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PRECIPITATION-RUNOFF CHARACTERISTICS OF THE CARIBOU CREEK RESEARCH WATERSHED NEAR FAIRBANKS, ALASKA

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THESIS

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PRECIPITATION-RUNOFF CHARACTERISTICS OF THE CARIBOU CREEK

RESEARCH WATERSHED NEAR FAIRBANKS, ALASKA

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ABSTRACT

Precipitation and runoff data for the Caribou Creek Watershed near Fairbanks, Alaska for the summers of 1970 and 1971 were analysed. Runoff during the 1971 summer season was much higher than that in 1970 because of a larger contribution from spring snowmelt in 1971.

The runoff process was found to be fairly stable. Storm runoff was linearly related to total precipitation amount. The response time of the hydrograph to storm rainfall, however, was not related to antecedent basin wetness. This indicated that runoff generation possibly was from a fairly stable source area.

Recessions were well modeled by the exponential decay relationship given by,

$$q_t = q_0 e^{-kt}$$

where q_0 and q_t are peak discharge and discharge t time units later, respectively, and k is the slope of the line when plotted on log-normal paper. With a regression equation for the time of rise of the hydrograph as related to total precipitation amount the hydrograph peak could be predicted by the following,

$$q_p = RO_s (0.5 T_r + k^{-1})^{-1}$$

where q_p is the peak flow, RO_s is storm runoff, T_r is the time of rise. Total storm hydrographs could be drawn by positioning the peak with regression equations relating time of rise and lag time to precipitation and assuming a linear increase in flow from the point of rise to the peak of the hydrograph. The recession can be calculated from the exponential decay relationship utilizing an appropriate k value.

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INTRODUCTION

Water is one of the most abundant compounds found on the earth. Because it is both a powerful life-sustaining force and an agent of destruction, it is imperative that its motions, loss, and recharge be understood. The global hydrologic cycle, while of great importance, involves the large scale movement of water. Of more practical immediate value is research in individual watersheds. Here, on a small scale basis, it is possible to study water relations from initial input to the exit from the local scene. In the face of larger demand for water in the future and a basically constant supply, the water relations of drainage basins that supply inhabited areas become of overwhelming importance.

The research watershed supplies a field laboratory where the natural processes of precipitation, evapotranspiration, detention, and transport of water to and through the channel can be studied. Hundreds of studies have been made on the various processes affecting water on the local level in many research watersheds throughout the United States. Sopper and Lull (1970) carried out extensive research on the hydrologic response of basins in the Eastern United States and the various factors which affect it. Several studies have focused on the sources of storm runoff (Betson, 1964; Hewlett and Hibbert, 1967; Dunne and Black, 1970a). Studies concerning groundwater relations and supply are also abundant in the literature (Urie, 1967; Black, 1970).

The runoff process is of particular interest from the standpoint of land management and water supply. Runoff can be violently destructive not only to the land surface but to human habitation as well. Rapid runoff from large rainstorms, then, must be modified or at least predicted adequately in order to protect human life and property. From a knowledge of the runoff phenomena of a region, gained by research at the watershed level, hydraulic structures and land management practices can be designed to lessen the threat of flood waters. Water for the carrying on of everyday human life is dependent on adequate reserves. Rapid storm runoff can be thought of as a loss to potential storage. Therefore, the timing and amount of streamflow from a given rainfall input becomes a problem of deep concern in arid areas where both input and storage are minimal and increasingly, in humid areas as populations increase.

The amount of runoff derived directly from storms is affected by the amount and intensity of storm precipitation (Osborn and Lane, 1969; Schreiber and Kincaid, 1967). The surface conditions of the basin play a large role in the partitioning of input into overland flow, interflow, and groundwater reserves and suggest several management alternatives including deforestation and reforestation (McGuiness and Harrold, 1971; Nakano, 1967; and Comer and Zimmerman, 1969). In the forefront, where storm runoff is concerned, has been the infiltration capacity concept developed by Horton (1933) which describes runoff as a function of rainfall excess and is dependent on the capacity of the soil to absorb rainwater. For heavily forested basins, however, this approach has been found to be somewhat unrealistic. In fact, the runoff process is so complex that a full description of it may be years in coming. Until then, largely empirical observation and quantification must be relied upon.

An understanding of the basic characteristics of water supply relations in a region is fundamental to management and may be the outstanding mission of workers in the research watershed. In interior Alaska, little is known about the runoff cycle. Since large destructive floods are known to occur, insight into the processes of precipitation-runoff must be gained. Up to the present, detailed studies of precipitation-runoff on interior basins have been lacking. The most complete was a study of the Glen Creek watershed near Fairbanks (Dingman, 1970).

An abundance of knowledge exists on the hydrologic behavior of basins in the contiguous States, but, as noted, very little has been written of interior Alaska watersheds. For this reason, information deduced in other parts of the country must be used as a starting point and a working knowledge of the similarities to and differences from Alaskan processes developed. In an effort to further the knowledge of water relations in the interior of Alaska, the Caribou-Poker Creeks research watershed was established in 1969. The hope was that models applicable to other ungaged basins in the region could be developed and that an understanding of the processes affecting precipitation-runoff, and the other phases of the hydrologic cycle, could be gained.

In this study, precipitation-runoff relationships were investigated on the Caribou Creek watershed. Indications were that watersheds of interior Alaska had properties much different from their counterparts in the humid and sub-humid States. Recessions were found to be considerably longer and the response times slower than basins of the same size in the conterminous states. The storage potential was found to be very high (Dingman, 1970). It was hoped that this study would yield further insight into these relationships and the precipitation-runoff relationship of an interior Alaska basin.

LITERATURE REVIEW

The investigation of the hydrologic response of a drainage basin requires that several topics be examined in relation to one another. These include factors from not only hydrology, but meteorology and fluid physics as well. From a practical standpoint, many of the related areas of interest must be left as residual to the more important or discernible aspects of the problem. This review of literature presents discussions on the hydrologic cycle, precipitation, runoff, hydrograph separation, and precipitation-runoff relationships. These are considered fundamental processes in this precipitation-runoff study.

The Hydrologic Cycle. The waters of the earth may be distributed into the two major categories of fresh and salt water. Salt water comprises 3.17×10^8 cubic miles of 97.2 per cent of all the water on the earth. The remaining 9.1×10^6 cubic miles or 2.8 per cent is partitioned as follows: 2.15 per cent locked in polar ice and snow, 0.625 per cent as subsurface water, 0.017 per cent in inland surface waters and 0.001 per cent in the atmosphere (Leopold and Davis, 1969). Clearly, watershed hydrology deals with a very small portion of the world water supply (Hewlett and Nutter, 1969).

On a global basis the annual hydrologic cycle processes may be represented numerically (billions of acre-feet) as follows: 1) 327 as evaporation from the oceans, 2) 53 as evaporation from the land surface, 3) 300 as precipitation back to the oceans, 4) 80 as precipitation onto land surface, and 5) 27 as river discharge to oceans (Hewlett and Nutter, 1969).

Over a continental land mass such as North America, the various processes making up the hydrologic cycle can be represented by the basic input-output equation

$$Pg = ET + Q + \Delta S$$

where Pg is gross precipitation, ET is evapotranspiration, Q is streamflow, and S is change in storage. The latter term approaches zero over a long time period; consequently, for the United States the general magnitude of the components are: Pg = 30, ET = 21, Q = 9, and S = 0 inches (Hewlett and Nutter, 1969).

On a drainage basin the hydrologic cycle can begin with precipitation. The delivery characteristics of precipitation will govern, in part, how much may be intercepted by surface materials or may reach the soil. The intercepted precipitation may be absorbed, evaporated, or dripped to the soil. Precipitation reaching the soil will comprise runoff over the land surface directly as surface runoff (overland flow), be detained in depression storage, and/or be infiltrated into the ground. Infiltrated water may percolate to the water table or seep laterally beneath the soil surface as interflow. Both interflow and surface runoff reach the stream channel rapidly and form that component of streamflow known as stormflow. Ground water enters the channel much more slowly and composes that part of streamflow known as ground water flow or baseflow (Bruce and Clark, 1966; Linsley, Kohler, and Paulhus, 1958). Crawford and Linsley's (1964) flow diagram developed for computer applications (figure 1), follows this general pattern. It should be noted here that each successive

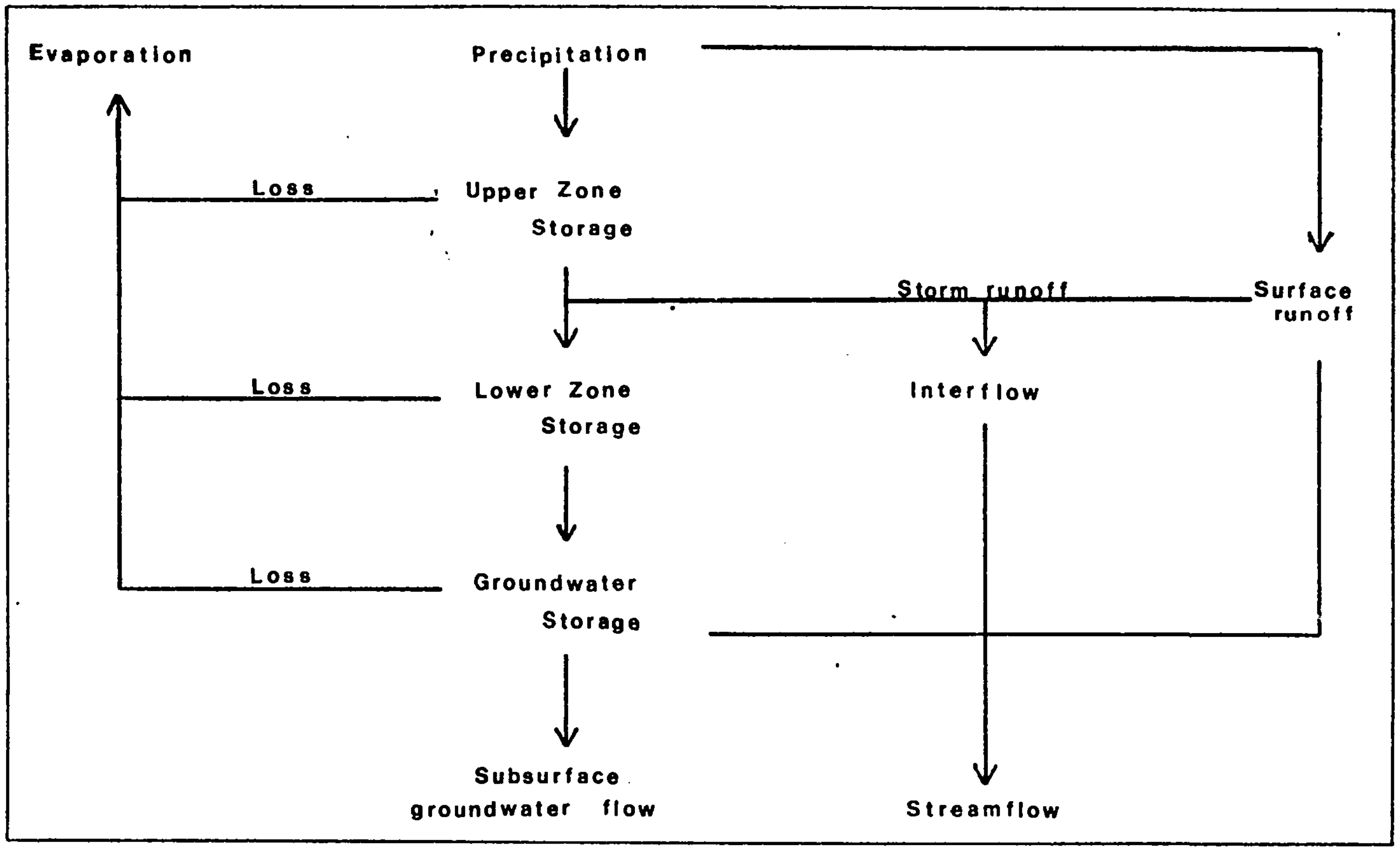

Figure 1

Flow diagram from the hydrologic cycle on the watershed level (Crawford and Linsley, 1964) step leading through a stream discharge experiences a loss to evaporation.

The hydrologic cycle other than on a global basis becomes an academic matter. Of interest to those delving into water relations problems are the extreme events of flood and drought or specific components such as precipitation and runoff (Linsley, Kohler, and Paulhus, 1958). These factors will be discussed in more detail in the following sections.

Precipitation. Precipitation in solid or liquid form is the result of condensation or sublimation of atmospheric water vapor (Flohn, 1969). The water vapor component of the hydrologic cycle is derived from transpiration from vegetation and evaporation from wet surfaces. Oceans and luxuriant vegetation provide much atmospheric water vapor, while often in continental areas large lakes heated by the summer sun become local sources of water vapor (Mackay, 1970).

The actual processes involved in precipitation are complex and poorly understood. Before precipitation can take place condensation, generally associated with ascending air masses, must first occur (Critchfield, 1966). As air ascends it cools adiabatically and its ability to hold water vapor decreases. Consequently, excess water vapor condenses and forms tiny cloud droplets (Critchfield, 1966). The droplets are so small that they would evaporate long before reaching the ground when falling through unsaturated air below the condensation level (cloud base). To produce drizzle from clouds composed mostly of water droplets the depth of the rising air layer must be several hundred meters thick and to produce rain the depth must be at least one kilometer. Therefore, the formation of raindrops is primarily by coalescence (Mason, 1962).

There are several mechanisms by which air is uplifted and subsequently cooled. Horizontal convergence, where an inflow of air concentrates in a low pressure center, causes air to rise. Orographic uplift occurs where topographic barriers force air to rise over them. Convective uplift results from surface heating and the warmed air will often rise to tens-of-thousands feet. Finally, frontal uplift occurs when an advancing air mass causes the intruded air to rise over it, resulting in widespread lifting and general rains over vast areas (Mackay, 1970).

Each of the precipitation producing mechanisms listed above has its own characteristic type of rainfall. Horizontal convergence and weather fronts produce the most widespread rainfall events, due to extensive air mass involvement. Warm fronts, generally of greatest areal extent, are often characterized by drizzle because air is lifted gently over a long period of time, precluding the possibility of the formation of giant raindrops; whereas, cold fronts, resulting in an abrupt uplift, are characterized by precipitation of shorter duration and greater intensity. Fronts in general involve large areas and tend to display an orderly pattern of precipitation, giving relatively uniform amounts over wide areas (Flohn, 1969).

Orographic enhancement of rainfall is a function of the degree of slope of the land, elevation, and moisture content of the air mass. Exposed slopes generally show greatest increases in precipitation with elevation (Mackay, 1970).

Convective storms are of relatively small dimensions and are not long lived. These, however, are the major source of summer rainfall for continental climates (Mackay, 1970).

Point rainfall variations occur for many reasons. Orographic enhancement of precipitation has been shown to occur. Studies in Oregon and Washington noted an increase in annual precipitation with an increase in elevation for the Olympic Cascades and coastal ranges (Schermerhorn, 1967). The same has been found to be the case in southeastern Alaska. Rain gages placed at 100 foot intervals from sea level to 1100 feet showed an average 2 per cent increase in rainfall with every 100 foot rise in elevation (Walkotten and Patric, 1967). On dissected terrain in southwestern Wisconsin there also was an apparent increase in annual precipitation between valley bottom stations at 800 to 1000 feet and ridge top stations some 200 to 300 feet higher (Whitson and Baker, 1912).

There have been studies which indicate orographic influences to be of lesser importance in some areas. Sartz (1966), working in the same area of Whitson and Baker (1912) found no appreciable differences in valley bottom and ridge precipitation amounts. However, ridge gages did catch more precipitation than valley gages. Between the valley bottom and the top of the highest ridge there was a 2 per cent increase in precipitation amount, which was attributed to normal storm variation. Hendrick and Comer (1970), working in Northeastern

United States, also found no significant correlation between elevation and precipitation amount. They noted variability with storm size and direction of movement. The reason for the lack of topographic influence is unclear, but may be related to less readily discernible factors such as terrain, wind velocity, and aerodynamic characteristics of the rain gages.

The physical factors which account for the variation in point rainfall catchment from gage to gage over a small area are of great importance in network design. Where relief is rugged with large sharp boulders or high mounds dominating, turbulence and eddies may markedly change the rainfall recorded between gages in the same general area (Helmers, 1954; Hamilton, 1954). A gage placed between mounds or on the leeward side of large rocks will tend to collect more rain because of water carried over the barrier by eddy currents and dropped on the other side in the relative calm. On a larger scale, gages on exposed hillsides, especially are subject to the same type of phenomenon. Gages on the windward side of ridges and mountaintops collect less rain than those on the leeward side, according to Hovind (1965). He found that the distribution of rainfall on a windy peak in California was markedly asymmetric. The magnitude of difference and asymmetry increased with increasing wind speed. The catchment on the windward slopes under high winds was as little as one third that of the leeward side, and true distribution of rainfall could not be determined with unshielded gages. Sartz (1966) found very little difference between catchment on the windward and leeward sides, noting only a 3 per cent decrease on the windward. This however, may be due to the fact that wind velocities were not high enough to account for a greater variation. As the wind speed decreased, the difference between windward and leeward decreased likewise (Hovind, 1965).

The effect of wind is not restricted to mountain areas. Flat smooth terrain will have wind-induced errors due to lack of surface roughness. Brooks (1938) found a 75 per cent decrease in rainfall catchment due to high winds on relatively flat terrain. This has been widely noted in the literature by such reviewers as Sartz (1966) and Helmers (1954).

The problem on low relief areas and a compounding factor in the mountains is turbulence and eddies created by the edges of the rain gage itself. Some of the precipitation that should fall into the gage is carried up and over it by wind currents (Helmers, 1954; Brooks, 1938). Merely tilting the gage in the direction of the wind has been seen to increase catch by up to 21 per cent (Storey and Hamilton, 1943). However, for a truly representative measure of rainfall, a method to eliminate the eddy effect is needed. As early as 1878, Nipher designed and experimented with various types of shields. His most successful shield is in wide use and consists of a trumpet-like extension up from the orifice of the rain gage with a flat rim. While this has proven to be of value for rain, snow is a more difficult matter. The measurement of snow, especially in windy areas, is quite uncertain and requires a different type of shield. One type that has proved exceptionally good in practice employs metal slats loosely attached to a ring around the gage orifice. No snow can collect on these and blow away later, and air is directed downward to overcome the eddying effect of the gage rim (Helmers, 1954). A further improvement to this design is the stereo orifice and slats described by Storey and Hamilton (1943). Both the rain gage orifice and the slats are oriented with respect to the wind (or slope of the ground surface). This is especially useful on tipping bucket type rain gages that cannot be tilted themselves.

Variations in rainfall from point to point also occur by other than mechanical means. The nature of the storm and storm size are important factors. Huff (1970a; 1970b) noted that large variations in rainfall catch occur during small showers. As the areal involvement increases, rainfall depths become more uniform with frontal storms giving the most evenly distrubuted rainfall. Osborn, Lane, and Hundley (1972) found that in Arizona the variation in rainfall due to convective thunderstorms was great, even over a few acres. McGuiness (1963) found the same general pattern to hold in the midwestern United States. Huff and Neil (1957), working on a level 19 acre plot in Illinois tested paired rain gages against storm size. They noted significant differences in catch between gages only six feet apart during showers. The overall differences averaged 6.1 per cent for storms of 0.1 inches or less to 1.3 per cent for storms of 1 inch and over. This is in general agreement with Huff (1970a) who found decreasing uniformity with events of decreasing areal extent and decreasing depth.

Precise areal estimates of precipitation become a problem of major importance because of the various factors influencing precipitation characteristics. The design of the rain gage network can be crucial. Initially, the objectives of the data should be considered. Huff (1970b) noted that sampling requirements for seasonal and monthly data are not as stringent as that for individual storm measurements. That is, fewer gages are required for a given area for long term precipitation characterization because for a long term statistical representation the recording of every event by every gage is not required. Linsley and Kohler (1951) have found that widely spaced gages, even on flat terrain, occasionally have large errors because of localized storms missing a gage altogether. This does not seem to affect the long term average, however.

If data from each storm are required, the area covered and the characteristic type of precipitation must be closely considered. Huff (1970a) noted that more dense networks are needed for events of small areal distribution. Large storms require fewer gages per unit area. Therefore, he found that roughly two times as many gages were needed during the summer, when most precipitation was convective, than in the winter when most precipitation was due to air masses, in order to maintain the same degree of accuracy. Osborn, Lane and Hundly (1972) found that it was necessary to place rain gages every 1000 feet to adequately sample thundershower rainfall in southeastern Arizona. For a watershed situation, however, one gage for a 120 acre basin was sufficient if it was centrally located. As the catchment area increased in size so did the number of gages required. For a circular 10 square mile basin, 5 evenly spaced gages were needed or one for every 1280 acres. This increase in gage number requirement is not linear because the isolated events which closely spaced gages sample become less important as watershed size increases. The response of a 120 acre basin to a small thunderstorm may be marked while it would be negligible on a 10 square mile basin. Large storms with more uniform distribution of rainfall will be sufficiently sampled on a small basin by one gage while 5 gages should be enough to sample its variation on the larger basin. Eagleson (1967) makes special note of this fact and maintains that even on moderate watersheds as few as two well placed gages may be required.

While absolute error in rainfall depth sampling is related to density and size of the catchment (McGuiness, 1963), Hendricks and Comer (1970) have found that in the northeastern United States gage spacing should take into account direction of the network. They found that gages in a NNW-SSE line of progression should be twice as dense as WSW-ENE gages. The reason is unclear, but it may be due to direction of storm movement or prevailing winds.

The three basic methods for determining average areal distribution of rainfall are the arithmetic mean, Thiessen method, and isohyetal method (Chow, 1964; Gray and Wigham, 1970; Wisler and Brater, 1959). The arithmetic mean is the easiest to use and is accomplished by summing catchment in all gages and dividing by the number of gages. This procedure does not allow for unevenly spaced gages, orographic influences, or differences in precipitation due to storm morphology. It is best used in flat country where the network is intense and evenly distributed (Wisler and Brater, 1959).

The Thiessen method of weighted polygons accounts for an uneven distribution of rain gages by weighting portions of the watershed to each gage. It involves interconnecting the gages with straight lines and forming perpendicular bisectors to these lines. This results in polygons containing one rain gage each. The area of each polygon is weighted against the total area of the catchment. Once again, no consideration is given to orographic influences (Linsley, Kohler, and Paulhus, 1958).

The isohyetal method is considered by some to be the most accurate method of determining average rainfall depth (Chow, 1964). Isohyets, contours of equal precipitation, can be drawn if an adequate network exists. The area between the isohyets is determined and multiplied by the average precipitation between them. These products are summed and divided by the area of the catchment. This procedure may give weight to both orographic influences and storm morphology if handled properly. Huff (1970b) has criticized the isohyetal method because many samples must be taken in order to evaluate a rainstorm and many factors influence the precipitation amount. This may be especially true where extensive urban areas complicate the rainfall pattern (Atkinson, 1971).

Runoff. Chow (1964) defines runoff as the part of precipitation "which appears in surface streams of either perennial or intermittent form." This implies that not all precipitation falling to the earth finds its way to the channel. Indeed, precipitation experiences losses through interception, absorption, evaporation, transpiration, and infiltration. In the overall hydrologic cycle the losses reduce to evaporation and transpiration, with infiltrated water eventually reaching the groundwater resevoir and becoming a subsequent part of stream discharge. This latter process may be very slow, requiring up to two years or more (Linsley, Kohler, and Paulhus, 1958).

Runoff is usually divided into four components based on the pathway to the channel. These include surface runoff, interflow, groundwater flow, and channel precipitation. Hewlett and Hibbert (1967), in their survey of literature definitions of runoff types, found only a general consensus as to what each really is.

Surface runoff, or overland flow, is that part of the runoff phenomenon that reaches the channel by flowing over the ground surface. The distance travelled by overland flow is generally quite short and occurs only from impervious areas of the basin during moderate to heavy storms. It is the form of runoff that reaches the channel most rapidly. (Linsley, Kohler, and Paulhus, 1958).

Water percolating into the soil may move laterally to the channel. This source of runoff is less rapid than overland flow and is known as interflow. Interflow is favored where a relatively impermeable layer exists below a thin soil cover. In moderate storms it may be the major constituent of storm runoff. Soils that are homogeneous and deep generally promote percolation to the water table (Linsley, Kohler, and Paulhus, 1958).

Water that percolates through the soil profile to the watertable may eventually become runoff if the watertable intersects the land surface at some point. This portion of runoff is called baseflow, or dry weather flow. Groundwater discharge seldom causes rapid changes in runoff because of the slow transfer rate through the aquifer. It is the component that sustains streams through dry periods (Chow, 1964).

Channel precipitation is that part of total runoff falling directly on the water surface, or splashing into it from nearby rocks and leaves (Hewlett and Hibbert, 1967). This component of runoff is generally either ignored or taken to be part of surface runoff. Its volume contribution to discharge is instantaneous but minimal, and even at high stages is never more than 5 per cent (Gray and Wigham, 1970).

In a broad sense, runoff is affected by the major categories of climate and physiography. Included in climatic factors are precipitation characteristics, interception, evaporation, and transpiration. The physiographic influences include the watershed characteristics such as infiltration capacity, soil type, vegetation, shape, and size (Chow, 1964).

Overland flow is closely related to the infiltration capacity of the soil. Horton (1933) maintained that the infiltration capacity, the maximum rate with which water enters the soil, could be characterized by:

$$f_p = f_c + (f_o - f_c)e^{-ct}$$

where f_p is the infiltration rate at time t, f_c is the infiltration capacity when the soil becomes saturated, f_o is the infiltration rate at t = 0, or, the beginning of the storm, c is a constant, and t is the time since the beginning of the storm. As $t \rightarrow \infty$, the exponential decay relationship approaches equilibrium and f_p equals f_c. Consequently, even on unsaturated soils, overland flow can occur when the rainfall rate exceeds the infiltration capacity.

Most overland flow has traditionally been conceived to originate from impervious areas of the watershed. Betson (1964) hypothesized overland flow as being derived from small but variable source areas. Initially, this area is insignificant but as the rainfall continues larger regions become saturated and surface runoff increases. Dunne and Black (1970a) found this to be the probable case on New England watersheds where the contributing source area was near the stream channel.

Overland flow, however, is noted to occur in the forest situation infrequently (Hewlett and Hibbert, 1967 ; Pierce, 1967). This is due to the fact that vegetative cover tends to increase permeability of the soil and decrease rainfall intensity via interception (Linsley, Kohler, and Paulhus, 1958). Pierce (1967), on the other hand, noted that matted leaves under the canopy can provide a surface over which rainwater can travel, but only for very short distances.

In arid areas overland flow is a major contributor to runoff even though the actual percentage of rainfall becoming runoff may be lower in the desert than in humid areas due to high evaporation rates. The high incidence of overland flow results from the lack of dense vegetation and significant areas of exposed impervious material (Slatyer and Mabbutt, 1964).

The degree of slope of a basin has an important controlling effect on runoff. Steep slopes may produce substantial runoff quantities while gentle slopes, none at all. This has been noted to be the case in both humid and arid regions (Slatyer and Mabbutt, 1964; Sopper and Lull, 1970). Some soil conditioning has been noted to occur in Arizona watersheds. Overland flow was initially very great on dry soil due to caking, but decreased as the soil became wet and capillarity increased. A decrease in the rivulet network was also noted (Schreiber and Kincaid, 1967). Maxey (1964) noted also that mountainous terrain yields higher runoff which was augmented when impervious or relatively impermeable surface conditions existed. In England, Hills (1971) observed that the capacity of the soils to absorb rainfall is great. The only notable places where overland flow may be seen to occur are on sparsely vegetated slopes where the soil has been compacted, or in the winter months on compacted soils. It has been seen to be a major contributor to quick stream rises only when the source area is very near the channel.

Overland flow has been shown to be the major contributor to runoff in urban and developed areas. Runoff to rainfall ratios of 50 per cent are common in large cities where the pervious to impervious surface area ratio is approximately the same (Viessman et al., 1970). This however, is a special case and is beyond the scope of the present study.

Interflow, or subsurface stormflow, is poorly understood. It is conceived of as occurring on basins in which the soil profile is highly differentiated. The rapid rate of transmission, although slower than overland flow, exceeds that of groundwater seepage. This is seen by some to be a function of biological pathways not present in subsoils (Hursh and Hoover, 1941).

The concept of interflow stems from the fact that little overland flow is noted on forested watersheds, but rapid rises in stream discharge are seen to occur (Hursh, 1945). The actual mechanics have defied quantification, however (Whipkey, 1967). It is usually viewed as being of one or two cases. First, interflow is the result of rapid spillage of an overcharged groundwater reservoir through highly porour surface soils. Second, interflow is due to shallow penetration of storm water into porous upper horizons with subsequent rapid lateral transport (Hursh, 1945).

Hewlett and Hibbert (1967) view subsurface flow in a slightly different manner. Instead of being rapid transport of recent rain, they suggest that water already stored in the soil profile is displaced. As water supplied to the basin increases this stored water is released in greater quantities. The process, called translatory flow, is operational at or above field capacity. All subsurface flow, then, may be thought of as a pulse of water moving downhill. Experimentally, a similar situation was witnessed in soil columns wetted to field capacity where isotopically labeled water added to the surface did not begin to run out the bottom until 87 per cent of the water already in the column had been displaced (Hewlett and Hibbert, 1967).

Interflow has been secn to be the largest contributor to direct runoff on gently sloping watersheds on the Georgia coastal plain. Here interflow was shown to account for 28 per cent of the precipitation loss while evapotranspiration accounted for 61 per cent of the precipitation loss. Very little overland flow was seen to occur (Asmussen and Ritchie, 1969).

Recently, as mentioned above, an alternative to the idea of subsurface flow has developed. The partial area concept introduced by Betson (1964) was found to be an attractive hypothesis on Vermont watersheds (Dunne and Black, 1970a). It is now necessary to consider overland flow once again. It was found that significant amounts of storm runoff were derived from hills containing areas on which the watertable had reached the surface, and that the importance a slope had in producing storm runoff was its ability to generate overland flow. This was the only form of runoff evidenced in large enough quantities to produce rapid rises in stream discharge. While subsurface flow had been seen to occur, it was in such small amounts as to be insignificant (Dunne and Black, 1970b). Dingman (1970) has also alluded to this possibility on an interior Alaska basin and carried out several hydrograph separations based on this concept. Clearly, this phase of the runoff process lacks sufficient explanation and much more study is required to gain a working knowledge of the processes involved.

Groundwater discharge and movement is somewhat less mysterious than overland flow and interflow. It is the net precipitation which has percolated to an aquaclude and begun to collect in the porous medium. Groundwater moves according to Darcy's law given by

$$v = -K \frac{dP}{dh}$$

where K is the hydraulic conductivity of the porous medium, $\frac{dP}{dh}$ is the slope of the hydraulic gradient, and v is the velocity of flow (Chow, 1964).

Once it is below the soil surface there is no assurance that water will reach the groundwater. In a forest, evapotranspiration by the vegetative cover can result in an enormous loss of potential groundwater. Studies by Urie (1971) indicate that clear cutting red pine (Pinus resinosa, Ait.) will result in a 20 per cent rise in the watertable, and that the amount of water reaching the watertable immediately after a major storm is increased by 30 per cent. Lewis (1968) found that the conversion of a basin from woodland to grassland markedly increased yield. Total runoff increased over a three year period by more than 13 inches. It was further shown that the consumptive use by the vegetative cover decreased by the value of the yield increase. Hibbert (1967) reports that on 39 watersheds studied there were significant increases in runoff either annually or for the growing season in all but a few cases. Nakano (1967) describes similar results, with annual runoff increases amounting to 8 to 24 per cent. In humid eastern watersheds deforestation increased streamflow substantially during the growing season. Little effect was seen during the dormant season, however. The reason for this, apparently, is that during the growing season transpiration is going on at nearly the potential rate, resulting in the loss of much infiltrated water. During the dormant season rainfall far exceeds transpiration and less effect is seen on streamflow (Patric and Reinhart, 1971). McGuiness and Harrold (1971) found that watershed reforestation not only reduces low and intermediate flows, but also reduces maximum annual flow volumes for all durations greater than one day.

Sopper and Lull (1970), in an extensive study of 137 small northeastern United States watersheds found that an increase in forest cover increased the runoff. They concluded that forests in this part of the country act to integrate several of the factors which account for higher runoff. Generally, the denser forests were associated with more northerly and mountainous areas. This means that they have steeper slopes, shallower soils, cooler temperatures, and more precipitation. Each one of these conditions may act to cause increased runoff. They did recognize, however, that if the same watersheds were deforested an increase in yield would occur.

In order to counteract the effects of transpiration and to otherwise increase annual runoff, a number of attempts have been made with the use of chemical antitranspirants. Greenhouse experiments have shown that phenylmercuric acetate reduced transpiration in the white pine (<u>Pinus strobus</u> L.), jack pine (<u>P. banksiana</u> Lamb.), and the red oak (<u>Quercus ruba</u> L.). Injury to the trees did result, however, possibly due to a volatilization of mercury (Waggoner, 1967). Higher order alcohols were thought to have considerable potential as antitranspirants. However, recent work by Gale, Roberts, and Hagan (1967) has shown that decreases in growth associated with the application of a monomolecular layer of alcohol outweighed any decrease in transpiration. Satterlund (1969) maintains that a successful antitranspirant in association with artificial enhancement of rainfall would be of enormous benefit in increasing yield both annually and for individual storms, especially in arid and semi-arid regions where water supplies depend on basin runoff to a large extent.

The groundwater contribution to total runoff is variable from one watershed to another. It is the major component of runoff in watersheds with surfaces that discourage the possibility of overland flow and encourage infiltration. This is especially true when evapotranspiration is minimal (Bay, 1969), and soils are homogeneous, permeable, and deep (Linsley, Kohler, and Paulhus, 1958). Bog and peatland watersheds of Minnesota show a very sluggish response to rainfall other than from very intense storms, and groundwater seems to dominate streamflow (Bay, 1969). When the depth to the impervious layer is shallow and the soils have a low permeability, groundwater contribution is usually minimal (Linsley, Kohler, and Paulhus, 1958). In a small (0.7 mi^2) interior Alaska basin, Dingman (1966) noted this to be the case. Here, streamflow was dominated by storm runoff. Baseflow accounted for under half of the annual runoff and the stream occasionally dried up during the extended rainless periods. An additional factor limiting the effective depth of the soil was presence of both seasonal frost and permafrost underlying much of the basin for substantial portions of the summer months when discharge normally occurred.

Comer and Zimmerman (1969) found base flow to be very low on northern Vermont watersheds with extensive fragipan-type soils. The nearly impermeable B horizon prevented deep percolation, and promoted rapid lateral seepage at the A-B interface. High organic content of the superficial layers further reduced baseflow due to high water retention capacity. On basins with deeper, well-drained soils in the same climatic region, they found higher baseflow that was sustained through dry spells. Black (1970) reported that layered soil horizons have more rapid runoff and yield smaller minimum streamflow volumes, whereas deep soils not only retard runoff but produce lower flood peaks, and yield higher minimum streamflows.

Aside from evapotranspiration and surface geomorphology, characteristics of the aquaclude itself will affect the magnitude of groundwater runoff. Singh and Stall (1971) report that leakage loss through the confining layer causes baseflow to fall off more rapidly with time and add less flow than would otherwise be expected. The reverse case can also exist where the leakage is upward from a confined aquifer. Groundwater contribution in this case is more stable with time and may cause more baseflow than would normally be expected.

Runoff volumes in humid and subhumid areas were not found to be affected by watershed size by Sopper and Lull (1970). In their study of small watersheds in northeastern United States, they found no significant relative decreases in runoff volume with increasing size. The most influential factors were latitude and topography, with the highest runoff volumes found in northern mountain basins. Osborn and Lane (1969) found that the unit area runoff decreased as watershed size increased, for small semi-arid basins in Arizona. They reported 90-95 per cent of the storm events to be convective with intense and short duration characteristics. Gray and Wigham (1970) assert that the largest flows are produced by thundershower activity on small basins and by synoptic storms on large basins.

Runoff can be graphically represented by means of a hydrograph relating the rate of discharge to time (figure 2). It is the integrator of all processes within and on the watershed. Any point

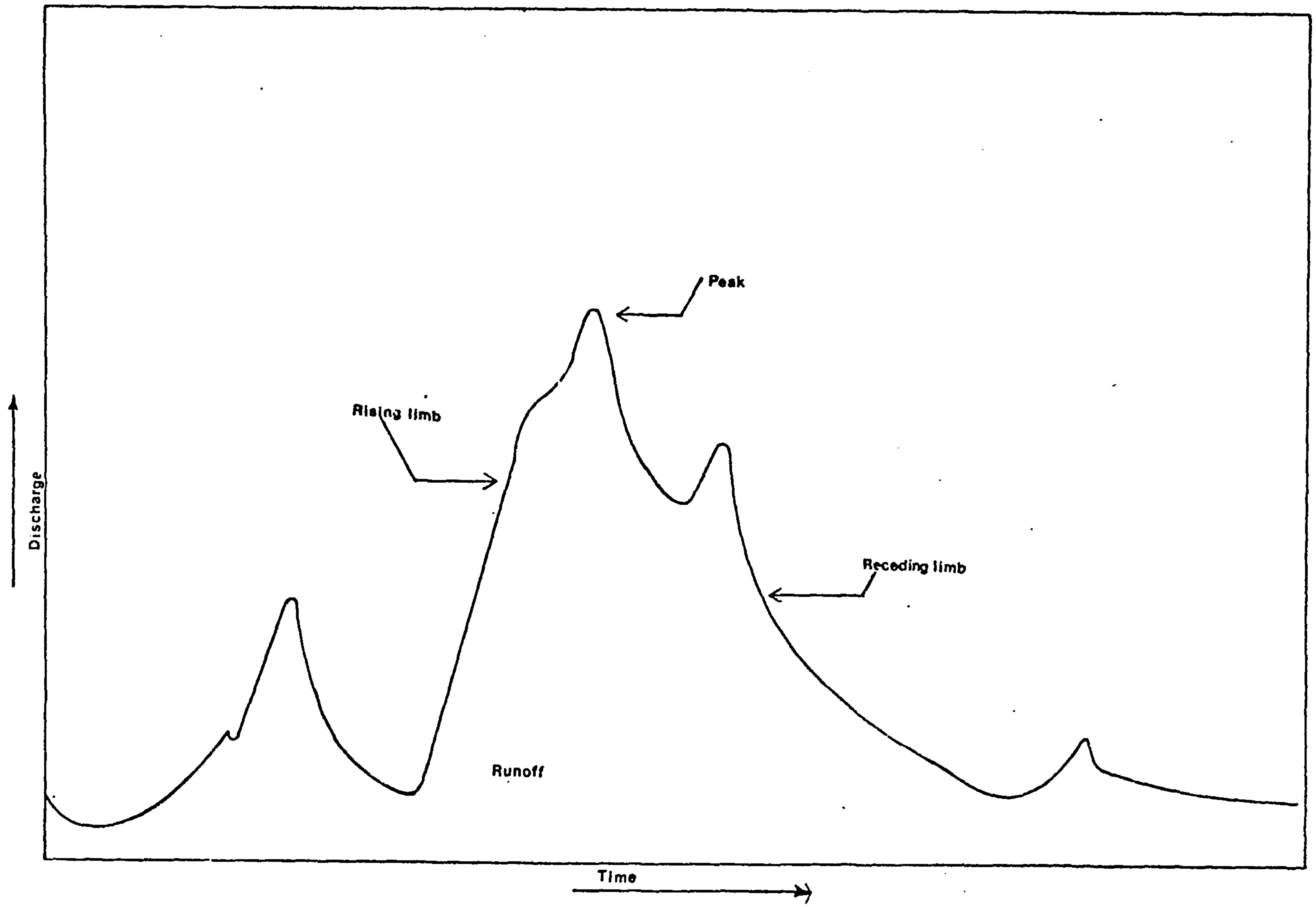

Figure 2

A hypothetical hydrograph

on the hydrograph represents the discharge rate at the corresponding time increment. The hydrograph can be divided into four major segments: 1) the rising limb, 2) the peak, 3) the receding limb, and 4) the runoff volume--represented by the area under the curve (Chow, 1964). The shape of the hydrograph is influenced by direction of storm movement and distribution, intensity and duration of rainfall, and basin topography.

The direction of storm movement over a basin can have a profound effect on both peak flow and time period of surface (or storm) runoff. On very long basins the effect is accentuated. Storms moving upstream tend to produce runoff hydrographs with lower peaks and longer time bases than comparable storms moving downstream (Gray and Wigham, 1970). This can be explained in that a storm entering at the mouth and proceeding toward the headwaters will first drop rain at the outlet. This produces a quick rise in the hydrograph. As the storm moves toward the far end of the basin, runoff from the source area near the mouth decreases and most of the runoff is from up-basin regions. Channel storage abstracts and modifies it, causing a drawn out recession. On the other hand, storms moving downstream first produce runoff in head water regions. The travel time, being similar to the time of concentration, will yield a broad rising limb, and high peak due to the simultaneous arrival at the outlet of up-basin and down-basin runoff. The falling limb, then, will be rapid (Roberts and Klingman, 1970).

The distribution of rainfall has been seen to have an effect on the shape of the hydrograph. A heavy rain at the mouth of the watershed may cause a short duration runoff event with rapid rise, high peak, and rapid recession. If the same storm occurred at the headwaters of the basin, a dampening effect would be seen and a broader time base and lower peak would result (Gray and Wigham, 1970). Watt and Kennedy (1969) report that as lag time decreased peak flow values increased. This would tend to confirm the above assertion, since the travel time to the outlet from the far reaches of the watershed would, in a relative sense, greatly exceed that from the lower regions. The volume contribution to runoff from the distant storm would be modified not only by the channel, but also by contributions from other sources along the path of travel.

Duration of rainfall has also been seen to affect the shape of the hydrograph. Storms of long duration show peak flows lower than equivalent storms of shorter duration (Bell and Kar, 1969). This implies that an increase in intensity causes higher peak values of the hydrograph. Schreiber and Kincaid (1967) found that the most highly correlated parameter in predicting runoff from convective storms in Arizona was intensity. As the intensity increased so did the volume of runoff and peak flow. Potter (1949) noted the same effect on subhumid watersheds. Unit hydrograph theory proposed by Sherman (1932) shows that for a storm of given intensity as the time base is increased. the resultant peak flow decreases; and for a given duration, as the intensity increases so does the peak of the hydrograph. On experimental model watersheds Roberts and Klingman (1970) have demonstrated that the peak value of the runoff hydrograph for isolated events increases with increasing intensity. It was further noted that this is especially true on small basins where the variation in streamflow is most sensitive to fluctuations in intensity. They also found that increasing duration will increase the value of the peak flow to an equilibrium point. After this point is reached the size of the crest segment will depend on the length of the rainstorm. Both the rising and falling limbs of the hydrograph will remain the same for a given set of stable conditions with only the duration of precipitation varying.

The slope of the basin exerts an influence on the shape of the hydrograph. In general, greater slopes give more runoff, larger peaks, and shorter time base (Maxey, 1967). The exact nature of this effect is difficult to quantify because of interactions with soil moisture, vegetation, and rainfall characteristics. The relationship is further confused by watershed size. Commoner (1942) notes that on large watersheds the time involved in overland flow is small compared to the overall time period for flow in the stream channel. On small watersheds the peak is increased with slope, since the ratio of time of overland flow to the time of flow in the channel is closer to unity. Osborn and Renard (1969) found that in the southwestern United States the average annual runoff decreased per unit area as the watershed size increased due to transmission losses in the permeable alluvium stream beds. Storm runoff also decreased per unit area as size of the watershed increased due to the limited areal extent of runoff producing thunderstorms and transmission losses in stream beds (Osborn and Hickock, 1968).

Butler (1957) gave the following relationship which includes slope of the basin

$$q = ad^bS^c$$

where q is the rate of discharge per unit width, d is surface storage, S is the slope of the basin, and a, b, and c are constants. This shows that, for a given rainfall excess, the instantaneous flow volume is greater on sloping land and the time base is proportionately less (Gray and Wigham, 1970). It can be assumed from this that the corresponding peak flow is higher.

Hydrograph Separation. It has been said by at least one researcher and implied by many that, "Hydrograph separation is one of the most desperate analysis techniques in use in hydrology." (Hewlett and Hibbert, 1967). The variety and number of techniques seem to bear this out, since the separation involves the four streamflow components of channel precipitation, overland flow, interflow, and baseflow. These components, in turn, all have widely varying definitions. Indeed, most authors make no attempt to break streamflow down into these four parts. In an effort to simplify the situation they include channel precipitation, overland flow, and interflow into a lump term usually designated direct runoff or surface runoff (Chebatarev, 1966; Linsley, Kohler, and Paulhus, 1958; Chow, 1964). The effect of this does little, however, to make the definition of hydrograph components discrete units (Hewlett and Hibbert, 1967).

It is desirable to separate the storm runoff from the groundwater portion of the hydrograph in most analyses. This is because of the differing characteristics of each. The storm component is marked by relatively rapid fluctuations and is directly related to rainfall events. It is the portion of stream discharge that for the most part, results from surface and shallow subsurface flow immediately after rains and is responsible for damaging floods and other forms of destruction such as accelerated erosion and sedimentation. Storm runoff is also considered a loss to the basin. It represents water that does not contribute to the moisture characteristics of the watershed. In other words, it is transient.

Groundwater is not subject to rapid fluctuations. It seldom is responsible for flood flows and rarely contributes to erosion of surface soils. It too, is representative of the moisture regime of the basin--a measure of wetness or dryness. The groundwater component, when it is the major contributor to streamflow, can be used to predict the storm runoff component for a given rainfall excess. This is because it is indicative of the recharge capacity of the basin at any given time. While there is little real basis for the separation of groundwater from surface runoff on a hydrograph, an indication, if only empirical, is a valuable tool in the prediction of flood flows, the characterization of a general moisture regime, and countless other engineering and forest management aspects (Linsley, Kohler, and Paulhus, 1958; Bruce and Clark, 1966; Chebatarev, 1962; Comer and Zimmerman, 1969).

Two distinct types of hydrograph events can be distinguished. They are simple and complex. The simple event is characterized by a single peak and a discrete block of rainfall contributing to it.

This usually implies that little or no rain falls after the recession has begun. If this is the case, the receding limb is fairly constant from one storm to another and analysis is greatly simplified. The complex hydrograph is characterized by more than one peak. The recession from the peak of one rainfall event is not completed before another precipitation event causes an additional rise; the peaks are so close together that the contribution to flow of discrete blocks of rainfall cannot be evaluated by the techniques used on simple hydrographs. This situation can be seen to occur with closely spaced rainstorms or with one continuous rain storm having variations in intensity (Linsley, Kohler, and Paulhus, 1958). Figure 2 illustrates both cases. Standard methods for separating baseflow from stormflow for each type of event will be described below.

Linsley, Kohler, and Paulhus (1949) describe three methods for the separation of baseflow from stormflow for simple hydrographs. The first and easiest is accomplished by drawing a straight horizontal line from the point of rise to its intersection with the falling limb. This method is used infrequently and can result in large errors due to the fact that it assumes that baseflow is constant. This is obviously not the case. If recessions during long rainless periods are examined it will be noted that groundwater yield follows an exponential decay and that it is recharged by rain storms. While it does not respond at the rate that storm runoff does, the assumption that there is no response at all will lead to exaggerated flow volumes for the latter. This is especially risky in areas with high infiltration capacity and well-drained porous surficial layers where groundwater recharge and subsequent response may be fairly rapid (figure 3a).

The second method consists of drawing a straight line from the point of initial rise of a hydrograph event to its intersection with the falling limb a given time period, N, after the peak. The value of N is constant for the watershed and represents an average time for flow to change from predominately storm runoff to predominately baseflow. N has been shown to be empirically related to area of the basin by

$$N = A^{0.2}$$

where A is the area in square miles. This, however, does not always hold and hence many hydrographs should be examined before N is set. This procedure assumes a slow linear increase in baseflow beginning with the onset of storm runoff. That baseflow is not linear is known; however, since the real situation is in doubt the method just described is usually a better estimator of a true separation than is a straight horizontal line (Linsley, Kohler, and Paulhus, 1949). Figure 3b illustrates this method.

The third method requires that a groundwater depletion curve be either estimated or constructed. A groundwater depletion curve is merely the characteristic streamflow recession of a basin during long dry periods. Two assumptions are made here. The first is that after a considerable rainless period stream discharge is largely due to groundwater, and its decay is due to the depletion of groundwater reserves on the basin. The second is that the relationship between outflow and storage is stable. In other words, true groundwater recession

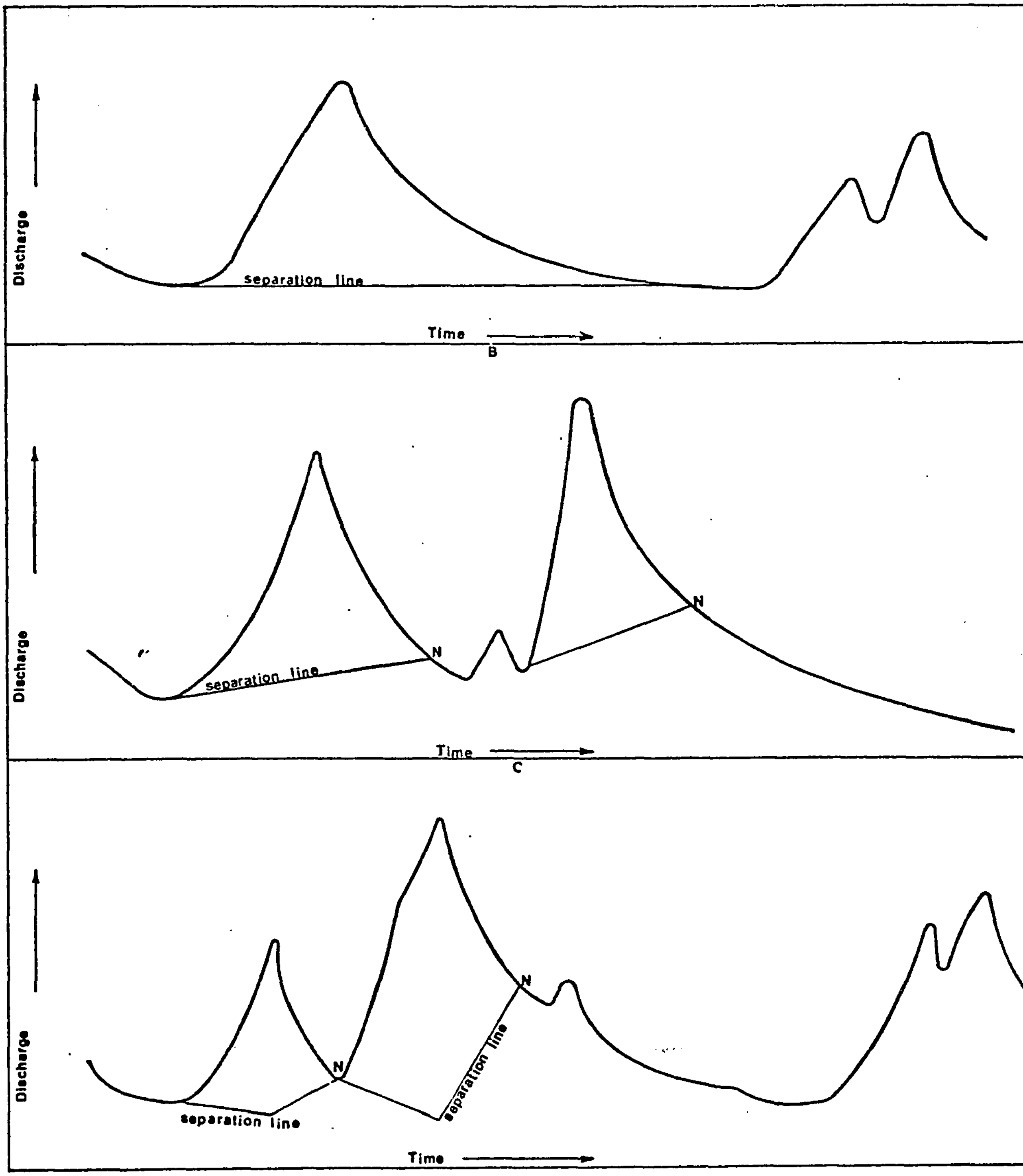

Figure 3

Various hydrograph separations: a) use of a horizontal separation; b) use of characteristic time period--straight line; c) use of ground water depletion curve.

is a characteristic of the basin and can be represented by a single curve (Holtan and Overton, 1963). The curve itself can be gotten by finding a suitable period of time on the discharge hydrograph upon which no rain has fallen and that covers all necessary ranges of flow. If sufficient rainless periods do not exist, a depletion curve can be constructed by piecing together bits of rainless recessions until a master curve covering all required flow values has been derived. Once the groundwater depletion curve has been obtained it is fitted to the recession immediately prior to the rise of the event to be separated and continued to a point under the peak. From here a straight line is drawn to its intersection with the recession at time N. The major assumption made by this technique is that as the stream rises onto unsaturated parts of the bank there is flow into the bank. Following this reasoning there should be a decrease in baseflow until the level of the water in the channel begins to fall, at which time bank storage returns to the stream. Linsley, Kohler, and Paulhus (1958) note that this may be true; however, if the amount of water stored during the stream rise is greater than the inflow of groundwater baseflow is effectively negative. For this reason, the method of separation just described is equally as arbitrary as the other methods (figure 3c).

The object of separation techniques as they relate to complex hydrographs is to divide the hydrograph into single peaked events where the contribution of each rainfall block can be evaluated, or to determine to what part of a rainfall event each peak can be attributed. The assumptions made here are basically the same as those for the simple events. The only addition is a total flow recession curve which may be constructed from rainless periods from any peak and to conform to the range of flow required. It is assumed that total flow recession, like groundwater recession, can be characterized for a given basin by a single curve. While there is less basis in fact for this, due to the variable factors of evapotranspiration, watershed soil condition, and condition of the vegetation, it is frequently used. The total flow recession curve is fitted to the small recession and continued into the area of the next rise (N days). The groundwater recession curve is fitted to the previous recession and continued to a point under the peak of interest. From there a straight line is drawn to its intersection with the total flow recession curve, N days later. The assumptions are the same as those for the simple hydrograph with the addition that groundwater acts as it would for a series of separate single-peaked events. This may not be true when the multiple peaks are due to variations in intensity or storage (Linsley, Kohler, and Paulhus, 1958). Figure 4 illustrates this case.

Unfortunately, these separation techniques are not always applicable to data collected from forested basins, or basins with only a few years of continuous data collection (Hewlett and Hibbert, 1967; Bay, 1969). The problem is acute when baseflow is not readily discernible; the hydrologist's judgement often determines where runoff ends and baseflow becomes the sole contributor to streamflow (Bruce and Clark, 1966; Linsley, Kohler, and Paulhus, 1949).

Bogs and forests complicate the situation since overland flow is minimal. Bay (1969) found that heavy moss layers on bogs detained

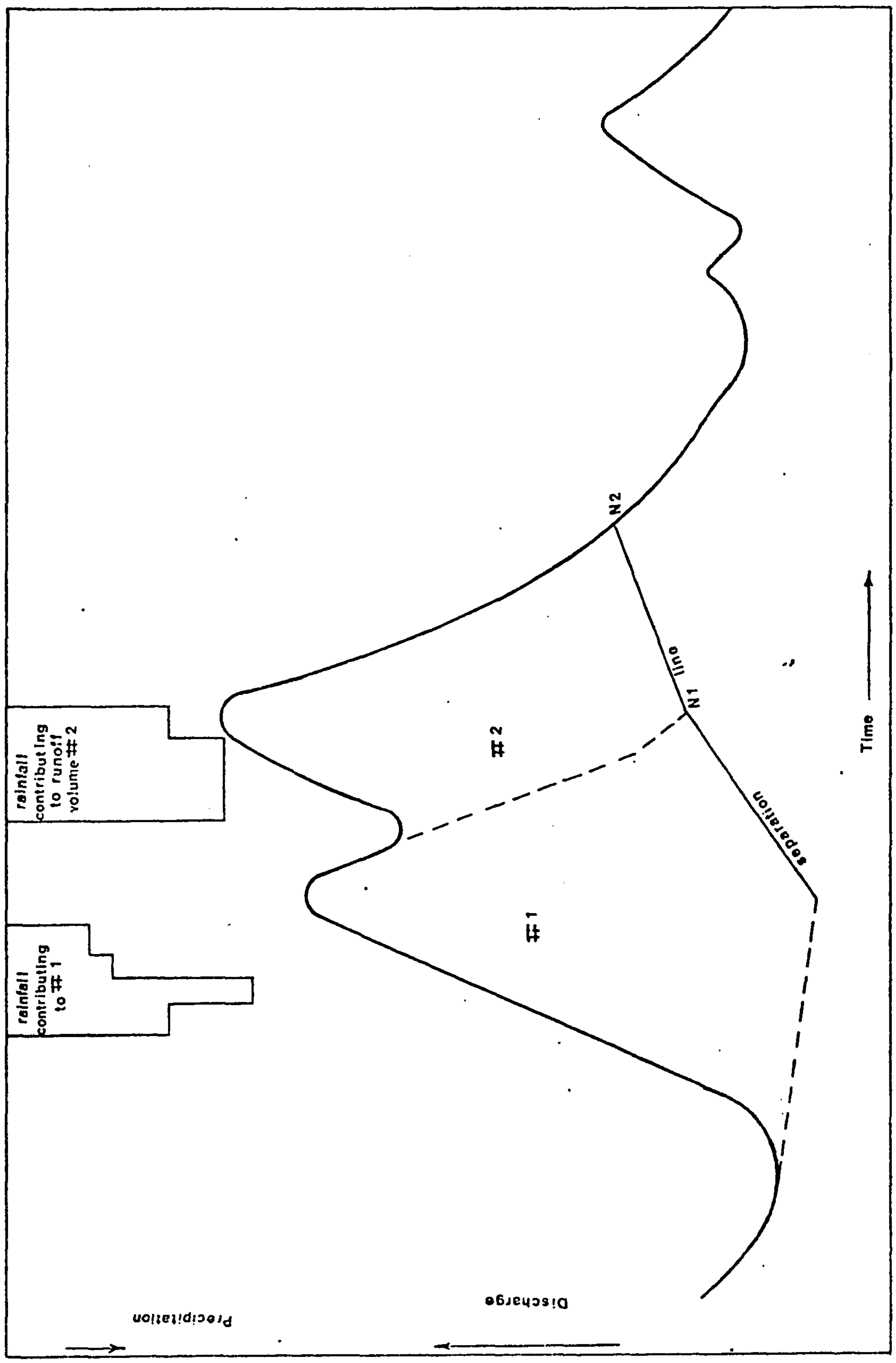

Figure 4

Separation of a complex hydrograph

most, if not all of the water that would otherwise become overland flow. Similarly, forest soils are considered optimum for infiltration. In fact, the infiltration capacity of forest soils may be exceeded only on very rare occasions (Pierce, 1967). Bay (1969) found that on Minnesota's bog-dominated basins hydrograph response to rainfall events was noted only under conditions of high antecedent soil moisture and large quantities of precipitation. Hewlett and Hibbert (1967) noted a similar situation on forested Appalachian watersheds. It can be seen then, that separating stormflow from the somewhat nebulous baseflow does require certain modifications.

Bay (1969) maintained that on Minnesota's bog watersheds the separation of baseflow from stormflow was difficult, if not impossible. Dingman (1966) was not as pessimistic. His study area not only lacked sufficient records for analytical separation techniques, but was a heavily forested watershed with a deep layer of moss on the soil surface. He reasoned that due to the presence of relatively impermeable soil over the basin groundwater contribution would be slow and insignificant. Therefore, he simply estimated the asymptotes to the recession portion of the hydrographs and defined this as baseflow.

Hewlett and Hibbert (1967) discovered that the separation of direct runoff from baseflow on several North Carolina watersheds was impossible. Faced with this dilemma and realizing that some separation was necessary, they defined the terms "quickflow" and "delayed flow." The difference between these two new components and their conventional counterparts lay in the fact that the only arbitrary

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1967; Linsley, Kohler, and Paulhus, 1958; Wisler and Brater, 1959). This is especially difficult in basins where no one method of separation seems more logically applicable than any other. In practice, it may come to nothing more than trial and error or invention.

Precipitation-runoff relationships. The relationships between precipitation and runoff are dependent on the factors discussed in the previous sections. The most readily discernible of these are precipitation intensity, duration, and amount. This is so since the visual display of precipitation-runoff events is by means of the hyetograph and hydrograph. From the hyetograph intensity, duration, and amount of storm rainfall can be gotten. Runoff information is obtained from the hydrograph. When the two are plotted together with the same time axis information concerning the interrelationships of the various parameters can be obtained. These include runoff variations with intensity, duration, and amount of precipitation and the integrating effects of the watershed on the temporal relationships as seen by lag times and recession characteristics.

According to unit hydrograph theory proposed by Sherman (1932), which assumes linearity of peak discharge with rainfall excess, as the precipitation intensity increases (larger amount over a constant duration) the runoff and peak flow will increase. The converse, that when the intensity decreases peak flow decreases also holds (Linsley, Kohler, and Paulhus, 1958).

Osborn and Lane (1969) found that intensity of rainfall when tested against amount was the most significant parameter in predicting peak rate of runoff for convective storms on southern Arizona watersheds. Specifically, the maximum 15 minute intensity was the most highly correlated to peak discharge. Other intensities tested were 5, 10, 20, and 30 minute depths. The reason for correlation with 15 minute maximum depth may be due to the fact that the time of concentration is nearest this value. Osborn and Lane (1969) noted that the total volume of runoff on their semi-arid watershed was most highly correlated to total precipitation amount. Schreiber and Kincaid (1967) working on small semi-arid rangelands found that precipitation intensity and total amount were both highly correlated with volume of runoff. Average runoff was most highly correlated with precipitation amount while intensity was most significantly related to storm runoff. Potter (1949) found that while the peak rate of runoff observed was influenced by intensity, the relationship was not strictly linear. He defined peak streamflow volumes as being dependent on the condition of wettness of the basin before storm rainfall commenced. Antecedent moisture in a maximum, minimum, or intermediate state, then, was an operator on intensity and peak flow would vary for each of these sets of conditions. In mathematical notation

$$R_p = f(M_a, I)$$

Where R_p is peak runoff volume, M_a is antecedent moisture condition of the basin, and I is the intensity of rainfall.

On basins outside of semi-arid regions a different situation exists. Whereas intensity was a highly correlated parameter on the watersheds in the southwestern United States, it does not always seem to be a major influence on basins in humid areas of the Northeast.

Dunne and Black (1970b) noted that no large areas of overland flow occurred on basins in Vermont under 5 and 30 minute intensities of 3.12 and 1.18 inches respectively. Streamflow was not seen to be sensitive to changes in intensity. Major changes in streamflow were, however, rapid during storm periods and may have been accounted for by the partial area concept already discussed. Overland flow production on a relatively small but already saturated part of the basin was the only type of flow that could account for the discharge increases noted. Betson (1964) and Betson and Maurius (1969) found the same situation to be operative on other Northeastern watersheds. In the bog watersheds of Minnesota large precipitation amounts occurred in the late summer and fall, but the corresponding streamflow was low. This was due to the ability of the vegetative cover and the soil to retain large amounts of water. Response to rainfall events by the stream was seen under high intensities, mostly in the spring and early summer when recharge due to snow melt caused a high watertable to exist (Bay, 1969). Bell and Kar (1969) noted that as duration of rainfall became very long the peak flow decreased. This, as was stated earlier, is due to the fact that for an equivalent storm, the intensity decreases as the duration increases and a decrease in peak is generally associated with a decrease in intensity.

Roberts and Klingman (1970) have shown with the use of model watersheds the theoretical influence of rainfall intensity. For a given basin with high antecedent soil moisture conditions, an increase in rainfall intensity is seen to markedly increase the size of the peak flow volume. The magnitude of the increase is proportional to the increase in intensity. Stated more simply, as the intensity becomes greater the increase in peak flow over that of some arbitrary standard intensity will also increase. Duration of rainfall has been seen to enlarge the flood volume and peak discharge if all other conditions remain identical on a permeable basin. On an impermeable basin the peak flow will not vary radically due to duration if it is long enough to impose equilibrium conditions. The width of the crest section, however, will be longer with increasing duration.

The dependence of peak flow volumes on intensity is widely accepted. The rational formula is given by

$$q_p = CiA$$

where q_p is peak discharge, C is the average ratio of runoff to rainfall, i is the average intensity of rainfall for a duration equal to the time of concentration of the basin, and A is the basin area. It not only takes into account the hydrologic response (C), but the intensity of rainfall. The assumption is that the peak flow will be proportional to the maximum intensity over a period of time sufficiently long so that the entire area is simultaneously contributing to streamflow at the outlet. The rational formula is valid only for small basins where modifying factors do not cause large variations in the hydrologic response from storm to storm (Linsley, Kohler, and Paulhus, 1958). The aspect of precipitation-runoff has already been discussed in a previous section.

Lag time has been a widely used parameter in the analysis and synthesis of unit hydrographs and for other analyses. There are two generally accepted definitions of lag time. The first is the time elapsed between the center of mass of the runoff producing-rainfall to the center of mass of the associated runoff. This definition is quite rigorous and difficult to apply. It does have the advantage however, of representing the storage delay time of the catchment. The second definition is the time from the center of mass of runoff-producing rainfall to the peak of the hydrograph. This definition has found wide acceptance in the literature because of its relative simplicity, and because it can be gotten directly from hydrograph-hyetograph displays (Linsley, Kohler, and Paulhus, 1958).

Lag times are most commonly associated with basin characteristics of slope, shape, area, and vegetation (Snyder, 1938). He gives the relationship

$$t_p = C_t(L_{ca}L)^{0.3}$$

where t_p is the lag time, C_t is a constant varying with slope and storage, L_{ca} is the distance from the center of area to the station, and L is the length of the basin. Taylor and Schwartz (1952) found the lag time in the North and Middle Atlantic States to follow the general form

$$t_{pr} = Ce^{mt_R}$$

where t_{pr} is the lag time, C is given by 0.6 s where s is the slope of a uniform channel having the same length and time of travel as the longest water course on the basin, m is $0.212/(LL_{ca})^{0.36}$, and t_R is the duration of runoff producing rainfall. Both of the above relationships take into account slope, distance of travel, and size of the watershed. Taylor and Schwartz also include terms for slope of the channel and the duration of rainfall. From Snyder's equation it can be seen that, holding all else constant, increasing the length of the basin will increase the lag time. This is due to the larger distance the most remote particle of water must travel before it contributes to the peak at the outlet. Also, increasing the circularity ratio will reduce the lag time. This is so since the length of time of travel in the channel would also be reduced. From Taylor and Schwartz' equation the additional effects of slope and rainfall duration are included. Holding all else constant an increase in the slope will cause a decrease in the lag time, since the speed of travel of runoff to the stream channel would be increased. Lengthening the duration of rainfall excess will increase the lag time. This aspect has already been discussed.

Bell and Kar (1969) include a term for area of the watershed and a constant taking into account the vegetation characteristics of the basin. Their relationship is

$$T_k = MA^{0.33}$$

where T_k is the critical lag time, M is a constant, and A is the area in square miles. They found that even on the same basin lag time is not constant from one storm to another. Rather, large floods were seen to have lags significantly shorter than the median lag for the basin. This time differential was on the order of 10 per cent for extreme floods. The critical lag, then, is an average value for lags of extreme events. There was little correlation between the lag and rainfall duration so physical factors were turned to. Correlation between the value of the constant, M, and several basin characteristics were attempted with vegetative cover being the only one strongly influential.

Mean values of M range from 2.05 for forests and woodland to 0.60 for very poor pasture and desert vegetation. This gives very strong evidence for the hypothesis that dense vegetation, especially in forests where soils are considered optimum for infiltration, will delay the surface runoff component considerably. Further evidence for the storage delay factor will be presented when recessions are discussed. It can also be seen that if area is increased with all other factors held constant, the lag time is increased because the time taken for all parts of the watershed to simultaneously contribute to discharge at the outlet increases. This is regardless of basin shape (Bell and Kar, 1969).

Dingman (1966) working on interior Alaska basins found excessively long lag times. On his 0.7 square mile basin, lag times were commonly 12 to 18 hours. Correspondingly, the vegetation was very dense and heavy layers of moss were present. Engman (1964) found lag times for basins in Vermont to be much shorter. A 3.25 square mile basin had an average lag time of 2 hours while basins up to 40 square miles and more had lag times of just over 5 hours. Holtan and Overton (1963) found lag times for Eastern watersheds from the centroid of rainfall to the centroid of runoff to vary greatly. The highest value was 102 hours for a 158 square mile basin. The shortest was 2.2 hours for a 61.1 square mile basin.

The recession curve, unlike the rising limb, can be mathematically analyzed. The reason for this is that while inflow-storage-outflow relationships that contribute to the rising limb are not stable with respect to one another, outflow from storage is (Holtan and Overton, 1963). They maintained that storage on the basin could be gotten by picking a discharge, q_t, at some time, t, and summing the area beneath the recession curve subsequent to q_t. If storage was plotted against q_t it was approximated by the relationship

$$S = mq_t$$

where S is storage, q_t is the discharge, and m is the slope of the line.

Chow (1964) gives stream discharge, q_t, at time, t, on the recession portion of the curve by

$$q_t = q_0 K_r^t$$

where q_0 is some flow before q_t on the recession and K_r is the recession constant. Noting that

$$q_t = -\frac{ds}{dt}$$

it can be shown that the storage left on the watershed at time, t, is

$$-\ ds = \quad q_0 K_r^t dt$$

or

$$S = -q_0 K_r^t / \ln K_r .$$

Since

$$q_t = q_0 K_r^t$$

we have

$$S = -q_t / \ln K_r .$$

Applying limits to integration we can obtain the change in storage between q_0 and q_t where

$$S = \frac{q_t - q_0}{\ln K_r} .$$

On the hydrographs of actual recessions this was found to be a very good approximation, however, the portion of the curve near the peak will still have as components both surface and subsurface flow. The variable lag times between the two will cause the value of K_r to be increasing with time (Barnes, 1940; Chow, 1964).

Dingman (1966) found that discharge, q_t, at some time, t, could be represented by

$$q_t = q_p e^{-ct}$$

where q_p is peak discharge, and c is the recession constant. He found that on his interior Alaska basin the average value of c was 0.027. This is quite small when compared to lag times from other parts of the country, indicating that there was a longer time of water detention on the slopes and a higher storage capacity.

It stands to reason that the storage capacity of the basin increases as the value of the recession constant decreases. Recalling some of the factors that produce rapid runoff, it can be seen that the arid basins of Osborn and Lane (1969), with little storage capacity, should have a large value of K_r. On the other hand, a forest situation such as that discussed by Hewlett and Hibbert (1967) will have an intermediate value of K_r. Sluggish basins like those of Bay (1969) and Dingman (1966), where detention and storage capacities (at least short term) are high, will have very small values of K_r, indicating that the time of delivery of rainwater to the watershed until its exit at the outlet is very long.

It should be noted that "constant" factors of permeability, vegetation, slope, and soil depth also affect the recession portions of the hydrograph and complicate the mathematical analysis. Some of these including storm direction of movement, and distribution of rainfall, will have a variable effect on the receding limb (Gray and Wigham, 1970). This is why the early portions of the recession frequently cannot be effectively analyzed (Chow, 1964). It is the groundwater recession, where no rain has fallen for long periods of time, that lends itself most accurately to the types of analyses listed above (Holton and Overton, 1963; Chow, 1964).

The magnitude of basin recharge resulting from a given storm is dependent upon two major factors: 1) the moisture deficiency of the basin when the rain commences, and 2) the amount and intensity of precipitation. Minshall (1960) found storm runoff to increase with increases in antecedent soil moisture. As evapotranspiration depletes soil moisture, more recharge is necessary to satisfy the soil moisture deficit. As indices of the moisture condition of a basin, several parameters have been used with the three most important being: 1) days between rainstorms, 2) antecedent stream discharge, and 3) antecedent precipitation (Kohler and Linsley, 1951). The National Weather Service has found this type of relationship to be so universal that empirically derived equations representing antecedent precipitation index (API) are used for runoff predictions throughout the United States (Sittner, Schauss, and Monro, 1969). Kohler and Linsley (1951) give the following equation form for an API:

$$P_a = b_1P_1 + b_2P_2 + \ldots + b_tP_t$$

where P_a is the precipitation index, P_1 is the precipitation one day before the storm, P_2 is the precipitation two days before the storm,

P_t is precipitation t days before the storm, b is an empirically derived constant relating the value of P_a to runoff.

Potter (1949) found that on a watershed where, from storm to storm, major topographic and vegetative influences did not change, for any given intensity and duration of rainfall the amount of runoff and size of the peak discharge will depend on the antecedent soil moisture conditions of the basin. The wetter the basin at the commencement of rainfall, the higher the resulting runoff volume and peak discharge. He also found that an absolute index was not necessary for runoff prediction, but that a general characterization was sufficient. The basin was classified as being in a maximum, minimum, or intermediate state of wetness and runoff could be adequately related to this.

THE STUDY AREA

Location. The Caribou Creek watershed is located in the west-central part of the Yukon-Tanana uplands of the interior of Alaska, at 65° 04' north latitude and 147° 33' west longitude, or approximately 35 miles northeast of Fairbanks. It is part of the Caribou-Poker Creeks Research Watershed which covers about 41.5 mi^2 (26,550 acres). The Caribou Creek drainage, located in the southwestern quadrant of the area, comprises 16 mi^2 (10,240 acres). This study was concerned with the upper 9.55 mi^2 (6080 acres) above the stream gaging site (Figure 5).

Physical description. The Caribou Creek basin is located on sloping terrain with a fairly narrow valley bottom. Approximately 85 to 90 per cent of the area exceeds 100 ft above the channel. The average slope of terrain as estimated from topographic maps is 17 per cent. Elevations range from 840 ft above msl at the gaging site to 2525 ft msl on top of Haystack Mountain.

Three subdrainages are identified. Subdrainage C-1 covers 3.20 mi^2 and has an eastern exposure. Its elevations range from about 1000 ft to 2310 ft above msl. C-2, the smallest subdrainage, represents southern aspect slopes and contains 1.95 mi^2. It ranges in elevation from 1000 ft to 2430 ft above msl. C-3, intermediate in size, has an area of 2.16 mi^2 and is of northern aspect. Its elevations range from 900 ft to 2525 ft. An elevation-area curve for each of these subdrainages is shown in Figure 6. The remaining 2.49 mi^2

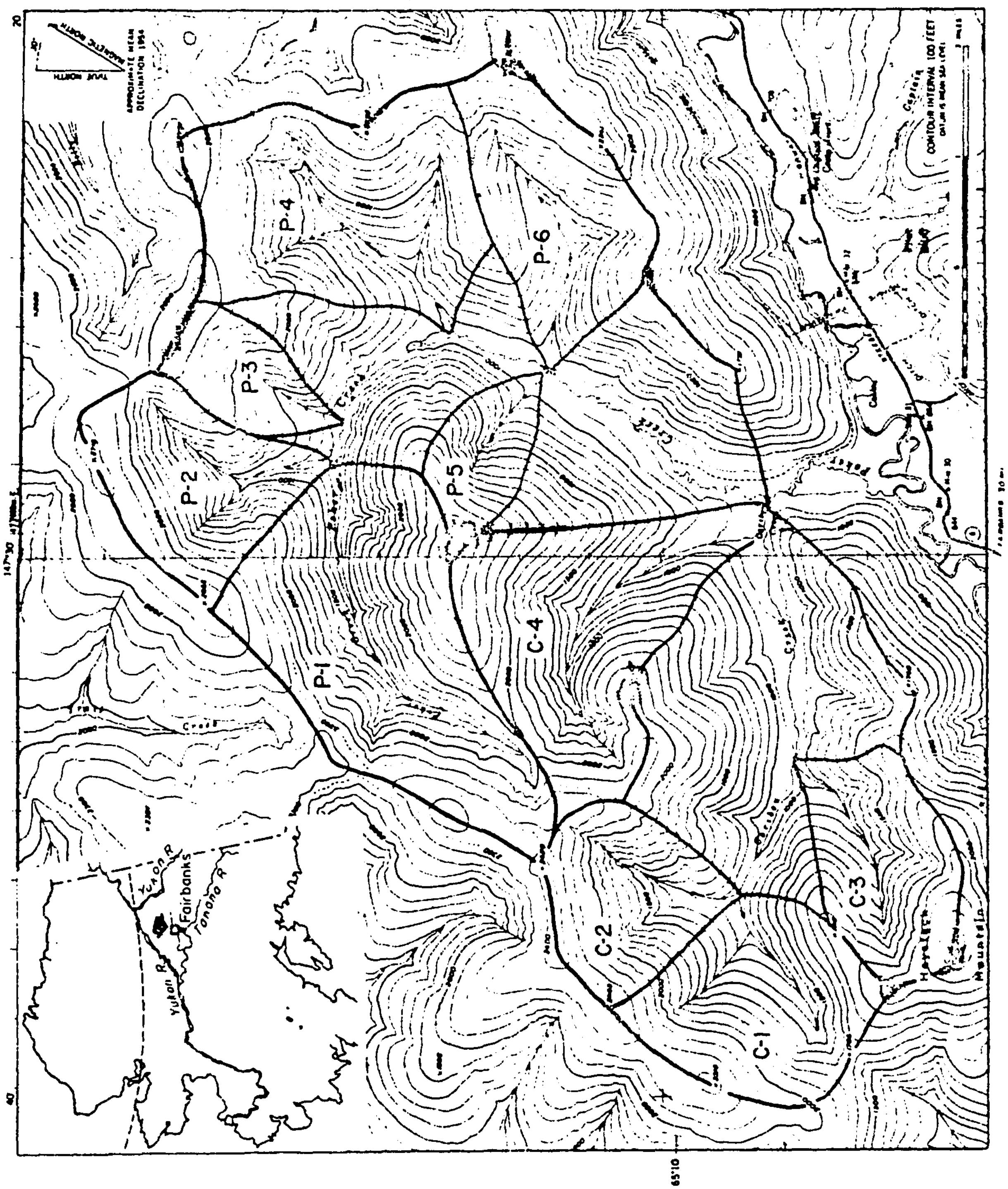


Figure 5

The Caribou-Poker Creeks research watershed near Fairbanks, Alaska

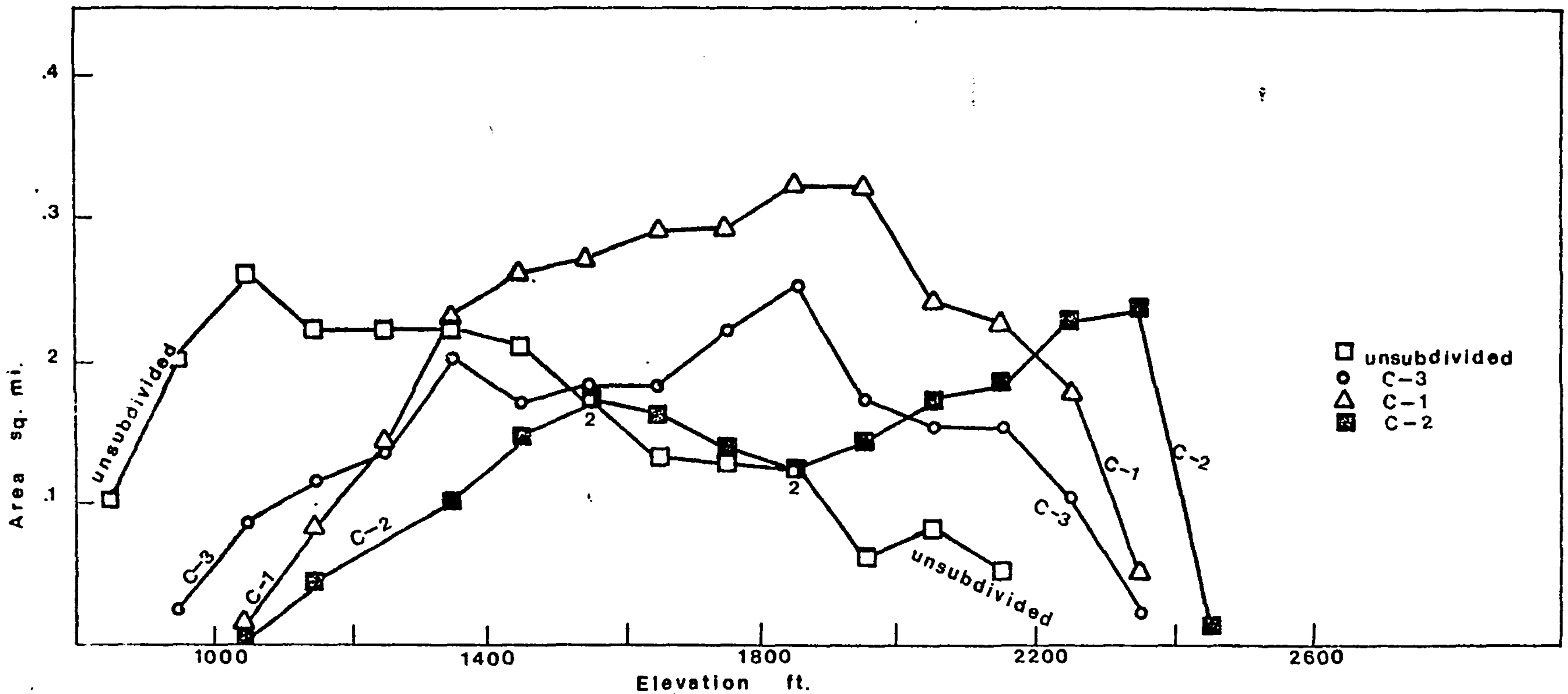


Figure 6

Area-elevation curves for the Caribou Creek watershed above the U.S.G.S. stream gaging station (840 ft above msl). Total area of this portion of the watershed is 9.55 mi^2.

are primarily southern aspect slopes and valley bottom ranging in elevation from about 840 ft to 2100 ft above msl.

Caribou Creek is a second order channel having its headwaters in the mid-reaches of C-1 at about 1300 ft. It is 3.35 mi long from the headwaters to the gaging site and flows generally eastward down an average channel slope of 4 per cent. Beyond the gaging site it continues eastward to its confluence with Poker Creek which flows south into the Chatanika River.

Draining into Caribou Creek above the gaging site are three major first-order tributaries. Progressing down stream, the first drains from about 1900 ft above msl from the northwest corner of C-1 and is 1.25 mi long, the second drains C-2 and 1.80 mi long, and the third drains C-3 and is 2.50 mi long. The total network is 8.90 mi long, giving a drainage density of 0.93, which is very low when compared to most other basins in the United States. Some basins of resistant sandstone in the Appalachian Mountains have drainage densities on the order of 3 to 4. Basins in the Rocky Mountains may have values as high as 50 to 100. Extreme cases, such as the Badlands of South Dakota, have drainage densities of 200 or more (Strahler, 1964). Dingman (1970) found drainage densities for interior Alaska watersheds to range from 1.8 to 9.8. He gives as reasons for the low values in Alaska the following: (1) low intensity precipitation favoring high evapotranspiration, (2) low amounts of rainfall, (3) highly absorptive moss cover, (4) spring runoff occurring over frozen ground precluding the possibility of erosion at that time, and (5) permeable soils.

The Caribou Creek basin is roughly circular with a circularity ratio of 0.92.

Soils. The predominating soil types found on the Caribou Creek watershed are Olnes silt loam, Fairplay silt loam, Ester silt loam, and Gilmore silt loam. The following descriptions are summarized from Reiger, et al. (1971).

The Olnes series dominates the lower south slopes of the Caribou Creek watershed, and consists of well-drained silty soils underlain by fragmented bedrock. It is found mostly on southerly slopes and has no permafrost. Aspen (*Populus tremuloides* Michx.) and paper birch (*Betula papyrifera* Marsh) are the dominant trees supported by the Olnes series. The profile is characterized by a shallow layer of litter overlying a dark grayish-brown silt loam. Schist fragments in the profile increase with depth. This soil series occurs on slopes of from 3 to 45 per cent.

The Ester series soils are silty and found on steep north slopes. They are poorly drained with thin permafrost. The depth to bedrock is usually under 20 in. The profile is characterized by a heavy moss layer over a shallow dark horizon with dark brown to olive gray horizons below. Gravel occurs within a few inches of the surface. The dominant trees are black spruce (*Picea mariana* Mill. Britt, Sterns, and Pogg.) and willow (*Salix* spp.), with *Sphagnum* spp. and other mosses, lichens and shrubs present on the ground. These soils occur on slopes of 12 to 45 per cent.

The Fairplay series soils are found on high ridges and slopes above tree line. They have little permafrost and are fairly well drained. The depth to bedrock is moderately deep. The profile is characterized by a thin organic layer underlain by mottled olive brown silt loam. Schist fragments occur below 18 in and bedrock is below 30 to 40 in. These soils are found on slopes of 3 to 30 per cent.

The Gilmore series is well drained, very shallow, and found primarily on southern aspect slopes. It is free of permafrost and has low moisture supplying capacity. It is similar to the Olnes series except that it is shallower and lighter in color. The profile is characterized by a thin layer of forest litter underlain by 10 to 20 in of dark yellowish-brown to dark brown silt loam containing rock fragments. As the depth increases the proportion of rock fragments increases. Aspen and paper birch dominate the sites despite a few white spruce. This series occurs on slopes ranging from 12 to 45 per cent.

Vegetation. With the exception of small areas of the valley bottom and mountain tops and ridges above tree line, the Caribou Creek watershed is forested. The north-facing slopes, with damp, cool, poorly drained soils, have as the major community type black spruce with associated willow, blueberry (*Vaccinium* spp.), high bush cranberry (*Vibernum edule* Michx.), and labrador tea (*Ledum groenlandicum* Oeder). About 95 per cent of the north-facing slopes are covered with a layer of moss.

South-facing slopes have aspen-paper birch as the principal community type. Associated plants include willow, mosses, and various grasses and forbs. On the upper slopes of C-2 several stringers of white spruce (Picea glauca Mill.) occur in the vicinity of springs. Mosses occur in small thin patches on south-facing slopes, or as heavy moss mats in the white spruce stringers.

The valley bottoms are dominated by blueberry, willow, and stunted black spruce. The soil surface is entirely covered by a thick carpet of mosses that ranges in depth from 8 to over 24 in. Aspen and paper birch occur on the northerly fringes of the bottom on the better drained soil.

Climate. The climate of the interior of Alaska is continental with long cold winters and short warm summers (Critchfield, 1966). The mean annual temperature in the interior ranges from 20° to 25°F. The coldest month is January with a mean of -10° to -20°F and July, the warmest month, has a mean of 50° to 60°F (Searby, 1969).

Caribou Creek is marked by significant variations in temperature from one point to another due to the mountainous terrain. Data from the top of Caribou Mountain at 2537 feet and from C-2 at 1100 feet indicate that inversion conditions exist throughout the winter (Table 1a). The mean temperature for January, February, and March on Caribou Mountain was 10°F, while at the lower station it was 7°F in 1970 (Slaughter, 1971). The extent of the inversion can be more clearly seen by comparing January temperatures. The mean on Caribou Mountain was -5°F while on C-2 at 1100 feet the mean was -12°F. Nearby, Fairbanks had a mean for January of -16.2°F, which was 5.1°F below normal.

TABLE 1a

Mean air temperature for Caribou Mountain (el. 2537 ft above msl), C-2 (el. 1100 ft above msl), Fairbanks (el. 454 ft above msl), and the Fairbanks departure from normal for January through March, 1970.

	Caribou Mt. El. 2537	C-2 El. 1100	Fairbanks El. 454	Fairbanks Departure from Normal
January	-5°	-12°	-10.2°	-5.1°
February	16°	10°	8.0°	+10.9°
March	20°	21°	20.9°	+12.0°

TABLE 1b

Mean air temperature for Caribou Mountain (el. 2537 ft above msl), C-2 (el. 1100 ft above msl), Fairbanks (el. 454 ft above msl), and the Fairbanks departure from normal for June through August, 1969.

	Caribou Mt. El. 2537	C-2 El. 1100	Fairbanks El. 454	Fairbanks Departure from Normal
June	52°	---	64.9°	+6.5°
July	42°	56°	59.4°	-0.3°
August	37°	45°	49.8°	-4.5°

The Fairbanks mean for the three winter months, however, was $4^{o}F$ which was $6^{o}F$ above normal (U.S. Dept. of Commerce, 1970).

Fairbanks temperatures are possibly lower than those on Caribou Creek during the winter because, in the broad, flat Tanana valley, the inversion caused by intense radiative cooling at the surface during the long hours of darkness is among the strongest in the world (Benson, 1970). Caribou Creek is dominated by raised relief in which the air drainage is somewhat better. It is removed from the strong inversion which develops on the Tanana valley floor and temperatures are more moderate.

In the summer, temperatures in the Tanana valley lowlands are higher than those on Caribou Creek due to their lower elevation. The sloping terrain, especially on northern aspects, also provides topographic sheltering on Caribou Creek. From existing data, summertime temperatures on Caribou Creek for the months of June, July, and August of 1969 averaged $43^{o}F$ (Table 1b). This compared to $59^{o}F$ in Fairbanks for the same period which represented a $+2.4^{o}F$ departure from normal (U.S. Department of Commerce, 1970).

The mean annual precipitation ranges from 10 to 15 in in the Alaskan interior. Over half occurs in June, July, and August as heavy rainshowers, due to the formation of convective cells in the late afternoon (Dingman, 1970). Frontal and air mass activity during this time of the year is conspicuously lacking. During the long winter months, moderate amounts of snow, averaging 50 to 60 in, falls on the interior (Searby, 1969). Snow covers the ground in the Fairbanks area from October into mid April with an average duration of about 214 days (Dingman, 1970).

There are no complete records of the annual precipitation on the Caribou Creek watershed. Nearby Fairbanks receives an annual average of 11.29 in, most of which falls in the summer as rainshowers and thunderstorms (Searby, 1969). Tables 2 and 3 give day by day comparison of precipitation totals from Caribou Creek and the Fairbanks office of the National Weather Service. The monthly totals appear to be greater on Caribou Creek. It could be expected that Caribou Creek may receive more precipitation than Fairbanks due to orographic enhancement.

Snow course data indicate that on February 4, 1970, there was an average of 10 in of snow on the ground with a water equivalent of 1.4 in. By March 3, the total had increased to 12 in on the ground and a water equivalent of 1.8 in (Slaughter, 1970b). In Fairbanks on February 4, 10 in of snow with a water equivalent of 1.1 in were on the ground and on March 3, this had decreased to 9 in on the ground with a water equivalent of 1.1 in (U.S. Department of Commerce, 1970). Dingman (1970) found that snow remained on the ground for an average of 214 days in the vicinity of Glen Creek, mid-way between Fairbanks and Caribou Creek. The mean length of the frost-free period in Fairbanks is 97 days, but is probably less on Caribou Creek. In Fairbanks the average date of the last spring freeze is May 24, and that of the first autumn freeze is August 29 (U.S. Department of Commerce, 1967).

TABLE 2

A comparison of precipitation at the Fairbanks office of the National Weather Service and average precipitation on the Caribou Creek watershed for June, July, August, and September 1970

	JUNE		JULY		AUGUST		SEPTEMBER		
	Fair-banks	Caribou Creek	Fair-banks	Caribou Creek	Fair-banks	Caribou Creek	Fair-banks	Caribou Creek	
1	0.02				0.07	0.30			1
2	0.02				0.04				2
3			0.53	0.17	0.14	0.56			3
4				0.02	0.01	0.01	0.01		4
5	0.03		0.01	0.50	0.18	0.59	0.01	0.06	5
6						0.05		0.01	6
7			0.01		0.17	0.24		0.02	7
8			0.08			0.01			8
9									9
10	0.11	0.36	0.12	0.38	0.02	0.12			10
11	0.06				0.09	0.02			11
12					0.05	0.03	0.09	0.28	12
13							0.13	0.67	13
14	0.05	0.30	0.08	0.05				0.40	14
15	0.07	0.08	0.03	0.23		0.02	0.01		15
16		0.06	0.13	0.23	0.01	0.21	0.06	0.02	16
17	0.01	0.04					0.13	0.44	17
18			0.24	0.25					18
19			0.11	0.11					19
20			0.05	0.05	0.01	0.02		0.01	20
21	0.44	0.77**			0.05	0.10	0.02		21
22	0.43	0.05**			0.02	0.28	0.07		22
23	0.02	0.03				0.05	0.01	0.02	23
24	0.38				0.27	0.47		0.05	24
25	0.09			0.02			0.05		25
26	0.23						0.04	0.03	26
27	0.09	0.03							27
28	0.34		0.07				0.01		28
29	0.16	0.16	0.02				0.01		29
30	0.02	1.10	0.06	0.06	0.72	0.36			30
31	----	----	0.26	0.09	0.13				31
Tot	2.57	2.98	1.81	2.16	1.98	3.44	0.65	2.01	Tot
Dep	+1.18	----	-0.03	----	-0.22	----	-0.45	----	Dep

** = Estimated
Voids = No Data

TABLE 3

A comparison of the precipitation at the Fairbanks office of the National Weather Service and average precipitation on the Caribou Creek Watershed for June, July, August, and September 1971

	JUNE		JULY		AUGUST		SEPTEMBER		
	Fair-banks	Caribou Creek	Fair-banks	Caribou Creek	Fair-banks	Caribou Creek	Fair-banks	Caribou Creek	
1			0.27	0.14	0.20	0.08	0.06		1
2			0.01	0.06	0.03	0.18	0.44	0.44	2
3				0.03	0.05	0.28	0.34	0.42	3
4				0.04	0.11	0.09	0.21	0.26	4
5				0.05					5
6					0.04	0.25	NO DATA		6
7				0.01	0.04	0.51			7
8					0.27	0.78			8
9					0.46	0.63			9
10					0.10	0.35			10
11			0.12	0.68					11
12	0.04		0.75	1.17		0.02			12
13	0.08	0.01	0.24	0.41		0.01			13
14	0.08	0.05	0.08	0.20					14
15				0.02					15
16				0.09					16
17				0.04					17
18	0.01	0.01		0.18					18
19		0.44	0.04	0.05	0.20	0.14			19
20	0.03	0.03	0.03	0.18	0.41	0.12			20
21	0.02	0.36		0.01	0.01				21
22									22
23					0.32	0.35			23
24				0.03					24
25						0.03			25
26									26
27			0.28	0.47					27
28			0.05	0.02					28
29	0.01	0.03	0.10	0.10		0.01			29
30	0.04	0.11	0.02	0.09	0.08	0.07			30
31	----	----	0.09	0.20					31
Tot	0.31	1.04	2.08	4.27	2.32	3.90			Tot
Dep	-1.08	----	+0.24	----	+0.12	----			Dep

Voids = No Data

MATERIALS AND METHODS

Precipitation and streamflow data for 1970 and 1971 were used in this study. The precipitation gages were installed by the National Weather Service, and the United States Army Corps of Engineers Cold Regions Research and Engineering Laboratory (CRREL) in 1969. They were serviced throughout the data collection period by personnel from CRREL, Fort Wainwright, Alaska. The United States Geologic Survey (U.S.G.S.) installed the stream gaging station in 1969 and have maintained it since that time. The raw data for use in this project were gathered from the respective agencies.

The three basic categories of the data analysis are: (1) precipitation, (2) runoff, and (3) the precipitation-runoff relationship. Precipitation presents the least difficulty within the bounds of the available data and will be discussed first. Runoff presents myriad variations and requires more thorough treatment. Precipitation-runoff will be dealt with last.

Precipitation data were gathered from a network of three Belfort weighing-bucket recording rain gages each equipped with a one-week revolving drum. Strip charts were analyzed for daily and hourly rainfall amounts. Standard eight-inch rain gages were located with each recording gage. The total increments in these were measured at the same time strip charts for the recording gages were collected. These weekly totals were checked against the strip chart weekly totals to verify catch.

The rain gage sites were at the 2100 ft contour of C-2 and C-3 and at the main gaging site at 840 ft above msl (Figure 5). The C-3 gage was 2.0 mi from the main site and 3.1 mi from the C-2 gage. The C-2 gage was 2.3 mi from the main site. One-third of the watershed, C-1, was ungaged. In addition, only 19 per cent of the watershed was composed of valley bottom, yet it was represented by one-third of the gage network. To avoid underestimating the contribution of the slopes it was determined, by calculating the area of the valley bottom, that the gage at the main site should not be weighted more than 20 per cent and that the remaining percentage should be divided between the two gages on the slopes. For these reasons, a Thiessen network was drawn and each gage weighted as follows: (1) 0.44 to C-3, (2) 0.37 to C-2, and (3) 0.19 for the main site. The total precipitation for a given time period caught by each gage was multiplied by its weight and the weighted precipitation from all the gages summed, giving the weighted total precipitation over the watershed. When one or more of the recording gages was not functioning the average rainfall recorded by the remaining one or two gages was used as the average precipitation on the basin.

The time continuity of each strip chart was checked by comparing the time-on and time-off recorded on each chart. Any variation between times recorded on the charts and times indicated by the pen was considered linear and the time scale was adjusted accordingly. Very little difficulty was encountered in this phase of the data analysis.

The strip chart divisions were bi-hourly. For daily precipitation amounts, the level at the day's beginning was subtracted from that at the day's end and the result recorded. For hourly precipitation, each division on the chart was divided in half and the level at the beginning of the time period was subtracted from the level at the end, resulting in hourly intensities.

In the event that no recording gage data existed for a time period, the increments in each storage gage were averaged giving a precipitation amount for the period. From National Weather Service records for Fairbanks, College, Gilmore Creek, and the University Experiment Station, the percentage of total rainfall for each day for the omitted interval was calculated. These percentages were applied to the total from the storage gages. This gave a precipitation amount for each day lacking recording gage data. Fortunately, only one storm period required this procedure (June 21-22, 1970). The storm was not used in any subsequent analyses except for storm runoff versus total precipitation. On the hyetographs the display of precipitation obtained as described above is indicated by an asterisk. A double asterisk appears by these data in the appropriate tables.

Seasonal records began in June, 1970 and 1971. It was necessary to compile the average-depth-of-rainfall data from continuous strip charts obtained from each rain gage. The weighted daily and hourly rainfall amounts were plotted as hyetographs.

Hourly rainfall rates were obtained for use with the hourly discharge hydrographs. The rates were derived from the continuous charts by subtracting the initial hourly reading from the final hourly reading for each chart. If readings for all three gages were available they were weighted according to their assigned Thiessen weights. If only one or two gages were operational, the hourly intensities were obtained by an arithmetic mean.

Numerous rainfall events during both summers caused no appreciable rise in stream discharge. Rainfall events were defined as either "episodes" or "storms" for the purposes of this study. An episode was a period of rainfall for which there was no well-defined hydrologic response. A storm was a rainfall event which resulted in a hydrograph event with a well-defined rising limb, crest segment, and falling limb. A single storm period, as identified by the mean daily discharge hydrograph, could actually consist of two or more isolated storm periods when examined on an hourly basis. This resulted because, on a daily basis, discontinuous blocks of rainfall were lumped together in a gross daily sum. Likewise, instantaneous discharge characteristics are masked by the mean daily figure because two or more rises occurring in the space of 24 hours are lumped together in the daily mean figure.

It was necessary to obtain mean intensities for storms plotted on an hourly basis. This was accomplished by summing the mean hourly rates and dividing by the storm duration. Mean intensities were used in early phases of precipitation-runoff analyses and will be discussed later. For the mean daily hydrograph intensities were not utilized due to the ambiguity resulting from irregular bursts of rainfall throughout the day.

Streamflow. Stream level for the summer of 1970 was monitored by a float-operated Fisher-Porter digital recorder in a walk-in shelter with a six by six foot stilling well. In September, 1970, the recorder was replaced with a Stevens A-35 continuous level recorder and was maintained throughout the summer of 1971. Rating tables for each were developed to relate stage to discharge. While mean daily discharge was supplied by the U.S.G.S., the hourly values were read directly from the strip charts and converted to discharge in cubic feet per second (cfs) with the appropriate rating table. The data plotted against time resulted in both mean daily and hourly discharge hydrographs. This phase of the data collection was carried out by the U.S.G.S.

The runoff analyses were accomplished in the two phases of mean daily flow for both summer periods and hourly discharge for isolated storms. The major analysis techniques for each phase were essentially the same.

General runoff characterization. Mean daily hydrographs were used for general characterization of runoff on Caribou Creek. These data were graphically represented by a smooth curve.

Eight major rises were noted in 1970 with four showing multiple peaks, complex rises or complex recessions. There were three principal events in 1971, all of which were complex. Total daily runoff was expressed in acre-feet (ac-ft). The total hydrographs for the summer periods were then examined for a general separation of baseflow from stormflow.

During rain-free periods the recessions from the peaks appeared to approach an equilibrium. Due to the presence of areas with moderately deep, well-drained soils on the watershed, it was assumed that the groundwater contribution would sustain streamflow through dry periods. In addition, the thick moss cover on the valley bottom and north aspect slopes would conceivably contribute to dry weather flow. For a general characterization, therefore, the asymptotes to the recessions were estimated and all flow below the resulting line was considered groundwater discharge and that above as storm runoff.

To separate baseflow from storm runoff for individual storms more detail was required. One-half of the events in 1970 and all of the events in 1971 were complex, while 1971 was further complicated by the fact that no events produced a rain-free recession. Also, in 1971 there were only three rain-free periods longer than 3 days each. The 1970 hydrographs, however, contained several rainless periods and had a number of recessions at least part of which were without rain.

The recession relationships of time to the logarithm of discharge approximated straight lines. This indicated a stable relationship between storage and outflow.

Recessions. The form commonly given for outflow from storage (Chow, 1964) is:

$$-\frac{dS_t}{dt} + q_t = 0 \tag{1}$$

where S_t is storage and q_t is discharge. If q_t is expressed as,

$$q_t = -kS_t \tag{2}$$

where k is constant, substitution of (1) into (2) results in,

$$\frac{dS_t}{dt} = -kS_t \tag{3}$$

and,

$$\frac{dS_t}{S_t} = -kdt \tag{4}$$

Integrating,

$$\ln S_t = -kt + C \tag{5}$$

At $t = 0$, $\ln S_0 = C$ so,

$$\ln S_t - \ln S_0 = \ln \frac{S_t}{S_0} = -kt \tag{6}$$

Thus,

$$\frac{S_t}{S_0} = \exp(-kt) \tag{7}$$

or,

$$S_t = S_0 \exp(-kt) \tag{8}$$

Recalling eq. (2), solving it for q_t and q_o, and substituting into eq. (8) results in,

$$q_t = q_0 \exp(-kt) \tag{9}$$

Graphically, with $\ln q_t$ as the ordinate and t as the abscissa $\ln q_0$ would be the y-axis intercept and $-k$ would be the slope of the line.

Flow separation -- Mean daily discharge. The recessions of 1970 showed two distinct curves differentiated by a change in slope at an average discharge of 8 cfs. The average k for each segment of the essentially rain-free recessions was computed. Two segments of a master recession curve for 1970 were derived from the average k values for each of the corresponding segments of the individual recessions. The point where the slope changed was assumed to be where storm runoff dominated above and groundwater discharge below. Where possible, recessions complicated by rainfall were extended by matching the master curve to the peak and extending the observed recession beyond its point of departure to the value of the average change in slope. Groundwater could then be separated by continuing the recession prior to the point of rise horizontally to the point under the hydrograph peak and then to its intersection with the recession at the change-in-slope point. In a few cases the master recession curve was inadequate and recessions were approximated.

Sufficient rain-free recessions were not available in 1971 to warrant using the same procedure as in 1970. However, it was possible to construct a master recession curve from recession segments that appeared to be rain-free or not seriously disturbed by rain. A log-normal plotting revealed that a marked change in slope occurred. Both segments had a smaller slope than their counterparts in 1970, and the change in slope came at a higher rate of discharge. Despite all events in 1971 being complex, they were separated in a manner similar to that of 1970. Storm runoff for both years was then obtained by planimetry.

It was thought that the above procedure would yield some insight into the hydrologic response of the basin when analyzed with rainfall data. Assumptions for this analysis were basically that baseflow decreases as the stream level rises due to abstraction by the unsaturated bank and that the recession follows the derived exponential decay if uncomplicated by rainfall (Linsley, Kohler, and Paulhus, 1958).

The rainy 1971 summer season was marked by high sustained flows and very slow recessions, even during rainless periods. The presence of thick moss mats on the basin's bottom and north aspect slopes lead to speculation that this might be a source of delayed flow that was not truly groundwater discharge. Similar situations have been noted in other boggy areas and forests (Bay, 1969; Dingman, 1970; Hewlett and Hibbert, 1967). In an attempt to account for this, complete recessions were not altered or estimated with the master depletion curve. The discharges at which a change in slope was observed for the three 1971 events were averaged and subtracted from the discharge at the point of initial rise. A straight line was then drawn from the latter to the former. The assumption was that the recession would show a change in slope where discharge changed from predominately storm runoff to predominately contribution from goundwater and moss. Discharge above the separation line was termed quick flow and that below, delayed flow (Hewlett and Hibbert, 1967). Results of this separation technique were of limited value.

The third separation technique employed a horizontal delineation of baseflow from storm runoff. Since the computation of peak discharge assumed triangular flow from the point of rise to the peak, the separation assumed that baseflow remained constant from the point of rise. A straight horizontal line was drawn from the initial rise to its intersection with the receding limb. Where necessary the master recession curve for the appropriate year was used to extend recessions complicated by rainfall so as to intersect with the horizontal line.

The objective of the mean daily discharge analyses was to develop a separation that could be used on all simple events uniformly and that could be applied to complex events if the proper modifications were made (normalizing the recessions and separating multiple peaks). As long as the same separation is used for each rise the true separation of baseflow from stormflow is not critical, since in any event the actual contribution of each is not known. The method employing continuation of the prior recession to the point under the peak and then intersection with the receding limb at the separation point, and the horizontal separation method described immediately above come closest to satisfying these demands. The first method involves the use of a separation point based on the average value of the change in slope of observed recessions. The derived separation point, therefore, occurs above or below the observed value in most cases. The method employing the horizontal separation line has the advantage that the baseflow and the length of the recession are fixed by the discharge at the time of the initial rise and the peak, respectively. One drawback is the assumption that no change in baseflow occurs throughout the event.

Hourly discharge. The hourly discharge hydrographs for both 1970 and 1971 were analyzed in much the same manner as the mean daily flow hydrographs. The 1970 events were simple while half of those in 1971 were highly complex. The 1970 recession characteristics were derived by averaging the slopes obtained by plotting discharge versus time on log-normal paper. The hourly hydrographs showed a definite change in slope so the point of occurrence was averaged with the result taken as the point of change on the master recession curve.

The 1971 hydrographs were marked by very few rainless periods.

Those that were available were plotted on log-normal paper. A difference in slope was noted between the early and late segments of the recessions. Because it was not possible to determine the actual turning point from the available data it was assumed to be the same as that for 1970.

Flow separation for hourly hydrographs. Separation of groundwater discharge from stormflow was accomplished in two ways. First, the number of hours after the peak that the change in slope of rainless recessions occurred was averaged, with the result being taken as the point of separation and marked on each event. For complex events the master recession curve was used to find this point. Next, the recession prior to the point of initial rise was extended to directly under the peak and then a straight line drawn from there to the separation point marked on the receding limb. For the storm of August 6-12, 1971, the separation point was set after the last "bump" on the receding limb. The latter modification was necessary due to the extreme complexity of the event.

The second technique employed a horizontal separation. Since the computation of peak discharge required that triangular flow be assumed from the point of rise to the peak, the separation assumed horizontal baseflow throughout the event, as for mean daily discharge. As will be seen, in the computation of peak flow it is assumed that storm runoff is all runoff above a straight line extended from the point of initial rise to its intersection with an exponential discharge decay. If the rise began from zero baseflow, this would take place in infinity, but since there was some prior flow it takes place when the decay reaches a value equal to flow immediately before the rise began if there is no change in baseflow.

Precipitation-runoff relationships. A primary objective of this study was to deduce discernible relationships between the precipitation falling on the watershed and the runoff caused by it. This was accomplished by analyzing corresponding discharge hydrographs and hyetographs.

Hydrologic response. The hydrologic response for both mean daily and hourly events was obtained for each storm by dividing the total storm runoff by the total precipitation. Precipitation occurring after the peak was discounted except for the storm of August 6-12, 1971, when all precipitation occurring during the event was included due to the modification in separation technique. It was hoped that this traditional parameter would be useful in predicting storm runoff from precipitation data alone.

To test further a precipitation parameter against total storm runoff regression analyses of total rainfall amount versus storm runoff, as calculated by the horizontal separation method, were used for both mean daily and hourly storms. These relationships are critical for hydrograph modeling as will be seen below.

Hydrograph modeling. Modeling the peak of the hydrograph is possible if the following are known, derived, or assumed: storm runoff, recession constant, time of rise, and linear flow increase from the point of rise to the peak. Holtan and Overton (1963) gave the following general form for the computation of the peak:

$$q_p = P_e(0.5\ D + m)^{-1} \tag{10}$$

where q_p is the peak flow (cfs), P_e is the precipitation excess or storm runoff (cfs-hr), D is the duration of the hydrograph rise (hr), and m is the slope of a linear relationship between storage and discharge.

A modification is needed to fit eq. (10) to the observed hydrographs (Dingman, 1970). A linear increase in flow is assumed from the point of rise to the peak (triangular flow). The volume of runoff contained therein is:

$$V_r = 0.5T_r\ q_p \tag{11}$$

where V_r is the runoff (ft^3), T_r is the time of rise (hr), q_p is discharge at peak flow in (cfs).

Runoff during the last half of the event is given by:

$$V_f = \int_0^\infty q_p e^{-kt}\ dt \tag{12}$$

where V_f is the volume of runoff after the peak (ft^3), k is the recession constant (hr^{-1}), and t is the time (hr). With q_p constant:

$$V_f = \lim_{C \to \infty} q_p \int_0^C e^{-kt}\ dt \tag{13}$$

$$= q_p\ k^{-1}$$

Total runoff from the storm then is given by:

$$V_t = V_r + V_f \tag{14}$$

or,

$$V_t = 0.5\ T_r q_p + q_p k^{-1} \tag{15}$$

Solving for q_p yields:

$$q_p = V_t(0.5\ T_r + k^{-1})^{-1} \tag{16}$$

Furthermore, the actual peak observed would require that baseflow discharge be added.

To position the peak, the lag time was obtained by regression analysis. Lag time was defined as the time lapse in hours between the centroid of the rainfall contributing to the peak and the peak itself. The time of rise of the hydrograph from initial discharge increase to the peak was related to the rainfall duration from its initiation to either the end of rainfall or the peak of the hydrograph, whichever came first.

With regression equations for runoff due to a given amount of storm rainfall, time lag, time of rise, and with a recession constant and expression for peak flow due to a given runoff volume, it is possible to model the entire storm hydrograph in a way similar to that used by Holtan and Overton (1963), with the only parameter needed being the hourly hyetograph and if actual total discharge is desired, baseflow at time t=0. The procedure can be followed in steps. First, the hyetograph must be plotted and total rainfall amount calculated. Next, the total volume of runoff from a rainfall event can be obtained from the regression equation relating runoff to storm rainfall. The peak may be positioned by determining the time of rise and the lag time, expressions for which are given in the Results and Discussion, and then the peak may be calculated from eq. (17). From the peak, the recession may be calculated directly from the recession equation. Total volume of runoff can be obtained by planimetry.

RESULTS AND DISCUSSION

<u>Precipitation</u>. From the averages of two summers' data it appeared that seasonal precipitation was quite substantial on the Caribou Creek watershed. An average of 8.90 in of rainfall was observed for June through August, 1970 and 1971. Averages for the two summers indicate that more rain fell in August than any other summer month, with a two year mean of 3.67 in. The monthly pattern, however, was inconsistent with July, 1971, having the largest individual total of 4.27 in. June had the lowest mean rainfall during the two summers, with 2.03 in. June, 1970, also had the smallest single month total, of 1.08 in. For the two summer periods the largest 24 hour total was 1.17 in on July 12, 1971. The frequency of days with rain in June, July, and August averaged 10, 18, and 17.5 days respectively. July, 1971, experienced 23 days with rain.

Rainfall intensities were generally quite low. For the seven storms in 1970 and 1971 with hourly rainfall data available, the average intensity was only 0.10 in hr^{-1}. The highest intensity observed was 0.21 in hr^{-1} for a five hour storm while the lowest was 0.03 in hr^{-1} for a 70 hour storm. Most storms were of short duration and characterized by a central burst of relatively high intensity precipitation which trailed off rapidly--the exception being the low intensity 70 hour storm of August 6-9, 1971.

The distribution of rainfall during the two years of data collection was inconsistent. The daily precipitation catch for each gage (Tables 4 and 5) indicated no general pattern.

TABLE 4

Daily recording gage catch (in) at C-2, C-3, and the main site and Thiessen weighted monthly means for June through September, 1970, on the Caribou Creek Watershed.

	JUNE			JULY			AUGUST			SEPTEMBER			
	C-2	C-3	Main Site	C-2	C-3	Main Site	C-2	C-3	Main Site	C-2	C-3	Main Site	
1	*	*	*	*	*	0.00	*	0.40	0.20	0.00	0.00	0.00	1
2			*			0.00	*	0.00	0.00	0.00	0.00	0.00	2
3			0.00			0.17	*	0.55	0.57	0.00	0.00	0.00	3
4			0.00			0.02	*	0.02	0.00	0.00	0.00	0.00	4
5			0.00			0.50	0.55	0.60	0.71	0.10	0.04	0.00	5
6			0.00			0.00	0.05	0.04	0.03	0.02	0.00	0.00	6
7			0.00			0.00	0.35	0.18	0.17	0.02	0.02	0.10	7
8			0.00			0.00	0.00	0.02	0.00	0.00	0.00	0.00	8
9			0.00		*	0.00	0.00	0.01	0.00	0.00	*	0.00	9
10			0.36		0.44	0.32	0.11	0.04	0.09	0.00	*	0.00	10
11			0.00		*	0.00	0.02	0.02	0.00	0.00	*	0.00	11
12			0.00		0.00	0.00	0.05	0.02	0.00	0.09	*	0.47	12
13			0.00		0.00	0.00	0.00	0.00	0.00	0.80	*	0.58	13
14			0.30		0.05	0.05	0.00	0.00	0.00	0.80	*	0.00	14
15			0.08		0.20	0.25	0.05	0.00	0.00	0.00	*	0.00	15
16			0.06		0.20	0.26	0.25	0.20	0.14	0.00	0.03	0.04	16
17			0.04		0.00	0.25	0.00	0.00	0.00	0.40	0.50	0.36	17
18			*		0.26	0.05	0.00	0.00	0.00	0.00	0.00	0.00	18
19					0.17	0.05	0.00	0.00	0.00	0.00	0.00	0.00	19
20					0.00	0.11	0.00	0.04	0.00	0.03	0.00	0.00	20
21				*	0.00	0.00	0.10	0.08	0.10	0.00	0.00	0.00	21
22				0.00	0.00	0.00	0.31	0.28	0.25	0.00	0.00	0.00	22
23		*		0.00	0.00	0.00	0.00	0.11	0.00	0.00	0.00	0.11	23
24		0.00		0.00	0.00	0.00	0.45	0.48	0.48	0.00	0.11	*	24
25		0.00		0.00	0.05	0.04	0.00	0.00	0.00	0.00	0.00	*	25
26		0.00		0.00	0.00	0.00	0.00	0.00	0.00	0.07	0.00	*	26
27		0.04		0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	*	27
28		0.00		0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	*	28
29		0.11		*	0.00	0.00	0.00	0.00	0.00	0.00	0.00	*	29
30	*	1.05	*	*	0.07	0.05	0.18	0.47	0.40	0.00	0.00	*	30
31	----	----	---	*	0.08	0.10	0.00	0.00	0.00	----	----	----	31
Tot		1.20		0.00	1.52	2.17	2.47	3.56	3.14	2.33	0.70	1.56	Tot
T.M.		*			2.16			3.44			2.01		T.M.

* = no data
T.M. = Thiessen Mean

TABLE 5

Daily recording gage catch (in) at C-2, C-3, and the main site and Thiessen weighted monthly means for June through August, 1971, on the Caribou Creek watershed.

	JUNE			JULY			AUGUST			SEPTEMBER			
	C-2	C-3	Main Site	C-2	C-3	Main Site	C-2	C-3	Main Site	C-2	C-3	Main Site	
1	0.00	0.00	0.00	0.13	0.16	0.12	0.08	0.08	0.06		NO DATA		1
2	0.00	0.00	0.00	0.08	0.02	0.10	0.17	0.22	0.11				2
3	0.00	0.00	0.00	0.03	0.03	0.00	0.22	0.33	0.32				3
4	0.00	0.00	0.00	0.08	0.03	0.08	0.12	0.08	0.05				4
5	0.00	0.00	0.00	0.13	0.00	0.00	0.00	*	0.00				5
6	*	0.00	0.00	0.00	0.00	0.00	0.27	*	0.23				6
7	*	0.00	0.00	0.04	0.00	0.00	0.57	*	0.45				7
8	*	0.00	0.00	0.00	0.00	0.00	0.80	*	0.77				8
9	*	0.00	0.00	0.00	0.00	0.00	0.77	*	0.49				9
10	*	0.00	0.00	0.00	0.00	0.00	0.42	*	0.29				10
11	*	0.00	0.00	0.70	0.60	*	0.00	*	0.00				11
12	*	0.00	0.00	0.97	1.37	*	0.00	0.00	0.00				12
13	*	0.00	0.08	0.40	0.42	*	0.04	0.08	0.10				13
14	*	0.09	0.00	0.12	0.28	*	0.00	0.00	0.00				14
15	*	0.00	0.00	0.00	0.02	0.02	0.00	0.00	0.00				15
16	*	0.00	0.00	0.00	0.15	0.10	0.00	0.00	0.00				16
17	0.00	0.00	0.00	0.12	0.00	0.00	0.00	0.00	0.00				17
18	0.04	0.00	0.00	0.18	0.17	0.17	0.00	0.00	0.00				18
19	0.29	0.40	0.80	0.12	0.03	0.00	0.15	*	0.13				19
20	0.02	0.02	0.04	0.20	0.16	0.19	0.17	*	0.07				20
21	0.22	0.55	0.20	0.00	0.03	0.00	0.00	*	0.00				21
22	0.00	0.00	0.00	0.00	0.00	0.00	0.00	*	0.00				22
23	0.00	0.00	0.00	0.00	0.00	0.00	0.60	*	0.11				23
24	0.00	0.00	0.00	0.03	0.04	0.00	0.00	*	0.00				24
25	0.00	0.00	0.00	0.00	0.00	*	0.05	*	0.05				25
26	0.00	0.00	0.00	0.00	0.00	*	0.00	*	0.00				26
27	0.00	0.00	0.00	0.54	0.41	*	0.00	0.00	0.00				27
28	0.00	0.00	0.00	0.04	0.00	*	0.00	0.00	0.00				28
29	0.04	0.04	0.00	0.10	0.07	0.14	0.00	0.00	0.00				29
30	0.12	0.12	0.10	0.10	0.07	0.08	0.01	0.02	0.00				30
31	----	----	----	0.30	0.16	0.12	0.06	0.08	0.06				31
TOT	0.73	1.22	1.22	4.41	4.22	1.16	4.50	0.89	3.24				TOT
T.M.		1.08			4.27			3.90					T.M.

* = no data
T.M. = Thiessen Mean

Data continuity was not good and restricted comparison between the gages. T-tests of differences in the July and August, 1970, catches between C-2 (2100 ft) and the main site indicated no significant difference at the 0.05 level. It was not possible to compare June data due to data discontinuity between C-3 and C-2. Evidence indicated that the C-3 rainfall during July and August, 1970, was slightly higher than that for C-1 and C-2.

Data continuity was better in 1971, but significant omissions existed (Table 5). A paired difference t-test revealed no significant difference in catch at the 0.05 level between the main site and C-3 for June and July. The catch during the same period on C-2 was significantly less in June and more in July at the 0.05 level. There was no difference in catch between C-2 and the main site in August but C-3 revealed a 0.22 in greater catch than the other two.

There did not appear to be a pronounced orographic influence on the watershed. From examination of the available data it did seem that C-3 received more precipitation than either the main site or C-2. The total catch for C-3 in June, July, and August, 1970 and 1971, for the days it had data in common with the other gages was 5.78 in. The respective catches for the same days for C-2 and the main site were 5.55 in and 5.37 in.

Runoff. Runoff for June through September, 1970, was 1907 ac-ft (Table 6; Figure 7). June had the least runoff and precipitation (Table 4). There was negligible snowmelt effect noticed during June since the total discharge due to spring snowmelt was slight and oc-

TABLE 6

Monthly runoff (ac-ft) for June through September, 1970, and June through August, 1971, from Caribou Creek.

Year	June	July	August	September
1970	373	554	500	480
1971	859	843	1203	---

TABLE 7

Recession constant (k) for four storms in July and August, 1970, from mean daily discharge hydrographs.

Date	K (day^{-1})
July 5-7, 1970	0.416
July 11-13, 1970	0.306
August 5-7, 1970	0.557
August 25-26, 1970	0.383

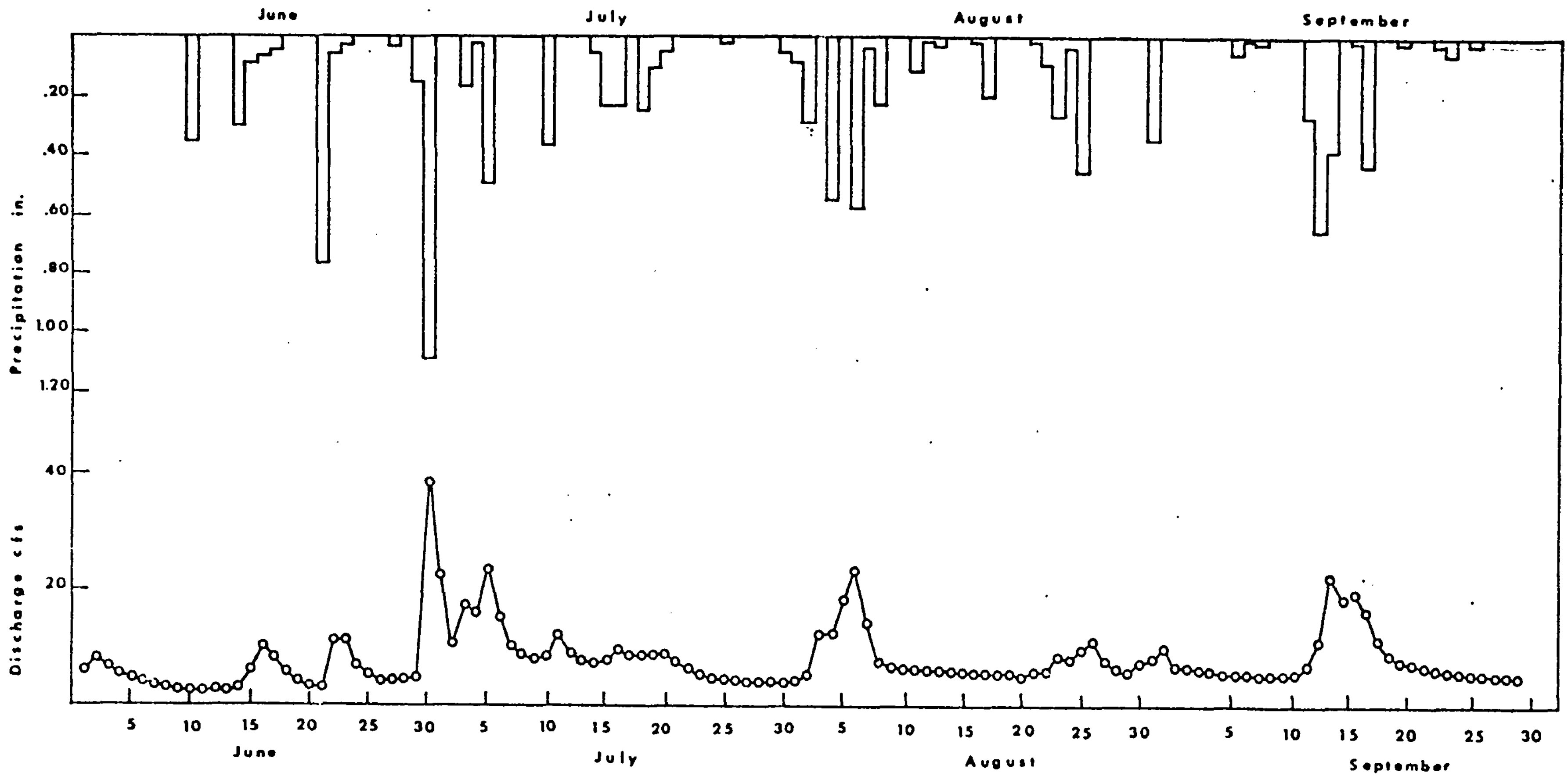


Figure 7

Mean daily discharge and precipitation from June through September, 1970, for the Caribou Creek watershed.

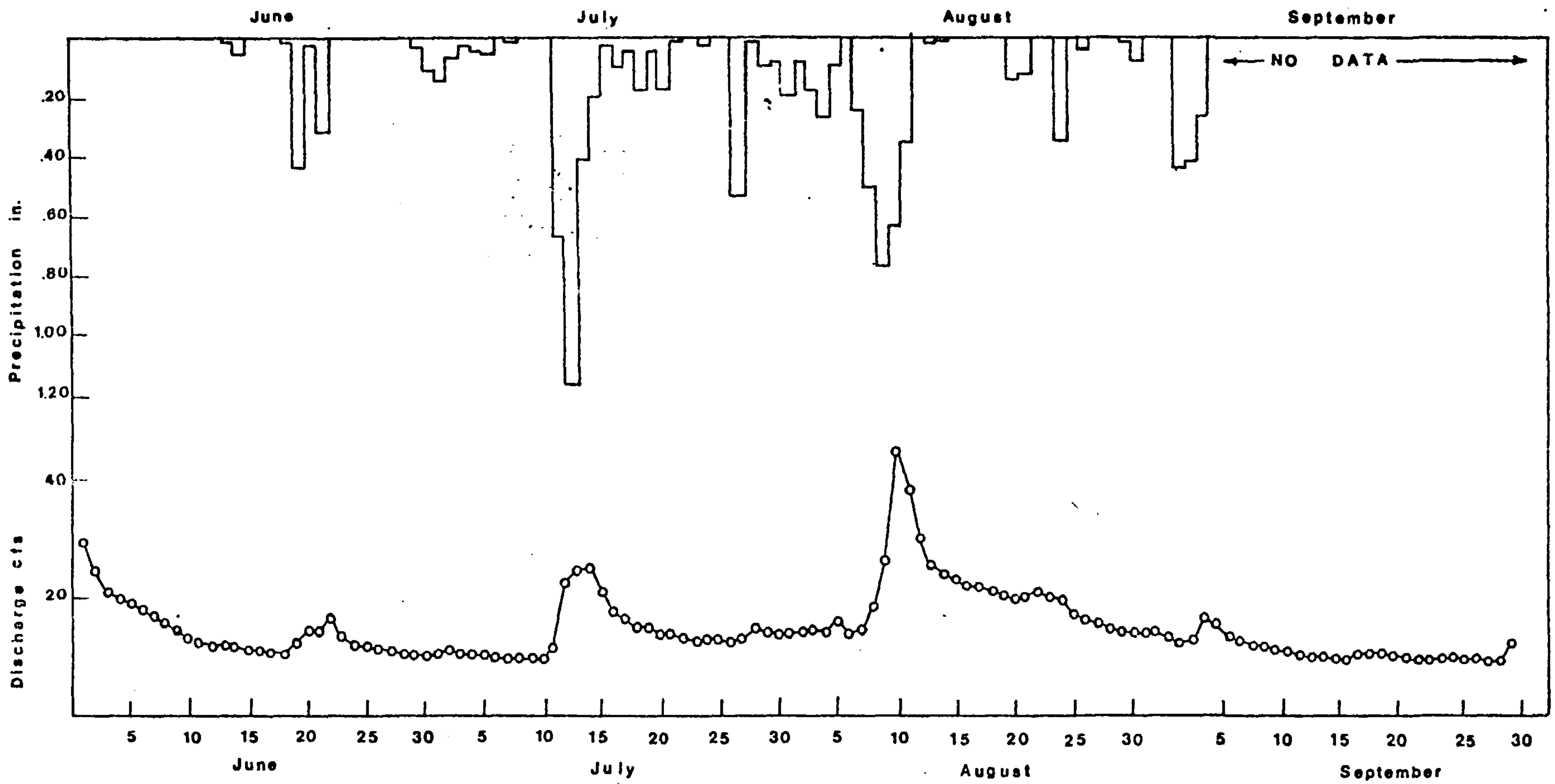


Figure 8

Mean daily discharge and precipitation from June through September, 1971, for the Caribou Creek watershed.

curred during mid-May (Figure 9). July showed the highest runoff volumes despite higher rainfall in August. The total runoff in June, July, August, and September, was 373 ac-ft, 554 ac-ft, 500 ac-ft, and 480 ac-ft. respectively.

For the months of June through August, 1971, the runoff was 2905 ac-ft as compared to 1427 ac-ft for the same three month period in 1970 (Table 6). September is excluded due to the lack of precipitation records. Runoff in June, July, and August was 859 ac-ft, 843 ac-ft, and 1203 ac-ft, respectively. Again, the month with the highest precipitation, July (Table 5), did not have the highest runoff while June, with the lowest precipitation, had the second highest runoff volume. Examination of the mean daily hydrograph for 1970 (Figure 8) revealed that a large part of the June runoff was due to spring snowmelt which was still noticeable as late as June 18.

The 1970 and 1971 rainfall from June through August was 8.58 in and 9.21 in, respectively. The volumetric precipitation differential, 340 ac-ft, cannot account for the 1478 ac-ft difference in runoff between the two years. The amount of runoff, however, from spring snowmelt may yield some insight into the difference. The snow present for the 1970 spring runoff was possibly at a record low, as was the case in Fairbanks, and contributed only 68 ac-ft to stream discharge all of it occurring in May (Figure 9). In contrast to 1970, 1971 had a very large amount of snow cover at the time of spring snowmelt. Excessive runoff began early in May and extended into June. From May 3 until June 18, when the first storm occurred, spring snowmelt had contributed 2062 ac-ft to stream discharge. Consequently, two factors

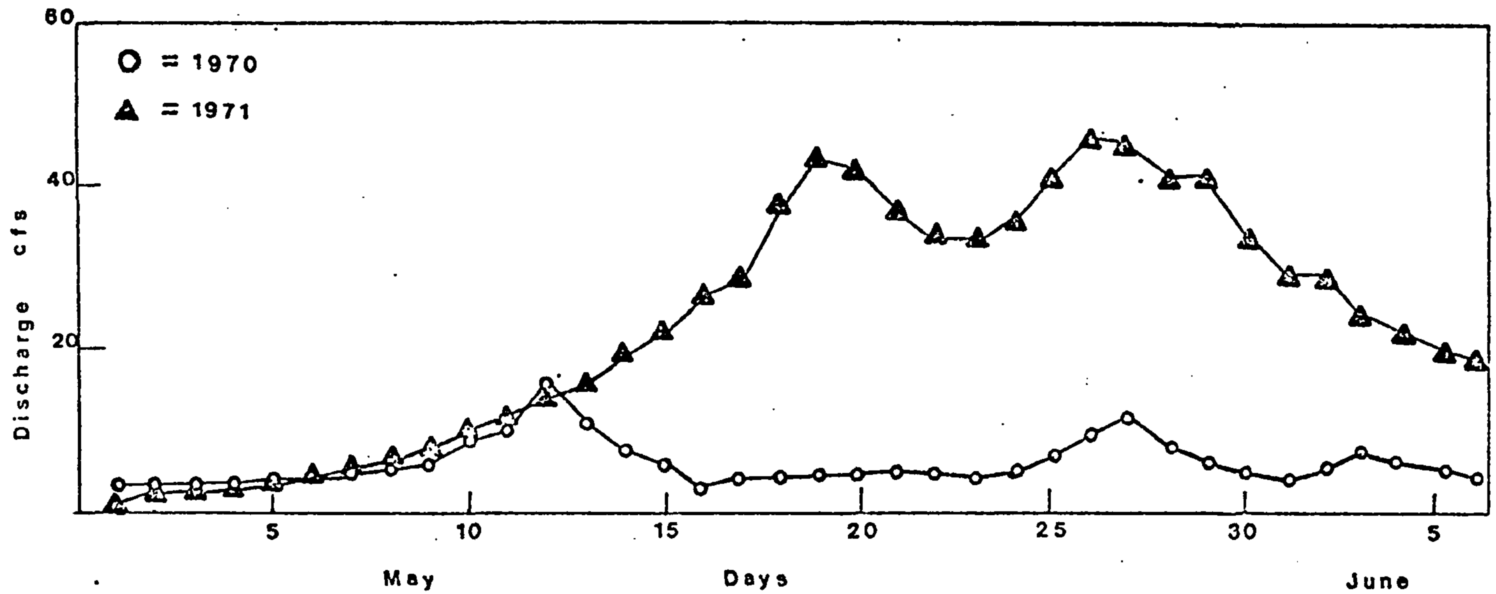


Figure 9

Mean daily discharge from May 1 to June 5, 1970 and 1971
for Caribou Creek watershed.

probably account for higher 1971 runoff. First, basin recharge from spring runoff minimized storage deficits. Second, individual storms in 1971 had longer durations and larger precipitation amounts which may combine to produce higher runoff volumes.

In summary, during the two years of record, runoff was highest for the year with the largest recharge due to spring snowmelt. The month with the greatest precipitation during each year, however, was found not to have the corresponding highest runoff volume.

This discrepancy of highest monthly runoff in July, 1970, and August, 1971, when the greatest precipitation occurred in August, 1970, and July, 1971, can possibly be explained by the equation for discharge from an aquifer (Chow, 1964),

$$\frac{dS_t}{dt} + q_t = 0 \tag{17}$$

where S_t is storage at time t and q_t is runoff at time t. From the exponential-decay discharge relationship given by eq. (10),

$$q_t = q_0 \exp(-kt) \tag{10}$$

k is the recession constant, or the slope of eq. (10) when plotted on log-normal paper. Substituting eq. (10) into eq. (18) gives

$$\frac{dS_t}{dt} + q_0 \exp(-kt) = 0 \tag{18}$$

Separating variables in eq. (18) gives,

$$-dS_t = q_0 \exp(-kt)dt \tag{19}$$

Solving for S_t by integration,

$$-S_t = k^{-1} q_0 \exp(-kt) \tag{20}$$

substituting eq. (20) into eq. (10) yields,

$$-S_t = k^{-1} q_t \tag{21}$$

The recession constant is affected by the abstraction rate of rainwater (input) by deep seepage, evapotranspiration, and if channel storage is considered, rate and amount of overland flow. Any of these or other factors which increase the rate of removal of water from storage will increase k. If, on the other hand, basin storage is high with low water loss, k will be minimal; i.e., a nearly horizontal line with a slope close to zero would result. Consequently, as k increases the storage left on the basin decreases. If q_t is considered to be the mean discharge of the receding limb for any k, large values of k for the Caribou Creek basin would imply little storage and the subsequent ability to abstract rainwater for recharge would increase, due to high storage deficit. The net result would be a decrease in runoff.

Recession constants for July and August of 1970 (Table 7) show that (on the basis of mean daily discharge), for two storms, the average July k value was 0.371 day^{-1}, and, for two storms, the August k value was 0.470 day^{-1}. Although rain fell during the August recessions, the k value was much larger than July, possibly indicating a storage deficit. Therefore, August could conceivably show a lower runoff total than July. This may account for the actual difference observed.

The mean daily hydrograph for 1971 showed no rain-free recessions for comparison; but on the basis of the hourly discharge hydrograph the k value for July (0.05 hr^{-1}) and August (0.02 hr^{-1}) yields some insight into the storage factor, S_t. Despite rain falling during both recessions, the indication was that basin storage in July appeared to be lower than August. This was supported by less observed runoff in July. As will be seen later, the hydrologic response, or, the ratio of rainfall to runoff during the August storm was also greater than July. August flow was possibly sustained by the slow release of snowmelt and rainwater detained earlier in the season. In qualitative terms, recharge from spring snowmelt and rains in June and July produced a basin in a low state of storage deficit by August and more rainfall was available for storm runoff during August.

Examination of the 1971 hydrograph showed that all but one of the mean daily, and two of the hourly events were highly complex. Four mean daily events in 1970 and 1971 were sufficiently complex to warrant special analyses. These occurred during the following periods: June 29-July 8, 1970; September 11-22, 1970; July 10-12, 1971; and August 5-15, 1971. The recessions were extended by a total flow recession curve to the appropriate separation point as derived and discussed in the previous section. The vast difference in the recession characteristics of the two years made it necessary to use separate recession curves. The curve characteristics are summarized in Table 8 and are shown in Figures 10a and 10b. It can be noted that the recession constants are larger in 1970 and thus, the slopes of the recessions steeper. This will be discussed later.

On the basis of the mean daily discharge hydrographs, there were ten storms identified in 1970 and three in 1971 (Figures 7 and 8). Only four major rises were identified from the hourly discharge hydrograph in 1970 and four in 1971. The hourly data from 1970 were not complete and omitted rises that occurred after mid-July (Figures 11, 12, 13, and 14). Table 9 gives total runoff volumes for each storm period by two separation techniques employed in 1970 and three in 1971.

TABLE 8

Caribou Creek's master recession curve characteristics for hourly and mean daily discharges in 1970 and 1971.

	CHANGE IN SLOPE (cfs)	K PEAK TO CHANGE IN SLOPE	K CHANGE IN SLOPE TO END
1970 Mean Daily Discharge	7.0	0.479 (day^{-1})	0.171 (day^{-1})
1970 Hourly Discharge	23.5	0.07 (hr^{-1})	0.03 (hr^{-1})
1971 Mean Daily Discharge	12.5	0.222 (day^{-1})	0.045 (day^{-1})
1971 Hourly Discharge	23.5	0.03 (hr^{-1})	0.01 (hr^{-1})

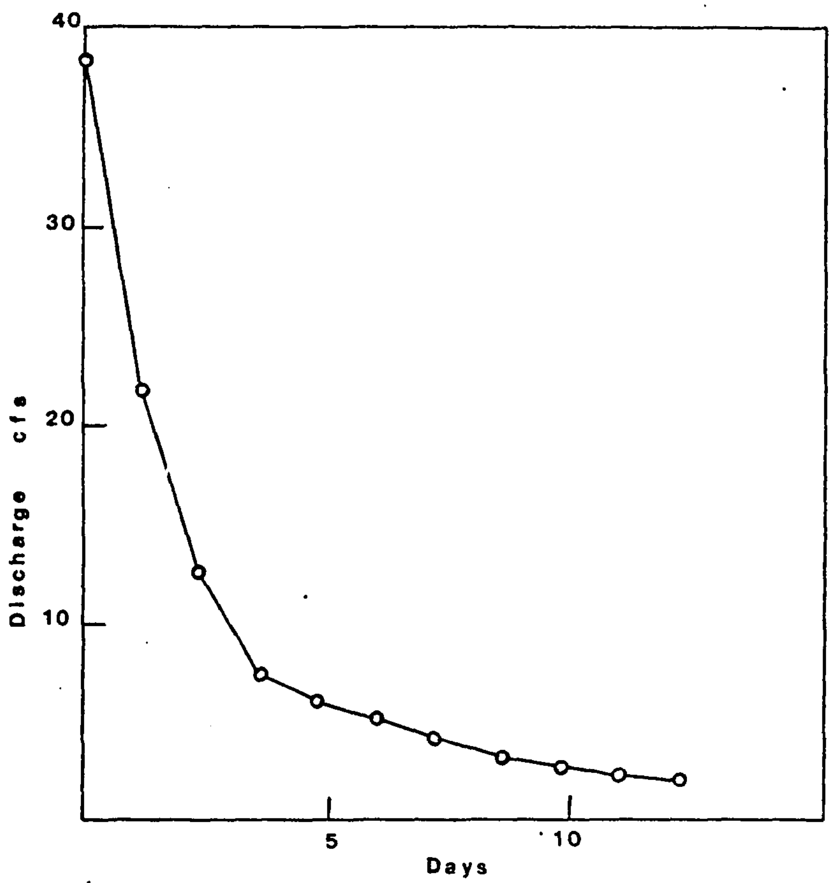


Figure 10a

Master mean flow recession curve for 1970 for Caribou Creek.

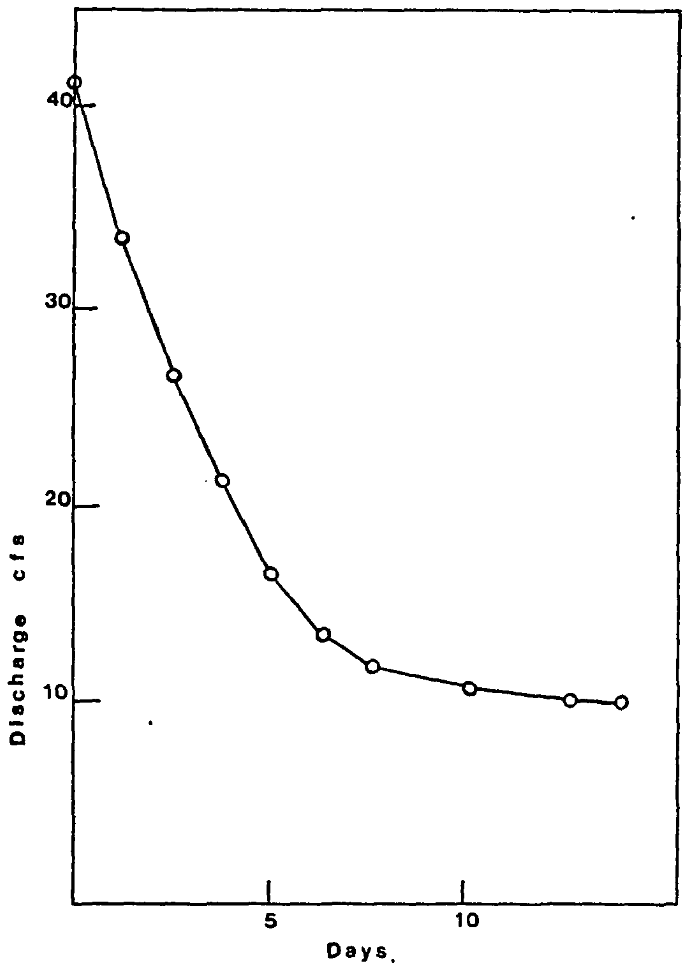


Figure 10b

Master mean flow recession curve for 1971 for Caribou Creek.

TABLE 9

Runoff volumes (ac-ft) for hourly and mean daily storms using two separation techniques for 1970 and three for 1971.

MEAN DAILY DISCHARGE	EXTENSION OF PRIOR RECESSION	HORIZONTAL SEPARATION	QUICK FLOW
1970 June 13-18	23.80	39.67	----
June 21-24	35.70	51.57	----
June 29-July 2	111.74	111.07	----
July 2-8	79.33	----	----
July 9-13	16.66	7.93	----
July 30-Aug 7	126.94	130.91	----
Aug 19-27	43.63	55.54	----
Aug 28-Sept 1	19.83	15.88	----
Sept 11-17	79.34	122.97	----
Sept 15-21	39.66	----	----
1971 June 18-24	31.73	39.65	----
July 10-18	115.04	126.96	287.20
Aug 5-15	230.08	202.32	----

HOURLY DISCHARGE	EXTENSION OF PRIOR RECESSION	HORIZONTAL SEPARATION	QUICK FLOW
1970 June 21	13.22	----	----
June 30	72.72	128.92	----
July 3	19.83	36.36	----
July 5	24.79	29.75	----
1971 June 19	7.44	13.22	0.39
June 21	7.43	11.57	----
July 12	44.63	59.50	67.77
Aug 6	182.23	143.80	185.12

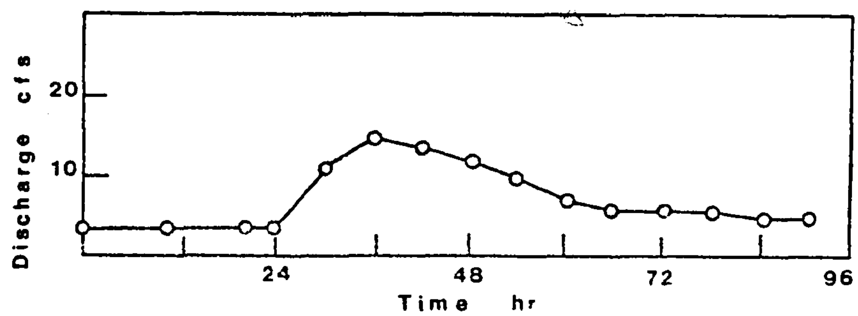


Figure 11

Hourly discharge for the storm of June 21-24, 1970. No hourly precipitation data was available.

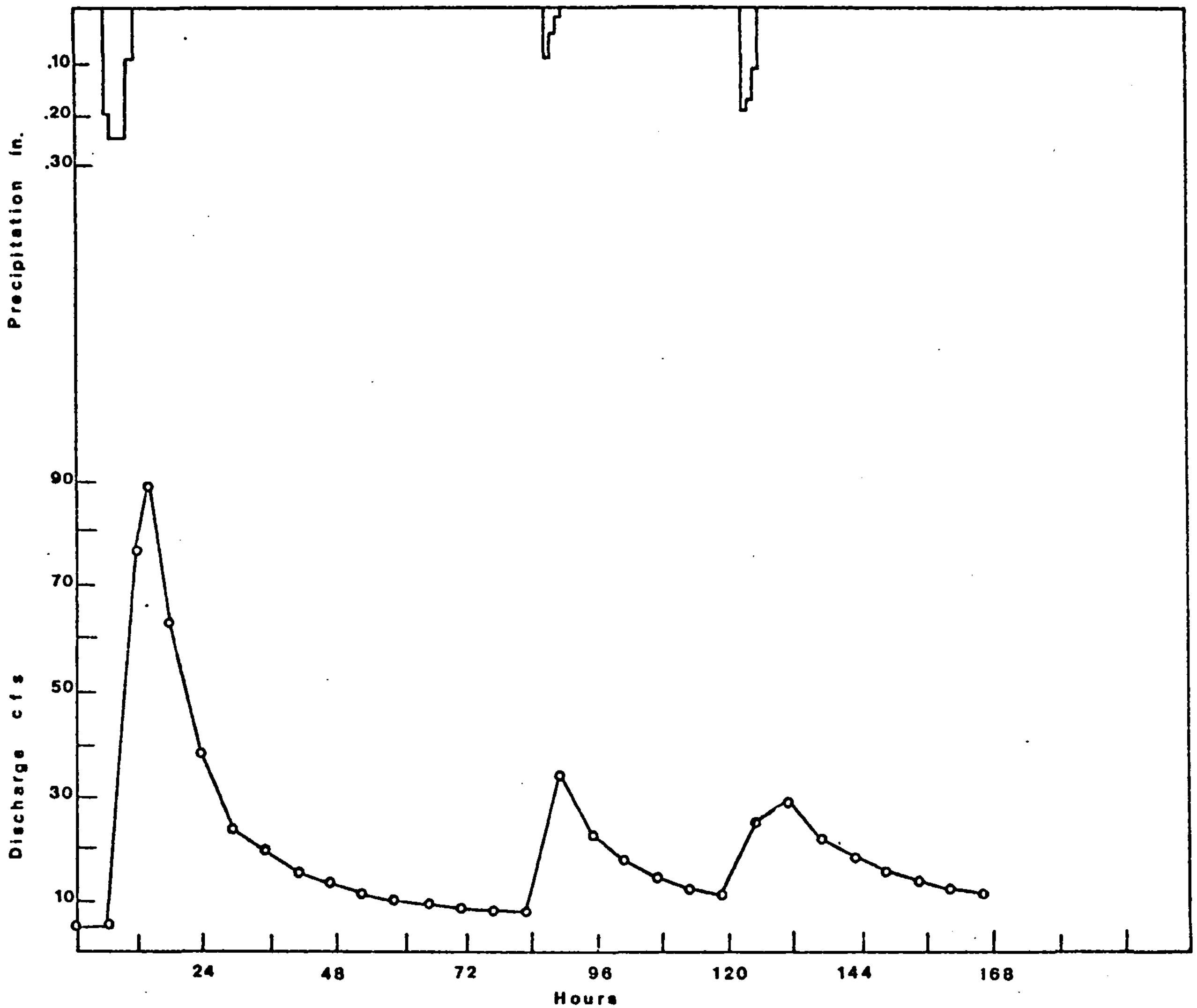


Figure 12

Hourly discharge and precipitation for the storms of June 30-July 6, 1970, on the Caribou Creek watershed.

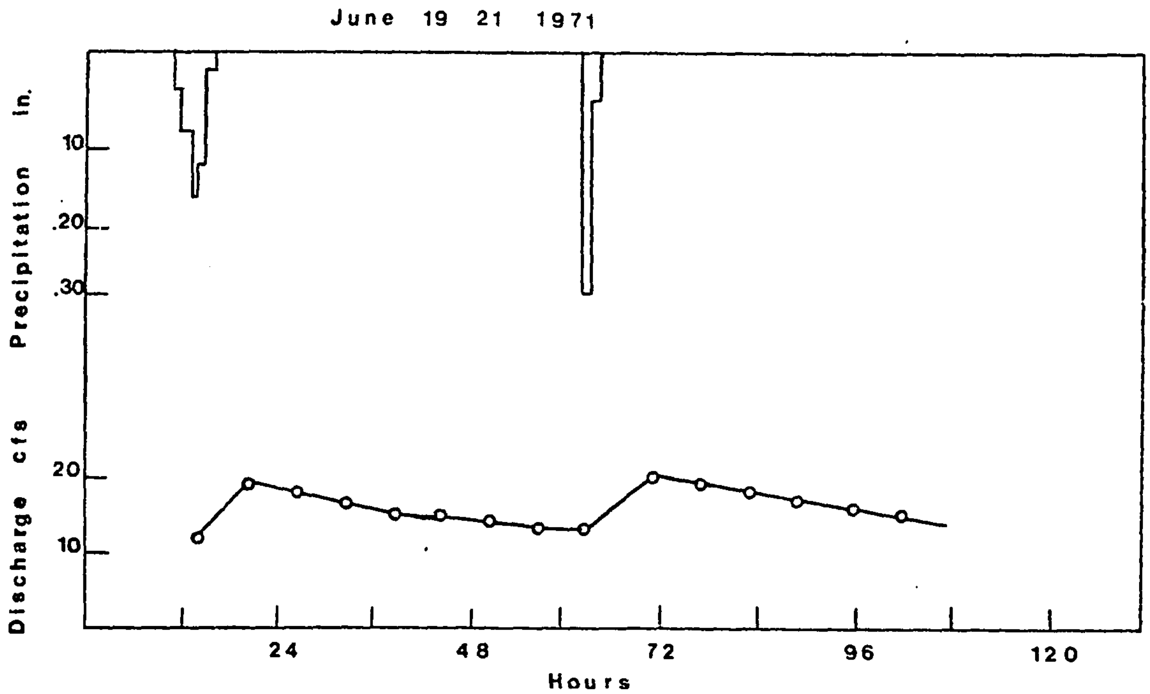


Figure 13

Hourly discharge and precipitation for the storms of June 19-21, 1971, on the Caribou Creek watershed.

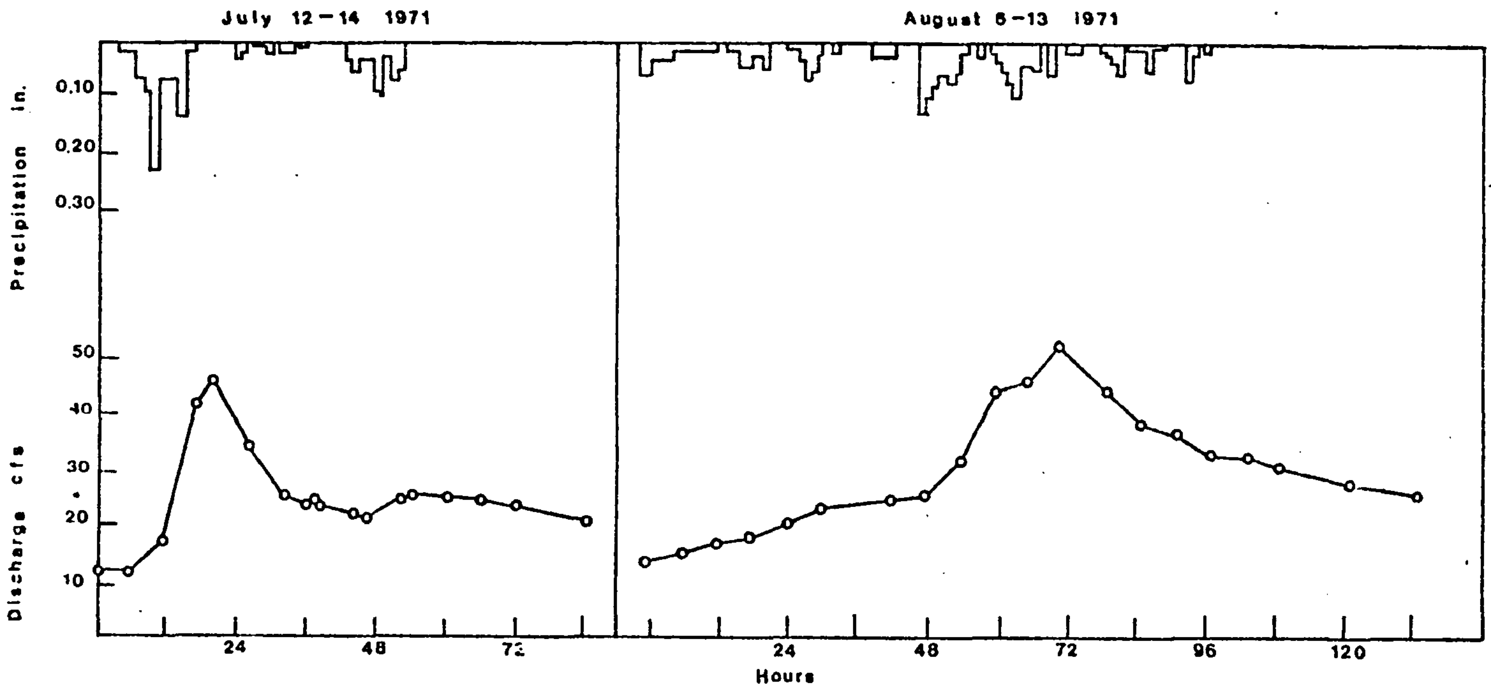


Figure 14

Hourly discharge and precipitation for the storms of July 12-14, and August 6-13, 1971, on the Caribou Creek watershed.

The most practical method applicable from storm-to-storm and year-to-year extended the prior recession to the point under the peak and then to the predetermined separation point.

A general picture of the flow regime of Caribou Creek can be obtained by the rough separation of stormflow and baseflow by estimating the asymptotes to the recessions. Storm runoff, then, accounted for 49 and 30 per cent of the total runoff during the summers of 1970 and 1971, respectively. A larger proportion of runoff from Caribou Creek was baseflow and indicated that the basin detention was very high. The June through August ratio of baseflow to total flow for 1970 and 1971 was 0.51 and 0.70, respectively. Both ratios are high when compared to influent streams and streams or basins with shallow, relatively impermeable surfaces where there is little opportunity for infiltration and storage of rainwater. The 1971 ratio reflects the large charge during spring snowmelt.

It is probable that the soils were not responsible, in total, for the high baseflow since there are large areas of both shallow and poorly drained soils on the basin. The influence of the moss cover is unknown, but it may be a substantial contributor to delayed flow. Dingman (1970) found the moss cover on nearby Glenn Creek to substantially delay runoff and sustain streamflow through dry periods. Bay (1969) found that runoff could be delayed by a moss cover in a bog watershed, but that it was not effective in long term water storage. The deep and extensive moss cover on the Caribou Creek watershed probably had a delaying effect on runoff.

Precipitation-runoff relationships. The traditional hydrologic response (HR) was investigated from mean daily discharge and rainfall records. It is the ratio between storm runoff and storm rainfall. Mean daily storm runoff was found by continuing the recession prior to the rise to the point directly under the peak and then projecting a straight line from there to the intersection with the receding limb at the average value of the change in the recessions slope. For 1970 and 1971 these values were 7 and 12 cfs, respectively.

Several studies have found the hydrologic response to be a function of antecedent discharge (Minshall, 1960; Dingman, 1970). The antecedent discharge has been taken as a measure of basin wetness at the time of the storm and should indicate, at lease in relative terms, the storage capability of the basin. Low antecedent discharge should indicate that the basin is in a state of low storage and capable of abstracting large amounts of precipitation input. High antecedent discharge, on the other hand, should indicate the reverse. Dingman (1970) found that on Glenn Creek in interior Alaska, hydrologic response was a linear function of antecedent discharge and was given by

$$HR = 0.085 + 0.734_q \tag{22}$$

As the discharge at the beginning of the storm increased so did the percentage of storm rainfall contributing to storm runoff.

On the Caribou Creek basin, hydrologic response for the mean daily storms ranged from 0.23 for the event of July 2-8, 1970, to 0.07 for the event of June 18-24, 1971. The average for the 13 storms during the two summers was 0.13 (Table 10). The hydrologic response for each individual storm was plotted against its corresponding antecedent

TABLE 10

Storm date precipitation amount, runoff, and hydrologic response for the mean daily hydrographs in 1970 and 1971, on the Caribou Creek watershed.

DATE	PRECIPITATION (ac-ft)	RUNOFF (ac-ft)	HR
1970 Jun 13-18	412.80	23.80	0.11
Jun 21-24	430.64	35.70	0.08
Jun 29-Jul 2	638.40	111.74	0.17
Jul 2-8	349.60	79.33	0.23
Jul 9-13	192.49	16.66	0.09
Jul 30-Aug 7	814.70	126.94	0.16
Aug 19-27	466.09	43.63	0.09
Aug 28-Sept 1	182.44	19.83	0.11
Sept 11-17	684.00	79.34	0.11
Sept 15-21	238.09	39.66	0.17
1971 Jun 18-24	425.60	31.73	0.07
Jul 10-18	1246.40	115.04	0.09
Aug 5-15	1094.40	230.08	0.21
STORM TOTALS	7175.65	953.48	0.13

discharge (Figure 15). There was no definite linear relationship. The regression equation,

$$HR = 0.07 + 0.007\ q_a \tag{23}$$

where n=13, HR is the hydrologic response, and q_a is stream discharge immediately before the rise, (cfs) had a correlation coefficient of 0.521. This r value was not significant at the 0.05 level. Elimination of all multipeaked, complex events provided no improvement in the correlation. The four simple storms had a correlation coefficient of 0.310 which was not significant at 0.10.

It was recognized that other factors will govern the percentage of storm runoff derived directly from storm rainfall. Evapotranspiration, infiltration capacity of the soil, rainfall intensity, and rainfall amount all have important roles. A convenient parameter from the available data was total rainfall amount, which was taken as that rainfall up to the peak of the mean daily hydrograph (Table 11). A plot of storm runoff versus storm rainfall amount indicated a definite linear relationship (Figure 16). The statistical linear regression of

$$RO_s = -0.02 + 0.15\ P_s \tag{24}$$

where n=13, RO_s is storm runoff (in) and P_s is storm precipitation (in) had a correlation coefficient of 0.894 which was significant at the 0.01 level. Apparently the factors governing the hydrologic response of Caribou Creek to rainfall input are complex. The relationship with storage cannot be discounted, however, as was already indicated by the monthly flow volumes (Table 6). It might be concluded that for short term storm periods runoff relationships are relatively

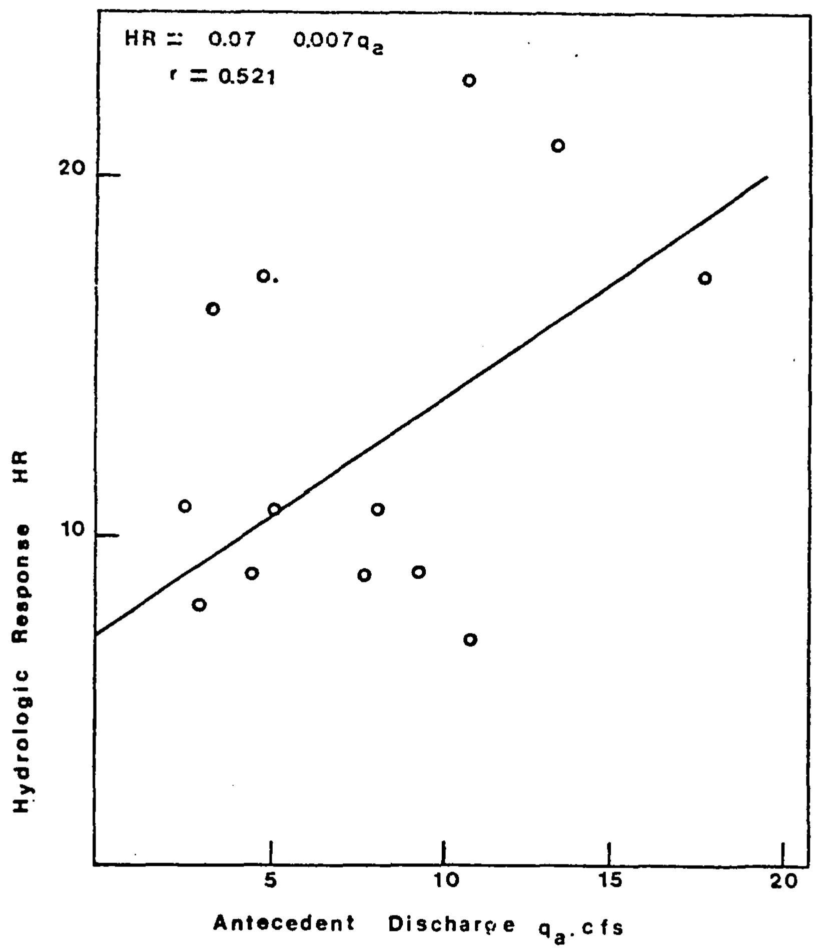


Figure 15

Relationship between hydrologic response (HR) and antecedent discharge (q_a) for mean daily storms on Caribou Creek in 1970 and 1971.

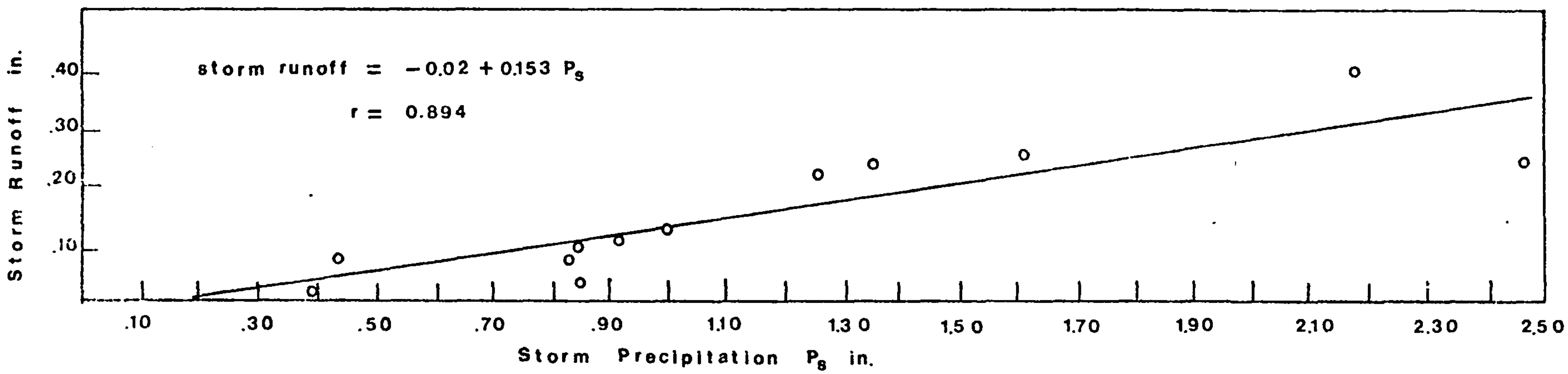


Figure 16

Relationship between storm runoff (in) and storm precipitation (in) for the mean flow hydrograph of Caribou Creek in 1970 and 1971.

TABLE 11

Storm date, precipitation (in), and runoff (in) based on a straight horizontal flow separation on mean daily hydrographs for the Caribou Creek watershed in 1970 and 1971.

DATE		PRECIPITATION (in)	RUNOFF (in)
1970	June 13-21	0.44	0.08
	June 21-26	0.85	0.10
	June 30-July 6	1.26	0.22
	July 9-11	0.39	0.01
	July 30-Aug 11	1.61	0.26
	Aug 19-28	0.92	0.11
	Aug 28-Sept 4	0.85	0.03
	Sept 11-19	1.35	0.24
	Sept 15-16	0.02	0.03
1971	June 18-26	0.84	0.08
	July 10-22	2.46	0.25
	Aug 5-15	2.17	0.40

stable. Examination of the hourly discharge hydrographs should give further insight into the nature of the runoff process on Caribou Creek.

Instantaneous hydrograph-hyetograph plots should give a more sensitive picture of precipitation-runoff relationships than the mean daily plots. Whereas the actual time distribution of both runoff and precipitation is obscured by combined terms on the mean daily hydrograph and daily hyetograph, the instantaneous hydrograph together with the hourly hyetograph will give a meaningful temporal view of rainfall and runoff relationships.

Hourly discharge hydrographs revealed a similar situation as that for the mean daily hydrographs. The hydrologic response for the instantaneous plots varied from 0.03 for the storm of June 19, 1971, to 0.22 for the storm of July 3, 1970. Although no significant relationship was found between the hydrologic response and antecedent discharge, a definite linear relationship was found between total storm runoff and rainfall (Table 12, Figure 17). The relationship was expressed by,

$$RO_s = 0.01 + 0.14\ P_s \qquad (25)$$

where n=7, RO_s is storm runoff (in), P_s is precipitation (in) and the correlation coefficient of 0.905 is significant at the 0.01 level. The hydrologic response--that is, the slope of the line--appeared to be very consistent as indicated by the high correlation coefficient, especially if the storm of July 3, 1970, is discounted. Indications for this storm were that the rainfall measurement was inaccurate. Examination of the hydrograph (Figure 10) will show that only 0.17 in

TABLE 12

Total storm rainfall versus total storm runoff on the hourly discharge hydrograph of Caribou Creek for 1970 and 1971 based on a horizontal separation.

DATE	PRECIPITATION (in)	RUNOFF (in)
1970 June 30	1.05	0.25
July 3	0.17	0.07
July 5	0.49	0.06
1971 June 19	0.44	0.03
June 21	0.34	0.02
July 12	1.12	0.12
August 5	2.02	0.28

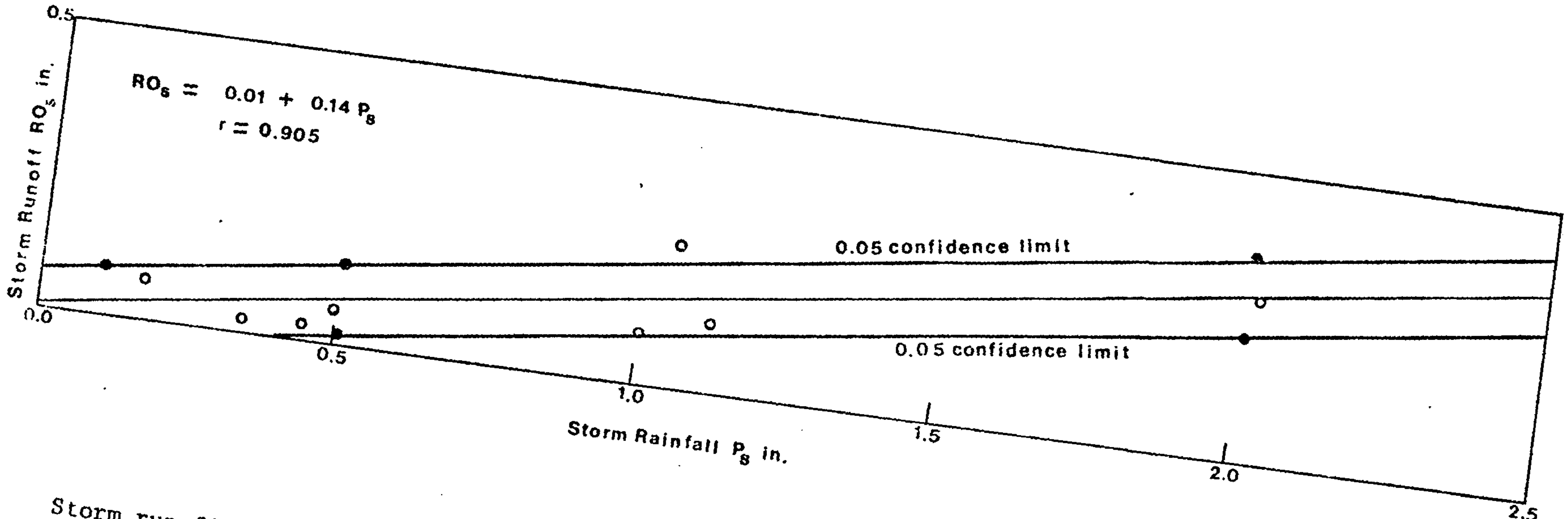


Figure 17

Storm runoff (in) versus storm rainfall (in) for the hourly discharge hydrograph of Caribou Creek for 1970 and 1971.

of rainfall was recorded. The corresponding hydrologic response was 0.22. The rises occurring before and after this event had hydrologic responses of 0.14 and 0.10, respectively.

A stable runoff relationship on the Caribou Creek basin was indicated by the linear increase in storm runoff amount with rainfall, as indicated by the similar equations 24 and 25. The range in hydrologic response, however, while not great when compared to nearby Glenn Creek (Dingman, 1970), suggests a complex of interactions governing the percentage of storm rainfall contributing to storm runoff (Table 13).

Studies have noted that as the rainfall intensity and amount increase the hydrologic response increases due to the decrease in infiltration capacity (Minshall, 1960). Furthermore, the concept of partial area contribution to overland flow has been recently brought forth by Betson (1964). According to his hypothesis small portions of a watershed where the watertable is consistently at or near the surface contribute most of the storm flow, stabilize the baseflow, and stabilize hydrologic response. Consequently, antecedent discharge would be insensitive for prediction of the amount of storm runoff. Total storm rainfall would be a better indicator since the increase in runoff would vary linearly with rainfall amount. This was found to be a possibility on Caribou Creek where the valley bottom has a high watertable underlain by permafrost. Any hydrologic response variability, then, may be due to evapotranspiration, or necessary recharge for fluctuating watertables.

The stable relationship between storm rainfall and runoff, has not generally been noted on most drainage basins (Linsley, Kohler,

TABLE 13

Variance and standard deviation for hydrologic response of seven storms on the Caribou Creek watershed and 16 storms on the Glenn Creek watershed (Dingman, 1970).

CARIBOU CREEK

HR	Variance
0.22	0.0169
0.14	0.0025
0.10	0.0001
0.10	0.0001
0.09	0.0000
0.04	0.0025
0.03	0.0036
0.03	0.0036

STD DEV of POP = 0.064

GLENN CREEK

HR	Variance
0.42	0.0484
0.39	0.0361
0.36	0.0256
0.34	0.0196
0.32	0.0144
0.27	0.0049
0.21	0.0001
0.20	0.0000
0.19	0.0001
0.15	0.0025
0.14	0.0036
0.03	0.0289
0.03	0.0289
0.03	0.0289
0.03	0.0289

STD DEV of POP = 0.136

and Paulhus, 1958). If this relationship does hold true, a dampening factor on antecedent basin wetness must be operative. It has already been suggested that the storage capacity of the Caribou Creek basin is great and that high levels of delayed flow are encountered. The thick moss carpet on the valley bottom may be implicated in this. Water movement through the moss layer when saturated conditions exist can be described by Darcy's law (Dingman, 1970),

$$q = \frac{k}{\mu} AS \tag{26}$$

where q is discharge in suitable flow rate units, k is the permeability constant in units complementary to q, A is the cross-sectional area of saturated flow, μ is the dynamic viscosity of the liquid, and S is the slope of the hydraulic gradient in ft ft^{-1}. If water at a given temperature is considered and q is specified temporally, Darcy's law can be written more simply as

$$q_t = PI_tA_t \tag{27}$$

where q_t is discharge in suitable units at time t, P is the permeability in units complementary to q_t, I_t is the slope of the hydraulic gradient at time t, and A_t is the cross-sectional area of saturated flow at time t. In the case of a drainage basin, I_t can be taken as the slope of the basin itself, since the vertical change in depth of the saturated layer with time will be very small when compared to the actual slope of the land. Dingman (1970) found permeabilities (P) for the moss on Glenn Creek to be 16.5 ft sec^{-1}.

Basin storage in the moss layer will be a function of the porosity and volume of the saturated layer (Butler, 1957). It would expressed by,

$$S_t = lA_t p \tag{28}$$

where S_t is storage at time t, l is the length of the aquifer, and p is the porosity. The porosity of the moss can be given by,

$$p = 1\frac{p_b}{S.G.} \tag{29}$$

where p_b is the bulk density of the moss in g cm^{-3} and S.G. is the specific gravity of the moss in g cm^{-3}. Farnham and Finney (1965) have found bulk densities for Sphagnum to be approximately 0.07 g cm^{-3} and specific gravity to be 1.50 g cm^{-3}. This gives a porosity of 0.95 which was the same as that found for the moss cover on the Glenn Creek watershed (Dingman, 1970).

Flow from the moss, described by the flow equation from a reservoir can be given as (Chow, 1964),

$$\frac{dS_t}{dt} = q_t = 0 \tag{30}$$

The substitution of eq. (27) into eq. (29) gives,

$$q_t = - \frac{dA_t}{dt} lp \tag{31}$$

Now substituting eq. (26) into eq. (30) gives,

$$- \frac{dA_t}{dt} lp = PA_t I_t \tag{32}$$

or,

$$- \frac{dA_t}{A_t} = \frac{PI_t}{lp} dt \tag{33}$$

Integration results in,

$$\ln A_t = PI_t(lp)^{-1}t \tag{34}$$

where f is a constant. At t=0 the area, A_0=f, a constant, so,

$$\ln A_t - \ln A_0 = -PI_t(lp)^{-1} + f - f \tag{35}$$

and,

$$\ln A_t(A_0)^{-1} = -PI_t(lp)^{-1}t \tag{36}$$

and,

$$A_t(A_0)^{-1} = \exp\left[-PI_t(lp)^{-1}t\right] \tag{37}$$

Recalling eq. (27), solving for area and substituting in eq. (35) we get,

$$q_t(PI_t)^{-1} = q_0(PI_t)\exp\left[-PI_t(lp)^{-1}t\right] \tag{38}$$

or,

$$q_t = q_0 \exp\left[-PI_t(lp)^{-1}t\right] \tag{39}$$

The recession constant for flow from the moss would, therefore, be $PI_t(lp)^{-1}$. Dingman (1970) obtained a similar expression which yielded a numerical value of 0.0013 hr^{-1}. Using Dingman's permeability (P) and porosity (p) of 16.5 ft hr^{-1} and 0.95, respectively, length (l) equal to one-half the width of the valley bottom on Caribou Creek (660 ft) and a slope (I_t) of 0.07 for the valley bottom on Caribou Creek, a recession constant for the moss of 0.0018 hr^{-1} was obtained.

Loss to deep seepage and evapotranspiration would tend to increase this constant (Bay, 1969; Dingman, 1970). This can be seen by considering q_t and k as the known independent and S_t as the unknown dependent variables. In reality, S_t represents not only storage on the basin, but also the various losses that act on gross precipitation. However, since both q_t and k represent a final integrated product, and since storage is given as a function of the two parameters, an increase in storage with its associated decrease in k ~~implies a lower loss~~ rate and vice versa. Regardless, the moss with a very small recession constant may indeed be a significant delay factor contributing to the high levels of baseflow on Caribou Creek.

The physical significance of the derived recession constant due to the moss can be investigated by recalling eq. (30) and substituting eq. (39) into it,

$$-\frac{dS_t}{dt} = q_0 \exp\left[-PI_t(lp)^{-1}t\right] \tag{40}$$

Separating variables and integrating gives,

$$-S_t = q_0 \exp\left[-PI_t(lp)^{-1}t \; lp\,(PI_t)^{-1}\right] \tag{41}$$

Resubstitution of eq. (38) gives,

$$-S_t = q_t \; lp(PI_t)^{-1} \tag{42}$$

As the length of the aquifer (l) and the porosity (p) increase storage on the basin increases. As the permeability and/or the slope (I_t) of the basin increases, the storage decreases. If any of the latter two parameters increase, storage will be depleted more rapidly. The moss with a very high porosity will then encourage large amounts of storage and become a dampening source on antecedent stream discharge. High discharge between storms may be due to delayed flow from the moss. With its large storage capacity the moss may be able to hold, in detention, most additional input. According to Darcy's law, a linear increase in storm runoff with increasing rainfall amount would result. This possibility seemed most likely for small storms such as those of June 19 and 21, 1971. The recessions of these storms showed no change in slope and were characteristic of delayed flow recessions with a k of

0.01 hr^{-1}.

The response time from the beginning of rainfall to the initial rise of the hydrograph is quite rapid on Caribou Creek (Table 14). The average value for five storms was two hours which strongly suggests a source of overland flow. The July 3 and July 5, 1970, rises appeared to begin before precipitation commenced. There was no basis for shifting the time scale of the strip charts recording rainfall and discharge, so the response time was taken to be 2 hours--the average response time of the other five storms.

There was no significant relationship found between antecedent stream discharge and response time at the 0.10 level. A similar situation was noted by Dingman (1970) for another interior Alaska basin. The slope of the regression line, with a correlation coefficient of 0.471, given by,

$$T_s = 0.84 + 0.11\ q_a \tag{43}$$

where n=7, T_s is the response time (hrs), and q_a is the antecedent discharge (cfs) which was not significantly different from zero at the 0.05 level based on a t-test. This suggested that the response time was nearly constant. This indicated that the antecedent moisture conditions of the basin do not significantly affect the response time and lends further support to the hypothesis that the runoff process on Caribou Creek is fairly stable and independent of the antecedent moisture of the basin.

The large peak flows, then, possibly result from a combination of delayed flow from the moss and overland flow from the valley bottom. The smaller peaks probably originate from moss layer discharge as

TABLE 14

Response times (hrs) and antecedent discharge (cfs) for the seven hourly storms during the summers of 1970 and 1971 on the Caribou Creek Watershed.

DATE	q_a (cfs)	RESPONSE TIME (hrs)
June 30, 1970	6	1
July 3, 1970	7.5	*
July 5, 1970	11.0	*
June 19, 1971	11.0	3
June 21, 1971	12.5	2
July 12, 1971	11.7	2
August 6, 1971	14.0	2
AVERAGE	10.5	2

*Assumed to be 2

rainfall enlarges the saturated volume.

Peak flow volumes for specific events, regardless of their source, must be predicted in order to effectively predict the discharge hydrograph from storm rainfall. Holtan and Overton (1963) found that the peak discharge can be predicted if the storm runoff, time of rise, and recession constant are known. Dingman (1970) modified their approach using antecedent discharge as a predictor of storm runoff and time of rise equal to storm duration.

In this study the peak flow volume was given by eq. (16). Its application requires the knowledge or predictive capability of runoff volume due to a rainfall event, the hydrograph's time of rise, and the recession constant. The required runoff input can be fulfilled by eq. (25) which relates runoff to specific amounts of storm rainfall for hourly events. The derivation of the time of rise and the recession will be discussed below.

On the Caribou Creek basin the time of rise was found to be a linear function of the storm duration. Storm duration was taken as either the time period from the beginning to the end of the rainfall, or, if the rain continued past the hydrograph peak, that time period from the beginning of the rain to the time of the peak. Therefore, only precipitation contributing to the peak was considered (Figure 18). The relationship was,

$$T_r = 3.94 + 0.926T_p \qquad (44)$$

where n=7, T_r is the time of rise, T_p is the duration of the storm (hrs). The correlation coefficient was 0.998 and was significant at the 0.01 level. The slope was not significantly different from unity

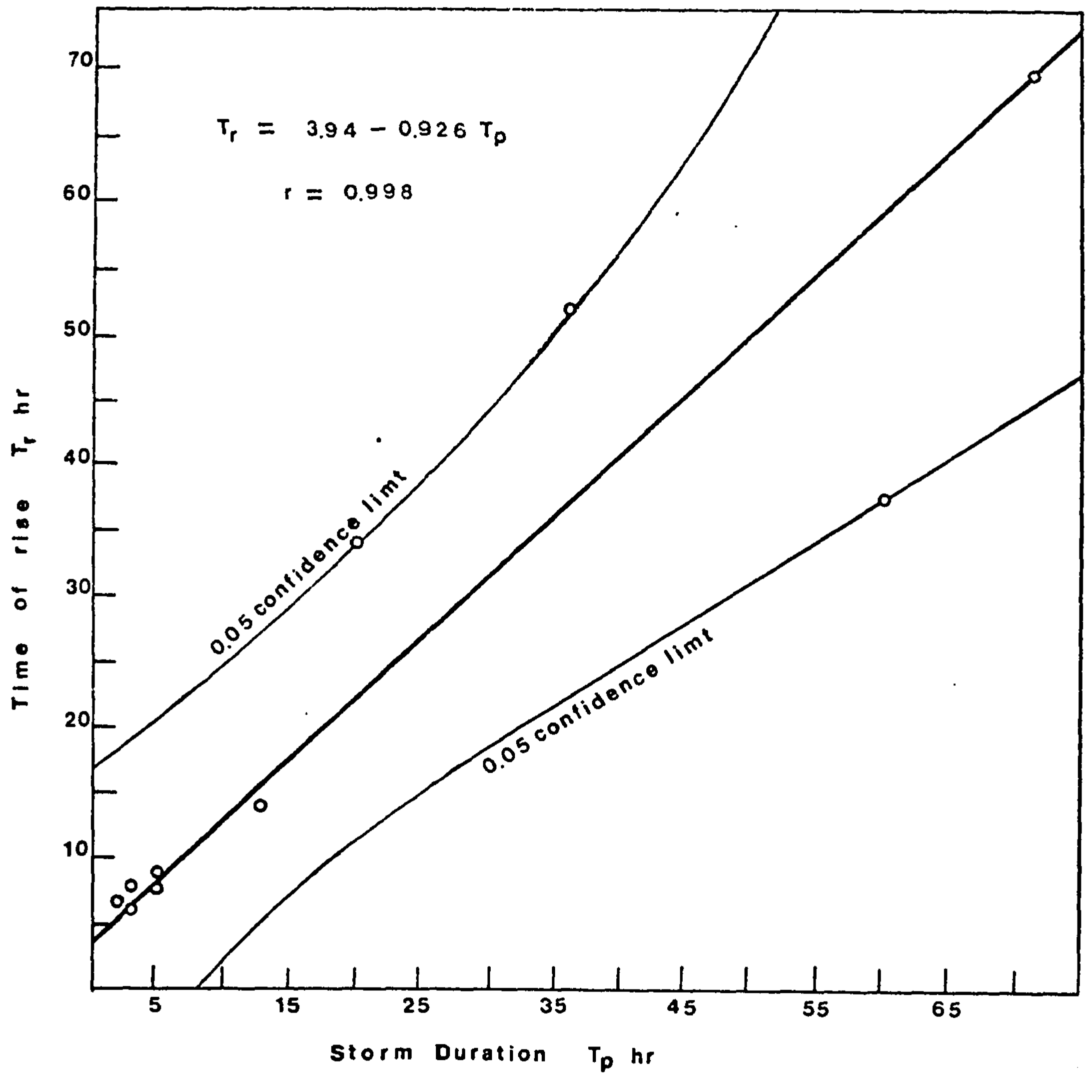


Figure 18

Time of rise (T_r) versus storm duration (T_p) in hr for seven storms on the Caribou Creek watershed during the summers of 1970 and 1971.

at the 0.05 level and the constant indicated a delay factor. The discrepancy between duration of the storm (T_p) and the time of rise (T_r) is apparent for the events of short duration and with small rainfall amounts (Table 15). The effect may be due to a delay in the activation of the source area.

To complete positioning of the hydrograph's peak the lag time must be determined. It is defined as the time span from the centroid of the rainfall to the peak of the hydrograph. The lag time was found to be linearly related to total storm amount (Table 15). The relationship,

$$T_1 = 1.70 + 9.66\ P_s \tag{45}$$

where n=7, T_1 is the lag time (hr), and P_s is the precipitation (in) had a correlation coefficient of 0.924 which was significant at the 0.01 level (Figure 19).

<u>Hydrograph shape</u>. Aside from the peak, the hydrograph is made up of a rising limb and a falling limb (recession). The characteristics of these two portions give the hydrograph its shape. Thus far, in the literature, the vast majority of work has been done with the recession while only limited attention has been paid to the rise.

Unfortunately, the variability between inflow, outflow, and storage that characterizes the rising limb of the hydrograph makes mathematical interpretations of that segment of the hydrograph very difficult. No scientist has yet deduced a rigorous expression for this part of the runoff hydrograph, although, by assuming a linear increase in discharge from the rise to the peak, the rising limb can

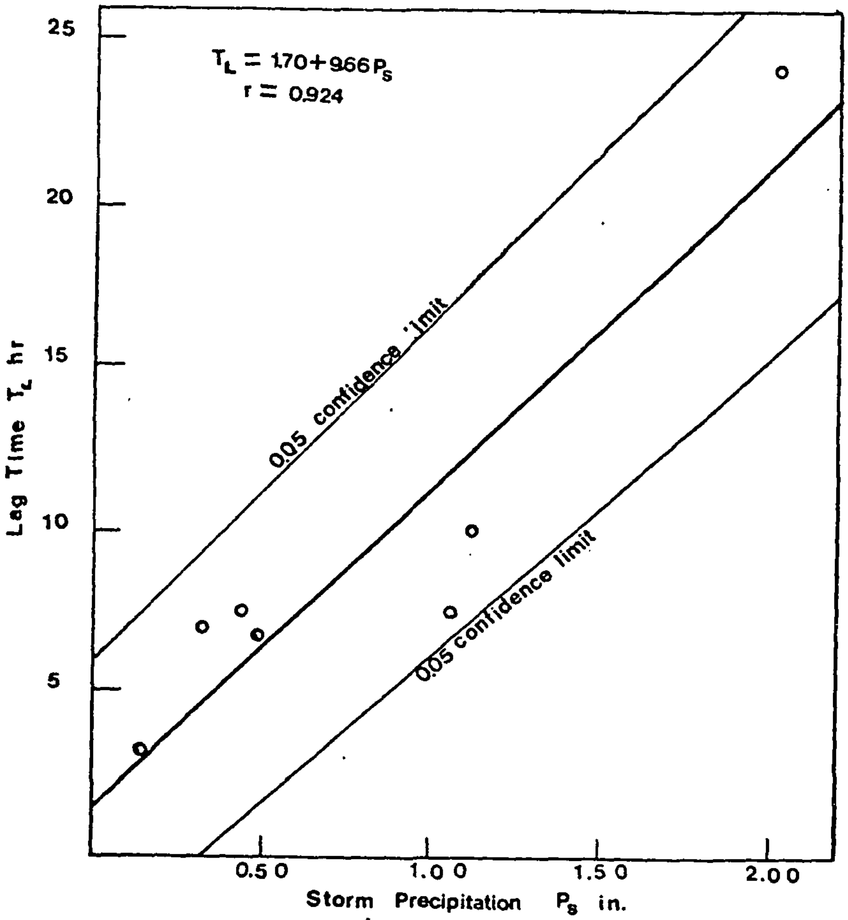


Figure 19

Lag time (T_l) versus total storm precipitation (P_s) in inches for seven storms on the Caribou Creek watershed during the summers of 1970 and 1971.

TABLE 15

A comparison of the storm duration (hr), time of rise (hr), lag time (hr), and total precipitation amount (in) from the hourly discharge hydrographs for 1970 and 1971 on the Caribou Creek watershed.

DATE	STORM DURATION (T_p)	TIME OF RISE (T_r)	LAG (T_l)	PRECIPITATION (P_s)
June 30, 1970	5	8	7.5	1.05
July 3, 1970	3	6	3.2	0.17
July 5, 1970	3	8	6.8	0.49
June 19, 1971	5	9	7.5	0.44
June 21, 1971	2	7	7.0	0.34
July 12, 1971	13	11	10.0	1.12
August 5, 1971	71	70	24.0	2.02

be approximated by a straight line (Holtan and Overton, 1963). Such an approximation is necessary if the runoff is to be predicted so as to validate the derivation of eq. (17).

The recession can be approximated by,

$$q_t = q_0 \exp(-kt) \tag{10}$$

which is a straight line with a slope of k on log-normal paper. Appendix I shows the hourly discharge recessions for portions of the hydrograph upon which no rain fell (except for the events of July 12-13, 1971 and August 9-10, 1971). The 1970 recessions had a change in slope (k) at an average of 23.5 cfs. The 1971 recessions did not indicate this change. This is probably because the recessions from the larger peaks did not have the opportunity to change flow source due to the occurrance of rain soon after the recession began. The small peaks were probably the result of a single source of flow. It was assumed that the 1971 recessions changed slope at 23.5 cfs. The complete recession relationships for 1970 and 1971 are as follows:

(1) 1970-- from the hydrograph peak to 23.5 cfs,

$$q_t = q_0 \exp(-0.07t) \tag{46}$$

where q_0 is the peak flow (cfs); from 23.5 cfs to the end of the recession,

$$q_t = q_0 \exp(-0.03t) \tag{47}$$

where q_0 is the turning point (cfs); (2) 1971-- from the hydrograph peak to 23.5 cfs,

$$q_t = q_0 \exp(-0.03t) \tag{48}$$

and from 23.5 cfs to the end of the recession,

$$q_t = q_0 \exp(-0.01t) \tag{49}$$

The recessions in 1971 were much more gradual than those in 1970. One explanation would be that rain occurred during the two recessions of the 1971 storms. Even without rain, however, they remained very gradual.

The recessions on Caribou Creek are very long when compared to those of basins in the contiguous States. Holtan and Overton (1963) present recession constants for individual storms for 37 basins in the eastern United States. The smallest basin, draining an area of 29.6 mi^2, had a recession constant of 0.20 hr^{-1} (considerably greater than those observed on Caribou Creek). A 158 mi^2 basin had a recession constant averaging 0.01 (comparable to late recessions in 1971 for Caribou Creek). A 764 mi^2 basin had a recession constant of 0.06 hr^{-1} which is roughly comparable to an early recession on Caribou Creek. The considerable delay noted on the Caribou Creek basin has been hypothesized as due to the moss cover. A similar effect was observed by Bay (1969) on Minnesota bog watersheds and alluded to by Dingman (1970) for an interior Alaska watershed. Dingman also noted that lower evapotranspiration rates in the interior of Alaska will tend to cause slower recessions. Consequently, it appears that a particle of water detained on the slopes of Caribou Creek watershed will take longer to reach the outlet than will a similar particle on a much larger basin in the humid and sub-humid conterminous States.

The small values of k on the Caribou Creek watershed indicate large detention on the slopes. This is evidenced by recalling eq. (20)

$$S_t = q_t\ k^{-1} \tag{20}$$

The graphical representation of storage (S_t) versus discharge (q_t) has been delineated for each summer's recession (Figure 20). It is apparent that S_t was smaller for corresponding q_t in 1970 because of the larger k value for that year.

With an equation for runoff, time of rise, and a recession constant, peak flow can be calculated from eq.(16) if storm rainfall is known. First, storm runoff and the time of rise is computed. Some subjectivity must be employed to determine what recession constant to use. Presently, with only two years of data, the constant for the appropriate year must be chosen. Hopefully, when sufficient years of record become available, either a generally applicable curve, or a compilation of a family of curves based on recharge due to spring snowmelt or antecedent precipitation could be generated. Nevertheless, many more years of data will be required before these possibilities can be adequately evaluated.

Table 16 compares the calculated peak due to the storm, that is, the peak which would be expected if baseflow was zero at the time of rise, and the observed peak due to the storm (actual peak observed minus baseflow). A linear regression resulted in a correlation coefficient of 0.910 which was significant at the 0.05 level (Figure 21). If antecedent discharge is added to the calculated peak due to the storm, the observed peak is estimated. Table 16 also compares these. The correlation coefficient of 0.900 was significant at the 0.05 level Figure 22).

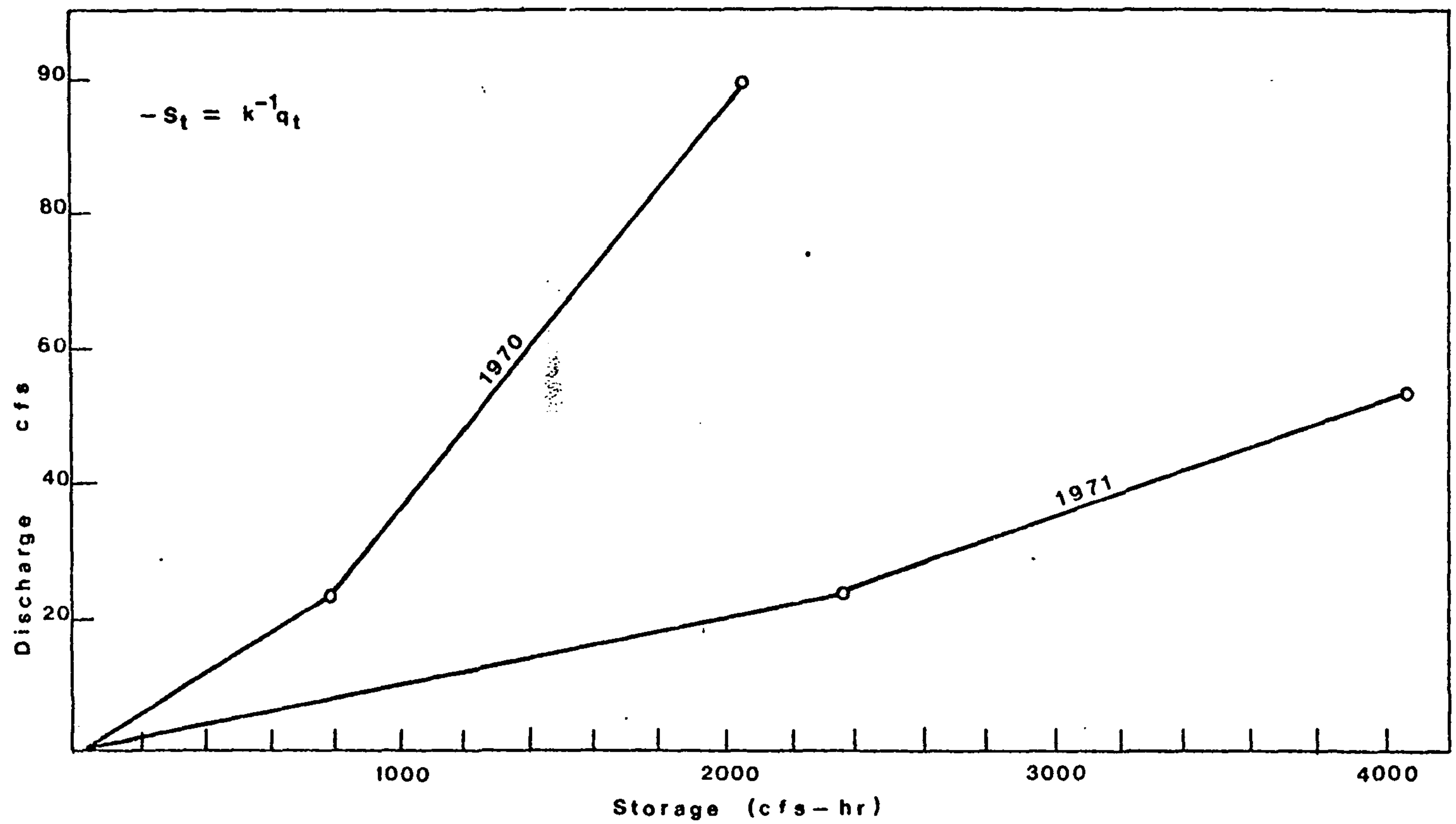


Figure 20

Storage (S_t) in cfs-hr versus discharge (q_t) in cfs for Caribou Creek in 1970 and 1971.

TABLE 16

A comparison of the calculated actual peak and the observed actual peak of the hydrograph and the calculated peak due to the storm and observed peak due to the storm for the hourly discharge hydrograph for 1970 and 1971.

DATE	CALC. ACTUAL PEAK (cfs)	OBS. ACTUAL PEAK (cfs)	CALC. PEAK DUE TO STORM (cfs)	OBS. PEAK DUE TO STORM (cfs)
June 30, 1970	64.0	87.0	63.0	76.0
July 3, 1970	22.5	33.5	15.5	26.5
July 5, 1970	42.0	28.0	31.0	17.0
June 19, 1971	22.4	17.5	11.4	7.5
June 21, 1971	22.7	18.5	10.2	6.0
July 12, 1970	34.7	46.0	23.7	35.0
Aug 6, 1971	40.0	52.3	26.0	38.3

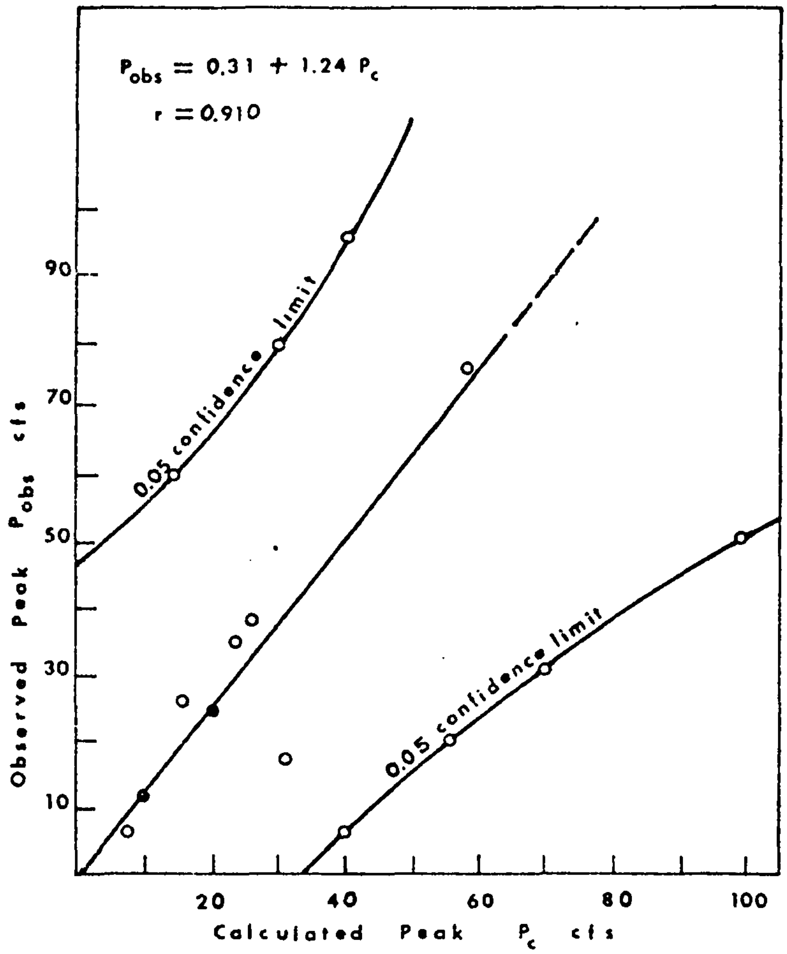


Figure 21

Observed versus calculated peak due to the storm for seven storms on the Caribou Creek watershed in 1970 and 1971

Of possibly more long term interest to those delving into water relations is how well the model can predict total runoff for a storm period. While it must be realized that it was not developed for complex events, due to the small amount of data at hand, the model has the advantage of simplicity. The only parameters needed are total storm rainfall and associated hourly rates. If actual peaks and runoff are to be estimated, antecedent discharge at time t=0 is needed. The total predicted hydrograph can be drawn by positioning the peak with expressions for lag time and time of rise. The recession can be computed (Figures 23, 24, 25). Rain falling before the change in slope is ignored since it was not considered in construction of the model, however, that after the change is considered. The runoff volumes so estimated can be obtained by planimetry. The observed versus computed runoff volumes compare very closely (Table 17). The correlation coefficient of 0.986 was significant at the 0.05 level (Figure 26).

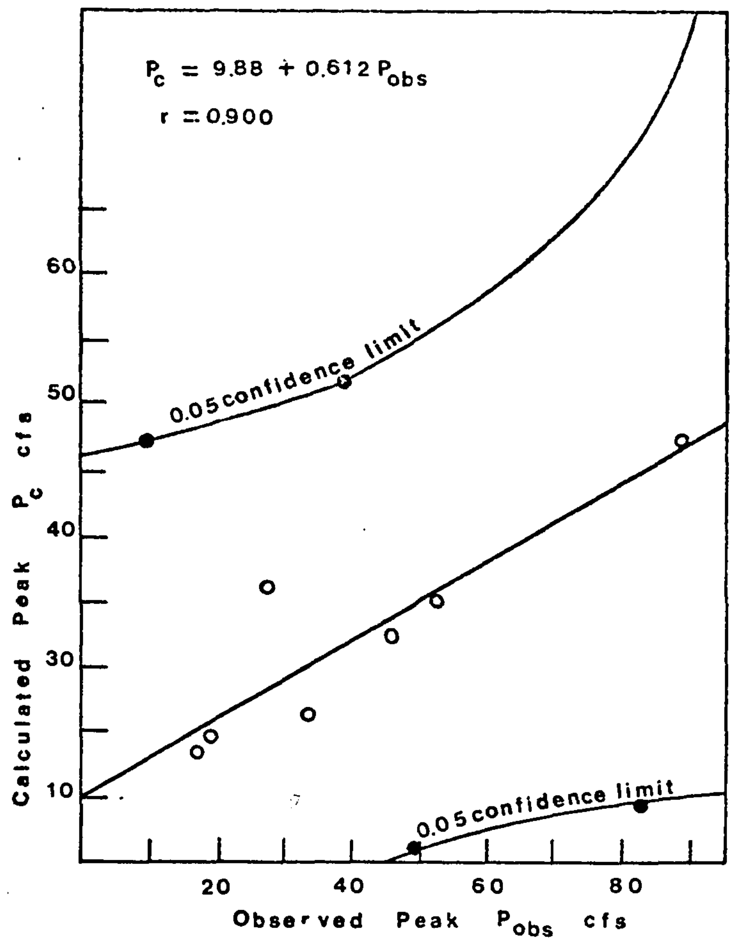


Figure 22

Calculated actual peaks versus observed actual peaks for seven storm on the Caribou Creek watershed in 1970 and 1971.

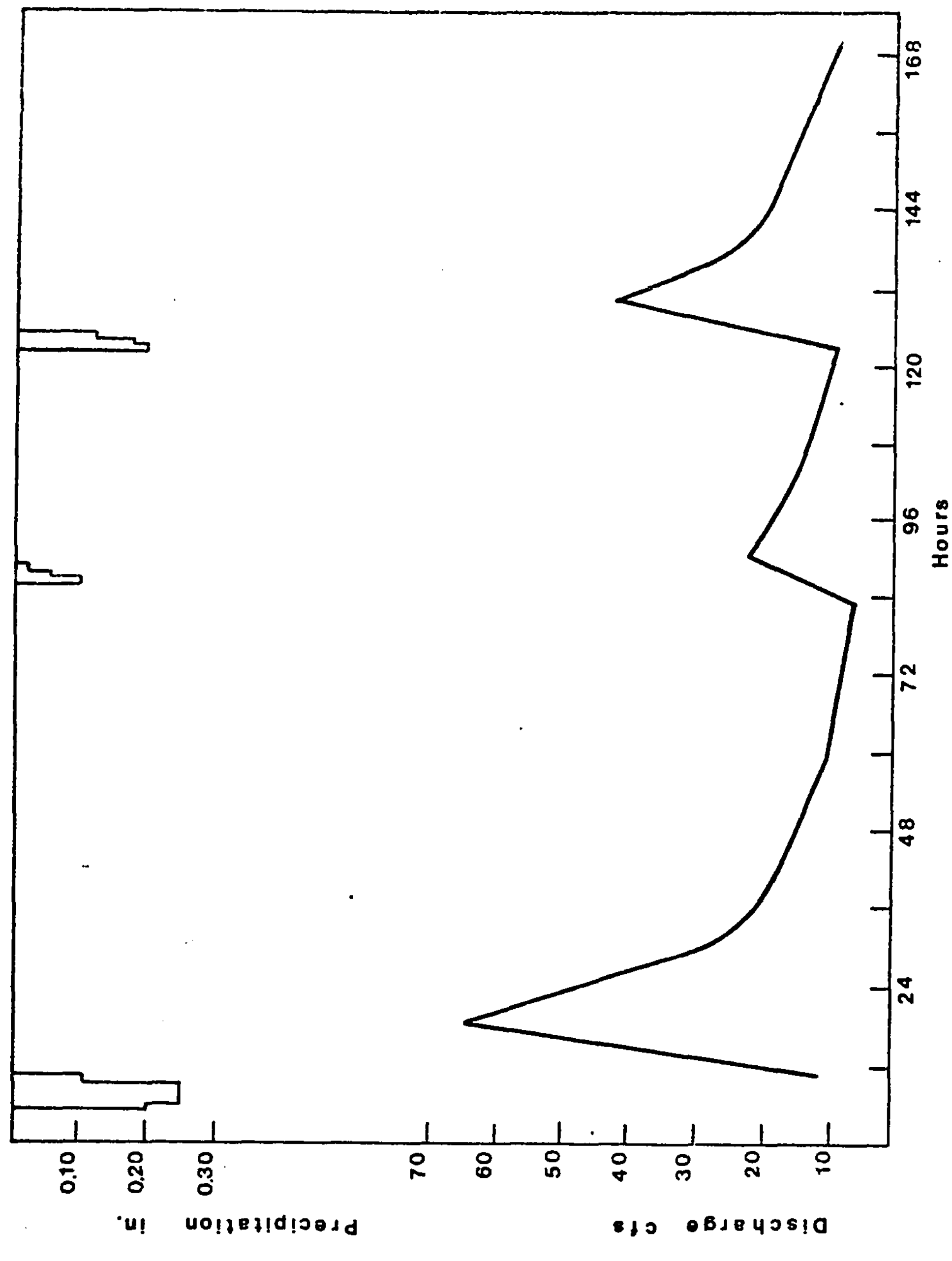


Figure 23

Actual precipitation and the calculated hourly hydrograph for June 30-July 6, 1970, for Caribou Creek.

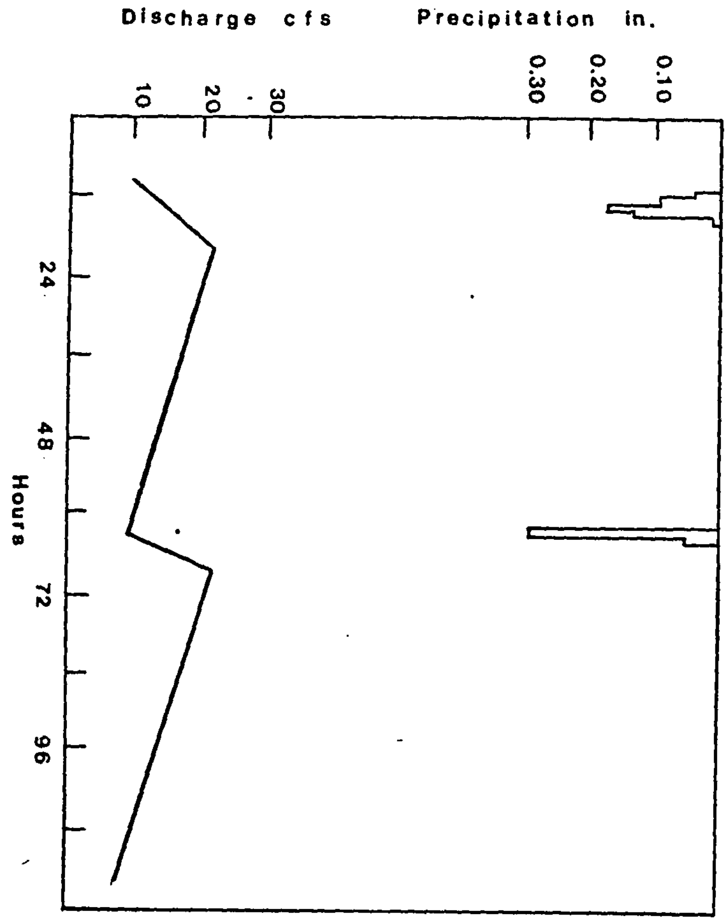


Figure 24

Actual precipitation and the calculated hourly hydrograph for June 19-June 23, 1971, for Caribou Creek.

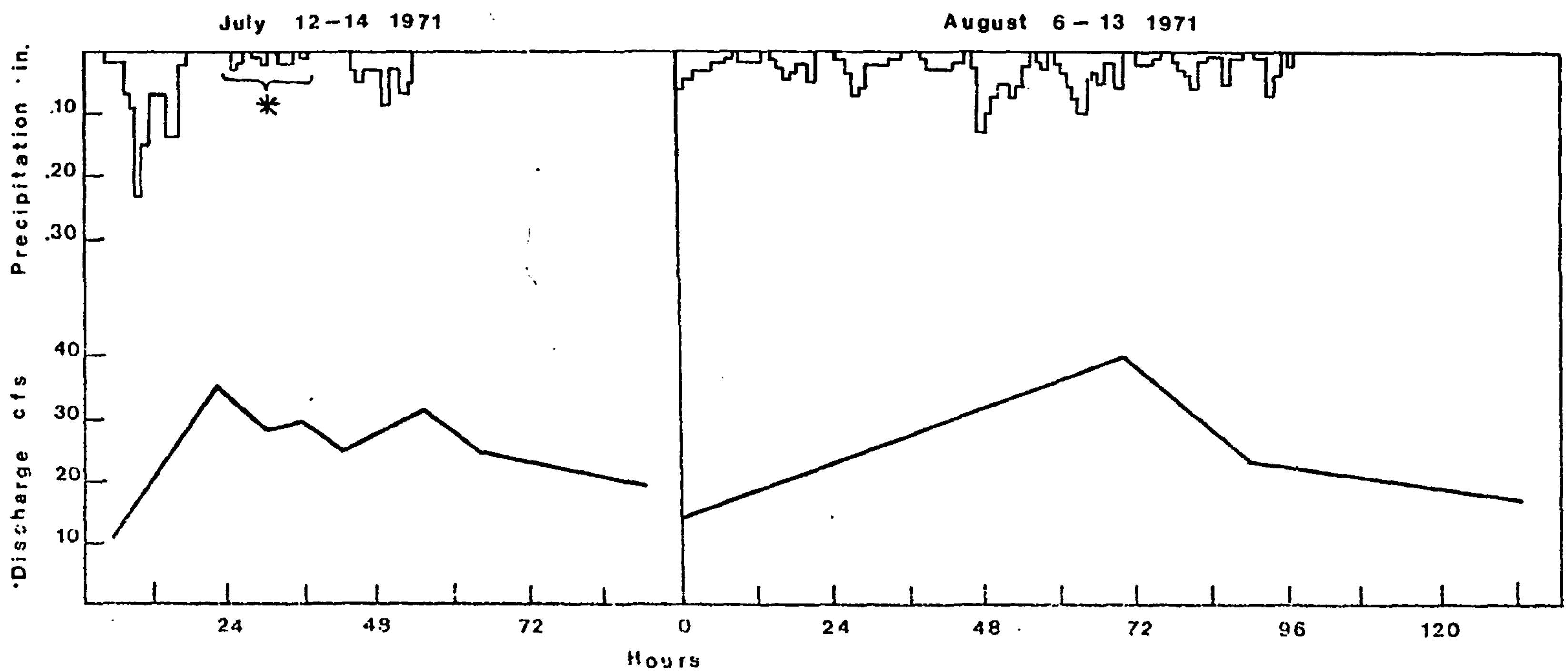


Figure 25

Actual precipitation and the calculated hourly hydrograph for July 12-14 and August 6-12, 1971, for Caribou Creek.

TABLE 17

A comparison of calculated runoff and observed runoff for indicated storm periods for the hourly discharge hydrograph.

DATE	CALCULATED RUNOFF (in/ac-ft)	OBSERVED RUNOFF (in/ac-ft)
June 30-July 2, 1970	0.252/127.93	0.305/154.58
July 2-4, 1970	0.084/42.64	0.105/53.30
July 3-6, 1970	0.147/74.62	0.137/69.24
June 19-21, 1971	0.131/66.63	0.116/58.63
June 21-23, 1971	0.121/61.30	0.121/61.30
July 12-14, 1971	0.331/167.91	0.358/181.24
August 6-13, 1971	0.516/261.20	0.652/330.50

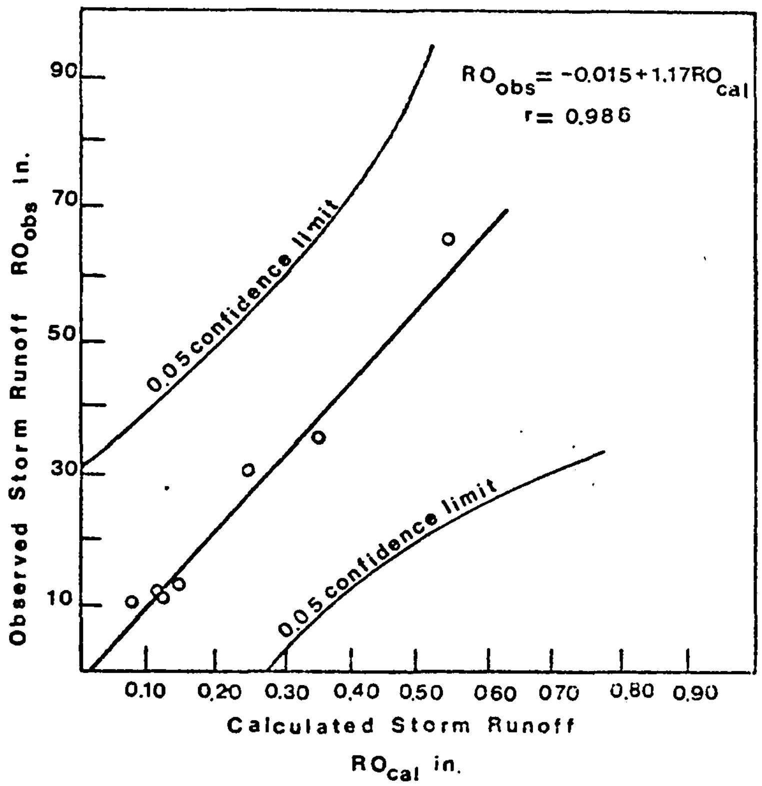


Figure 26

Observed runoff (in) versus calculated runoff (in) for seven storm periods on the Caribou Creek Watershed in 1970 and 1971.

SUMMARY AND CONCLUSIONS

Precipitation and runoff data for the Caribou Creek watershed, located near Fairbanks, Alaska, were analyzed. Moderate summer precipitation amounts fell during the two years of data collection and averaged 8.90 in for the months of June, July, and August. August had the highest two year mean precipitation followed by July and June, respectively. While June, July, and August averaged 10, 18, and 17.5 days with rain respectively, intensities ranged from 0.03 in hr^{-1}, to 0.21 in hr^{-1}.

There was little evidence to indicate a marked orographic influence, however the C-3 gage at 2100 ft above msl caught more rain during the two years of data collection than either of the other two recording gages.

The largest seasonal runoff was observed in 1971 and was a result of the tremendous contribution to streamflow from spring runoff and greater precipitation in 1971.

Storage on the basin can be expressed as,

$$S_t = \frac{1}{k} q_t$$

where S_t is storage at time t, k is the slope of the recession when plotted on semi-log paper, and q_t is discharge at time t. A plot of q_t versus S_t for the summers of 1970 and 1971 revealed that storage on the basin during the latter year was much higher. This was possibly due to the excessive recharge resulting from record amounts of snow in 1971.

The storage deficit, indicated in relative terms by k, was seen to affect the monthly runoff percentage. It was found that as the value of k decreased runoff volume increased. This is a reflection of the rate of evapotranspiration, deep seepage and other losses.

A rough separation of stormflow and baseflow showed that storm runoff accounted for 49 and 30 per cent of the total runoff in 1970 and 1971, respectively. The high percentage of baseflow may be due not only to permeable soils, but also to detention of rainwater in the moss layer--especially on the north slopes and in the valley bottoms.

The runoff process was found to be somewhat unique in that it was much more stable than is generally noted. The hydrologic response was found to be independent of antecedent moisture conditions on the basin. Storm runoff, however, was adequately predicted from storm rainfall by a simple linear regression model,

$$RO_s = 0.01 + 0.14\ P_s$$

where RO_s is the storm runoff in in, and P_s is the storm rainfall in in. A similar situation was found for the mean daily hydrograph where,

$$RO_s = -0.02 + 0.15\ P_s$$

Significant correlation coefficients of 0.905 and 0.894, respectively indicated that the percentage of storm rainfall becoming storm runoff does not vary greatly from storm to storm. The variability that was observed may have been due to evapotranspiration, deep seepage, and other undetected sources of water transfer as well as inaccurate rainfall sampling.

The response time of the hydrograph to storm rainfall averaged two hours. While the response time (T_s) itself, was not significantly related to antecedent wetness of the basin, the slope of the regression line was not significantly different from zero at the 0.05 level. This lent further support to the assertion that runoff on the Caribou Creek basin came from a fairly stable source. It was speculated that an area of rather constant size contributed most of the storm runoff in the form of overland flow. Some variation in hydrologic response may, then, be due to delay in activation of the source area of runoff.

The time of rise (T_r) was found to be related to storm duration (T_p) by the regression equation,

$$T_r = 3.94 + 0.926\ T_p$$

The slope of the line was not significantly different from one, however the intercept indicates a delay factor which may be attributed to detention in the moss. The deviation between the time of rise and the storm duration was more apparent for smaller events.

The lag time was found to be related to total precipitation amount by the regression equation,

$$T_l = 1.70 + 9.66\ P_s$$

where T_l is the lag time in hr and P_s is the storm precipitation in in.

Recessions were well modelled by the exponential decay function,

$$q_t = q_0 e^{-kt}$$

where q_t is the instantaneous discharge at time t, q_0 is the discharge at time t=0, and k is the recession constant. The value of k for

Caribou Creek was found to be very low when compared to values for basins in the other humid and sub-humid states. Reasons for this may be low evapotranspiration in the interior of Alaska, and the thick carpet of moss found on the valley bottoms and on north aspect slopes of the Caribou Creek watershed.

A recession constant due to the moss was derived and given as,

$$k = -PI_t(lp)^{-1}$$

where P is the moss permeability in ft hr^{-1}, I_t is the slope of the valley bottom, l is the length of the moss layer in ft, and p is the porosity of the moss. This replaces k in the recession equation. The recession constant calculated for the moss on the Caribou Creek basin was 0.0013 hr^{-1} which was smaller than any recession noted during the study and indicated a significant delay source.

The peak of the hydrograph can be adequately predicted by the equation,

$$q_p = RO_s(0.5T_r + k^{-1})^{-1}$$

where q_p is peak flow in cfs, RO_s is the storm runoff in cfs-hr, T_r is the time of rise, and k is the recession constant (hr^{-1}). All of the necessary parameters were predicted from the appropriate regression equations relating them to a precipitation characteristic. The value of k was obtained from an appropriate recession equation; the position of the peak on the time axis was obtained by placing it with expressions for the time of rise and lag time; the rising limb of the hydrograph was approximated by a straight line; and, the receding limb was calculated using the exponential decay equation and appropriate recession constant.

Volumes of storm runoff were obtained from the calculated hydrograph. Good correlation was obtained between calculated and observed runoff volumes. The entire runoff model has the advantage of simplicity. The only parameter needed, other than an average recession constant are precipitation amount and hourly distribution.

Indications were that the results of this data analysis technique have considerable potential for application to small basins in interior Alaska. With adequate results from the small amount of data at hand, it was felt that more data would improve results and reveal greater insight into the actual runoff processes.

After analyzing the data four recommendations could be made for future efforts relative to the data and data collection. First, the data lacked continuity, with only a few periods existing during which all three recording gages supplied data simultaneously. This complicated the Thiessen weighing technique. To improve continuity, it may only be necessary to change the recording system from weekly to monthly periods. This would permit greater leeway in changing charts as logistics would demand.

Second, the rain gage network could be altered to provide better representation of the precipitation on the basin. A minimal but reliable network could be achieved by distributing rain gages evenly over the watershed if it can be assumed that orographic influences are negligible. By installing a recording gage at the center of the watershed above the stream gaging site and using it as a reference point, four others could be placed in the center of the four quadrants formed by perpendicular bisectors drawn through the reference point.

These evenly spaced gages would have similar areal coverage, each representing about 1.8 mi^2. The valley bottom would have one gage representing 20 per cent of the network, while the four gages on the slopes would represent 80 per cent of the network. This would be approximately the proportion of slopes to valley bottom noted on the watershed.

Third, of great importance in the understanding of some phenomena in the interior of Alaska, such as recessions, is a knowledge of evapotranspiration. It is unfortunate that no such data existed for use in the present study. As Dingman (1970) has shown, the effects of evapotranspiration may well account for some of the differences noted between precipitation and runoff in interior Alaska and the conterminous States. The installation of an evaporation pan, then, on Caribou Creek watershed would seem a necessity.

Fourth, the responsibility for the installation of equipment and collection of the various data used for studies on the Caribou-Poker Creeks watershed should rest with the individual researcher. A definite handicap is encountered when only the raw data are supplied and the methods of collection and the reliability of techniques are unknown to the analyst. If any recommendation presented here should be stressed, this last one must be chosen. Proper feeling for the data cannot be obtained from tables and figures compiled elsewhere.

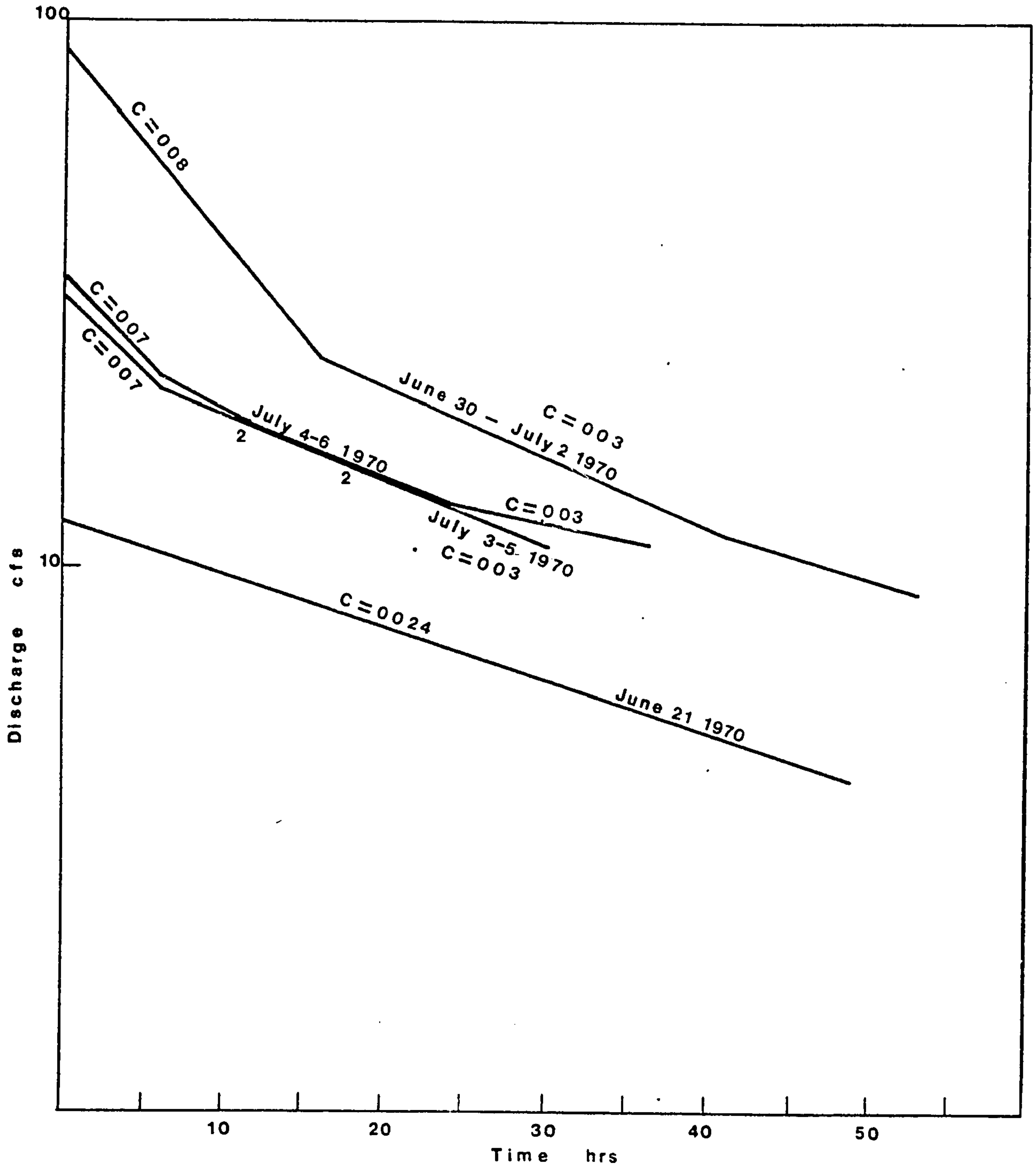
The results of this study were based on a small back-log of data. With the raingage network suggested and continuous water level records which are now available, a more precise relationship between precipitation and runoff could be obtained. Further insight into the apparent independence of hydrologic response from antecedent moisture conditions of the basin could be obtained through more detailed analyses.

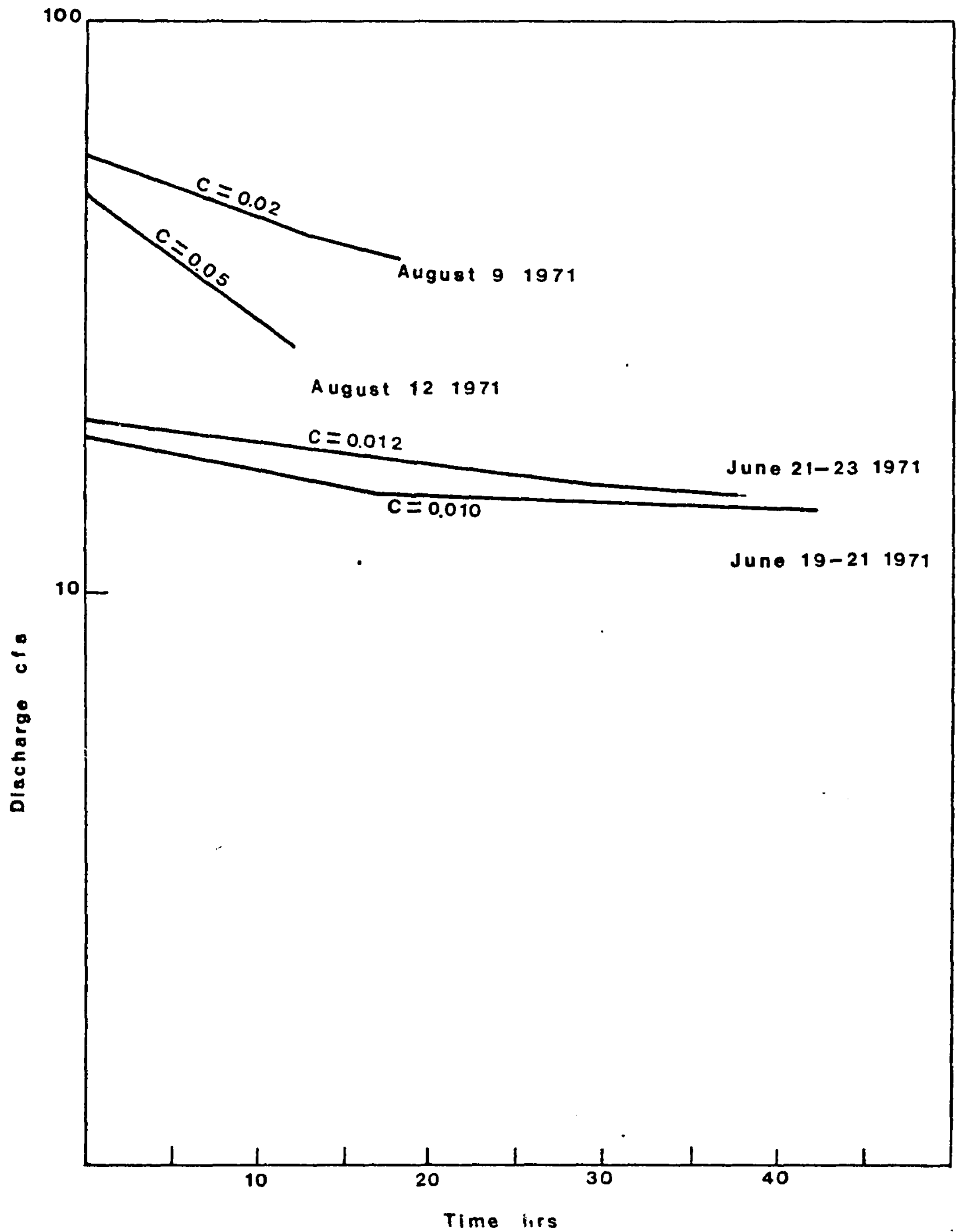
An experimental watershed program, such as that for Caribou-Poker Creeks, is extremely valuable, especially in an area such as interior Alaska where little is known of the hydrologic regime and a vast majority of the drainage basins are ungaged. If, as the present study indicated, basins in interior Alaska are subject to modelling with fairly simple techniques, the research watershed concept is all the more valuable and should be intensively developed.

APPENDIX

Hourly Recessions on Caribou Creek

for 1970 and 1971


100
10
Discharge cfs
C=008
C=007
C=007
June 30 – July 2 1970
C=003
July 4-6 1970
2
2
C=003
July 3-5 1970
C=003
C=0024
June 21 1970
10
20
30
40
50
Time hrs


100
10
C = 0.02
C = 0.05
August 9 1971
August 12 1971
C = 0.012
June 21–23 1971
C = 0.010
June 19–21 1971
Discharge cfs
10
20
30
40
Time hrs

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