaining stations in each of these cases are shown in Appendix I. Precipitation readings are made at 0800 local standard time; the time periods do not, therefore, conform precisely to discharge records, which utilize the calendar day from midnight to midnight, or the hour. This discrepancy is unfortunate, especially as there is no way to correct for it, and data are used as they are, ignoring the eight-hour difference.

In the analysis of the yearly data the problem of finding precipitation stations becomes much more difficult. Of the five stations mentioned previously, only East Angus recorded for the period 1933-1964, and the records have gaps affecting seven years. Fortunately, Lennoxville, located nine miles west of the catchment, has complete records for this period. The correlation between the two stations for yearly snow-free season precipitation is +0.82, and the least squares regression is

$$P_{ea} = 1.22\ (P_l) - 3.48'' \qquad (4.5)$$

where P_{ea} is total precipitation in the snow-free season at East Angus, and P_l is total precipitation in the snow-free season at Lennoxville; no closer estimate of basin rainfall, for the yearly data, could be obtained. Even in the years when complete records were available for East Angus, the error in using this station's rainfall as a measure of the basin's rainfall is probably large. Some attempt to determine this error was made by comparing, for the 38 individual storms, rainfall at East Angus with basin rainfall as computed by the Thiessen polygon method. The least squares equation is:

$$p_b = 0.99\ (p_{ea}) - 0.06'' \qquad (4.6)$$

where p_b is the basin rainfall for a single storm, and p_{ea} is the rainfall at East Angus for a single storm; the correlation coefficient is only 0.72. Precipitation values used in the analysis of the yearly data are, therefore, considerably less accurate than those used in analysis of the individual storm data.

Hydrologic Response

The hydrologic response of an individual storm event is the ratio of storm runoff and rainfall associated with that event. Quantitatively it is defined by either

$$r_{h(h)} = (s_h)/(p_b) \tag{4.7}$$

or

$$r_{h(b)} = (s_b)/(p_b) \tag{4.8}$$

depending upon which hydrograph separation procedure was used. Actually, one should note that the two p_b values are not necessarily identical; the duration of the flood event was usually greater when the Barnes separation procedure was used and, in 11 of the 38 storms, rainfall occurred after the end of the flood event as designated by the Hewlett criterion but before the end as indicated by the Barnes method.

Using the yearly data, the average hydrologic response of the basin is given by

$$R_{h(h)} = S_h/P_{ea} \tag{4.9}$$

and

$$R_{h(b)} = S_b/P_{ea} \tag{4.10}$$

where R_h indicates average hydrologic response for an entire snow-free season. It is directly comparable to the term employed by Hewlett and Hibbert (1967) except in as much as it relates to a six to eight month average rather than to a full 12-month average.

Storm Intensity

Precipitation data were too crude tc give more than a very generalized estimate of storm intensity; records for periods of time shorter than 24 hours, for the events examined in this work, simply do not exist. Data are also complicated by an observational day beginning at 0800 hours rather than at midnight. To circumvent these problems as far as possible, three different variants in storm intensity were used for the individual storm data; these were (1) average daily rainfall amount (inches per day); (2) maximum one-day rainfall amount recorded at Sawyerville (inches per day); (3) number of days of rainfall. The first and third values differed for the same flood event, depending on whether the duration of stormflow was determined by the Barnes or the Hewlett method. Two measures of storm intensity were used for the yearly data; these were: (1) average rainfall amount per rainy day (total rainfall in the snow-free season divided by the number of days in which rainfall amounted to, or exceeded, 0.01 inches); (2) percentage of rainy days during the snowfree season (the number of days on which rainfall amounted to, or exceeded, 0.01 inches divided by total number of days in the snow-free season, and expressed as a percentage).

Antecedent Moisture Conditions

No actual antecedent soil moisture data exist for the events examined here. Instead surrogates for this effect were used based on the available meteorologic and hydrologic data. Three measures were used in the study of the individual storm data.

The first measure was the time, in days, since the last storm producing a rise of 500 cfs or greater in one day, using daily discharge figures since hourly figures were not available on a continuous basis. The value of 500 cfs was chosen because this, using daily discharge data, was the one-day rise of the smallest of the 38 storms. A figure much smaller would have resulted in including almost every fluctuation in the hydrograph. In giving an indication of the length of time to the last major event, this index was hoped to be inversely related to antecedent moisture conditions in the basin, and, in turn, to the effect these conditions would have on the flood.

The parameter just described probably is the simplest antecedent rainfall index available; a more complicated index is often derived from the formula

$$AP = \sum_{n=1}^{N} k^n P_n \tag{4.11}$$

given by Bruce and Clark (1966), and derived from a similar formula by Kohler and Linsley (1951). In this formula AP represents the average antecedent rainfall over a period of N days prior to the beginning of the particular storm in question; P_n is the actual precipitation on the day that is n days before the storm. Both k and N are chosen by the investigator. In this study N was chosen equal to 30; an examination of response in relation to days since the last storm (as defined above) indicated that rainfall more than 30 days prior to a storm had little effect on the associated flood event. Previous work by Bruce and Clark (1966) in southern Ontario suggested that k = 0.84 produced maximum correlation between storm runoff and antecedent precipitation in that area. There is, however, no justification for a blind adoption of this value for the Eaton basin. Instead, antecedent precipitation values, for each of the 38 storms, were constructed for each value of k ranging from 0.82 to 1.01, and these were correlated with both storm runoff and hydrologic response; in the subsequent analysis only those values of k which maximized the correlation with s_b, s_h, $r_{h(b)}$, and $r_{h(h)}$ were used. The differences between these four values (Table 4.2)

Response	Storm Runoff	k	Storm Runoff	Response
0.322	0.142	0.82	0.147	0.416
0.330	0.154	0.83	0.158	0.419
0.340	0.168	0.84	0.170	0.423
0.350	0.184	0.85	0.184	0.426
0.360	0.201	0.86	0.198	0.428
0.370	0.219	0.87	0.212	0.429
0.379	0.237	0.88	0.223	0.430
0.388	0.257	0.89	0.246	0.429
0.396	0.277	0.90	0.262	0.426
0.401	0.296	0.91	0.278	0.421
0.406	0.315	0.92	0.294	0.415
0.408	0.332	0.93	0.307	0.405
0.408	0.348	0.94	0.320	0.394
0.404	0.361	0.95	0.330	0.380
0.398	0.371	0.96	0.336	0.363
0.389	0.379	0.97	0.341	0.345
0.377	0.384	0.98	0.343	0.325
0.362	0.386	0.99	0.342	0.302
0.346	0.385	1.00	0.338	0.280
0.328	0.381	1.01	0.332	0.257

Left-hand columns indicate Hewlett data;
right-hand columns indicate Barnes data

Maximum correlation coefficients are underlined

Table 4.2

Correlation coefficients linking hydrologic response and storm runoff with antecedent precipitation for different values of k is puzzling, although the issue was not explored further.

The third index of antecedent moisture conditions was the level of hourly baseflow (cfs) prior to the beginning of each flood event. As pointed out in the description of the "partial area" models of runoff production in Chapter II, this might be expected to play a prominent role, directly, in determining basin response.

An antecedent moisture index for the yearly data proved to be much more difficult to visualize and define in terms of easily measured variables. Finally an average was taken of the minimum discharge preceding each storm recorded by the Hewlett separation method. It was hoped that by averaging the fluctuations in "wetness" over the year a meaningful estimate of moisture deficiency might be obtained. A second index of average basin wetness was provided by the number of major storms that occurred in the snow-free season. A major storm was, for this purpose, defined as any storm capable of producing a hydrograph rise greater than 1.2 csm/day, and thus capable of detection by the Hewlett separation method.

Temperature and Evapotranspiration

These two variables were included in order to obtain an indication of any moisture lost during the flood period after the storm by evapotranspiration, and therefore unavailable for flood production. The temperature index was simply the average maximum temperature for the period of the storm. An evapotranspiration index was constructed describing the number of hours of daylight and the number of degrees above 32°F for each storm. No comparable indices were constructed for the yearly data since the variance among different years would almost certainly have been very small. Even in the individual storm data the contribution made by these two parameters was completely insignificant.

The Analysis

The simple correlations between all variables for the individual storm data are shown in Table 4.3 with the Barnes data and the Hewlett data separately given in the upper and lower triangles of the matrix respectively. The similar correlation matrix for the yearly data is given in Table 4.4.

Stepwise multiple linear regressions were undertaken for hydrologic response on the predictor variables for both individual storm data and yearly data. The results are summarized in Tables 4.5 and 4.6. A similar regression analysis was also used with storm runoff as the dependent variable (and including storm rainfall as a predictor) for the individual storm data and the significant results are given in Table 4.7. Actually it is misleading to describe this analysis as "stepwise" since, in effect the program that was used examines all combinations of independent

	0	1	2	3	4	5	6
0		-.175	<u>.434</u>	<u>.629</u>	.143	.130	-.060
1	-.122		-.252	-.092	-.058	.078	.090
2	<u>.400</u>	-<u>.362</u>		<u>.525</u>	-.062	-.050	.044
3	<u>.617</u>	-.092	<u>.460</u>		-<u>.342</u>	-.248	.182
4	.211	-.059	.012	-<u>.342</u>		.739	-.071
5	.312	.041	.043	-.234	<u>.880</u>		-<u>.514</u>
6	.039	-.065	.088	.278	-.169	-.457	

0: hydrologic response
1: days to last major flood
2: antecedent precipitation index
3: preceding baseflow
4: maximum one-day rain
5: average daily rainfall
6: storm duration

Upper triangle denotes Barnes data; lower triangle denotes Hewlett data.

Only those values that are underlined are statistically significant at the 0.95 level.

Table 4.3

<u>Sample correlation matrix linking hydrologic and meteorological variables for 38 storms</u>

	0	1	2	3	4
0		$\underline{.493}$	.280	.216	.795
1	$\underline{.457}$		.218	-.047	$\underline{.572}$
2	.313	.218		-.542	.248
3	.225	-.047	$-\underline{.542}$		.277
4	$\underline{.725}$	$\underline{.572}$	.248	.277	

0: hydrologic response
1: average preceding baseflow
2: average daily rainfall
3: percentage rainy days
4: number of storms

Upper triangle denotes Barnes data;
lower traingle denotes Hewlett data.

Only those values that are underlined
are statistically significant at the
0.95 level.

Table 4.4

Sample correlation matrix linking hydrologic and
meteorological variables for 32 snow-free seasons

	Hewlett		Percent Reduction In Variance Of Response (X_0)		Barnes
X_1	Preceding Baseflow	38.0	39.5	X_1	Preceding Baseflow
X_2	Average Daily Rainfall	22.1	14.6	X_2	Maximum One-Day Rain
	Total	60.1	3.0	X_3	Storm Duration
			1.6	X_4	Average Daily Rainfall
			58.7		Total

Left-hand columns indicate Hewlett data; right-hand columns indicate Barnes data.

Final regression equations are:

Hewlett: $X_0 = 0.007 + 0.001\ X_1 + 0.123\ X_2$

Barnes: $X_0 = 0.046 + 0.001\ X_1 + 0.143\ X_2 - 0.038\ X_3 - 0.073\ X_4$

Table 4.5

<u>Summary of stepwise multiple linear regression with hydrologic response as dependent variable</u>

		Percent Reduction In Variance Of Response (X_0)			
X_1	Number of Storms	52.6	63.2	X_1	Number of Storms
X_2	Percentage of Rainy Days	2.4	63.2		Total
X_3	Average Daily Rainfall	1.9			
	Total	56.9			

Left-hand columns indicate Hewlett data; right-hand columns indicate Barnes data.

Final regression equations are:

Hewlett: $X_0 = -0.184 + 0.009\ X_1 + 0.298\ X_2 + 0.279\ X_3$

Barnes: $X_0 = 0.010 + 0.013\ X_1$

Table 4.6

Summary of stepwise multiple linear regression with hydrologic response as dependent variable: yearly data

		Percent Reduction In Variance Of Storm Runoff (X_0)			
X_1	Storm Rainfall	49.0	52.6	X_1	Storm Rainfall
X_2	Preceding Baseflow	14.7	12.3	X_2	Preceding Baseflow
X_3	Average Daily Rainfall	7.1	5.5	X_3	Storm Duration
X_4	Antecedent Rainfall Index	1.4	1.7	X_4	Antecedent Rainfall Index
	Total	72.2	72.1		Total

Left-hand columns indicate Hewlett data; right-hand columns indicate Barnes data.

Final regression equations are:

Hewlett: $X_0 = -0.694 + 0.423\ X_1 + 0.001\ X_2 + 0.180\ X_3 + 0.109\ X_4$

Barnes: $X_0 = -0.489 + 0.581\ X_1 + 0.001\ X_2 - 0.083\ X_3 + 0.199\ X_4$

Table 4.7

Summary of stepwise multiple linear regression with storm runoff as the dependent variable variables, although, for any level of combination (e.g. a pair of predictors), only the particular combination that maximizes the reduction in the variance of the dependent variable for that level is shown in the output. The program is therefore similar to the "sequential" multiple linear regression analysis described by Krumbein, Benson and Hempkins (1964), and is free from general criticisms levelled at "stepwise" programs made by mathematical statisticians (e.g. Beale, Kendall and Mann, 1967). All computations were performed on the IBM 360/75 computer at the McGill University Computing Centre.

CHAPTER V

DISCUSSION

The results for both sets of data present a fairly simple picture and, moreover, one that is independent of the method of separating storm flow from baseflow.

In the correlation matrix for the individual storm data (Table 4.3) there are only three really significant sets of association among the variables: the correlation among the variables representing storm intensity (variables 4, 5, and 6); the correlation between two of the parameters representing antecedent moisture conditions (2 and 3); and the correlation between response (0) and these two antecedent moisture variables (2 and 3). Perhaps more important is the complete lack of association between response (0) and the storm intensity parameters (4, 5, and 6); this fact may be taken as support for the rapidly growing belief in hydrologic circles that the Horton model of overland flow, which depends so much on storm intensity, is simply not a successful interpretation of runoff production in humid areas such as the northern Appalachians. Admittedly one might question the accuracy of the storm intensity data used in this study, but they probably contain less error than some of the other data employed. A similar pattern emerges in the correlation matrix for the yearly data (Table 4.4); the relation between response (0) and the storm intensity parameters (2 and 3) is dwarfed by the association between response and the measures of average basin "wetness" (1 and 4). There is a significant decrease in the importance of preceding baseflow in moving from the individual storm data to the yearly data, but this is not surprising; it presumably results from the reduced variance of this parameter in the latter set of data.

In turning to the results of the multiple correlation analysis, there is little extra information provided by the yearly data (Table 4.6). The only significant predictor of response is the number of storms per year; the disappearance of the other basin "wetness" parameter (average preceding base flow) follows directly from its own close correlation with the number of storms per year ($r_{1,4} = 0.572$). On the other hand, perusal of the individual storm data (Table 4.5) reveals that storm intensity, as represented by average daily rainfall (for the Hewlett data) and maximum one-day rainfall (for the Barnes data) are now significant, although subsidiary to antecedent moisture conditions.

Overall, the failure to account for more than about 60 percent of the inter-storm, or inter-year, variability in response is rather disheartening, but not entirely unexpected. Most of the parameters employed are, in fact, no more than approximate surrogates for the elements they are meant to represent (Carson, 1968), and for this reason alone it would be surprising if much higher figures of prediction had been obtained. Moreover, the existence of random error, or noise, in the data, will also tend to reduce the correlation coefficients. Table 4.7 shows, for comparison, the combined predictive powers of these variables, along with storm rainfall, for runoff rather than response. The overall reduction in the variance of the dependent variable (72 percent) is increased slightly, but the actual contribution of parameters representing antecedent moisture conditions and storm intensity is reduced relative to Table 4.5 Both changes are due to the inclusion of storm rainfall as an independent variable; indeed the dominant position of storm rainfall in Table 4.7 justifies the creation of the response parameter, eliminating storm size from the group of predictor variables and allowing more attention to be paid to the others. The predictive ability of storm runoff by the two equations given in Table 4.7 is comparable in magnitude to those provided by Lee and Bray (1969) for five New Brunswick watersheds.

Actually the purpose of this study was not so much to maximize the prediction of hydrologic response for the Eaton basin as to examine the <u>relative</u> importance of different parameters as predictors. In this way some indication may be obtained of the dominant controls of runoff production in the basin, and, perhaps, some indication is provided of the relative merits of different models of runoff production here. Unfortunately, all that the study shows clearly is that antecedent moisture conditions are the dominant control and that storm intensity and post-storm evapotranspiration are relatively minor factors. This is a useful conclusion, but, on its own, is hardly a basis for speculating on the relative merits of the different models outlined in Chapter II. Nevertheless, a closer look at this statistical data, along with other information acquired during the study, does permit a certain amount of theoretical consideration, especially as to the contribution made by the "partial area" model of runoff production.

It was pointed out above that two of the three parameters representing antecedent moisture conditions were significant predictors of inter-storm variability in response. No comment was made on the failure of the third variable to correlate with response, nor was attention directed to the discrepancy of the first two variables in their powers of prediction. The relative ranking of the three variables (1, 2, and 3 in Table 4.3) in terms of correlation with response (0) may, in fact, simply reflect the varying adequacy of the variables as measures of antecedent soil moisture conditions. It is not surprising, for instance, that days to the last major flood (1) ranks lowly; during the period between two major floods the amount of rainfall may very considerably and soil moisture conditions differ accordingly. One would, therefore, expect that an antecedent precipitation index (2) would rank more highly than the number of days to the last major flood. Whether the same type of argument can be advanced to explain the greater importance of preceding baseflow (3) in comparison to the antecedent precipitation index (2) is more dubious. Hewlett (1961) has pointed out that, in the steep mountainous terrain of the southern Appalachians, the major source of baseflow in between-flood periods is, in fact, subsurface seepage of moisture downslope through the soil mantle above the bedrock contact. In this situation, baseflow is presumably an almost perfect indicator of soil moisture conditions. On the other hand, in more subdued terrain, such as the Eaton basin, it seems likely that a large part of baseflow is derived from the main ground water system. In this type of situation, fluctuations in baseflow are not a simple index of soil moisture conditions; they must also reflect water table levels for the main ground water system. Although water table levels will, to some extent, indirectly correlate with soil moisture conditions, it is doubtful, in this situation, whether baseflow is a more accurate index of soil moisture conditions than is an antecedent rainfall parameter. If this argument is valid, one must turn elsewhere in an attempt to explain the higher correlation of response with preceding baseflow (3) than with the antecedent rainfall index (2) in Table 4.3.

One obvious point is that water table levels play a <u>direct</u> role in runoff production in the "partial area" model, so that stormflow should be expected to correlate well with water table levels, irrespective of moisture conditions in the soil mantle above the ground water system. This is true for both the variable source area model (Hewlett and Hibbert, 1967) and the more static partial area model presented by Ragan (1968), although the former model, as put forward by Hewlett, contains a certain amount of interdependence between water table levels and soil moisture conditions. The role of water table levels in the partial area model may be recalled by reference to Figures 2.6 and 2.7; as the water table gets nearer the surface (or further upslope if it is already at the surface). There is an increase in the area through which subsurface water emerges, and in the area of direct interception of rainfall by a water surface.

In the Eaton basin, it seems unlikely that direct throughbank seepage as depicted in Figure 2.6 is, on its own, sufficient to generate actual runoff amounts. This is deduced from the following simple equation:

$$S = (a/A)P + 2 \int_{t\ start}^{t\ end} (D_d) q_s dt \qquad (5.1)$$

where S is amount of storm flow in inches; a is area of perennial channel network; A is area of the catchment; P is storm precipitation in inches; t is a unit of time between the start and the end of the flood; D_d is perennial drainage density; and q_s is discharge of through-bank seepage per unit length of channel.

The first element on the right side of the equation is simply the amount of storm runoff due to direct interception of rainfall by the perennial channel system. The second is the total amount of through-bank seepage; as the equation shows, this is given by the product of throughbank seepage per unit length of channel (q_s), the amount of channel bank per unit area of the basin ($2D_d$), and the duration of the flood.

On the basis of maps at 1:50,000, Cartier and LeClerc (1964) report the drainage density of the Eaton Basin to be 1.24mi/mi^2; Map information indicates that (a/A) is about 0.0025. With a median figure for the individual storm data for storm rainfall equal to 1.36 inches, direct interception of rainfall by the perennial channel network alone would, therefore, produce only about 0.0034 inches. Actual runoff amounts in the 38 individual storms averaged 0.32 inches (maximum 1.90 inches; minimum 0.08 inches) using the Hewlett separation procedure and 0.38 inches (maximum 1.96; minimum 0.14 inches) using the Barnes method. Direct interception of rainfall by the perennial channel network is, therefore, two orders of magnitude smaller than total storm runoff, and can almost be neglected. Attention can now be focussed on the second term in Equation 5.1 and, by substituting for D_d and t, an estimate can be made of the amount of through-bank seepage necessary to produce actual runoff amounts. Median flood duration for the 38 individual storms was 44 or 78 hours, depending on the separation procedure. The calculations are as follows.

S = 0.32 inches (Hewlett); 0.38 inches (Barnes)

t = 44 hours (Hewlett); 78 hours (Barnes)

$$D_d = 1.24 mi/mi^2 = 1.24\ mi^{-1}$$

$$= 1.24/(5280 \times 12)\ in^{-1}$$

$$D_d = 0.00002\ in^{-1}$$

Therefore, converting the second element in Equation 5.1 to the form

$$S = 2(D_d)(q_s)t \qquad (5.2)$$

or

$$q_s = S/2(D_d)(t) \qquad (5.3)$$

and inserting the above values, we have

$$q_s = 0.32/2(0.00002)(44)$$

$$= 0.32/0.00176$$

$$= 190\ in^2/hr \quad \text{(Hewlett)},$$

and $$q_s = 0.38/0.00004(78)$$

$$= 0.38/0.00312$$

$$= 120\ in^2/hr \quad \text{(Barnes)}.$$

It is difficult to comment on these figures without some knowledge of either the velocity of subsurface seepage in the valley bottoms or the extra thickness of the bank through which this subsurface storm flow is seeping. Much of the till in the Eaton basin contains a high proportion of silt material (Table 3.1), and this is especially true of the valley bottom areas. It is likely therefore that the hydraulic conductivity of the channel bank material is rather low; Todd (1959), for instance, indicates that, for saturated seepage through silt soils, maximum values of hydraulic conductivity are about 3/4 inch per hour. The actual velocity of subsurface seepage depends on the hydraulic gradient as well as the hydraulic conductivity as given by the Darcy formula (Darcy, 1856).

$$v = ki \tag{5.4}$$

where v is velocity, k is the hydraulic conductivity of the soil, and i is the hydraulic gradient.

In view of the rather low hydraulic gradients that occur in valley bottom areas, it seems very unlikely that the velocity of subsurface seepage exceeds 0.1 in/hr. For the sake of comparison, one may note that Dunne and Black (1970b) measured subsurface velocities on the lower parts of hillslopes in similar terrain (but sandy soil) in Vermont of about 10 in/hr.

If one accepts these values of q_s equalling 120-190 in^2/hr and v equal to about 0.1 in/hr, this means that subsurface storm flow must emerge through a depth of 1200-1900 inches (100-150 feet) at the side of the stream channel. This, of course, is based on the assumption that subsurface flow contributes all the storm runoff in the typical Eaton flood. Clearly, this conclusion, as applied to the model depicted in Figure 2.6, is untenable. On the other hand, if one assumes that the water table rises sufficiently to reach the ground surface, the model depicted in Figure 2.7 is the more appropriate one to consider. The condition stated above is now no longer unrealistic; only a small rise in the water table would be necessary to extend the outcrop of the water table at the surface a distance of 100 feet from the channel side, especially in a floodplain area of gentle slope.

If it is assumed that the water-table does emerge at the ground surface then the effective amount of direct interception of rainfall by the surface water system (the true channel network and water table outcrop) will also increase. In addition, it is now unrealistic to assume that the true channel network itself will not expand. Tischendorf (1969) reported a three-fold increase in drainage density on a tiny basin in the southeastern Piedmont during a 4.14 inch storm. If this effect is accommodated into Equation 5.3, the average length of ground surface through which subsurface water must break out would now be reduced to only 30-50 feet, even before extra "channel interception" of rainfall had been included.

One must conclude therefore that the partial-area model of runoff production seems applicable to the Eaton basin. The relative importance of the headward expansion of the drainage net (Figure 2.5) as emphasized by Hewlett and Hibbert (1967) and the lateral expansion of the surface water system (Figure 2.7) as suggested by Dunne and Black (1970 a,b) cannot fairly be determined from the meagre data presently available. Moreover, it is probably very misleading to pose this particular problem as a simple combination of these two "models". Notwithstanding the plausibility of the reasoning just presented, certain evidence suggests that direct interception of rainfall by a surface water system is greater than given by a figure of about 0.25 percent of the basin area. Of course, it has already been noted that lateral and headward extension of the drainage network would increase this figure, although probably not by much. Tischendorf (1969), for instance, noted that at its maximum extension the drainage net in his area did not exceed one percent of the basin area.

One piece of evidence suggesting that direct interception of rainfall might be greater than this is the fluctuation in the dissolved solids concentration of channel water with changes in discharge. Many workers have noted that the concentration of dissolved solids in channel water decreases as discharge increases. Kunkle and Comer (1969) have noted this in a basin not unlike the Eaton's in Vermont, and Pesant (Ministère des Richesses Naturelles du Québec, In Correspondence) has reported it in the Eaton basin. It is difficult to explain this if the storm runoff is almost entirely subsurface seepage. The subsurface water that is displaced into the channel network during storms is essentially old water being moved out by entry of new moisture into the ground. Hewlett and Hibbert (1967) term this process "translatory flow"; they report experimental justification for the idea in the work of Horton and Hawkins (1965). This has important implications. The old water has had time to achieve chemical equilibrium, during the between-storm period, with its surroundings; its dissolved solids content should therefore be very similar to the concentration in base-flow. As a further corollary, subsurface storm runoff should produce little change in the dissolved solids concentration of channel flow during floods. Presumably this dilution of the channel water is due to incorporation of rainfall that has been intercepted directly by the surface water system.

Unfortunately there is very little data available to indicate the amount of dilution that takes place in a typical flood in the Eaton basin. There is sufficient, however, to indicate that the percentage amount of dilution of the dissolved solids load is appreciably greater than the percentage of the basin area taken up by the channel network. At this stage, one can add little more without venturing completely into the field of speculation, but one point seems important. Inspection of the 1:50,000 maps of the area indicate that between 3 and 12 percent of the basin area is designated "swamp or marsh". If surface water drains from the whole of this area into the channel network, as distinct from providing surface detention, then there exists a route via which rainfall directly intercepted by the swamp area can be conveyed from the basin as storm runoff in sufficient amounts to produce the observed dilution of dissolved solids in the stormflow. Clearly, this specific component of the general partial area model is one which merits serious attention in future studies of the hydrology of the Eaton basin.

APPENDIX I

PERCENTAGES OF BASIN AREA FOR THIESSEN POLYGONS, WITH MISSING PRECIPITATION DATA

East Angus Data Missing (4 cases)

Sawyerville	56.9%
La Patrie	21.2
St. Malo	15.6
Chartierville	6.3

St. Malo Data Missing (5 cases)

Sawyerville	62.2%
La Patrie	21.2
East Angus	8.4
Chartierville	8.2

Chartierville Data Missing (2 cases)

Sawyerville	48.6%
La Patrie	26.2
St. Malo	16.8
East Angus	8.4

Chartierville and St. Malo Data Missing (1 case)

Sawyerville	64.5%
La Patrie	27.1
East Angus	8.4

Chartierville, St. Malo and East Angus Data Missing (1 case)

Sawyerville	72.9%
La Patrie	27.1

APPENDIX II

INDIVIDUAL STORM DATA

Key to Tables on Following Pages:

A	stormflow (inches)
B	precipitation (inches)
C	Hydrologic response (dimensionless)
D	days since last major storm
E	preceding baseflow (cfs)
F1	antecedent precipitation index (used for response)
F2	antecedent precipitation index (used for stormflow)
G	largest one-day precipitation (inches)
H	average storm intensity (inches per day)
I	storm duration (days)
J	evapotranspiration index (°F-days above 32°F)
K	average maximum temperature

	A	B	C	D	E	F1
July 7, 1953	0.52	2.17	0.24	75	46.8	1.43
Nov. 26, 1953	0.16	0.64	0.25	27	272.0	0.72
Aug. 31, 1954	0.42	1.30	0.32	20	108.0	1.34
Sept. 12, 1954	0.85	2.34	0.36	10	271.0	1.83
Oct. 4, 1954	1.27	2.18	0.58	11	386.0	1.72
July 2, 1956	0.28	1.33	0.21	40	84.4	1.01
Oct. 7, 1956	0.21	0.89	0.24	34	171.0	1.19
Nov. 22, 1956	0.31	1.34	0.23	46	146.0	0.72
June 19, 1957	0.37	1.49	0.25	34	50.0	0.70
Sept. 23, 1957	0.22	1.63	0.13	70	81.1	1.17
Dec. 11, 1957	0.99	2.63	0.38	18	230.0	1.38
Oct. 17, 1958	0.52	1.29	0.40	6	186.0	1.04
June 9, 1959	0.12	0.80	0.15	18	129.0	1.22
June 15, 1959	1.30	2.43	0.54	6	143.0	1.46
June 29, 1959	1.90	2.76	0.69	13	129.0	1.80
Oct. 7, 1959	0.32	1.19	0.27	92	82.1	0.84
May 15, 1960	0.67	2.17	0.31	27	176.0	0.80
June 25, 1960	0.23	1.19	0.19	7	156.0	1.71
Oct. 25, 1960	0.22	1.01	0.22	42	156.0	1.12
June 2, 1961	0.35	1.81	0.19	4	325.0	1.78
June 30, 1961	0.11	0.69	0.16	16	110.0	1.90
Dec. 5, 1961	0.64	1.51	0.42	158	247.0	0.99

E	F1	F2	G	H	I	J	K
46.8	1.43	2.90	2.17	2.17	1	1232	76.0
272.0	0.72	1.63	0.64	0.64	1	100	42.0
108.0	1.34	4.00	1.30	1.30	1	1365	66.7
271.0	1.83	4.67	1.37	0.78	3	1740	57.8
386.0	1.72	4.83	1.29	0.44	5	1800	61.8
84.4	1.01	2.67	1.33	1.33	1	1360	77.0
171.0	1.19	2.67	0.89	0.89	1	660	62.5
146.0	0.72	1.00	1.04	0.67	2	260	40.3
50.0	0.70	2.01	1.47	0.75	2	1320	86.0
81.1	1.17	2.57	1.51	0.82	2	1260	67.0
230.0	1.38	3.47	1.37	0.53	5	135	28.3
186.0	1.04	1.60	1.04	0.43	3	858	58.3
129.0	1.22	2.83	0.54	0.40	2	765	83.0
143.0	1.46	3.23	1.85	0.81	3	1530	65.7
129.0	1.80	4.57	2.76	2.76	1	1980	76.0
82.1	0.84	1.47	1.08	0.59	2	638	61.0
176.0	0.80	1.99	1.80	0.72	3	728	58.5
156.0	1.71	3.36	1.05	0.40	3	1260	73.0
156.0	1.12	2.01	0.76	0.34	3	308	44.5
325.0	1.78	3.76	0.85	0.36	5	2475	65.4
110.0	1.90	4.56	0.69	0.69	1	574	73.0
247.0	0.99	1.57	0.61	0.50	3	126	39.0

	A	B	C	D	E	F1
June 11, 1962	0.08	1.54	0.05	21	59.8	0.88
Aug. 8, 1962	0.12	1.00	0.12	20	115.0	2.30
Aug. 21, 1962	0.16	1.04	0.15	14	101.0	1.57
Oct. 7, 1962	0.36	2.26	0.16	46	61.4	0.44
June 21, 1963	0.12	1.88	0.06	37	47.1	0.70
July 3, 1963	0.16	1.91	0.08	12	42.7	0.99
July 9, 1963	0.53	1.65	0.32	6	171.0	2.08
Aug. 14, 1963	0.28	1.42	0.20	9	196.0	2.04
Aug. 24, 1963	1.02	2.11	0.48	9	137.0	1.97
Nov. 19, 1963	0.31	0.93	0.33	11	320.0	1.20
Nov. 20, 1964	0.11	0.74	0.15	121	184.0	1.14
Nov. 27, 1964	0.46	0.99	0.46	6	200.0	1.15
Oct. 7,1965	0.56	0.82	0.68	32	670.0	2.37
June 25, 1966	0.11	1.40	0.08	32	56.2	1.03
Sept. 23, 1966	1.91	0.08	0.08	32	48.1	0.75
Oct. 20, 1966	0.36	1.34	0.27	32	176.0	1.15

APPENDIX II.1

DATA FOR HEWLETT SEPAI

E	F1	F2	G	H	I	J	K
59.8	0.88	2.23	1.12	0.77	2	1290	75.0
115.0	2.30	5.62	0.94	0.33	3	910	67.0
101.0	1.57	4.69	0.78	0.52	2	1040	71.5
61.4	0.44	1.50	1.76	0.75	3	759	54.7
47.1	0.70	1.30	1.23	0.63	3	840	60.0
42.7	0.99	2.39	1.86	0.96	2	532	70.0
171.0	2.08	4.20	1.21	0.83	2	896	64.0
196.0	2.04	3.62	1.11	0.47	3	780	61.5
137.0	1.97	4.79	1.39	1.06	2	728	60.0
320.0	1.20	2.26	0.65	0.47	2	280	46.5
184.0	1.14	2.01	0.66	0.25	3	200	42.0
200.0	1.15	2.04	0.66	0.50	2	220	42.0
670.0	2.37	4.79	0.65	0.41	2	319	61.0
56.2	1.03	2.06	1.24	0.47	3	1290	75.5
48.1	0.75	2.06	0.89	0.64	3	576	56.5
176.0	1.15	3.11	0.83	0.67	2	374	49.0

APPENDIX II.1

FOR HEWLETT SEPARATION

	A	B	C	D	E	F
July 7, 1953	0.62	2.17	0.29	75	46.8	0.[illegible]
Nov. 26, 1953	0.23	0.64	0.36	27	272.0	0.[illegible]
Aug. 31, 1954	0.49	1.30	0.38	20	108.0	0.[illegible]
Sept. 12, 1954	0.89	2.34	0.38	10	271.0	1.[illegible]
Oct. 4, 1954	1.19	2.57	0.46	11	386.0	0.[illegible]
July 2, 1956	0.36	1.33	0.27	40	84.4	0.[illegible]
Oct. 7, 1956	0.27	0.89	0.30	34	171.0	0.[illegible]
Nov. 22, 1956	0.37	1.34	0.28	46	146.0	0.[illegible]
June 19, 1957	0.46	1.49	0.31	34	50.0	0.[illegible]
Sept. 23, 1957	0.32	1.74	0.18	70	81.1	0.[illegible]
Dec. 11, 1957	0.57	2.15	0.27	18	230.0	0.[illegible]
Oct. 17, 1958	0.65	1.29	0.50	6	186.0	0.[illegible]
June 9, 1959	0.18	0.87	0.21	18	129.0	0.[illegible]
June 15, 1959	1.33	2.46	0.54	6	143.0	0.[illegible]
June 29, 1970	1.96	2.95	0.66	13	129.0	0.[illegible]
Oct. 7, 1959	0.43	1.19	0.36	92	82.1	0.[illegible]
May 15, 1960	0.81	2.17	0.37	27	176.0	0.[illegible]
June 25, 1960	0.35	1.19	0.29	7	156.0	1.[illegible]
Oct. 25, 1960	0.38	1.01	0.38	42	156.0	0.[illegible]
June 2, 1961	0.29	0.91	0.32	4	325.0	1.[illegible]
June 30, 1961	0.14	0.69	0.20	16	110.0	1.[illegible]
Dec. 5, 1961	0.73	2.06	0.35	158	247.0	0.[illegible]

E	F1	F2	G	H	I	J	K
46.8	0.88	2.54	2.17	2.17	1	1232	76.0
272.0	0.50	1.38	0.64	0.64	1	100	42.0
108.0	0.58	3.31	1.30	1.30	1	1365	66.7
271.0	1.02	3.92	1.37	0.78	3	1740	57.8
386.0	0.78	4.04	1.29	0.43	6	1800	61.8
84.4	0.55	2.22	1.33	1.33	1	1360	77.0
171.0	0.70	2.31	0.89	0.89	1	660	62.5
146.0	0.55	0.95	1.04	0.67	2	260	40.3
50.0	0.33	1.67	1.47	0.75	2	1320	86.0
81.1	0.59	2.19	1.51	0.44	4	1260	67.0
230.0	0.71	2.95	1.37	0.54	4	135	28.3
186.0	0.68	1.51	1.04	0.43	3	858	58.3
129.0	0.64	2.44	0.54	0.29	3	765	83.0
143.0	0.87	2.80	1.85	0.62	4	1530	65.7
129.0	0.88	3.89	2.76	1.48	2	1980	76.0
82.1	0.56	1.32	1.08	0.60	2	638	61.0
176.0	0.42	1.69	1.80	0.72	3	728	58.5
156.0	1.05	2.98	1.05	0.40	3	1260	73.0
156.0	0.63	1.77	0.76	0.34	3	308	44.5
325.0	1.13	3.25	0.85	0.23	4	2475	65.4
110.0	1.04	3.90	0.69	0.69	1	574	73.0
247.0	0.71	1.45	0.61	0.41	5	126	39.0

	A	B	C	D	E	F
June 11, 1962	0.17	1.54	0.11	21	59.8	0.4
Aug. 8, 1962	0.20	1.47	0.14	20	115.0	1.1
Aug. 21, 1962	0.24	1.04	0.23	14	101.0	0.7
Oct. 7, 1962	0.46	2.26	0.20	46	61.4	0.1
June 21, 1963	0.23	1.88	0.12	37	47.1	0.4
July 3, 1963	0.21	2.02	0.10	12	42.7	0.4
July 9, 1963	0.67	1.65	0.41	6	171.0	1.3
Aug. 14, 1963	0.42	1.42	0.30	9	196.0	1.2
Aug. 24, 1963	1.08	2.11	0.51	9	137.0	1.0
Nov. 19, 1963	0.35	0.93	0.38	11	320.0	0.7
Nov. 20, 1964	0.16	0.74	0.22	121	184.0	0.7
Nov. 27, 1964	0.54	1.10	0.49	6	200.0	0.7
Oct. 7, 1965	0.61	0.82	0.74	32	670	1.4
June 25, 1966	0.20	1.40	0.14	32	56.2	0.5
Sept. 23, 1966	0.32	1.91	0.17	32	48.1	0.2
Oct. 20, 1966	0.36	1.34	0.27	32	176.0	0.5

APPENDIX II.2

DATA FOR BARNES SEPARATI

E	F1	F2	G	H	I	J	K
59.8	0.42	1.90	1.12	0.77	2	1290	75.0
115.0	1.17	4.81	0.94	0.37	4	910	67.0
101.0	0.71	3.86	0.78	0.52	2	1040	71.5
61.4	0.18	1.21	1.76	0.75	3	759	54.7
47.1	0.45	1.16	1.23	0.63	3	840	60.0
42.7	0.49	2.06	1.86	0.51	4	532	70.0
171.0	1.32	3.68	1.21	0.83	2	896	64.0
196.0	1.29	3.28	1.11	0.47	3	780	61.5
137.0	1.00	4.11	1.39	1.06	2	728	60.0
320.0	0.70	2.10	0.65	0.47	2	280	46.5
184.0	0.71	1.76	0.66	0.25	3	200	42.0
200.0	0.72	1.85	0.66	0.37	3	220	42.0
670	1.41	4.23	0.65	0.41	2	319	61.0
56.2	0.58	1.83	1.24	0.47	3	1290	75.5
48.1	0.29	1.71	0.89	0.64	3	576	56.5
176.0	0.59	2.59	0.83	0.67	2	374	49.0

PPENDIX II.2

BARNES SEPARATION

APPENDIX III

YEARLY (SNOW-FREE SEASON) DATA

Year	1	2	3	4	5	6	7	8	9
1933	0.87	1.67	17.81	0.05	0.09	7	34	0.28	168.6
1934	2.30	3.10	21.32	0.11	0.15	8	35	0.26	296.0
1935	2.87	4.07	23.87	0.12	0.17	12	42	0.28	372.7
1936	7.85	9.55	31.89	0.25	0.30	17	54	0.25	376.3
1937	5.35	7.05	30.32	0.18	0.23	17	44	0.30	246.7
1938	3.02	4.52	31.17	0.10	0.14	15	39	0.38	240.2
1939	4.28	5.28	27.47	0.17	0.22	10	36	0.38	185.1
1940	2.44	3.14	28.12	0.09	0.11	7	42	0.33	156.4
1941	2.43	3.33	23.85	0.10	0.14	9	41	0.27	128.9
1942	5.75	7.05	28.65	0.20	0.25	13	44	0.30	278.7
1943	5.32	6.52	25.33	0.21	0.26	12	41	0.34	239.2
1944	2.58	3.28	22.25	0.12	0.15	7	33	0.38	153.9
1945	7.44	9.14	30.56	0.24	0.30	17	40	0.34	388.1
1946	2.89	4.59	25.78	0.11	0.18	17	42	0.28	292.1
1947	3.54	4.54	25.88	0.14	0.18	10	31	0.45	352.1
1948	4.09	5.39	25.56	0.16	0.21	13	36	0.28	326.1
1949	0.98	1.78	22.35	0.04	0.08	8	42	0.26	294.1
1950	1.10	2.00	23.68	0.05	0.09	9	48	0.24	125.3
1951	0.96	1.66	17.35	0.06	0.10	7	44	0.21	154.6
1952	1.79	2.49	24.06	0.07	0.10	7	40	0.31	221.2
1953	2.24	3.34	28.31	0.08	0.12	11	38	0.29	419.1
1954	7.32	9.12	38.74	0.19	0.24	18	38	0.47	411.4
1955	1.50	2.00	24.26	0.06	0.08	5	37	0.30	272.1
1956	1.09	1.79	21.86	0.05	0.08	7	41	0.27	230.0
1957	2.94	4.34	30.42	0.10	0.14	14	40	0.33	262.3
1958	1.40	2.30	26.35	0.05	0.09	9	42	0.29	167.6
1959	5.64	7.04	26.40	0.21	0.27	14	42	0.31	242.2
1960	1.67	2.37	26.72	0.06	0.09	7	34	0.34	127.6
1961	1.45	2.65	20.69	0.07	0.13	12	32	0.36	351.5
1962	1.54	2.44	22.60	0.07	0.09	9	44	0.30	176.3
1963	4.61	6.71	24.96	0.18	0.27	21	41	0.29	329.1
1964	0.86	1.36	20.09	0.04	0.07	5	41	0.21	252.0

1 Stormflow - Hewlett Separation
2 Stormflow - Barnes Separation
3 Precipitation
4 Response - Hewlett
5 Response - Barnes
6 Number of Storms
7 Days with rain (%)
8 Rain per rainy day
9 Average annual minimum preceding discharge

BIBLIOGRAPHY

THE EATON BASIN AND SURROUNDING AREA

Bowker, G.L., 1968. An analysis of the first five years of climatological records at Sawyerville-North (International Hydrologic Decade Station). Feuillet Meteorologique, 7, 238-249.

Braun, E.L., 1950. Deciduous Forests of Eastern North America. Hafner, New York.

Cartier, L. and A. LeClerc, 1964. Rivière Eaton: Caracteristiques Topographiques du Bassin Versant. Ministère des Richesses Naturelles du Québec Publication HP-5.

Ferland, M., 1967. Rivière Eaton: Données Météorologiques, 1965. Service de Météorologie , Ministère des Richesses Naturelles du Québec, 89pp.

Kunkle, S.H. and G.H. Comer, 1969. Suspended, Bed and Dissolved Loads in the Sleepers River, Vermont, U.S.A. Mimeographed. Sleepers River Research Watershed, Northeast Branch, Soil and Water Conservation Research Division, Agricultural Research Service, U.S. Department of Agriculture.

McDonald, B.C., 1967. Pleistocene Events and Chronology in the Appalachian Region of South-Eastern Quebec, Canada. Unpublished Ph.D. dissertation, Yale University, 161 pp.

McDonald, B.C., 1969. Surficial Geology of LaPatrie - Sherbrooke Area, Quebec, Including Eaton River Watershed. Geological Survey of Canada Paper No. 67-52.

Ministère des Richesses Naturelles du Québec, 1967. Rivière Eaton: Debits Journaliers 1932-1964. Service d'Hydrometrie, Ministère des Richesses Naturelles du Québec, 53 pp.

GENERAL HYDROLOGY

Amorocho, J. and W.E. Hart, 1964. A Critique of Current Methods in Hydrologic Systems Investigation. Transactions of the American Geophysical Union, 45, pp. 307-321.

Bruce, J.P., and R.H. Clark, 1966. Introduction to Hydrometeorology. Pergamon Press.

Chow, V.T. (Editor), (1964), Handbook of Applied Hydrology, McGraw-Hill, New York, 1418 pp.

Darcy, H., 1856. Les Fontaines Publiques de la Ville de Dijon. V. Dalmont, Paris.

Linsley, R.K., Kohler, M.A. and J.L.H. Paulhus, 1958. Hydrology for Engineers, McGraw-Hill Co., New York, 340 pp.

Meinzer, O.E. (ed.), 1942. Hydrology, Physics of the Earth Series, vol. 9. McGraw-Hill Book Company, New York 728 pp.

Thiessen, A.H., 1911. Precipitation for Large Areas. Monthly Weather Review, 39, 1082-1084.

Todd, D.K., 1959. Ground Water Hydrology. John Wiley, New York.

MECHANISMS OF RUNOFF PRODUCTION

Amerman, C.R., 1965. The use of unit-source watershed data for runoff prediction. Water Resources Research, 1, 499-508.

Barnes, B.S., 1939. The Structure of Discharge-Recession Curves. Transactions of the American Geophysical Union, 20, pp. 721-725.

Benson, M.A., Factors influencing the occurrence of floods in a humid region of diverse terrain, U.S. Geological Survey Water Supply Paper 1580-B, 1962.

Betson, R.P., 1964. What is Watershed Runoff? Journal of Geophysical Research, 69, pp. 1541-1552.

Betson, R.P., Marius, J.B. and R.T. Joyce, 1968. Detection of Saturated Interflow in Soils with Piezometers. Proceedings of the Soil Science Society of America, 32.

Bodman, G.B., and E.A. Colman, 1943. Moisture and Energy Conditions during Downward Entry of Water into Soils. Proceedings of the Soil Science Society of America, 8, pp. 116-122.

Dunne, T., and R.D. Black, 1970a. An experimental investigation of runoff production in permeable soils. Water Resources Research, In Press.

Dunne, T., and R.D. Black, 1970b. Partial-area contributions to storm runoff in a small New England watershed. Water Resources Research, In Press.

Hack, J.T., and J.C. Goodlett, 1960. Geomorphology and forest ecology of a mountain region in the Central Appalachians, U.S. Geological Survey Professional Paper.

Hansen, V.E. (1955). Infiltration and soilwater movement during irrigation. Soil Science, 79, 93-105.

Hewlett, J.D., 1961. Soil Moisture as a Source of Base Flow from Steep Mountain Watersheds. Southeastern Forest Experimental Station Paper No. 132, 11 pp.

Hewlett, J.D., and A.R. Hibbert, 1963. Moisture and energy conditions within a sloping soil mass during drainage. Journal of Geophysical Research, 68, 1081-1087.

Hewlett, J.D., 1967. A Hydrologic Response Map for the State of Georgia, Water Resources Bulletin, 3, pp. 4-20.

Hewlett, J.D., and A.R. Hibbert, 1967. Factors Affecting the Response of Small Watersheds to Precipitation in Humid Areas. Proceedings of the International Symposium on Forest Hydrology September 4-18, 1965 (Pennsylvania State University), pp.275-290. Pergamon Press.

Hewlett, J.D., 1969. Tracing storm and base flow to variable source areas on forested headwaters. Completion Report, OWRR Project A-114-Ga.

Hoover, M.D., and C.R.Hursh, 1943. Influence of topography and soil depth on runoff from forest land. American Geophysical Union, Transactions, 24, 692-698.

Horton, J.H., and R.H. Hawkins, 1965. Flow path of rain from the soil surface to the water table. Soil Science, 100, 377-383.

Horton, R.E., 1933. The role of infiltration in the hydrologic cycle. American Geophysical Union, Transactions, 14, 446-460.

Horton, R.E. (1940). An Approach toward a physical interpretation of infiltration capacity. Soil Science Society Proceedings, 5, 399-417.

Kursh, C.R., 1936. Storm-water and absorption. In Discussion on list of terms with definitions, report of the Committee on Absorption and Transpiration. American Geophysical Union, Transactions, 17, 301-302.

Kirkby, M.J. and R.J. Chorley, 1967. Throughflow, Overland Flow and Erosion. Bulletin of the International Association of Scientific Hydrology, 12, pp. 5-21.

Kirkby, M.J., 1969. Infiltration, Throughflow and Overland Flow, in Water, Earth, and Man (Edited by R.J. Chorley), Methuen.

Lee, J., and D.I. Bray, 1969. The estimation of runoff from rainfall for New Brunswick watersheds. Journal of Hydrology, 9, 427-437.

Nixon, P.R., and G.P. Lawless, 1960. Translocation of moisture with time in unsaturated soil profiles. Journal of Geophysical Research, 65, 655-661.

Ragan, R.M., 1968. An Experimental Investigation of Partial Area Contributions. Proceedings of the Fourteenth General Assembly (1967), International Union of Geodesy and Geophysics (Bern, Switzerland), Publication No. 76, 241-249.

Richards, L.A. and S.J. Richards, 1957. Soil Moisture, In Yearbook of Agricultural Soil, pp. 49-60.

Schumm, S.A. and G.C. Lusby, 1963. Seasonal Variation of Infiltration Capacity and Runoff on Hillslopes in Western Colorado. Journal of Geophysical Research, 68, pp. 3655-3666.

Swartzendruber, D., 1960. Water flow through a soilprofile as affected by the least permeable layer. Journal of Geophysical Research, 65, 4037-4042.

Tischendorf, W.G., 1969. Tracing Stormflow to Varying Source Areas in a Small, Forested Watershed in the Southeastern Piedmont. Unpublished Ph.D. thesis, University of Georgia.

Whipkey, R.Z., 1965. Subsurface Stormflow from Forested Slopes. Bulletin of the International Association of Scientific Hydrology, 10, 74-85.

Woodruff, J.F. and J.D. Hewlett, 1969. Determining and mapping the average hydrologic response of Eastern United States. Completion Report, OWRR Project A-013-Ga.

STATISTICAL METHODS IN HYDROLOGY

Beale, E.M.L., Kendall, M.G., and D.W. Mann, 1967. The discarding of variables in multivariate analysis. Unpublished manuscript.

Carson, M.A., 1968. Some problems with the use of correlation techniques in morphometric studies, Occasional Paper No. 4 of British Geomorphological Research Group, pp. 49-67.

Carson, M.A. 1969. On the use of models in geography. Geografisch Tijdschrift, 3, 432-443 .

Ezekiel, M., and K.A. Fox, 1959. Methods of Correlation and Regression Analysis. John Wiley, New York, 548 pp.

Krumbein, W.C., Benson, B. and W.B. Hempkins, ~~1964.~~ Whirlpool, A Computer Program for Sorting Out Independent Variables by Sequential Multiple Regression. Office Naval Research Technical Report 14, ONR Task No. 389-178, Northwestern University.

Krumbein, W.C. and F.A. Graybill, 1965. Statistical Models in Geology, McGraw-Hill, 475 pp.

Snyder, W.M., 1962. Some possibilities for multivariate analysis in hydrologic studies. Journal of Geophysical Research, 67, 721-729.