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HYDROLOGY OF THE GLENN CREEK WATERSHED TANANA RIVER BASIN, CENTRAL ALASKA

S. Lawrence Dingman

September 1971

CORPS OF ENGINEERS, U.S. ARMY
COLD REGIONS RESEARCH AND ENGINEERING LABORATORY
HANOVER, NEW HAMPSHIRE

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PREFACE

This report was prepared by Dr. S. Lawrence Dingman, formerly a Research Hydrologist of the Earth Sciences Branch, Research Division, U.S. Army Cold Regions Research and Engineering Laboratory. The work was performed under 1) DA Project 1T061102B52A, *Research in Military Aspects of Terrestrial Sciences*, Task 02, *Military Aspects of Cold Regions Research*, Work Unit 006, *Cold Regions Hydrology* (FY 1969); and 2) DA Project 4A062112A894, *Engineering in Cold Environments*, Task 01, *General Engineering Support*, Work Unit 001, *Hydrology of Rivers and Lakes* (FY 1970).

This report was submitted to Harvard University in partial fulfillment of requirements for a Ph.D. in Geology.

Acknowledgements are due to a number of USA CRREL personnel for contributions to this study. Dr. R.W. Gerdel, formerly Chief of the Environmental Research Branch, encouraged and supported the work in its early stages. Dr. P.L. Johnson, formerly Research Ecologist, assisted in identifying several plant species. Mr. R.K. Haugen, Research Geographer, provided estimates of potential evapotranspiration by the Thornthwaite method. Mr. P.V. Sellmann, Research Geologist, directed the exploratory drilling; and discussions with him and with Dr. Jerry Brown, Research Soil Scientist, were invaluable in conducting the research and interpreting the results.

Specialists G.M. Font-Jimenez, H. Galle, H. Natvig, R. Burton, H. Csergei, and F. Banfield provided able field assistance. Specialist R.V. Musselman conducted the laboratory work summarized in Appendix B. Mr. J.V. Tedrow directed construction and repair of the stream-gaging weir. Mrs. Madonna Hartley cheerfully helped with many logistic details during the field work. Mr. H. Larsen very competently handled the drafting of the figures.

The U.S. Army Meteorological Team at Ft. Wainwright, Alaska, collected routine meteorologic data in 1965-1967. The Surveys and Mapping Branch, Alaska District, U.S. Army Corps of Engineers, provided invaluable help in overseeing and monitoring the topographic mapping. The cooperation of Messrs. T.C. Freeman and R. Beaumont, U.S. Soil Conservation Service, Anchorage, Alaska, in establishing a snow course in the watershed is gratefully acknowledged.

Discussions with Dr. F.R. Hall, Department of Soil and Water Science, University of New Hampshire, greatly strengthened the author's understanding of many of the hydrologic questions arising during the study.

Drs. M.P. Billings and R. Siever, Department of Geological Sciences, and Dr. H.A. Thomas, Division of Engineering and Applied Physics, Harvard University, critically reviewed the manuscript and made several suggestions which improved the presentation.

This work is dedicated to the late Dr. John P. Miller, Department of Geological Sciences, Harvard University, who initially inspired the author to study hydrology and geomorphology.

Manuscript received 22 June 1970.

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ABSTRACT

The results of a four-summer (1964-1967) hydrologic study of the watershed of Glenn Creek, about 8 miles north of Fairbanks, Alaska, in the Yukon-Tanana uplands physiographic province, are presented. This work was initiated to provide initial base line hydrologic data for a small subarctic watershed, the first of its kind in North America. Standard hydrologic and meteorologic instrumentation was used, and streamflow characteristics were analyzed by standard hydrograph-analysis techniques. The stream is second-order, and drains an area of 0.70 square mile. Basin elevations are from 842 ft to 1618 ft. In regard to topography, geology, soils, permafrost, vegetation, and climate, the watershed seems to be representative of low-order, low-elevation drainage basins in the province. Analysis of rainfall-runoff data indicates that about half the 12.9-in. normal annual precipitation is runoff. The remainder is the actual evapotranspiration, which equals only about 30% of estimated potential evapotranspiration. For individual storms, runoff/rainfall proportions were from 0.03 to 0.42, and were positively correlated with antecedent discharge of the stream, which is a measure of watershed wetness. The stream responds rapidly to rainstorms except when the basin is very dry, and has markedly slow recessions compared with temperate-region streams of similar size. Rate of recessions is apparently controlled by concurrent evapotranspiration rates. Analysis of hydrographs and knowledge of the physical characteristics of the basin indicate that storm runoff occurs initially as surface runoff from bare soil areas adjacent to the stream, while recessions are dominated by a combination of tunnel flow beneath moss-covered parts of the basins and typical groundwater flow through the moss and soils. Peak discharges for individual storms could be well estimated by an equation including antecedent discharge, total precipitation and storm duration, and average recession constant. These results represent the first detailed hydrologic data from the discontinuous permafrost zone of the North American taiga and should be of significance to the International Hydrological Decade and International Biological Program.

HYDROLOGY OF THE GLENN CREEK WATERSHED, TANANA RIVER BASIN, CENTRAL ALASKA

by

S. Lawrence Dingman

INTRODUCTION

The objective of this study was to achieve an understanding of the hydrologic behavior of a small watershed typical of a significant portion of the discontinuous permafrost zone of the subarctic.

There can be little doubt of the need for such knowledge. Because the constant land area of the world will be required to support a rapidly increasing population, more intensive development of the subarctic and other presently sparsely populated areas will occur. The water on this area is a crucial resource; and the consequences of mismanaging it are becoming increasingly apparent in many parts of the world. In the subarctic, where precipitation is generally low, perennially frozen ground (permafrost) is present, and surface water bodies are frozen for much of the year, rational management of water resources is especially important. It is hoped that this study will contribute to the base of knowledge required for such management.

Little information on hydrologic processes in the Arctic and subarctic exists. Straub and Johnson (1950) wrote a chapter on hydrology for Stefansson's *Encyclopedia Arctica*, but the chapter was based on very limited field data and was never formally published. Ellsworth and Davenport (1915) presented streamflow data from 122 sites in central Alaska, along with precipitation data from 20 stations, for the period 1907-1912; they also presented some discussion of hydrologic processes.

The U.S. Geological Survey began its comprehensive stream-gaging program in Alaska in 1946; there are presently 117 regular gage sites in that state. The National Weather Service maintains 176 precipitation stations in Alaska (roughly one station per 2200 square miles), most of which report precipitation on a daily basis only. Most of these precipitation stations are in populated lowland areas; and the extensive mountain and upland regions, where precipitation is generally highest, are virtually unsampled. The sparseness and low elevation of precipitation stations and the generally large size of the drainage basins gaged make an investigation of hydrologic processes by examination of normally published rainfall and streamflow records unfruitful. For these reasons, the study of Glenn Creek watershed was initiated.

Four factors influenced the selection of Glenn Creek as a study site: representativeness, degree of disturbance by man, accessibility, and size. It was first decided that a watershed of about 1 to 5 square miles would be small enough to allow reasonably detailed measurements of streamflow, precipitation, and other meteorological factors, without excessive cost and inconvenience, and yet would be large enough to integrate any extremely local factors affecting hydrologic processes. Good accessibility required that the watershed be located near Fairbanks and fairly close to the highway network. Consideration of these two factors allowed selection of several possible sites from studies of maps and air photos before field examination.

Because of the extensive placer-gold mining operations in the Fairbanks district, the requirement that the study area be essentially undisturbed by man was extremely limiting. Field inspection showed that the watershed of Glenn Creek was one of the very few watersheds whose slopes were not ditched to collect water for hydraulic mining operations. While evidence of human occupation and three small mine shafts have since been discovered in the basin, it is safe to say that the hydrologic regime is not affected by these features. In places, trees were cut by early miners, but there is no indication that the area was extensively logged. Charred stumps and tree-ring studies* show that at least a portion of the basin has been burned within the last 60 years, but this is true of virtually all of interior Alaska.

Because little was known of the hydrologic regime of the area, the hydrologic representativeness of Glenn Creek could not be determined. A general knowledge of geologic, permafrost, vegetational, and topographic conditions was relied on in assuming that Glenn Creek watershed was generally similar to a large portion of interior Alaska. Certainly there were no factors leading one to believe that the area was significantly atypical.

In determining representativeness, one must first state the area of interest,or "target" area, about which one wants information. Is it a watershed typical of Alaska, the subarctic, interior Alaska, or some other geographical division? It seems clear that a small watershed cannot be expected to be approximately quantitatively representative of a region of generally similar climate, topography, geology, and vegetation. A physiographic province of moderate extent can be expected to exhibit such general similarities, and in addition it provides a generally convenient basis for analogy to other parts of the world. Thus, Glenn Creek watershed might be considered to be representative of the Yukon-Tanana uplands physiographic province as defined by Wahrhaftig (1965), an area of some 35,000 square miles (see Fig. 1). Qualitative, and even quantitative, similarities may exist between certain hydrologic processes in Glenn Creek Basin and in areas outside the target area.

It is never possible to state that a given watershed is completely representative of a larger area, given the areal and elevational variations in climate, vegetation, and geology which are present in nature. Several facets of the watershed in question, which together should determine its hydrologic behavior, must first be considered separately. This is done in the section of the report describing the Glenn Creek Basin (see p.3). The degree of similarity between the watershed and the larger target area for each of these facets must then be considered together to obtain an overall picture of its representativeness.

This study was begun in June 1964, and continued through that summer and subsequent summers until August 1967. Periods of collection of the various types of data are described in detail in the chapter on *Data collection* (p.37).

The general plan of this report is to attempt to form conclusions by the inductive method about hydrologic processes (principally rainfall-runoff relations) operating in an upland watershed representative of at least a large portion of interior Alaska. Thus, the characteristics of topography, geology, soils, permafrost, climate, and vegetation of Glenn Creek Basin are first described in some detail. Previous studies of these characteristics in central Alaska, and direct observations and measurements in the basin, are the bases for these descriptions. These characteristics establish certain boundary conditions on the hydrologic processes. Next, the types of hydrologic data collected during this study and the periods of collection are presented. Finally, these data and the boundary conditions imposed by the physical characteristics of the basin are used to *infer* the proximal sources of streamflow.

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* Personal communication (Haugen, 1966).

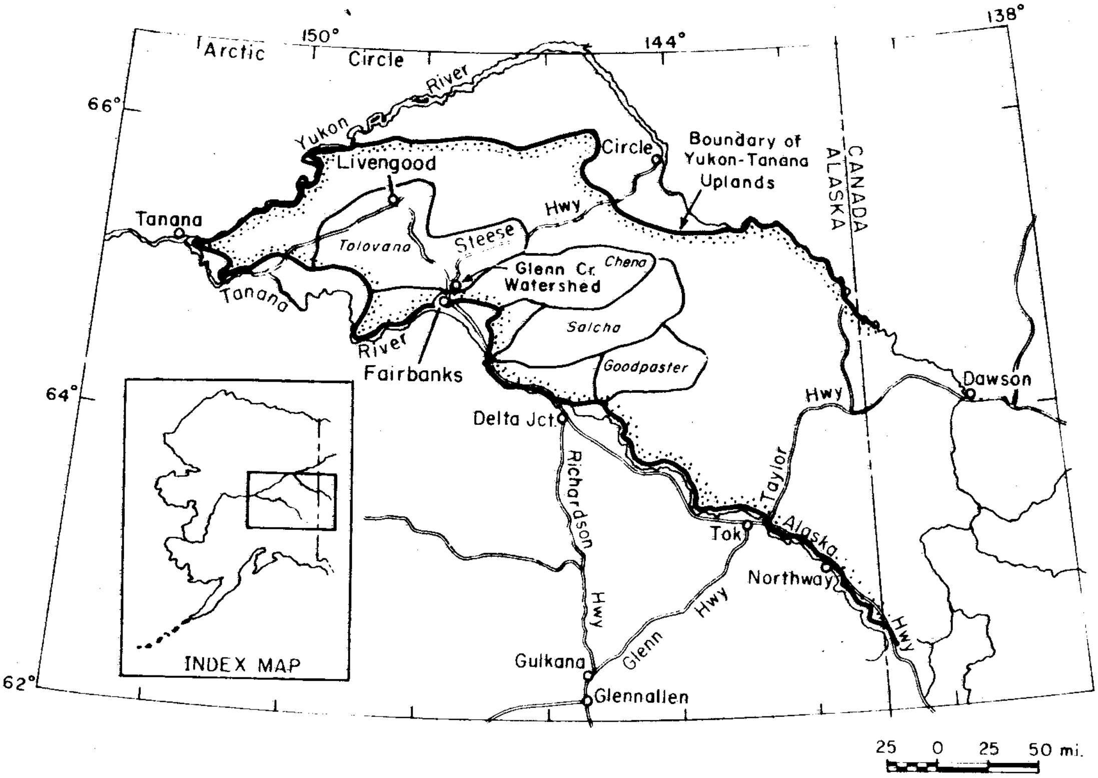

Figure 1. Map of Alaska showing boundaries of the Yukon-Tanana uplands physiographic province, and locations of Glenn Creek watershed, and the Tolovana, Chena, Salcha, and Goodpaster watersheds.

GLENN CREEK WATERSHED

Location

The watershed of Glenn Creek is located in Fox, Alaska, 8 miles north-northeast of Fairbanks, at latitude 64°57'N, longitude 147°35'W (Fig. 1). It drains northwestward to Goldstream Valley, which in turn drains westward to the Tanana River at Minto, and thence to the Yukon River.

Topography

Topographic data in the basin were taken from a map of the watershed made from aerial photographs, prepared by Walker and Whitford, Inc., of Seattle, Washington. This map was made to national map standards, with third-order (1 in 5000) horizontal control and vertical control error less than 1 ft. The original map scale was 1:2400 with a 5-ft contour interval. To facilitate mapping of various features in the watershed, a 500-ft grid system was surveyed in the basin, with each grid point marked by a metal stake. The base line of the grid was laid out along the maximum dimension of the basin parallel to the lower reaches of the main channel, at a bearing of 120° true from the origin at the weir. The origin was identified as point 0, and subsequent points as A through M. Points on lines at right angles to the base line were identified by a letter according to the base-line point of intersection, a letter indicating whether they were north (actual bearing 30°) or south of the base line, and a number indicating the distance from the base line. For example, point BN2 was 1000 ft from point B along the bearing 30°, and point ES3 was 1500 ft from point E along the bearing 210°. Figure 2 shows the grid-point locations.

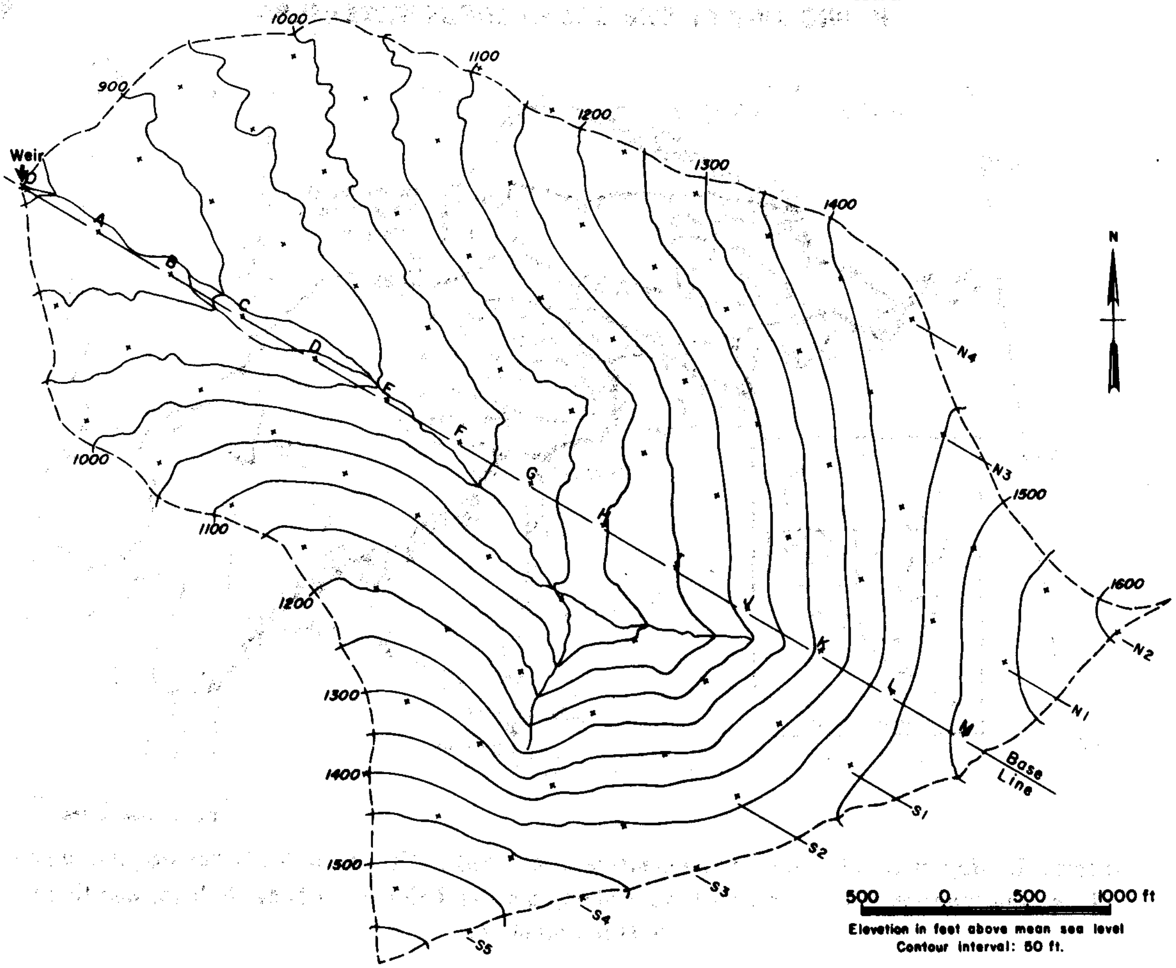

Figure 2. Topographic map of Glenn Creek watershed showing plan of mapping grid.

The area of the basin, measured by planimeter, is 0.70 square mile. Elevations are from 842 ft at the weir (basin outlet) to 1618 ft at the eastern perimeter. The area-elevation curve, determined by the grid-sampling technique described by Wallis and Bowden (1962), is shown in Figure 3. The distribution of slopes in the basin, also determined by grid-sampling, is shown in Figure 4. Average basin slope is 0.184. Figure 5 shows the profile of Glenn Creek and its major tributary. Throughout the lower 4000 ft of its course, the channel gradient varies little, and averages 0.049 (259 ft/mile).

Glenn Creek watershed is located near the southern edge of the west-central portion of the Yukon-Tanana uplands (Fig 1). Wahrhaftig (1965, p. 24) gave the following description of this physiographic province:

> "Rounded even-topped ridges with gentle side slopes characterize this section of broad undulating divides and flat-topped spurs. In the western part these rounded ridges trend northeast to east; they have ridge-crest altitudes of 1,500-3,000 feet and rise 500-1,500 feet above adjacent valley floors. The ridges are surmounted by compact rugged mountains 4,000-5,000 feet in altitude but have some domes as high as 6,800 feet, and rise 1,500-3,000 feet above adjacent valleys.... Valleys in the western part are generally flat, alluvium floored, and ¼-½ mile wide to within a few miles of headwaters. Streams in the eastern part that drain to the Yukon flow in narrow V-shaped terraced canyons.... Most streams in the western part follow courses parallel to the structural trends of the bedrock.... Drainage divides are very irregular."

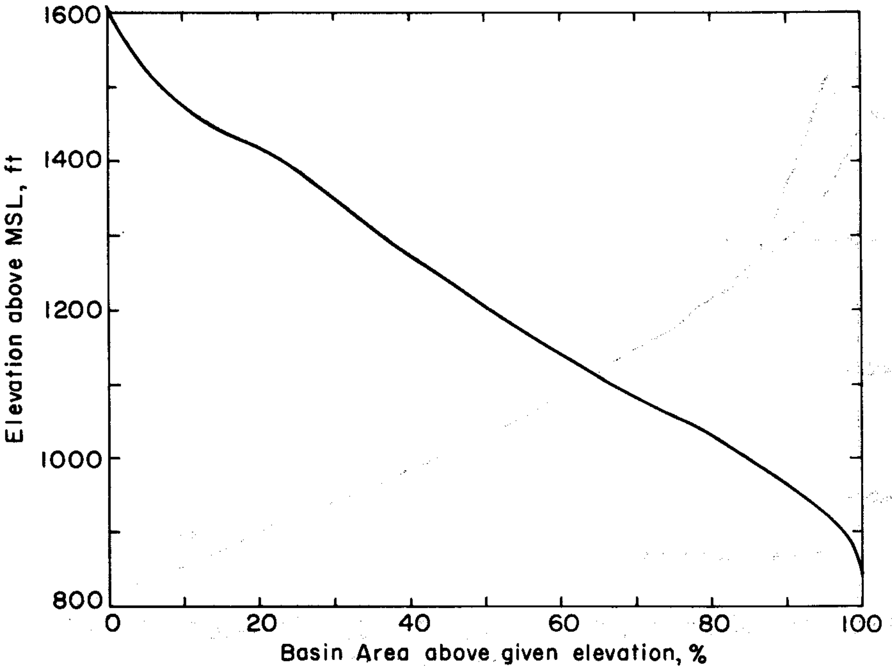

Figure 3. Area-elevation curve for Glenn Creek watershed.

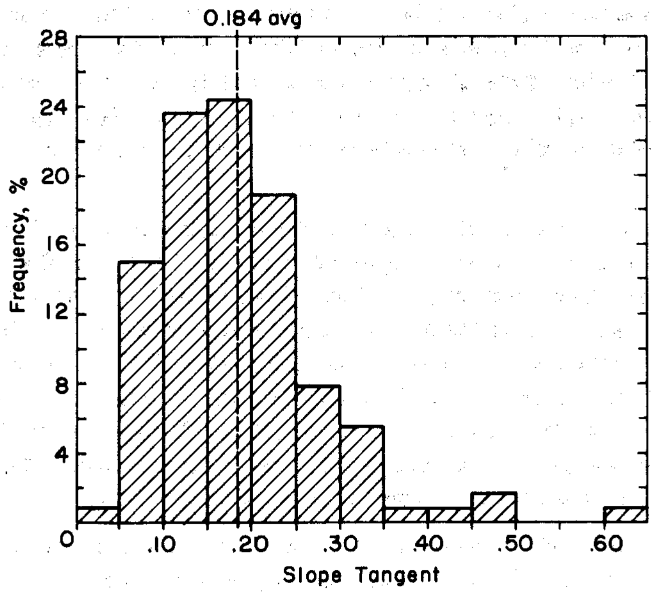

Figure 4. Distribution of ground slopes in Glenn Creek watershed.

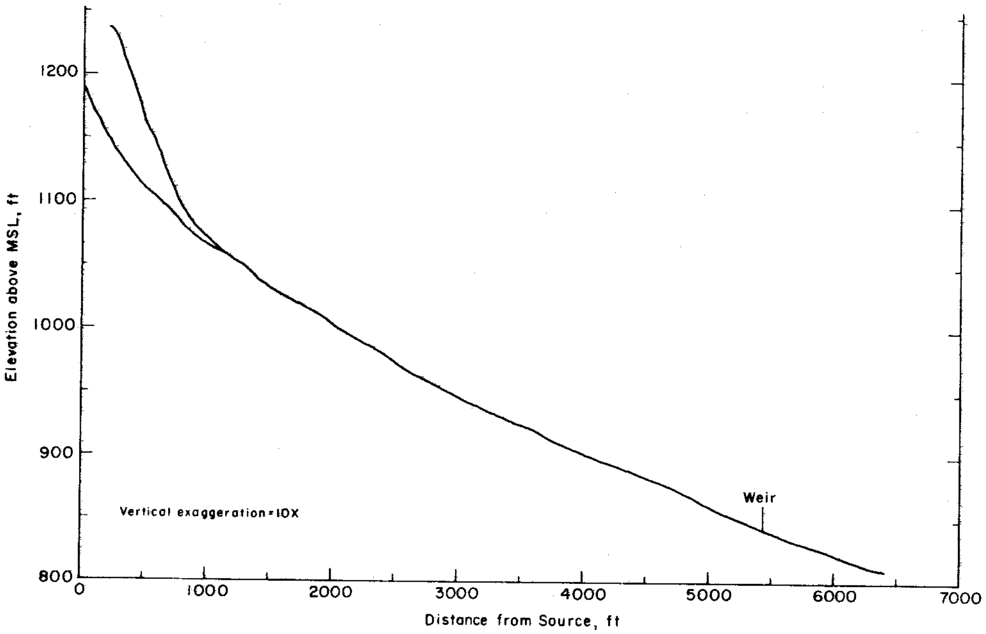

Figure 5. Longitudinal profile of Glenn Creek.

In terms of basin relief, ridge-crest elevation, and valley-side slopes, Glenn Creek appears to be typical of the western Yukon-Tanana uplands. A conventional topographic description of the region near the basin would characterize it as maturely dissected; the area-elevation curve is typical of the "mature" or "equilibrium" form of Strahler (1957). Figure 6 is a combined area-elevation curve for the basins of the Tolovana, Chena, Salcha, and Goodpaster Rivers, which together drain 8570 square miles (about 25%) of the Yukon-Tanana uplands. Assuming that this curve represents the entire province, over 80% of the Yukon-Tanana uplands lies above Glenn Creek Basin.

Based on field and air-photo identification of stream channels as well defined linear depressions lacking a continuous ground cover of grass or moss, and associated with water-loving vegetation such as sedge tussocks, willows, and alders, Glenn Creek is a second-order stream for 4800 ft of its 5450-ft course (actual stream channels are shown in Fig. 2). However, considering only the channels shown as blue lines on the 1:63,360 U.S. Geological Survey topographic map (Fairbanks D-2 Quadrangle), Glenn Creek is first-order. The average drainage area of such blue-line first-order streams in the basin of the Little Chena River, to the east of Glenn Creek in the Yukon-Tanana uplands, is 0.97 square mile. Thus, in terms of drainage area, Glenn Creek appears to be typical of streams of its order in the region.

Drainage density is defined as the total length of stream channels draining an area divided by the area. However, a *stream channel* in practice may be defined in a number of ways. On small-scale maps, when no corroborating field or photographic evidence is available, an objective measure of drainage density is generally made by constructing a channel network on the basis of contour crenulations, a channel being delineated where the contours indicate a continuous linear depression connected to another channel. When this is done for Glenn Creek, using a topographic map of a scale of 1:63,360, with a 50-ft contour interval, drainage density is 5.1 miles/square mile (18,850 ft of channel). Using the larger-scale map (1:2400) and the same criterion for identifying channels, drainage density is 5.8 miles/square mile (21,600 ft of channel). If actual stream channels are identified from field and air-photo inspection as described above, the true drainage density is 2.6 miles/square mile (9760 ft of channel).

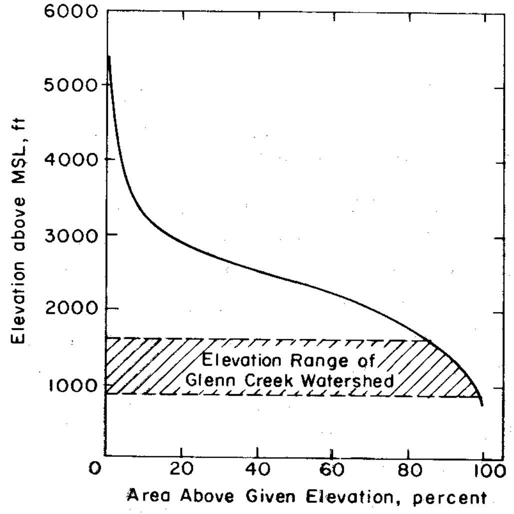

Figure 6. Combined area-elevation curve for the drainage basins of the Tolovana, Chena, Salcha, and Goodpaster Rivers (see Fig. 1 for locations).

Strahler (1964, p. 4-52) stated that the lowest drainage densities reported in the literature are between 3 and 4 miles/square mile, for the resistant sandstone terrains of the Appalachian Plateau. However, the very low value found for Glenn Creek Basin appears to be typical of much of interior Alaska. Drainage-density values, where channel length was determined from contours on 1:63,360-scale maps, were determined for some 80 basins in central Alaska (not including Glenn Creek Basin), ranging in area from 0.05 to 2199 square miles. The drainage densities ranged from 1.8 to 9.8 miles/square mile, and fit the regression equation

$$D = 5.06 \; A^{-0.13}$$

where D is drainage density in miles/square mile and A is area in square miles. With this relationship the average drainage density for a basin of 0.7 square mile is 5.3 miles/square mile, which compares with a measured value of 5.1 miles/square mile for Glenn Creek Basin. This indicates that Glenn Creek Basin is also typical of the region in regard to drainage density.

Probable reasons for the very low values of drainage density in central Alaska are: 1) summer rains are of generally low intensity, so that a high proportion of the precipitation evaporates or infiltrates; 2) total precipitation, at least at lower elevations, is low; 3) the moss carpet which covers a high proportion of the region (described later) acts as a very permeable soil, promoting infiltration; 4) the normal period of greatest runoff (spring snowmelt) occurs when the ground is frozen, reducing erosion; 5) where areas lacking perennially or seasonally solidly frozen ground and lacking a moss cover exist, soils are permeable, increasing infiltration.

It is also of interest to compare other quantitative geomorphic measures for Glenn Creek with those established in other regions. Strahler (1950) plotted average valley-side slope (θ_g, degrees) versus average second-order channel gradients (θ_c, degrees) for a wide range of geographical regions (not including permafrost areas) and found that these values could be fitted by the regression equation

$$\theta_g = 4\ \theta_c^{0.8} \quad .$$

For Glenn Creek Basin, $\theta_c = 2.8°$, so the calculated θ_g is 9.1°, compared with an actual θ_g of 10.4°. Thus Strahler's relationship seems to apply to Glenn Creek, and presumably to second-order streams in much of central Alaska, as well as to other geographic areas.

The *ruggedness number* is defined as the product of relief H and drainage density D, when both parameters are expressed in the same units. Strahler (1964, p. 4-67) stated that values of HD "range from as low as 0.06 in the subdued relief of the Louisiana coastal plain to over 1.0 in coast ranges of California or in badlands on weak clays."* For Glenn Creek Basin, H = 776 ft or 0.147 mile, and $HD = 0.38$. Strahler (1958) also defined a *geometry number* HD/S_g where S_g is the average ground slope in percent. He found that this parameter fell in the range 0.4 to 1.0 for six regions with markedly different values of H, D, and S_g. For Glenn Creek, $HD/S_g = 0.38/18.4 = 0.021$, well outside the range found by Strahler. This low value throws some doubt on his conclusion that "the geometry number tends to be conserved about a common value and that a change in any one of the three components is compensated for by changes in one or both of the other two, thus tending to keep the product constant" (Strahler, 1964, p. 4-68).* The value found for Glenn Creek could be interpreted as being due to anomalously low drainage density relative to the relief and slope of the drainage basin. By dimensional analysis, Strahler (1958) related drainage density to runoff intensity per unit area Q, the mass rate of sediment removal per unit area K_e, and basin relief H, such that

$$D = \frac{1}{H}\, f\,(Q \cdot K_e).$$

The product $Q \cdot K_e$ is clearly a function of fluvial (i.e. sheetwash) erosion intensity; thus the low value of HD/S_g for Glenn Creek would seem to indicate that such erosion is not predominant in determining the form of the basin. If this is true, some other mode of erosion, such as mass wasting or chemical erosion, is most important in the basin. This conclusion is corroborated by other considerations, to be discussed later.

Geology, soils, and permafrost

Introduction. Knowledge of the geology and soils of a region is important in formulating its complete hydrologic and morphologic picture. In addition, permafrost exists in the drainage basin of Glenn Creek as a special subsurface condition that can be expected to have important hydrologic and morphologic effects. In this section, the distributions and properties of the subsurface materials, especially those which are important hydrologically, are described.

Bedrock geology. Much of the Yukon-Tanana uplands, including the basin of Glenn Creek, is underlain by the Birch Creek schist. The geologic maps of Péwé (1958) and Péwé *et al.* (1966) show this formation in the upper parts of the watershed, where, according to these descriptions, it is covered by less than 3 ft of surficial materials. From radioactive-dating measurements, the formation is considered to be Precambrian or early Paleozoic in age. It is described as "light to dark-gray, reddish-brown to tan-weathering schists, predominantly quartz-sericite schist and micaceous quartzite," but includes "muscovite-biotite schist, garnet-mica schist, calcite- and dolomite-bearing schist, dark-gray to black chloritic and graphitic schist, amphibolite, light-gray to white impure marble, and light-brown ... gneiss" (Péwé *et al.*, 1966).

During a traverse in which grid points throughout the watershed were visited, outcrops were noted at only a few places in the upper portions of the basin (see Fig. 7a): 1) about halfway between grid points GN4 and HN4; 2) immediately west of point HN2; 3) about halfway between points HS2 and HS3; and 4) about halfway between points KN2 and KN3. Examination of a hand specimen from the outcrop near point HN2 indicates that the rock is tan-weathering micaceous quartzite.

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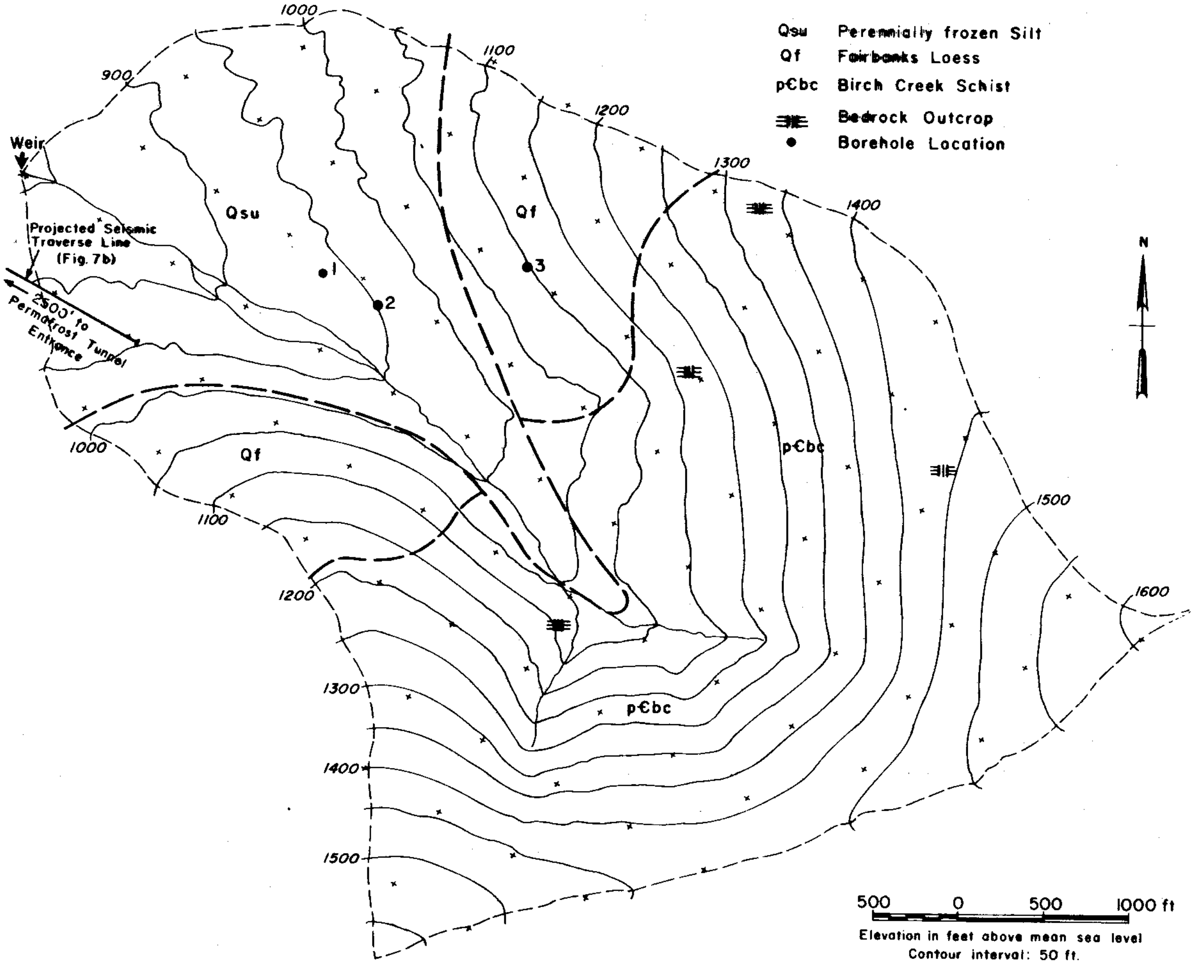
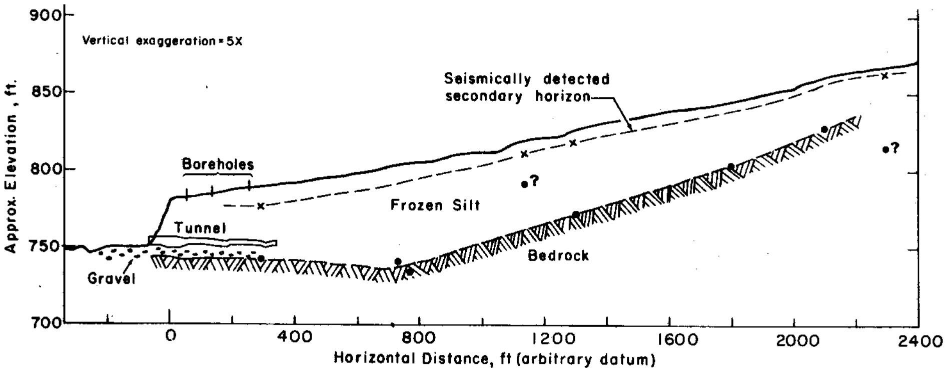


a) Geology of the Glenn Creek watershed, according to Pewe (1958). Location of exploratory boreholes and bedrock outcrops of present study also shown.



b) Sellmann's (1967) interpretation of seismic and borehole observations at lower end of watershed.

Figure 7. Geology of Glenn Creek watershed and interpretation of seismic and borehole observations at lower end of watershed.

Sellmann (1967), in connection with geologic studies of the Cold Regions Research and Engineering Laboratory (CRREL) Permafrost Tunnel, located about ½ mile downslope from the weir site, determined depth to bedrock along a 2300-ft upslope transect by means of drill logs and seismic refraction methods. His results in relation to Péwé's (1958) geologic boundaries and the positions of outcrops observed in the field are shown in Figure 7b.

Additional information is available from three auger holes drilled on the north side of the basin, at the locations indicated in Figure 7. Hole 1 was drilled near the bottom of the north slope, about halfway between lines CN and DN, with refusal at 13 ft. This refusal probably indicated bedrock, as fragments of completely weathered schist, with the consistency of clay, were recovered from the auger at this depth. Such clay has been reported as commonly occurring just above bedrock by several writers (see Taber, 1943, p. 1464-1465; Cederstrom, 1963, p. 14-15). Hole 2 was located about 100 ft from point DN1, with refusal at 6 ft. Indications from the behavior of the auger and materials recovered from it were that we had encountered a layer of coarse gravel, which may or may not have lain directly on bedrock. Hole 3 was drilled just west of point FN2, and refusal was encountered here at 14 ft, with completely weathered schist fragments again recovered from the auger at this depth.

In probing to determine depths to permafrost (a graduated ¼-in. steel rod was used), impenetrable gravel horizons were frequently encountered, almost exclusively on the upper portions of the south slope, at depths of less than 3 ft (see Fig. 7b). It is likely that in many cases this gravel represented a thin weathered horizon immediately overlying bedrock.

These data, taken together, indicate that the bedrock surface is, approximately, a replica of the ground surface of the watershed. This statement is most accurate for the upper portions of the basin, particularly on the south side, where bedrock is probably within a few feet of the surface everywhere. In the lower portions, the bedrock surface slopes more steeply than the ground surface, and there is a wedge of surficial materials above it which thickens downslope. The seismic and lower borehole records indicate that the maximum thickness of this wedge is between 30 and 60 ft.

This general picture coincides with the interpretations of most who have studied the geology in the Fairbanks area, as illustrated by Tuck (1940, Fig. 3), Taber (1943, Fig. 6), Péwé (1949, Fig. 2; 1954, p. 315; 1955b, Fig. 9, 10; 1965, Fig. 1-12), and Cederstrom (1963, Fig. 6).

With the exception of observations of the small and scattered outcrops, there are only indirect indications of the nature (*tightness*) of this bedrock surface within the watershed. Péwé (1958) described it as follows: "Surface drainage good to excellent. Joints, faults, fracture cleavage, and foliation result in poor to fair permeability." In the vicinity of Glenn Creek, Péwé *et al.* (1966) showed foliation striking east-northeast and dipping 15° to the north-northeast (toward the basin outlet). These indications, and the fact that the Birch Creek formation ultimately weathers to a residual clay up to 15 ft thick (Taber, 1943, p. 1464-1465; Cederstrom, 1963, p. 14-15), suggest that the drainage basin of Glenn Creek is essentially *tight*, with little likelihood of significant inflow or outflow through the bedrock, even in the absence of permafrost.

Surficial geology. Three principal types of surficial deposits are found in the Fairbanks area: 1) gravel, 2) silt, and 3) organic silt (muck). These have been studied most extensively by Tuck (1940), Taber (1943, 1953, 1958), Péwé (1955a, 1958) and, in the vicinity of Glenn Creek, by Sellmann (1967). The general relationships among these materials can be most quickly grasped by reference to Figure 8, taken from Péwé (1965). The Fairbanks area, including Glenn Creek watershed, was not glaciated during the Pleistocene.

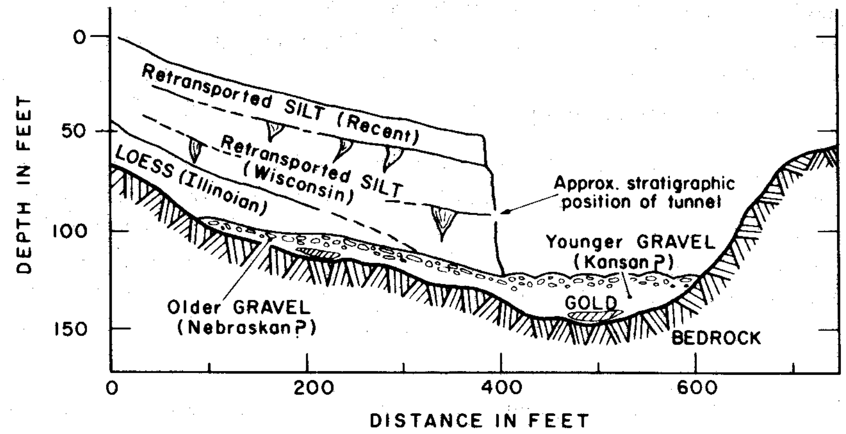


Figure 8. General stratigraphic relations in central Alaska. The retransported silt is organic-rich (muck) (after Péwé, 1965).

Gravels. Gravels in the uplands of the Fairbanks area are of two types. By far the greater in volume are the well known gold-bearing alluvial deposits of the larger creek valleys, which have been extensively mined by dredging. On the basis of their relationships to younger deposits and the presence of scattered fossils of Quaternary mammals, these are considered to be of early Quaternary age, possibly Kansan and Nebraskan (Péwé, 1965, p. 8). Péwé *et al.* (1966) described the creek gravels as "well stratified layers and lenses of poorly sorted angular to subrounded brown to buff, locally heavy iron-stained, auriferous sandy gravel, containing boulders of quartz, gneiss, and schist as much as 24 in. in diameter." They range in thickness from a few feet to more than 150 ft (Tuck 1940, p. 1298; Taber, 1943, p. 1467). Sellmann (1967, Fig. 3b) reported a thickness of about 13 ft beneath the CRREL Permafrost Tunnel, on the basis of borehole records (see Fig. 7b). If the top surface of these gravels is approximately horizontal, they must pinch out about 800 ft upslope from the tunnel portal (see Sellmann, 1967, Fig. 2a), and hence are not present beneath the watershed.

Lesser amounts of gravel, mixed with varying proportions of finer material, are present in 1 to 10-ft-thick layers immediately above bedrock throughout the uplands of central Alaska. These were thought to be solifluction deposits by Péwé (1965, p. 8). Cederstrom (1963, p. 15-16) reported material of this general type encountered just above bedrock in wells drilled in the area just south of Glenn Creek, and felt that at least some of these deposits are alluvial.

Several lines of evidence indicate that this material is present throughout the drainage basin of Glenn Creek. Reference has already been made to the apparent presence of an impenetrable gravel layer at the base of the hole 2, and the gravel layers encountered at shallow depths over much of the upper portions of the watershed during probings for permafrost. In addition, spoils adjacent to several small mine shafts (which, with one exception, are located in the permafrost-free upper parts of the north slope) contain abundant gravel. These shafts, which were probably one- or two-man mining operations, were most likely dug soon after "discovery" in the Fairbanks District in 1902 in an attempt to reach hopefully gold-rich gravel layers. It is unlikely that these mines were successful, as there are only four or five in the watershed, and there is very little associated evidence of mining equipment or human habitation. Futhermore, Tuck (1940, p. 1298) and Péwé (1958) both indicated that the emplacement of gold accompanied intrusions of igneous rock during the Mesozoic, and that significant secondary gold is found only down valley from such intrusive bodies, none of which occurs in the drainage basin of Glenn Creek.

In all of the gravel deposits observed within the watershed, the particles are angular, with no evidence of transportational wear. This suggests that the material represents weathered bedrock in place or transported short distances downslope by creep and/or solifluction, rather than the alluvial deposits found in some areas by Cederstrom (1963, p. 15-16).

Gravel layers or lenses are also present at various horizons throughout the generally silt-sized overburden above the bedrock. This was established by the behavior of the auger in drilling holes 1, 2, and 3, and by the presence of gravel-sized particles (up to 2 in., maximum dimension) on the auger when withdrawn for sampling. In holes 2 and 3, the gravel layers were encountered within 2 ft of the ground surface; that is, as much as 10 ft above the bedrock surface. The presence of these layers or lenses has implications concerning the depositional history of the entire overburden section, which is discussed below.

Silt. The origin of the extensive deposits of mainly silt-sized material that mantle the bedrock in the Fairbanks area, and indeed much of central Alaska, has been the subject of considerable controversy. Hypotheses of a fluvial, marine, or lacustrine origin of these deposits by some of the earlier writers have been largely discounted (see Péwé, 1955a). Taber (1943, 1953, 1958) maintained that they are largely residual, while Tuck (1940) and Péwé (1955a, 1965) held that they had an eolian origin. The latter explanation is now generally accepted (Sellmann, 1967), and the deposits are referred to as the "Fairbanks loess" on geological maps (Péwé, 1958; Péwé *et al.*,1966). However, Cederstrom (1963, p. 9-11) did not find the evidence for either eolian or residual origin compelling, and stated that the question remains moot.

Evidence from Glenn Creek Basin indicates that silts of both eolian and residual origin mantle the bedrock. As noted earlier, gravel was encountered within 3 ft of the ground surface at many points, particularly on the upper south-facing slope (see Fig. 7b). Also, gravel was present throughout the sections at the three auger holes. Certainly the presence of gravelly horizons well above bedrock must indicate either a completely residual origin or the incorporation of coarse bedrock fragments as the eolian silt mantle moved *en masse* downslope. Grain-size analysis of a composite of samples collected along the entire length of the CRREL Permafrost Tunnel (Sellmann, 1967, Fig. 4) revealed a distribution curve very similar to the curves for unretransported hilltop silts from the Fairbanks area (Péwé, 1955a, Fig. 7) (see Fig. 9). Both curves are similar to loess deposits found in other regions (Smalley, 1966, p. 669). These facts suggest that the valley bottom silts in Goldstream Valley are largely retransported eolian material, as originally stated by Tuck (1940, p. 1305).

However, the grain-size distribution curve of the < 2 mm fraction of samples recovered from drilling hole 3 is markedly dissimilar to curves for loess deposits (Fig. 9). The former shows poor sorting, with more than 50% of the material coarser than 0.1mm. Such a curve for an unglaciated upland area would, taken by itself, suggest a residual origin for the surficial material. However, the evidence of Péwé (1955a) and Sellmann (1967) strongly suggests significant eolian deposition in the general area, which the drainage basin of Glenn Creek must have experienced. Further, if the 0.1-mm fraction of the samples from hole 3 is plotted separately, the curve falls close to the curves of Péwé (1955a, Fig. 7) and Sellmann (1967, Fig. 4), making a partial eolian origin plausible.

Largely unretransported loess is apparently present in the northwestern one-third of the watershed, above the valley bottom. While not supported by grain-size analysis, the presence of this material is indicated by the presence of pronounced ridge-and-gully topography, which Péwé (1965, p. 10) states is characteristic of loess-covered areas in the vicinity of Fairbanks. The failure to encounter gravel while this area was being probed further supports this hypothesis.

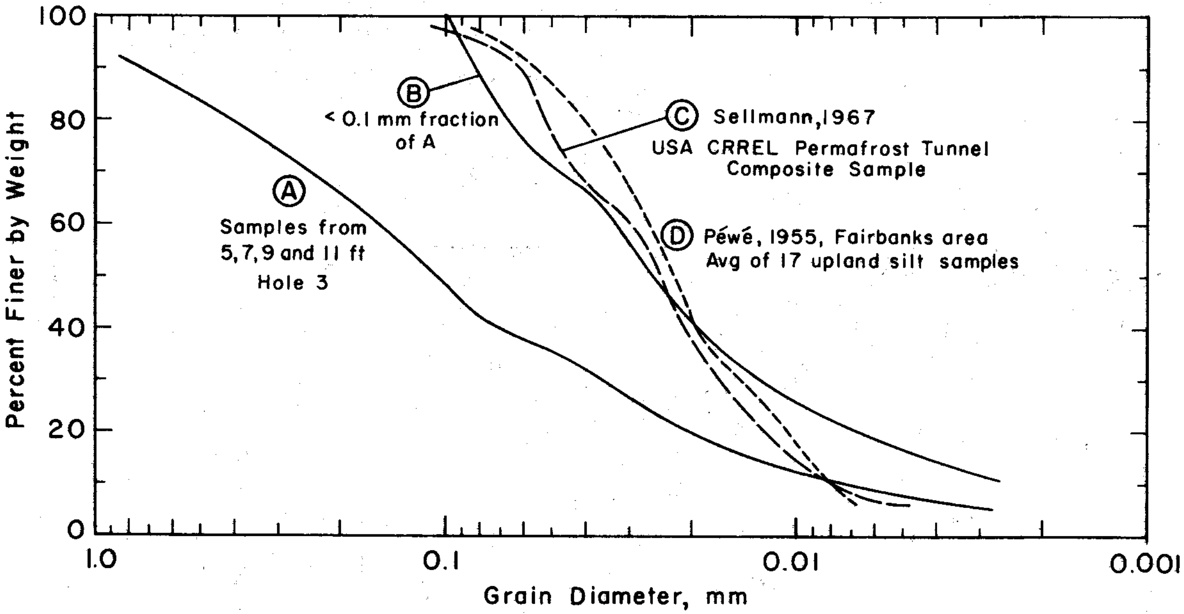


Figure 9. Grain-size distributions of silts in the Fairbanks area.

Muck. As observed by Tuck (1940, p. 1303) and Péwé (1965, Fig. 1-2), the organic-rich valley-bottom silts have a gradational contact with the generally organic-poor upland silts. Both those writers stated that these deposits have formed by downslope movement of originally wind-deposited silts, which incorporated organic materials as it occurred. This interpretation is supported by grain-size analyses of material from the CRREL Permafrost Tunnel (Sellmann, 1967), previously mentioned. The identifiable organic material is largely vegetation, and the flora is essentially the same as that presently growing in the area (Tuck, 1940, p. 1299; Taber, 1943, p. 1481). Considerable numbers of vertebrate fossils, including bison, mammoth, and horse, are also found (Péwé, 1965, p. 10). Radiocarbon tests of the organic material indicate that it is largely Wisconsin in age, but includes some post-Wisconsin (11,000-4,000 yr B.P.) material (Péwé, 1965, p. 10; Sellmann, 1967, Table 2).

The muck is virtually everywhere perennially frozen, and contains interstitial ice, thin ice bands, and massive ground ice in the form of vertical wedges and irregular bodies (Tuck, 1940, p. 1301; Taber, 1943, p. 1510-1528; Péwé, 1958; 1965; Sellmann, 1967, p. 22). Sellmann's (1967, Table I) analyses of material taken from the walls of the Permafrost Tunnel (exclusive of massive ground ice) showed ice volumes ranging from 54% to 79% and moisture contents between 45% and 128% by dry weight.

Exposures along the banks of Glenn Creek in its central and lower portions also reveal dark brown, organic-smelling silts with interstitial ice and ice segregations up to at least 2 ft thick. A short core sample of this material, taken from near grid point C, had a moisture content of 109% by dry weight. Indications are that the muck is present throughout the valley bottom of the drainage basin of Glenn Creek, which is defined by a pronounced break in slope (see profiles D, F, and H, Fig. 18). The high ice content of this material, which is typical of upland creek valleys in the area (Péwé, 1955b, p. 127), and the presence of massive ice bodies, make it highly susceptible to thermal erosion, a process which has recently been very active in the watershed of Glenn Creek.

Summary. Péwé's (1958) geological map indicates a band of "Fairbanks loess," with conjectured boundaries, extending across much of the basin in its central portions, with "perennially frozen, retransported eolian silt" at lower elevations (see Fig. 7). He mapped the upper

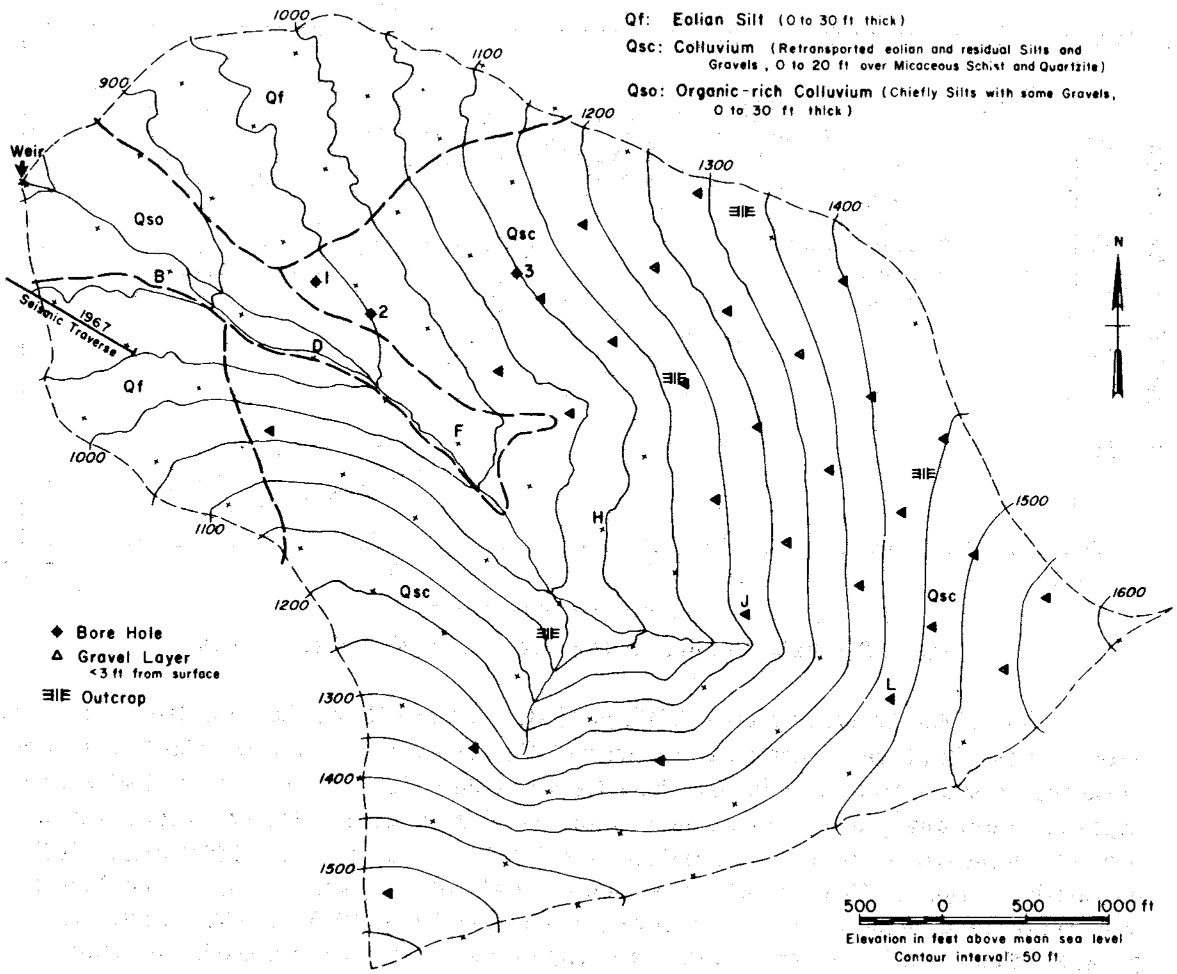


Figure 10. Geology of the Glenn Creek watershed, based on field observations during this study.

half of the basin as Birch Creek schist overlain by less than 3 ft of loess. On the basis of observations made during the present study, this geologic picture can be modified (Fig. 10).

Organic-rich retransported silts (Qso), essentially Péwé's (1958) "organic silt," are mapped here in the valley bottom adjacent to Glenn Creek in its lower portions. As noted later, this material is perennially frozen, with abundant ice segregations. Because of this, and because of its low permeability (Péwé, 1958), this material is poorly drained and transmits groundwater slowly.

The Fairbanks loess (Qf) is mapped in the northwestern one-third of the basin above the immediate valley bottom. This corresponds generally to Péwé's (1958) map, but the limits of the formation are modified based on the more detailed topographic map and subsurface observations.

Hydrologically, Péwé (1958) characterized the Fairbanks loess as follows: "Good surface drainage. Lateral permeability poor to fair; vertical permeability good. No permafrost. Water table generally deep." Except in regard to permafrost, these observations are apparently applicable to this material in the watershed of Glenn Creek. As noted later in this report, the loess to the south of Glenn Creek is perennially frozen below a depth of about 3 ft.

The remainder of the basin, roughly corresponding to Péwé's (1958) "Birch Creek schist . . . where a veneer of less than 3 ft of Fairbanks loess is present," is mapped here as 0 to 20 ft of colluvium, consisting predominantly of retransported residual and eolian silts and angular gravel, over Birch Creek schist (Qsc). Péwé (1958) stated that the material above the schist has good to excellent surface drainage, low permeability, and a deep water table. Permafrost is present on north-facing slopes. This characterization agrees with the field observations of the present study. That the water table is relatively deep in this material is indicated by the fact that perceptible moisture was observed only with auger samples taken from within 2 ft of refusal (11 ft) in hole 1, and not at all in holes 2 (6 ft) and 3 (13 ft). Cederstrom (1963, p. 37) stated that "the silt is a very poor water-bearing formation, although a few wells obtain seepage from it."

While the geology of the Yukon-Tanana uplands has not been mapped in the same detail as that of the Fairbanks area, the bedrock and surficial deposits of Glenn Creek basin appear to be representative of much of the province. The Birch Creek schist is the predominant bedrock formation of the Yukon-Tanana uplands. Other bedrock formations found in the province are "Paleozoic metamorphic rocks undifferentiated," "predominantly granitic intrusive rocks with associated extrusive rocks and metamorphic rocks," and "predominantly mafic intrusive rocks with associated greenstones" (Dutro and Payne, 1954). While there may be some differences in weathering characteristics of these formations and in the chemical compositions of waters draining them, the bedrock of Glenn Creek appears to be hydrologically similar to the crystalline rocks underlying virtually the entire Yukon-Tanana uplands.

Karlstrom's (1964) map of the surficial geology of Alaska shows Glenn Creek Basin as containing one formation: "coarse- and fine-grained deposits associated with moderate- to steep-sloped mountains and hills with bedrock exposures restricted to upper slopes and crest-lines." This formation is present on about 90% of the Yukon-Tanana uplands. The bedrock and surficial geology can thus be considered as essentially typical of almost all of the Yukon-Tanana uplands.

Soils. The soils of most of the watershed of Glenn Creek were mapped and described by Rieger *et al.* (1963); soils in the vicinity of Fairbanks were described in a general way by Kellogg and Nygard (1951).

According to Rieger *et al.* (1963), the Fairbanks silt loam underlies the south-facing slopes of the watershed and the northwest-facing slopes at the head of the basin. They stated that "Soils of the Fairbanks series are believed to be mature representatives of the Subarctic Brown Forest group. These soils occur principally on southerly slopes of hills and have developed in micaceous loess" (Rieger *et al.*, 1963, p. 34-35). These soils are described as well-drained, with permeabilities of about 0.2 to 0.6 in./hr, and depth to seasonally high water table greater than 15 ft. Permafrost is generally absent beneath these soils. Chemically, these soils have pH values from 5.5 to 6.5 and have high percentages of base saturation, with Ca and Mg the most abundant exchangeable cations (Rieger *et al.*, 1963, Table 9).

The Saulich silt loam is mapped on the north-facing slopes of the watershed. These slopes are Low-Humic Gley soils, which are characterized as "imperfectly drained and poorly drained soils that have thin, organic surface horizons and highly mottled mineral horizons" (Rieger *et al.*, 1963, p. 37). In their natural state these soils have permafrost at depths of 1 to 3 ft, and permeabilities from 0.2 to 0.6 in./hr above the permafrost table. Saulich soils are generally somewhat more acidic than those of the Fairbanks series (pH 5.0 to 6.0); no published data on chemical composition are available.

The valley bottom in the watershed, including a narrow strip adjacent to the main stream channel and extending well up toward the head of the basin, is mapped as Goldstream silt loam. This soil is similar in many respects to the Saulich soil in that it is poorly drained, with permafrost present at shallow depths. Its permeabilities are also from 0.2 to 0.6 in./hr. The principal difference between the two lies in the fact that the Goldstream soils are wetter, with the water table commonly at or above the ground surface; the Goldstream soils also contain considerable organic material even at depth. The pH is reported as 5.0 to 5.5 at the surface, increasing to 6.5 to 7.0 at depth.

The soil descriptions and boundaries presented by Rieger *et al.* (1963) correspond in general with the author's observations in the field, with one exception. As noted, these authors mapped the Fairbanks silt loam on the northwest-facing slopes at the head of the basin; however, there are a number of observations which indicate that the soils in that area belong to the Saulich series. Most important, the vegetation (discussed in detail later) is identical to that on much of the north-facing slopes, consisting generally of black spruce with a ground cover of thick mosses. Secondly, permafrost (discussed below) is present at shallow depths throughout the area. Both features are characteristic of the poorly drained Saulich soils, rather than the Fairbanks series. The Goldstream soil is probably limited to the Qso unit on the geological map (Fig. 10).

Soils throughout the Yukon-Tanana uplands have been mapped only on a reconnaissance basis. Kellogg and Nygard (1951) mapped the eastern half of the province as largely Mountain Tundra soils, with a high proportion of Subarctic Brown Forest soils and lesser areas of Lithosols, Bog and Half-Bog soils, and Alluvial soils. They showed the western portion, including Glenn Creek Basin, as dominantly Subarctic Brown Forest soils, with considerable areas of Half-Bog soils and lesser amounts of Alluvial, Mountain Tundra, Bog, and Podzol soils. On the basis of the areal proportions estimated by Kellogg and Nygard (1951), the Yukon-Tanana uplands consist of about 30% each of Subarctic Brown Forest soils and Mountain Tundra soils, 15% Half-Bog soils, 10% Lithosols, and lesser areas of Alluvial, Podzol, and Alpine Meadow soils. Since the Low-Humic Gley soils mapped by Rieger *et al.* (1963) are included as Half-Bog soils by Kellogg and Nygard (1951), two of the three most important Great Soil Groups of the Yukon-Tanana uplands are well represented in the Glenn Creek drainage basin. The Mountain Tundra soils occur at elevations above those of the basin.

The spatial arrangements of the soils of Glenn Creek watershed, in terms of slope degree and aspect, are characteristic of those of the lower uplands along much of the southern border of the Yukon-Tanana uplands province (Rieger *et al.*, 1963, Fig. 4; Schoephorster, personal communication). Thus, on the basis of present knowledge, it appears that the soils of Glenn Creek Basin are typical of the lower elevations of much of the Yukon-Tanana uplands.

Permafrost. The presence of permafrost, defined as earth material in which the temperature remains less than 0°C continuously for two or more years (Muller, 1947; Brewer, 1958) has been referred to in the foregoing discussion. Most of Alaska, including the drainage basin of Glenn Creek, lies in the zone of laterally discontinuous permafrost (Péwé, 1966, Fig. 1), where quite local climatic conditions and recent geological history determine the presence or absence of perennially frozen ground.

Because unbound water is commonly present in earth materials, the absence or presence of permafrost conditions is usually determined by noting the presence or absence of ice in the soil at a time when seasonal thawing has progressed to its maximum extent. In the present study, this was done by systematic probing with a 100-cm-long steel rod in late August, and from the auger holes, as described earlier. Vegetation types are also a generally reliable guide in determining the presence or absence of permafrost (see, for example, Péwé 1966, p. 13), and were used as a basis for inference where the probe encountered impenetrable gravel layers at shallow depths.

According to Pewe's (1958) map (see Fig. 7a), permafrost is present beneath the undifferentiated silts and absent where Fairbanks loess is mapped. Rieger *et al.* (1963) indicated that both the Saulich and Goldstream soils are underlain by permafrost at shallow depths (1 to 3 ft), while the Fairbanks silt loam is permafrost free. Figure 11, taken from Rieger *et al.* (1963, Fig. 4), illustrates the general relations of permafrost present in the valley bottoms and on the north-facing slopes in the uplands. Similar diagrams are presented in Péwé (1955b, Fig. 9) and Cederstrom (1963, Fig. 6).

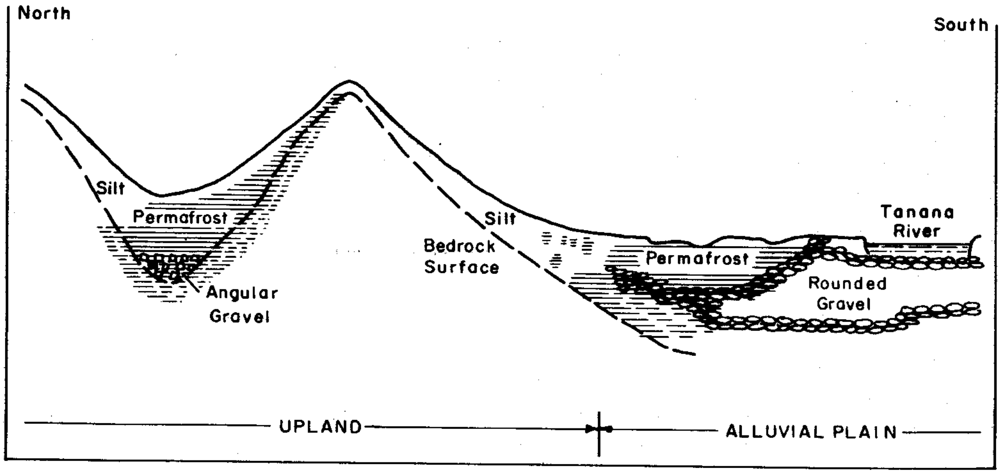


Figure 11. General permafrost relations in the Fairbanks area (after Rieger et al., 1963).

Field observations indicate that the watershed of Glenn Creek conforms in all essential aspects to this generalization (see Fig. 12). These observations are summarized in Appendix A. Permafrost was encountered at almost every grid point south of the grid base line, including points on the base line. The only exceptions were three points near the watershed boundary in the lower portion (points CS2, DS2, and ES2) and one in the upper portion (KS3), where no refusal was encountered at less than 100-cm depth. Impenetrable gravel layers were reached at a few other points in this area (DS1, HS3, HS5, J, JS2, and L). On the north side of the grid base line, gravel was encountered at most points east of and including line FN, with the exceptions of points HN1, and FN4, where permafrost was present. The latter point and point EN4 (which lies outside the basin), where permafrost was also found, lie very close to the northern divide, where permafrost is typically considered absent (see Fig. 11). Of the remaining points lying west of line FN, permafrost was present at points AN1 and BN1, but no refusal at less than 100 cm was encountered elsewhere.

These data, and supplementary probing to locate more closely the permafrost boundaries, along with inferences based primarily on vegetation, permit plotting of the contact between the permafrost and nonpermafrost areas of the watershed (Fig. 12). The vegetational inferences were most heavily relied upon in the northeastern portion, where shallow gravel layers were virtually ubiquitous. Use of vegetation types in locating the permafrost boundary is justified by Péwé's (1966, p. 13) observation:

> "A well-defined boundary between two vegetation types generally lies near the boundary between slopes underlain by permafrost and permafrost-free slopes. The boundary between black spruce scrub forest on permafrost areas and white spruce-birch-aspen forest of permafrost-free slopes is distinct and readily recognized. Generally, permafrost, with or without ice masses, extends a short distance upslope past the line marking the border between these two types of vegetation." *

Rieger *et al.* (1963, p. 4-5) described the same correspondence between vegetation and permafrost. The absence of ice in the auger holes and the presence of soils and vegetation typical of well-drained slopes indicate that most of the south-facing slopes are permafrost-free even at depth. Within the perennially frozen area, the data do not permit drawing isopleths of equal active layer (seasonal thaw zone) depth; depth to permafrost table may vary by 0.3 ft within a horizontal distance of 1 ft. At the time of measurement, active layer depths ranged from 0.82 to 3.0 ft, and averaged 1.6 ft.

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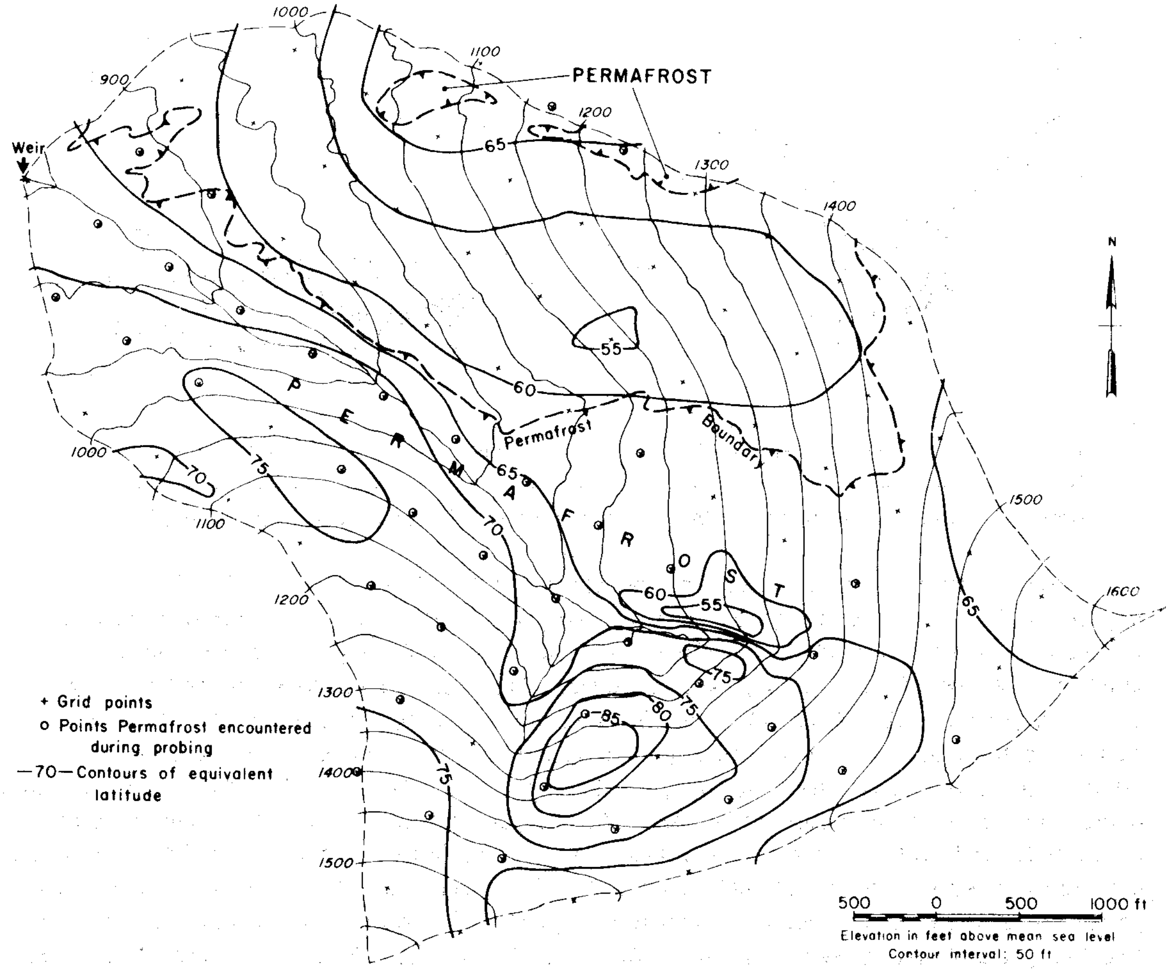

Figure 12. Permafrost boundaries and contours of equivalent latitude, Glenn Creek watershed.

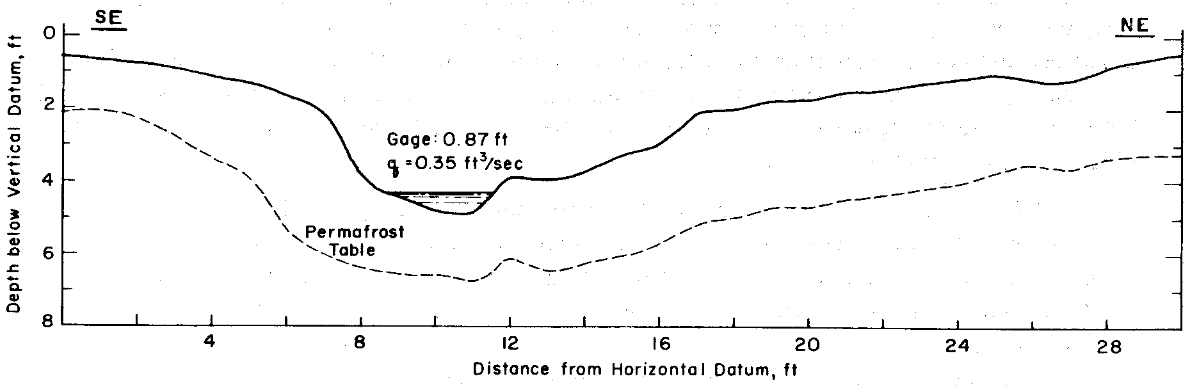

Figure 13. Cross section of Glenn Creek at the weir site, showing depth of permafrost, 24 August 1964. q = discharge.

Figure 13 is a surveyed cross section of Glenn Creek at the weir site, showing depth to permafrost (established by probing) at the time of maximum thawing in 1964. The permafrost table replicates very closely the ground surface, and appears to be little affected by the stream itself. All evidence indicates that Glenn Creek flows a few feet above permafrost throughout its length.

In the zone of discontinuous permafrost, the presence or absence of permafrost in an area is determined by the heat balance of the ground, and therefore by the geothermal heat flow, the activity of circulating groundwater, and the heat balance at the ground surface. Thus, in a restricted area of generally similar surficial materials, the surface heat balance is dominant in determining the presence or absence of permafrost. The general presence of permafrost on north-facing slopes and its absence on south-facing slopes, as found in the basin of Glenn Creek, suggest that the amount of solar irradiation on a surface is a major factor in determining ground temperatures. In central Alaska, it is true that a ground cover of thick mosses is commonly present above permafrost areas; and it could be argued that the insulating properties of this material led to permafrost formation. However, mosses require restricted drainage to flourish (Drury, 1956, p. 36), so that it is most logical to assume that the restricted drainage (permafrost) came first, with the presence of mosses probably causing the permafrost table to rise subsequently.

With the data from Glenn Creek watershed, it is possible to determine the relation between an index of the amount of solar radiation received on a surface and the presence or absence of permafrost. Lee (1962, 1964) cites Kimball's (1919) statement that "In the case of a slope facing *h* degrees in azimuth, the angle of incidence of the solar rays will be the same as on a horizontal surface at a point on a great circle passing through the slope at right angles to it and as many degrees removed as the angle of the slope." Thus, an "equivalent latitude" and "equivalent longitude" can be calculated for any slope, knowing the actual latitude and longitude and the slope inclination and azimuth. The equations presented by Lee (1964) show that the equivalent latitude is the major determinant of the amount of solar radiation received on a slope (neglecting atmospheric influences, which can safely be assumed constant over a restricted area, and topographic shading). This equivalent latitude θ' is calculated as

$$\theta' = \sin^{-1} (\sin k \cos h \cos \theta + \cos k \sin \theta)$$

where k is slope inclination, h is slope azimuth, measured from north = 0°, and θ is actual latitude. Slopes and inclinations were measured from the 1:2400, 5-ft contour-interval map for each grid point in Glenn Creek watershed. The value of θ' was then calculated for each point and plotted on the map. Isopleths of equal θ' were then sketched, guided by the contours and a specially-constructed nomograph relating θ' to k and h; the results are shown on Figure 12.

It is clear from the map that a close relationship exists between the distribution of permafrost, plotted from field evidence, and equivalent latitude. With one probable exception, discussed below, all areas where $\theta' < 60°$ are permafrost-free, and all areas where $\theta' > 65°$ have permafrost. The southern and eastern boundaries of the permafrost-free area correspond closely to the 60° isopleth, while the presence of permafrost at the northern watershed boundary is well reflected by the pattern of the 65° isopleth in that area. The probable exception to this generalization is found near the head of Glenn Creek, where southwesterly-facing steep slopes are found immediately north of the stream. Values of θ' in this area are as low as 54°, but the vegetation suggests that permafrost is present. The only point probed in that area (point J) revealed an impenetrable gravel layer at 1.8 ft-depth (see App. A). At nearby points I and K, permafrost was present at 1.1 ($\theta' = 64.2°$) and 1.4 ft ($\theta' = 65.5°$), respectively, suggesting that permafrost, if present near point J, is at greater depth. It is likely that topographic shading of this region has led to permafrost formation, even though the equivalent latitude suggests that it is absent.

A relationship between depth to permafrost and equivalent latitude is also revealed by the permafrost-probe data. At each grid point, probing was done at four locations, 3 ft north, south, east, and west of the point; the average of these values was taken as the depth to permafrost at that point. (At six of the 38 grid points where permafrost was found, measurement could be made

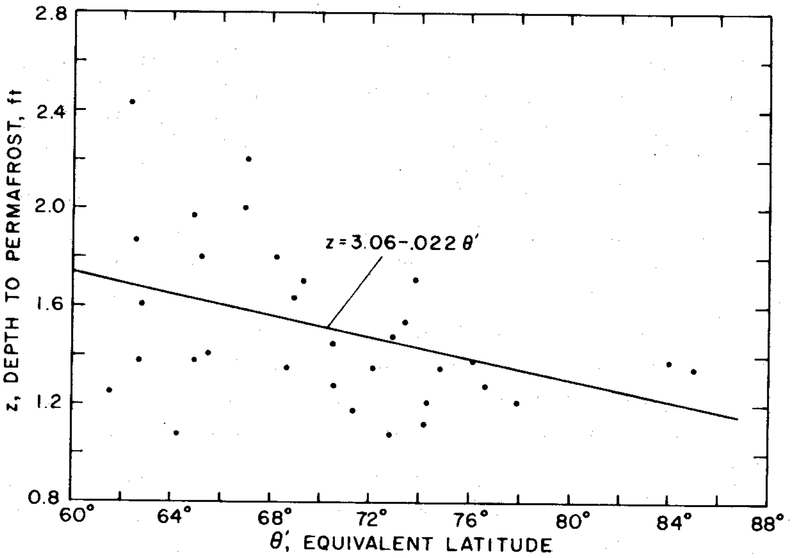

Figure 14. Relationships between depth to permafrost and equivalent latitude, Glenn Creek watershed.

at only one location because of impenetrable gravel layers at the other three locations; these six grid points were eliminated from the regression analysis below. Only two measurements were available at four grid points, and three measurements were available at three grid points; these were included in the total of 32 points used in the regression analysis.) The scatter diagram of depth to permafrost z vs θ' is shown in Figure 14. The correlation coefficient for a linear relation between z and θ' is -0.406, significant at the 0.025 level, and the regression equation is

$$z = 3.06 - 0.022\,\theta'$$

where z is in feet and θ' is in degrees.

The available data do not allow a determination of the effects of insulation and shading by vegetation on depth to permafrost. Presumably these factors, along with hydraulic and thermal conductivity of the soil, average snow depths and densities, and average wind speeds, account for variations in the relation between z and θ'. Still, the relationship between equivalent latitude and the presence of permafrost and thickness of active layer seems well established for Glenn Creek Basin, and indicates that this approach may be a useful quantitative tool for predicting permafrost relations elsewhere.

Péwé (1955b, p. 127) stated that "In the upland north of the Tanana Valley, silt and bedrock are frozen to unknown depths beneath north slopes" and that "The flat, swampy valley floors are underlain by frozen ground to depths of 50 to 200 ft." These observations, plus the fact that the complete silt and gravel section at the CRREL Permafrost Tunnel is frozen (Sellmann, 1967), make it highly probable that, within the permafrost area of Glenn Creek Basin, the ground is perennially frozen at least down to the bedrock surface.

Ground temperature measurements and the fact that permafrost is observed to form in recently deposited alluvial sediments indicate that permafrost in central Alaska is in equilibrium with the present climate (Pewe, 1966, p. 32). However, Hamilton (1965) concluded from a study of Alaskan temperature records that "There has been a probable net gain in the order of 1 – 1½°F from mean annual temperatures of the late 1800's to those of the present. Both the net change and its major components show a general agreement with world and regional trends determined by previous investigators."* A more nearly complete analysis by Haugen *et al.* (in press) indicates very little change in mean annual temperatures or mean July temperatures in the region since the early 1900's, but a pronounced increase (about 10°F) in mean January temperatures since that time. Such a trend would have the result of slightly decreasing the depth of seasonal freezing, but since that depth is severalfold greater than the depth of seasonal thaw over permafrost in the basin, there would be little effect on the general permafrost conditions.

The previous discussion of the relationship between permafrost and vegetation indicates that the distribution of permafrost in Glenn Creek Basin is typical of the general area. No detailed maps of permafrost distribution are available for the Yukon-Tanana uplands, but according to Ferrians' (1965) permafrost map of Alaska, Glenn Creek lies near the southern border of the area designated as "generally underlain by discontinuous permafrost." This designation includes the entire Yukon-Tanana uplands, as well as other large portions of interior Alaska.

Climate

General. The drainage basin of Glenn Creek lies in the interior climatic division of Alaska, a zone of marked continentality, with a large annual temperature range and low mean annual precipitation (Watson, 1959).

This continentality is due both to distance from the ocean (Glenn Creek lies 275 miles from the nearest portion of the ocean) and to the presence of topographic barriers in virtually all directions, which generally prevent maritime air masses from reaching the area. [The severe rains and flooding in the Fairbanks area in mid-August 1967 were due to unusual weather patterns that favored a flow of moist maritime air from the southwest to interior Alaska, along the only topographic *corridor* leading from the sea to the interior. This corridor lies between the Kuskokwim Mountains and the Alaska Range (Streeter, 1967).]

Because of the isolation of this area, its weather is largely determined by the areal radiation balance. In the winter, long periods of darkness lead to intense radiational cooling and the formation of a high pressure area over much of the interior, with calm winds and low temperatures. Cold air drainage and inversions are common at this time; Bilello (1965) found that surface-based inversions are present during more than 40% of the wintertime at Fairbanks. These have an average height of 1800 ft above sea level and an average gradient of +1.2 F°/100 ft.

In the summer, the long hours of daylight and general absence of moisture brought in by large circulatory air masses lead to solar radiational heating of the surface and the formation of thermal convective cells. Maximum daily temperatures generally rise to 75°F and above, and precipitation is showery. A typical summer day has clear skies in the morning, with increasing formation of cumulus clouds as the day progresses; moisture is supplied to the air by evapotranspiration. These clouds generally produce numerous light, local showers in the afternoon and evening. As Watson (1959, p. 6) pointed out, "To a considerable extent, the moisture in the interior basin goes through repeated cycles of evaporation, condensation, and precipitation in showers." Infrequently, the convective cells are intense enough to produce thunderstorms; Rieger *et al.* (1963, p.4) gave an average of eight thunderstorms per year at Fairbanks.

Precipitation and temperature. Precipitation records collected for the present study show measurable dewfall (generally 0.01 to 0.02 in.) on virtually every clear night, commencing about 1800-2000 hours and evaporating from the rain gage by 0600 hours the next morning. Over the summer, the total precipitation received from this process may amount to 0.50 in., an amount which is not recorded by standard observations nor included in the normal precipitation totals given for the area.

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The wide variation in temperatures from summer to winter results in rapid changes in mean daily temperatures in the spring and fall, and thus tends to fix the freezing and thawing seasons within fairly constant limits (Watson, 1959). (See Table I.)

Table I gives the normal monthly precipitation and temperatures for the University Experiment Station (UES), University of Alaska, which is 10 miles southwest of the basin at an altitude of 475 ft. Of the total precipitation, about 3.70 in., or 30%, falls as snow.

Table I. Normal monthly precipitation and temperature at University Experiment Station*

Month	*J*	*F*	*M*	*A*	*M*	*J*	*J*
Temperature (°F)	-7.3	0.9	12.6	30.7	46.9	57.9	59.7
Precipitation (in.)	0.83	0.51	0.42	0.24	0.80	1.48	2.10

Month	*A*	*S*	*O*	*N*	*D*	*Annual*
Temperature (°F)	54.7	44.5	27.5	6.2	-5.9	27.4
Precipitation (in.)	2.44	1.36	0.93	0.63	0.57	12.31

* Data from U.S. Weather Bureau (1965).

Table II gives the results of an analysis of thaw-season weather at UES for the years 1938-1967. The thaw season was defined as beginning on the first date when a sequence of days with mean temperature above 32°F was not followed by a sequence of the same or greater length with mean temperature less than 32°F. Its end was similarly defined as the date immediately preceding the beginning of the first sequence of days with mean temperature less than 32°F which was not followed by an equal or longer sequence of days with mean temperatures greater than 32°F.*

Over the 30 years ending in 1967, there had been considerable variation in the dates of the commencement of the thaw season, from 20 March (1965) to 7 May (1949) (49 days); but less in the dates of the ending, from 1 October (1947, 1956) to 27 October (1938) (27 days). As noted above, thaw season length tends to be fairly constant (average = 176 days), with a standard deviation of about 2 weeks. Total thaw-season precipitation is highly variable, as is the total precipitation divided by the thaw season length. However, mean daily thaw season temperature is remarkably constant from year to year.

Figure 15 is a plot of mean thaw-season temperature and total thaw-season precipitation for the 30 years analyzed. Points are rather well distributed in the four quadrants defined by the mean values, indicating only a slight tendency for wet years to be warmer and dry years cooler, as found by Brown (1967) at Barrow, Alaska. It is clear from Figure 15 and Table II that three of the four seasons of this study were cooler than normal (1964, 1965, 1967), and three were drier than normal (1964, 1965, 1966). The period of study included the driest (1966) and second-wettest (1967) as well as the coolest (1965) of the last 30 years.

Table III compares the precipitation totals at Glenn Creek and University Experiment Station for the thaw-season periods of record at Glenn Creek. It can be seen that the showery nature of summer precipitation results in rather poor correspondence between the two stations.

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* This definition is not the same as the more commonly used "freeze-free" or growing season, which is based on minimum temperatures, rather than averages. According to Rieger *et al.* (1963, p. 2-3), the growing season at UES is 88 days, with average beginning and ending dates of 29 May and 24 August, respectively.

Table II. Summary of thaw-season length, precipitation, and temperature, 1938-1967, University Experiment Station

Year	*Thaw season start*	*Thaw season end*	*Length of thaw season (days)*	*Thaw season pptn. (in.)*	*Mean thaw season pptn. intensity (in./day)*	*Mean thaw season temp (° F)*
1967	17 Apr	15 Oct	182	13.32	0.073	50.0
66	27 Apr	9 Oct	135	4.22	0.031	51.0
65	20 Mar	3 Oct	198	7.32	0.037	48.7
64	21 Apr	14 Oct	177	7.87	0.044	50.5
63	21 Apr	9 Oct	172	10.99	0.064	51.4
62	16 Apr	16 Oct	184	15.29	0.083	49.8
61	22 Apr	3 Oct	165	11.01	0.067	51.8
60	23 Apr	5 Oct	156	10.90	0.066	52.5
59	23 Apr	6 Oct	167	11.40	0.068	50.7
58	1 Apr	3 Oct	186	7.21	0.039	51.5
57	9 Apr	23 Oct	198	4.28	0.022	50.7
56	16 Apr	1 Oct	169	9.91	0.059	51.6
55	27 Apr	6 Oct	163	12.76	0.078	51.1
54	23 Apr	15 Oct	176	8.64	0.049	51.6
53	9 Apr	16 Oct	191	10.22	0.054	52.5
52	19 Apr	16 Oct	181	7.36	0.041	49.2
51	14 Apr	3 Oct	173	7.60	0.044	52.5
50	13 Apr	7 Oct	178	4.74	0.027	51.5
49	7 May	5 Oct	152	10.27	0.068	51.6
48	5 May	12 Oct	161	11.90	0.074	49.7
47	21 Apr	1 Oct	163	6.14	0.038	50.2
46	24 Apr	20 Oct	180	7.78	0.043	50.9
45	6 May	6 Oct	154	11.84	0.077	52.0
44	20 Apr	20 Oct	184	8.38	0.046	51.0
43	9 Apr	18 Oct	193	6.98	0.036	51.5
42	8 Apr	5 Oct	181	11.86	0.066	53.1
41	5 Apr	8 Oct	187	8.26	0.044	50.6
40	1 Apr	8 Oct	191	6.78	0.035	50.9
39	8 Apr	5 Oct	181	8.40	0.046	48.8
38	27 Apr	27 Oct	184	7.99	0.043	50.7
Avg	17 Apr	10 Oct	175.7	9.05	0.052	51.0
Std dev			14.72	2.73	0.017	0.90
Coef var			0.084	0.30	0.33	0.018
Max			198	15.29	0.083	53.1
Min			135	4.22	0.022	48.7

Daily correspondence between average temperatures and precipitation at the two stations was examined by randomly selecting 100 dates from within the average thaw season (17 April - 10 October) and randomly assigning a year (1964 to 1967) to each date. Duplication of dates so chosen and lack of record at Glenn Creek reduced the number of days available for comparison to 57. Since the observations at the Weather Bureau station are made at 1700 hours, this time was used in calculating daily precipitation P and temperature T from the recording rain gage and hygrothermograph records at Glenn Creek. The relationship between values is summarized in the following regression equations:

$$P_{GC} = 0.005 + 0.97\,P_{UES},\ r = 0.830$$

mean Glenn Creek = 0.0347; mean UES = 0.0304

$$T_{GC} = -9.3 + 1.11\,T_{UES},\ r = 0.940$$

mean Glenn Creek = 49.89, mean UES = 53.53

where r = correlation coefficient.

Table III. Comparison of precipitation totals at Glenn Creek and University Experiment Station.

Dates	Precipitation (in.) Glenn Creek	Precipitation (in.) UES
3 May - 3 Oct 1967	11.05	12.77
11 May - 9 Oct 1966	4.75	4.08
5 Aug - 29 Sept 1965	3.10	3.14
1 June - 14 Oct 1964	9.94	6.36

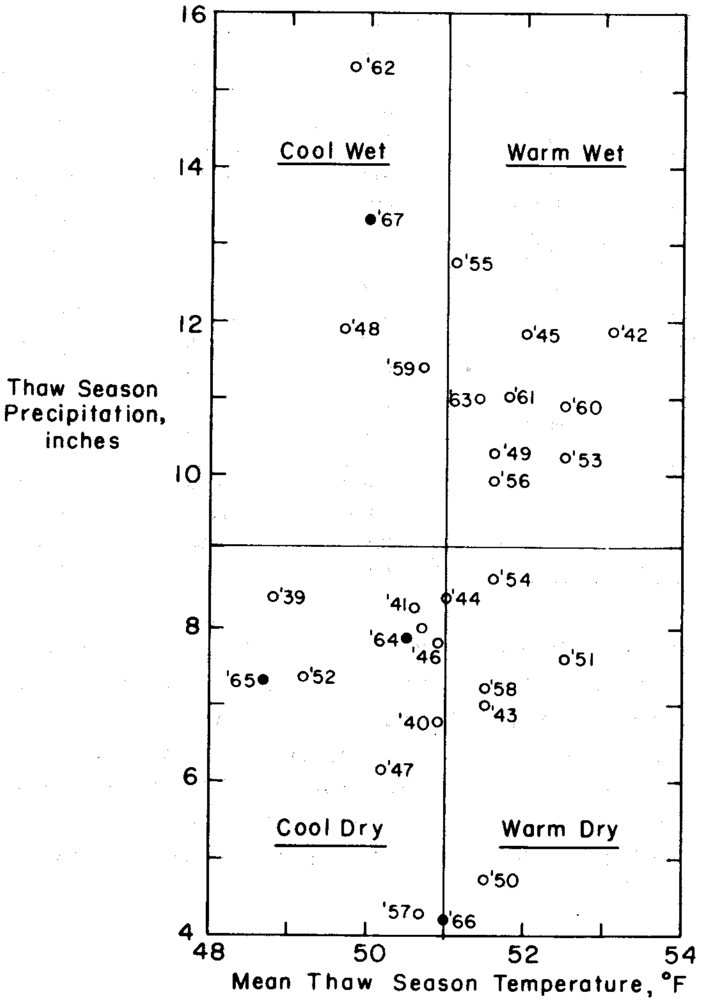

Figure 15. Mean temperature and total precipitation, thaw seasons 1938-1967, University Experiment Station.

Both correlations are significant at the 0.01 level. Although precipitation for individual days differs by as much as 0.18 in., the constant in the regression equation is not significantly different from zero, nor is the regression coefficient significantly different from unity ($\alpha = 0.05$). Further, there is no difference between the means at the two stations at the same significance level.

In the temperature relationship, the regression constant and coefficient are significantly different from zero and unity, respectively ($\alpha = 0.05$), and the means differ significantly at the same level. The difference in means, 3.64°F, is greater than can be accounted for by the difference in elevation between the two stations and a standard lapse rate. The comparison of daily temperatures and precipitation indicates that general trends in thaw-season values at UES, as shown in Figure 15, are applicable to Glenn Creek.

Rainfall intensity characteristics in the region are perhaps best summarized by examination of its rainfall-frequency characteristics. The data shown in Figure 16 are extracted from maps developed by Miller (1963), and show the amount of rainfall to be expected for durations of 30 minutes to 24 hours and recurrence intervals of 1 to 100 years. The values in Figure 16 may be compared with those on generalized rainfall frequency-duration maps of the conterminous United States given by Gilman (1964, p. 9-46 to 9-57). For the shorter duration storms (30 minutes to 1 hour), the amounts expected at Glenn Creek are less than for all but a few small areas of the "lower-48" at a given recurrence interval. At longer durations, the amounts expected at Glenn Creek are comparable with those for the arid region including Nevada, western Utah, eastern Oregon, and southern Idaho, and lower than for the rest of the United States.

Snow. About 30% of the annual precipitation in the area occurs as snow. Weather Bureau records for UES have reported snow-on-ground since 1951; these data are summarized in Table IV. The average first date of continuous snow cover is 8 October; the average last date is 9 May, for an average total of 214 days with snow cover. Corresponding to the dates of thaw-season beginning and ending, there is less year-to-year variability in the initial date of continuous snow cover than in the final date.

Table IV. Summary of snow-on-ground, University Experiment Station.

			Snow-on-ground (in.)					
Year	*Snow beginning*	*Snow ending*	*1 Nov*	*1 Dec*	*1 Jan*	*1 Feb*	*1 Mar*	*1 Apr*
1966-67	10 Oct 66	3 May 67	3	17	17	21	20	31
65-66	6 Oct 65	27 Apr 66	8	15	29	28	35	25
64-65	14 Oct 64	25 Apr 65	4	12	21	20	24	2
63-64	9 Oct 63	27 May 64	5	7	11	13	17	15
62-63	11 Oct 62	12 May 63	2	3	7	14	15	30
61-62	5 Oct 61	21 May 62	8	13	17	21	26	21
60-61	5 Oct 60	7 May 61	4	10	11	13	14	14
59-60	12 Oct 59	6 May 60	3	9	19	21	21	19
58-59	30 Sept 58	15 May 59	5	12	16	16	23	22
57-58	26 Sept 57	4 May 58	2	5	8	14	12	12
56-57	2 Oct 56	25 May 57	8	12	17	37	26	23
55-56	6 Oct 55	18 May 56	7	10	26	28	32	28
54-55	17 Oct 54	22 May 55	0	9	14	13	23	21
53-54	22 Oct 53	30 Apr 54	2	2	7	14	13	11
52-53	19 Oct 52	20 Apr 53	5	3	4	6	8	9
51-52	1 Oct 51	30 Apr 52	1	10	25	28	27	25
Avg	8 Oct	9 May	4	9	16	19	21	19
Avg total	214 days							

Table V. Summary of snow-course observations, Glenn Creek Basin.

Date	*Depth (in.)*	*Glenn Creek Water content (in.)*	*Density (%)*	*UES Depth (in.)*
1 Dec 65	13.5	2.2	16.3	15.0
11 Jan 66	21.5	3.7	17.2	25.0
2 Feb 66	20.5	3.8	18.5	27.0
1 Mar 66	28.0	5.1	18.2	35.0
1 Apr 66	26.0	4.9	18.8	25.0
5 May 66	12.0	2.8	23.3	0
15 Nov 66	7.0	1.1	15.7	7.0
15 Dec 66	15.0	2.4	16.0	16.0
3 Feb 67	19.0	3.2	16.8	20.0
13 Mar 67	26.5	4.4	16.6	25.0
4 Apr 67	29.0	3.6	12.4	28.0
13 Dec 67	10.0	1.7	17.0	10.0
8 Jan 68	15.5	2.9	18.7	17.0
15 Feb 68	21.0	4.0	19.0	24.0
4 Mar 68	19.5	4.0	20.5	22.0
2 Apr 68	20.0	4.5	22.5	18.0
6 May 68	7.0	2.0	28.6	

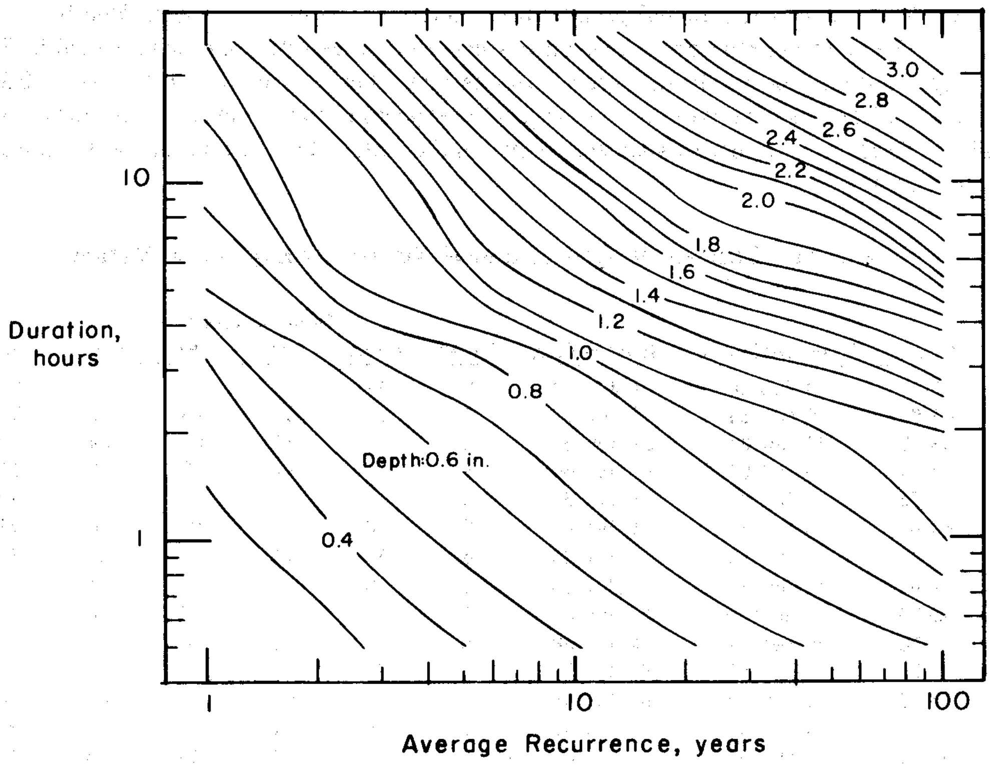

Figure 16. Relations among rainfall depth, duration, and frequency, Glenn Creek watershed (after Miller, 1963).

Table V summarizes the results of 17 snow-course measurements made in a clearing in Glenn Creek Basin. Depths at this site agree well with those reported at UES. Snow densities range from about 15% to 20% in the early winter and tend to increase to about 20% to 28% in April and May.

Water balance and evapotranspiration. Few data are available on potential and actual evapotranspiration in Alaska. Patric and Black (1968) used the Thornthwaite method (Thornthwaite and Mather, 1957) to estimate these parameters for all meteorological stations in the state. At University Experiment Station, their estimate of the mean annual Thornthwaite potential evapotranspiration (PET) was 18.31 in., and the mean annual Thornthwaite actual evapotranspiration (AET) was 11.93 in. The former value was calculated from an empirical formula based solely on mean monthly temperatures. This formula provides reasonable estimates for many parts of the world; in particular, Sanderson (1950) found a close correspondence between water loss from a well-watered grass lysimeter and Thornthwaite PET at Norman Wells, NWT, which is at the same latitude as Glenn Creek. This estimate of PET also corresponds well to average pan evaporation at UES, values of which are summarized in Table VI.

Thornthwaite (1944, p. 687) defined PET as "the water loss which will occur if at no time there is a deficiency of water in the soil for the use of vegetation." Kohler (1957, p. 3) noted that, since surface characteristics (vegetation type) affect water use, PET might better be defined as "the evaporation from a free water surface of extended proportions, but independent of any heat-storage effects." If an evaporation pan is considered to present such a surface, and if Sanderson's (1950) experiment is representative, it would appear that Thornthwaite's method of estimating PET can be applied to subarctic regions, such as the Glenn Creek watershed, no matter which definition of PET is used.

Table VI. Monthly class-A pan evaporation at Glenn Creek (1965-67) and University Experiment Station (1957-67).

				Pan evaporation (in.)					
	Year	*Apr*	*May*	*June*	*July*	*Aug*	*Sept*	*Oct*	*Annual*
Glenn	1967		3.62	5.79	6.90		1.08	0	
Creek	1966			5.46	5.13	3.23	1.71	0.06	
	1965					1.76	1.08	0	
UES	1967	0	3.73	6.13	4.09	2.77	1.54	0.02	18.28
	1966	0	3.11	5.55	5.99	3.95	2.04	0.23	20.87
	1965	0.37	4.12	4.02	4.37	2.45	1.48	0	16.81
	1964	0	2.98	5.59	3.90	2.88	1.29	0.01	16.65
	1963	0	5.14	3.79	4.18	2.03	2.21	0.09	17.44
	1962	0.19	3.62	5.50	5.02	3.21	1.07	0.20	18.81
	1961	0.39	4.65	5.05	4.27	2.64	1.25	0.03	18.28
	1960	0	5.91	5.12	5.16	2.67	1.14	0.39	20.39
	1959	0	4.25	6.10	3.49	3.13	1.18	0.05	18.20
	1958	0	4.29	6.31	5.66	3.27	1.33	0	20.86
	1957	0	4.43	6.60	5.23	3.91	2.14	0	22.31
	Avg*	0.09	4.20	5.43	4.67	2.99	1.52	0.09	18.99

* Averages of UES only.

On the other hand, the estimate of Thornthwaite AET presented by Patric and Black (1968) for UES (11.93 in.) is certainly too high. As noted in Table I, the normal annual precipitation at UES is 12.31 in.; this would leave only 0.38 in. average annual runoff. Table VII presents rainfall and runoff data for the two years of this study (1964, 1966) when stream-gaging problems (leakage around and under the weir) did not prevent collection of continuous runoff records at least for the months June through August. Both summers were drier than normal; 1966 was the driest thaw season in 30 years (see Fig. 15). Yet in both years, thaw-season runoff was considerably greater than 0.38 in.; presumably the normal thaw-season runoff for Glenn Creek is over 2 in. In addition to this, the data of Table V indicate that about 4 in. of water in the form of snow is present on the watershed in early April. Since virtually all of this must run off during the snowmelt period, a reasonable estimate for the mean annual runoff of Glenn Creek is 6 in., or about 50% of precipitation. This value, which is the only figure available from central Alaska for an area where both precipitation and streamflow have been accurately determined, must be accepted as the current best estimate of the water balance at low to moderate elevations in this region.

Table VII. Monthly rainfall and runoff, Glenn Creek watershed, 1964 and 1966

Month and year	*Rainfall (in.)*	*Runoff (in.)*	*Rainfall minus runoff (in.)*
Jun 64	2.45	0.41	2.04
Jul 64	2.90	0.40	2.50
Aug 64	3.75	0.89	2.86
Sep 64	0.68	0.59	0.09
Oct 64	0.26	0.10	0.16
total	10.04	2.39	7.65
Jun 66	2.13	0.58	1.55
Jul 66	1.24	0.03	1.21
Aug 66	1.16	0	1.16
total	4.53	0.61	3.92

Apparently, the low runoff figure obtained by Patric and Black (1968) is at least in part due to an overestimate of soil-water storage capacity, on which the estimate of Thornthwaite AET depends. Although these writers did not state the values assumed for this parameter in their calculations, a check computation for UES indicates that a value of 4 in. was used.

Vegetation

Introduction. In the present study, there are several reasons for considering in some detail the types and distribution of vegetation within the watershed. These reasons may be discussed in terms of *passive* and *active* relations of the vegetation to the hydrologic and morphologic characteristics of the basin.

Passively, the types of plants present in subareas within an area of essentially uniform geology and mesoclimate may reflect variations in: 1) soil drainage; 2) microclimate (principally insolation); and 3) history (especially geomorphologic, fire, and human interference). The first two factors plus geomorphologic history are directly relevant to the present study, but knowledge of the influences of fire and man is required to separate the man-caused disturbances from the hydrological and morphological information in the vegetation types and patterns.

Considering the active relations, the vegetation 1) forms the interface between the watershed and the atmosphere, affecting interception, evapotranspiration, and infiltration; 2) is a component of the physical structure of the immediate subsurface of the watershed, influencing water movement and soil strength; and 3) may influence the types and rates of chemical processes occurring in the watershed.

It is true that the above factors, active and passive, are complexly inter-related, but their consideration in this way provides a rationale for examining and mapping the vegetation of the drainage basin of Glenn Creek.

Mapping. Although the objectiveness of identifying vegetation communities or ecotones has been questioned by those who hold that plants are present as overlapping species ranges rather than as discrete communities (see Daubenmire, 1966), it is apparent that the plant community concept is useful and meaningful in interior Alaska. Several studies have discussed the relationships of such discrete communities to edaphic, climatic, and disturbance conditions in the region (Stoeckeler, 1949; Frost, 1950; Lutz, 1956; Holmes and Benninghoff, 1957; Sigafoos, 1958; Johnson and Vogel, 1966). Johnson and Vogel (1966, p. 40, Fig. 31) presented quantitative data demonstrating a strong tendency for forest types to occur as essentially pure stands in the Yukon flats region (about 100 miles northeast of Fairbanks), supporting more qualitative observations of earlier workers. Daubenmire (1966) argued that the ecotone concept is in general objective as well as useful.

The vegetation map (Fig. 17) of the watershed was made from aerial photographs at a scale of 1:4800, supported by systematic ground observations. The latter were accomplished by occupying 85 points on the 500-ft grid covering the entire basin. At each point, the species or types of the readily visible plants growing within an approximately 25-ft radius were noted, along with qualitative observations as to tree heights, spacing, abundance, and similar factors. Anderson's (1959) compendium of Alaskan flora was used as a guide for identification.

As noted above, previous studies of plant communities in interior Alaska have recognized several characteristic associations occurring in this general region. Further, there is general agreement among these studies as to the composition of these characteristic associations. With these as a guide, the vegetation of Glenn Creek drainage basin could be assigned to one of six major map units, based primarily on dominant tree species and ground-cover type:

1. Black spruce (***Picea mariana***) forest with continuous thick (6 in. to 1 ft) moss including ***Sphagnum*** spp. and ***Pleurozium schreberi***, and lichen ground cover. Mostly pure stands, sparse to dense, but locally with subordinate birch (***Betula papyrifera***), alder (***Alnus crispa***), and more rarely tall willow (***Salix*** sp.). The most common shrubs in this association are labrador tea (***Ledum*** sp.), blueberry (***Vaccinium uliginosum***), mountain cranberry (***Vaccinium vitus-idaea***), with bog rosemary (***Andromeda polifolia***), dwarf birch (***Betula nana***), and squashberry (***Viburnum edule***) rarer. Many herbaceous plants grow within the moss mat, including horsetail (***Equisetum*** spp.), wild rhubarb (***Polygonum alaskanum***), salmonberry (***Rubus chamaemorus***), grasses, and sedges (including ***Carex*** spp. and ***Eriophorum*** spp.).

1a. Black spruce - white birch forest (in approximately equal proportions), with thick moss ground cover. Generally dense stands with undergrowth generally similar to that of black spruce forest.

1b. Unforested with thick moss mat. Ground vegetation similar to that of black spruce forest.

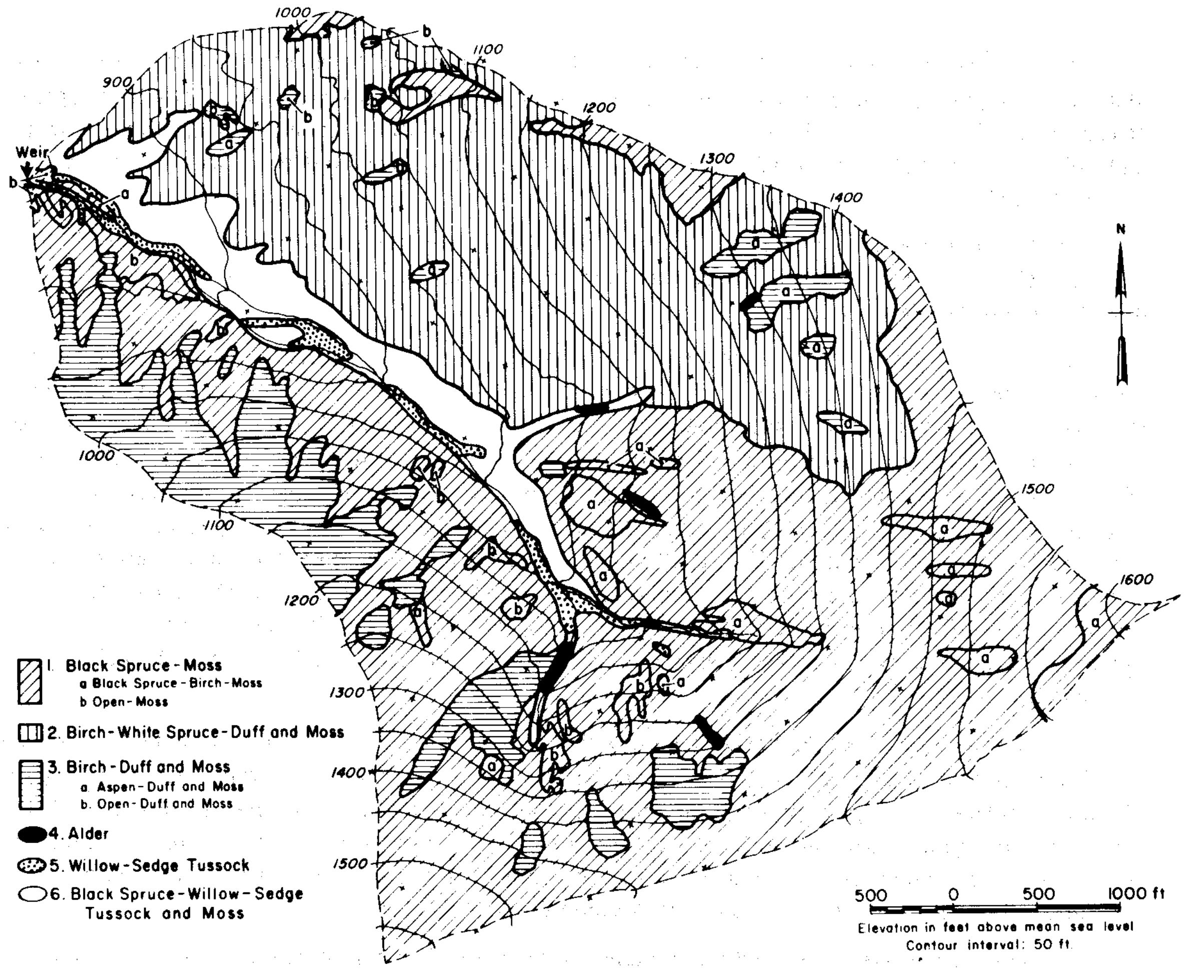

Figure 17. Vegetation map of Glenn Creek watershed.

2. Birch - white spruce (*Picea glauca*) forest with ground cover of litter and thin moss patches. Moderately dense to dense stands, trees generally taller than 30 ft. Birch predominates, with white spruce and, much less commonly, black spruce abundantly scattered. Locally, tall willow (including diamond willow) or alder dominate. The most common shrubs in this association are rose (*Rosa acicularis*), labrador tea, spiraea, and mountain cranberry. Grasses, fireweed (*Epilobium angustifolium*), horsetails, and bunchberry (*Cornus canadensis*) are common herbaceous plants.

3. Birch forest with ground cover of litter and thin moss patches. Moderately dense to dense stands, trees generally taller than 30 ft. Locally, other deciduous species; willow (including diamond willow), quaking aspen (*Populus tremuloides*), or alder are present in nearly pure stands. White spruce, black spruce, and balsam poplar (*Populus balsamifera*) are less common. Rose is the most common shrub, with labrador tea, squashberry, and raspberry (*Rubus strigosus*) also occurring. Fireweed, horsetails, bunchberry, and grasses are the dominant herbs.

3a. Aspen forest with ground cover of litter and minor moss patches. Pure, dense, even-aged stands, trees taller than 30 ft. Shrubs and herbs sparse, with grasses, fireweed, and bunchberry dominant.

3b. Unforested with ground cover of litter and thin moss patches or, locally, bare ground. Shrubs and herbs generally similar to those of birch forest.

4. Alder thicket with no ground cover. Shrubs lacking, with scattered grasses and horsetails.

5. Unforested with sedge tussocks and willow bushes, with no continuous ground cover. Scattered black spruce and birch are present in this unit; blueberry, labrador tea, dwarf birch, and mountain cranberry occur as shrubs. Herbaceous plants include cloudberry, grasses, and coltsfoot (*Petasites* sp.).

6. Black spruce forest with willow bushes; no continuous ground cover, moss patches and sedge tussocks. Generally sparse stands, trees less than 30 ft tall. Locally birch, alder, and scattered larch (*Larix laricina*) also present. Willows locally very thick; other shrubs and herbs as in unit 5.

The areal extent of each of these units in the watershed of Glenn Creek is given in Table VIII.

Table VIII. Areal distribution of vegetation units in Glenn Creek watershed.

	Unit	*Area (mi^2)*	*Area (%)*
1.	black spruce - moss	0.330	47.1
1a.	black spruce - birch - moss	0.018	2.5
1b.	open - moss	0.011	1.6
2.	birch - white spruce - duff and moss	0.210	30.0
3.	birch - duff and moss	0.066	9.4
3a.	aspen - duff and moss	0.008	1.2
3b.	open - duff and moss	0.002	0.3
4.	alder - bare	0.001	0.2
5.	willow - sedge - bare	0.008	1.1
6.	black spruce - willow - sedge - bare	0.046	6.6
	total	0.700	100.0

Measurements from Küchler's (1967) map of potential natural vegetation indicate that about 60% of the Yukon-Tanana uplands (including Glenn Creek Basin) is "spruce-birch forest," 30% is "dry as meadows and barrens," 9% is "muskeg," and 1% is "black spruce forest." It is clear from this map (scale 1:7,500,000) that the areas indicated as "black spruce forest" include only the very extensive black spruce forests that occur in major lowland areas (e.g., Yukon Flats, Minto Flats), and that the very common and areally more important smaller black spruce forest, such as found in Glenn Creek Basin, are included in the "spruce-birch forest." Sigafoos' (1958) vegetation map of Alaska shows only two units in the Yukon-Tanana uplands: the "interior spruce and birch forest," which makes up about two-thirds of the province (including Glenn Creek Basin), and an "undifferentiated treeless region" at the higher elevations. His text, and those of Stoeckeler (1949) and Lutz (1956) indicate the widespread occurrence of vegetation communities essentially identical to those mapped in the watershed of Glenn Creek. Thus, while vegetation mapping is of insufficient detail to determine the areal distribution in the province of the units mapped in Glenn Creek watershed, it is clear that the general forest types of the basin are widespread in the Yukon-Tanana uplands.

Hutchinson (1967) mapped the vegetation of Alaska from an economic point of view. His map indicates that the watershed of Glenn Creek is in an area of "medium- to poor-stocked non-commercial spruce-hardwood forest." This unit is present in about 50% of the Yukon-Tanana uplands, which also contains about 40% "non-forest" (generally at the higher elevations) and 10% "well-stocked commercial spruce-hardwood forest" (generally in major river valleys). Thus, Glenn Creek Basin is apparently generally typical of the low and moderate elevations of the Yukon-Tanana uplands in regard to stand density and tree size as well as species composition.

Passive relationships. The relations between vegetation and topography are illustrated in Figure 18. The black spruce - moss association (unit 1) and the black spruce-willow-sedge-moss association (unit 6) correspond most closely to Stoeckeler's (1949) "muskeg and swamp forest" and Holmes and Benninghoff's (1957) "black spruce muskeg type." Stoeckeler (1949, p. 61) reported that this type occurs generally where the ground is perennially frozen at a depth of 2 ft or less; Lutz (1956, Fig. 10), Holmes and Benninghoff (1957) and Sigafoos (1958, p. 173) stated more generally that this is characteristic of poorly drained areas.

Stoeckeler (1949, p. 53) stated that the white spruce-birch forest (unit 2) is the most widespread vegetation type encountered in interior Alaska, occurring on valley, slope, and hilltop sites on loess, residual, or water-laid soils. According to Lutz (1956, p. 41), this forest type is an advanced successional stage following forest fire, which, barring further fires, evolves to a pure white spruce forest. This was also essentially the conclusion of Holmes and Benninghoff (1957, p. 215). Sigafoos (1958, p. 172) stated that "White spruce and white birch forest are characteristic of some lowlands and well-drained uplands and river terraces throughout the forest region."

A more definite statement of the relationship between these two forest types and the presence of permafrost was made by Péwé (1966, p. 13) and cited earlier (see p. 17).

While Lutz (1956, p. 37) believed that the birch forest type (unit 3) is a relatively early stage in the succession following fires, evolving to the white spruce - birch and eventually to the climax white spruce forest, a recent study indicates that it is "very unlikely that more than an occasional white spruce can become naturally established beneath a birch stand" in interior Alaska (Gregory, 1966, p. 253). Holmes and Benninghoff (1957, p. 215x) recognized a deciduous forest type, which is analogous to the birch forest described above, and stated that birch tends to occur on well-drained sites. This was also essentially the conclusion of Sigafoos (1958, p. 174).

Stoeckeler (1949, p. 73) is the only one of the earlier writers who described an association similar to the alder thicket (unit 4) of Glenn Creek watershed. He believed vigorous stands of pure alder are best developed on moist peaty soils, which are commonly perennially frozen within 3 ft of the surface. In Glenn Creek Basin, tree-height alders were not observed on peaty soils, but rather on highly mineral soils having thin ground cover or lacking ground cover. Here, the area mapped as alder forest with bare soil lies in a small "draw," which is probably an ephemeral stream channel. Permafrost was encountered at relatively great depth (2.92 ft) beneath this area.

The sedge hummock - willow bush community (unit 5) is characteristic of valley bottoms and marshy areas in interior Alaska (Johnson and Vogel, 1966, Fig. 32).

The discussion above indicates the general information, principally in regard to soil drainage, provided by various vegetation associations in interior Alaska. However, forest fires, natural and man-caused, are frequent and often of large magnitude in the region. As much as 4.5 million acres have burned in a single year (Lutz, 1956, p. 14). The ecological effects of such fires are, of course, severe, and Lutz (1956, p. 5) has stated that "only when the influence of past fires is recognized can one begin to account for the seemingly haphazard mosaic of vegetation."

In the watershed of Glenn Creek, charred stumps, found principally in the birch-spruce forest of the south-facing slope, are definite evidence of recent fire. While there are no written records of fires in the area prior to 1940, a study of tree rings from the watershed has provided information

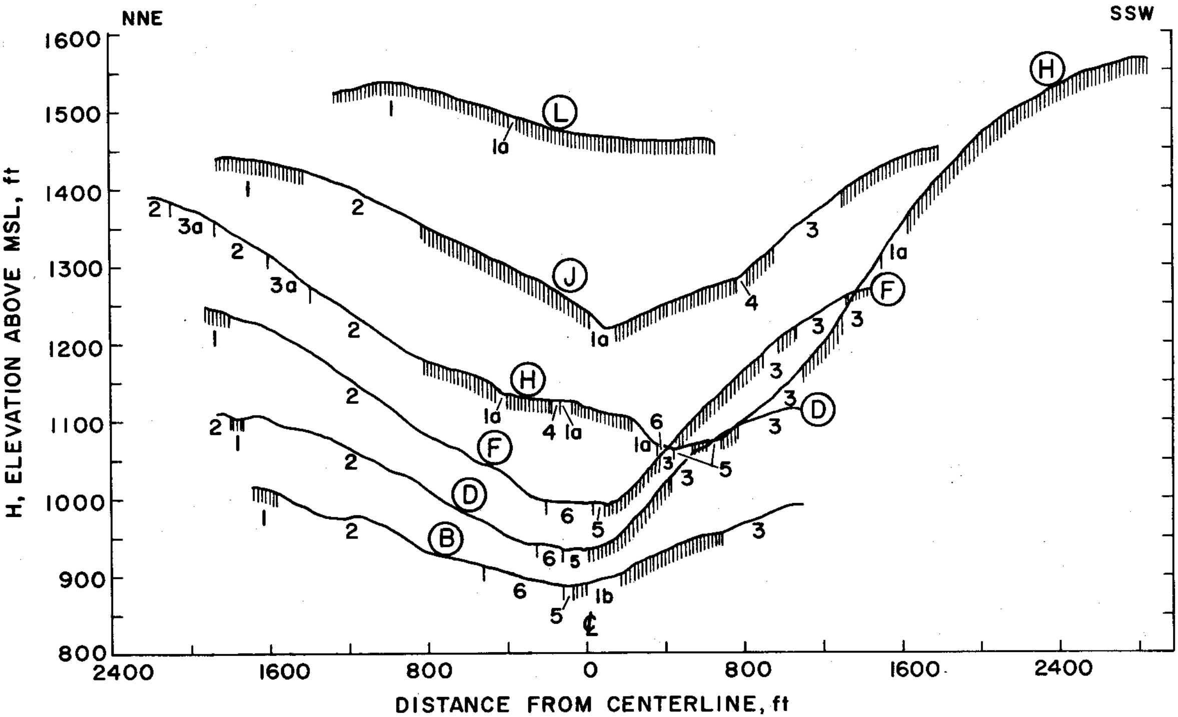

Figure 18. Cross sections of Glenn Creek watershed showing vegetation. Sections perpendicular to base line, through points indicated by letters. Numbers indicate vegetation associations of Figure 17.

on probable dates of forest fires. The general chronology of the trees, the presence of resin canals, and in a few cases direct signs of probable fire damage, all indicate a fairly extensive fire in the basin between 1900 and 1910. That this fire did not completely destroy the forest is indicated by the presence of living trees predating this period by as much as 60 years. A minor concentration of resin canals about 1937 may indicate a less extensive fire at that time, but there is little corroborating evidence.

Even though much of the watershed may have been burned some 60 years ago, its major effect would seem to have been to prevent the vegetation of the south-facing slope from evolving to the climax pure white spruce forest postulated by Lutz (1956). Variations of dominant tree types within the area mapped as birch and birch-spruce forests (aspen and balsam poplar stands) may be related to fires, but do not alter the general interpretation that these types are indicative of well-drained soils. Lutz (1956, Fig. 10) also indicates that black spruce may repopulate poorly-drained areas following fires, except where there is repeated severe burning. Therefore, there seems to be no reason for believing that forest fires have introduced significant disturbance of the normal plant-edaphic relations as discussed above.

Active relationships. No studies have been conducted on the relationships between vegetation types and interception, evapotranspiration, or infiltration in central Alaska. The present discussion must therefore be based largely on inferences from such studies conducted elsewhere on generally similar types.

Interception. Interception is both a process (the collection of precipitation on vegetation and litter and its subsequent evaporation) and a quantity (the amount of water so collected and evaporated). As a quantity its implications are different depending upon whether it is viewed on a seasonal or annual basis or on the basis of an individual storm.

Considered seasonally or annually, at least part of the "loss" of intercepted water is part of the evapotranspiration and need not be considered separately. There is a disagreement in the hydrologic literature concerning the degree to which intercepted precipitation is an alternative for, rather than an addition to, the amount of water transpired by plants. Thorud (1967) found that only 10-15% of intercepted water replaced transpiration in his experiments, and Rakhmanov (1958) and Schindel (1963) reported somewhat higher transpiration reductions. Leyton and Reynolds (1964) reported transpiration reduction generally less than interception, while Burgy and Pomeroy (1958) found that all interception was used to replace transpiration in grasses. Rutter (1967) shows a diagram of measured interception and transpiration vs calculated transpiration which definitely indicates that all intercepted water replaced water which would have been transpired by a pine forest. The subsequent discussion is based on the assumption that this is generally true, at least as a first approximation.

Assuming that precipitation and the climatic factors influencing evapotranspiration are the same over the watershed, and that climate, rather than vegetation type, is the dominant control of evapotranspiration, differences in interception by different tree types do not materially affect evapotranspiration. This is because the evaporative demand of the climate is satisfied by both intercepted water and soil moisture; although one type of vegetation intercepts more than another, the evapotranspiration of the former will be satisfied to a greater extent from intercepted water and to a lesser extent from soil moisture, the total water loss being equal in both vegetation types.

Accepting this, it is of interest to cite figures for seasonal interception which may be applicable to the dominant vegetation types in the watershed of Glenn Creek, the black spruce forest (map unit 1) and the predominantly birch forests (map units 2 and 3). No published data are available for these forest types in the subarctic. Table IX summarizes values which are probably approximately applicable to the study area.

On the basis of the data in Table IX, it is difficult to select values which are applicable to the vegetation of Glenn Creek. The spruce forest studied by Stålfelt (1963) seems to be most similar in terms of tree height and spacing to the black spruce forest of Glenn Creek. However, the percentage of precipitation intercepted by a forest is a function not only of vegetation type, size, and spacing, but also of precipitation characteristics of the area. Helvey and Patric (1965), after examination of all available interception studies, found that one regression equation could adequately represent the relationships between interception and gross precipitation for eastern hardwoods:

$$I = 0.083\,P + 0.036\,n \qquad \text{growing season} \qquad (1a)$$

$$I = 0.059\,P + 0.020\,n \qquad \text{dormant season} \qquad (1b)$$

where I is interception loss (in.), P is gross precipitation (in.), and n is the number of storms per season. Patric (1966) found an equation of the same form for the coniferous forests of southeastern Alaska, and cited others from other studies of conifers. An unweighted average equation for the reports cited by Patric (1966) is

$$I = 0.292\,P + 0.023 \qquad (2)$$

which applies to rainfall interception. The influence of number of storms and gross precipitation on the percentage of gross precipitation intercepted is shown in Figure 19, using eqs 1a and 2. Using the rainfall data from 1 June to 17 September 1964 (Dingman, 1966a) (9.68 in. of rainfall in 45 storms), Figure 19 indicates that the birch forests intercepted about 22% of this, and the spruce forest about 38%.

Table IX. Summary of interception study results applicable to Glenn Creek watershed.

		Percentage of gross precipitation intercepted			
	Vegetation	*Summer*	*Winter*	*Annual*	*Source*
1.	Spruce			23	cited in Horton (1919, Tab. 1)
2.	Spruce			33	
3.	Spruce-spruce-fir	32	35		Lull (1964, Tab. 6-2)
4.	Young spruce (*Picea abies*)	27			Reynolds and Leyton (1963)
5.	Sitka spruce ***(Picea sitchensis)*** and western hemlock ***(Tsuga heterophyla)***			35	cited in Patric (1966, Tab. 1)
6.	Western hemlock and sitka spruce			27	Patric (1966)
7.	Sitka spruce			33	Law (1957)
8.	Spruce (*Picea excelsis*)	46	47	47	Stalfelt (1963)
9.	Aspen-birch	10	4		Lull (1964, Tab. 6-2)
10.	Beech-maple-birch	20			Beall (1934)
11.	Beech-maple-birch	13	12	12	Leonard (1961)

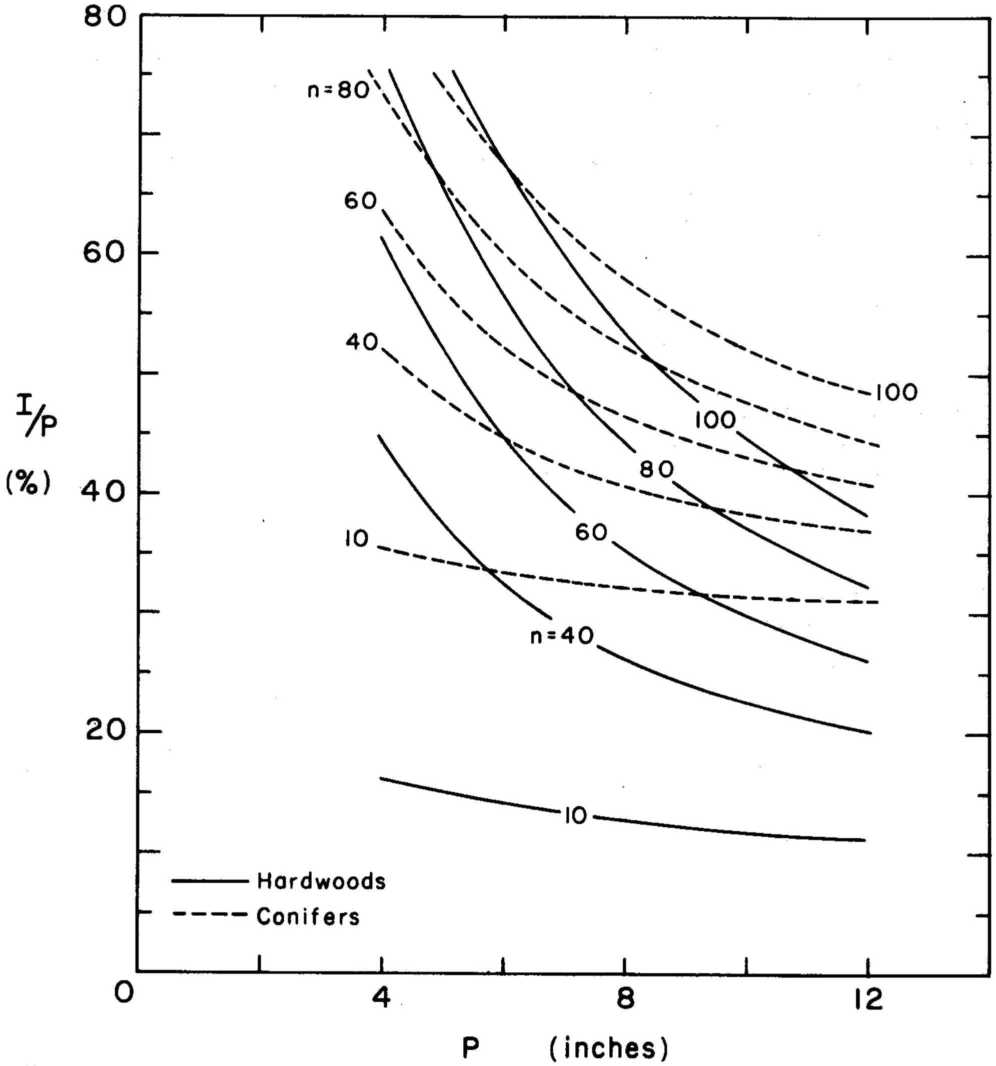

Figure 19. Fraction of gross precipitation intercepted I/P as a function of gross precipitation P and number of storms per season n, for typical hardwood and conifer forests.

Transpiration. Evapotranspiration, which is an upper limit for transpiration, has been discussed under the section on climate.

Information on the transpiration characteristics of subarctic forest vegetation types is extremely rare. Further, since transpiration is closely controlled by meteorologic conditions, it is risky to apply data from other regions to the watershed of Glenn Creek. Hare (1950) found that isopleths of Thornthwaite's potential evapotranspiration (which depends only on mean monthly temperatures) corresponded closely to forest-type boundaries in subarctic eastern Canada. He gave the following values of mean annual potential evapotranspiration: "open boreal woodland" (spruce-lichens-muskeg), 14 to 17 in.; "main boreal forest" (black spruce-white spruce-balsam fir-muskeg), 16.5 to 19 in.; and "boreal-mixed forest ecotone" (spruce-fir-pine-birch), 18.5 to 20 in. (Hare, 1950, Tab. 2). Gulidova (1958) studied evapotranspiration in the taiga forest in northern USSR (near 55.5 N, 40 E), and found average evapotranspiration values of about 12 in./yr in mature dominantly spruce forests, and 12 to 15 in./yr in spruce-birch forests. This amounted to about 60% of the average annual precipitation in the region.

It is clear that these values are considerably greater than those which must prevail in central Alaska (see p. 28). This is probably due to the increased temperatures and growing seasons at the lower latitudes where Hare (1950) and Gulidova (1958) worked.

Infiltration. To the author's knowledge, no infiltration studies have been done in central Alaska or areas of similar soils and vegetation. It seems safe to say, however, that, because of the ground cover, infiltration rates virtually always exceed rainfall rates in most of Glenn Creek Basin. This conclusion is based on a consideration of the watershed in three portions: 1) areas with a ground cover of duff and moss patches (vegetation map units 2, 3, 3a, 3b; 40.9% of the watershed); 2) areas with a thick moss ground cover (map units 1, 1a, 1b; 51.2% of the watershed); and 3) areas with bare ground (map units 4, 5, 6; 7.9% of the watershed).

As noted earlier, the permeability of the Fairbanks silt loam, the boundaries of which correspond closely to the portion of the basin with a duff - moss patch ground cover, was given as 0.2 to 0.6 in./hr (Rieger *et al.*, 1963, Table 9). Figure 16 indicates that storms of 0.2 in./hr are not rare in this area for short durations. However, examination of recording rain gage records for this study reveals only a few occasions when this intensity was reached.* Further, a thorough field examination of the north side of the basin immediately following the very unusual, long and intense storm of 8 to 12 August 1967 revealed no sheet erosion, gullying, leaf-and-twig debris piles, or other evidences of overland flow, indicating that infiltration capacity was never exceeded there.

In addition, there is mounting evidence that infiltration rates in the upper layers of forest soils are generally very high. From his studies in a forested watershed in Vermont, Engman (1966, p. 3) stated that "if you dig deep enough, you can find surface runoff." The reasons given for this lack of overland flow were: 1) low-intensity rains; 2) excellent vegetation cover (which both promotes soil permeability and reduces rain intensities by interception); and 3) few swelling colloids in the silt-loam soils which might restrict infiltration. All three of these factors would seem to be acting with at least equal force on the north side of Glenn Creek watershed. [Rieger *et al.* (1963, Table 6) gave the shrink-swell potential of the Fairbanks silt loam as low; this is corroborated by the absence of clay-sized material in the grain-size distributions shown in Fig. 9.] Thus there seems to be no reason to believe that infiltration rates are exceeded and that overland flow occurs in this portion of the basin.

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* This discrepancy is more apparent than real. The shortest duration considered by Miller (1963) was 1 hour; the high-intensity rains recorded at Glenn Creek were all of shorter duration than this. Miller (1963) gives a recurrence interval of 1 year for 0.2 in. of rain in 1 hour. This means there is a probability of unity that it will be equaled or exceeded in a given year.

Comments on the infiltration characteristics of the moss-covered portion of the basin stem from more direct information. Appendix B gives the results of a laboratory study of the permeability of moss from this area. Vertical permeability was measured at about 310 in./hr (0.218 cm/sec), obviating the possibility of surface runoff from this portion of the basin.

The Goldstream silt loam, which underlies most of the portions of the basin with no continuous ground cover, was described by Rieger *et al.* (1963) as being poorly drained, with permafrost at shallow depths and the water table commonly at or above the ground surface. These characteristics were confirmed by the author's observations in the field; standing water was observed several times in several locations in the valley bottom. In addition, this soil was described by Rieger *et al.* (1963, Table 6) as having a moderate shrink-swell potential, so that some reduction in infiltration capacity upon wetting can be expected. Where the water table is close to or above the ground surface, consideration of permeabilities becomes irrelevant, and infilitration capacity is essentially zero.

While the above reasoning with regard to infiltration characteristics is somewhat inferential, it will be seen later that such characteristics are consistent with the most reasonable hypothesis of runoff production in the watershed.

DATA COLLECTION

Figure 20 summarizes the periods during which data of various types were collected in the course of this study.

Precipitation

The number and location of rain gages at Glenn Creek were controlled to some extent by logistical considerations. The first measurements were made approximately daily in a standard U.S. Weather Bureau 8-in. nonrecording gage located about ⅓ mile down valley (northwest) of the discharge-measuring site, from 1 June to 24 June 1964. From 24 June to 24 August, a similar gage about ¼ mile northwest of the discharge-measuring sited was used. A standard U.S. Weather Bureau 8-in. weighing-recording gage was installed at the same location on 24 August, and provided records until freeze-up on 25 October 1964.

During the spring of 1965, arrangements were made with the U.S. Army Meteorological Team, Ft. Wainwright (Fairbanks), Alaska, to install and service a meteorological station adjacent to the stream-gaging station. However, the station could not be installed until 4 August. On 5 August, data collection began at the weir site, with the installation of a nonrecording rain gage and a weighing-recording rain gage. Measurements were discontinued as freeze-up began on 29 September 1965. During 1966, the recording and nonrecording gages were in operation at the weir site station from 6 May to freeze-up on 11 October, and in 1967 from 3 May to 3 October.

All rain gages were installed in accordance with instructions in U.S. Weather Bureau Circular B (U.S. Weather Bureau, 1955). In particular, no trees or other objects were closer to the gages than twice the height of the objects above the gages. Alter-type windshields were used in 1964, but were discontinued thereafter, as wind speeds are generally quite low (maximum 24-hour wind travel recorded at the weir site evaporation pan was 72.8 miles), and standard Weather Bureau and U.S. Army Meteorological Team stations in Alaska do not use windshields.

Hershfield (personal communication, 1965) stated that, where only one rain gage is to be placed in a drainage basin, the most representative location is at the geographic center of the

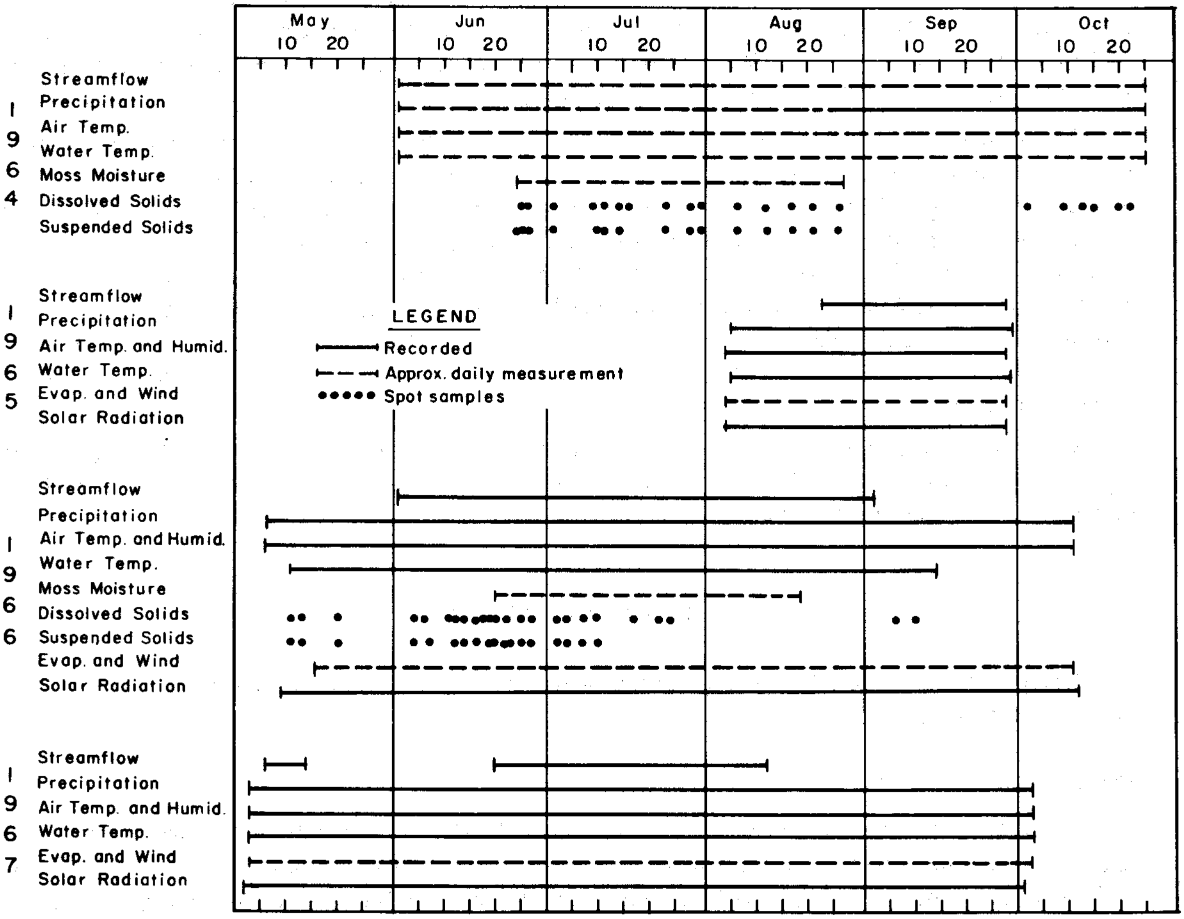

Figure 20. Periods of data collection, Glenn Creek watershed, 1964-1967.

basin. Because of logistic problems, this was not possible in the Glenn Creek watershed. However, a study of rain-gage spacing (Hershfield, 1965), though based on an analysis of only 15 sites in the conterminous United States, suggests that the one gage at the weir site provides adequate estimates of rainfall for the drainage basin of Glenn Creek.

Hershfield's analysis led to a nomogram relating 2-year, 24-hour rainfall and 2-year, 1-hour rainfall for the area of interest to the rain-gage spacing that gives an "adequate" measure of storm rainfall for a given area. The criterion for adequacy used was a correlation of 0.9 between amounts measured at a gage at the center of a watershed and amounts measured at some distance from this gage. As noted previously, Miller (1963) used the relatively sparse and poorly distributed rainfall data from Alaska to construct rainfall-frequency maps for the state. From these maps, the 2-year, 24-hour rainfall and 2-year, 1-hour rainfall at Glenn Creek are about 1.25 in. and 0.4 in., respectively (see Fig. 16). With Hershfield's nomogram, these values give an "adequate" gage spacing of about 2 miles. Since the distance from the rain gage to the farthest-removed part of the watershed is 1.5 miles the one gage should provide a good sampling for the entire basin.

Streamflow

During 1964, streamflow was measured on an approximately daily basis by one of two methods. Most measurements were made with a Price pygmy current meter attached to a rod graduated in tenths of feet. A straight reach of channel free from large obstructions was selected for the measuring site. This reach was cleared of channel vegetation, and a guide board was fixed across the channel, level and at right angles to it. This board was marked and provided with nails so that the current-meter rod could be accurately placed at 0.1-ft intervals across the channel (see Fig 21). For each discharge determination, a plumb bob was used to locate the edges of the water relative to the markings on the board. The current meter, which could be adjusted to any position on the rod, was then placed in the channel at the point closest to the left bank at which the water was deep enough to cover the meter at its lowest position (0.2 ft above the channel bottom). Channel depth and flow velocity were then noted at this point and at successive 0.1-ft intervals across the channel until the depth again became insufficient to cover the meter. Flow velocities were measured by counting the number of revolutions of the current-meter cups for 60 sec and then referring to the rating table for the meter. Where depths were 1.0 ft or greater, the meter was read at 0.2 and 0.8 of the distance from the water surface to the bottom, and the results were averaged to give an average velocity for the vertical. At depths between 0.5 and 1.0 ft, a single reading at 0.6 of the depth was taken.

These methods conform to standard stream-gaging procedures for determining average velocity in a vertical plane (see Corbett, 1943; Boyer, 1964). Where depths were less than 0.5 ft, but greater than 0.3 ft, velocity measurements were read at 0.2 ft above the channel bottom. Since at these small depths the diameter of the current-meter cups was a large percentage of total depth, it was assumed that the measured velocity was close to the average, and no corrections were made for these conditions.

At very low flows (generally when discharge was less than 0.1 ft^3/sec and average depth less than 0.3 ft), a volumetric method was used for discharge determination. A polyethylene-covered wire-mesh flume was installed in the stream channel about 40 ft upstream from the current-meter site. At the time of a discharge measurement, flows could be temporarily diverted through this flume into a bucket of known volume, and the time required to fill the bucket noted by means of a stopwatch. Five such measurements were made for each discharge determination and the results were averaged.

A staff gage was installed immediately downstream from the guide board, so that a rating curve (discharge vs gage height) could be developed (see Fig. 24). Once established, the rating curve was used for occasional discharge estimates. This was especially useful in conjunction with a peak-stage gage, also installed in the measuring reach. It was found that organic debris floating on the stream would adhere to this peak-stage gage, and a line of debris was left at the peak stage after flow diminished. Readings on it were graphically related to staff gage readings, and discharge at the peak was then estimated from the rating curve.

It is not possible to state with certainty the accuracy of these methods, but it is believed that individual determinations by current meter or by the volumetric method are within 10% of the true values. This belief is based on the following factors: 1) the stopwatch was checked against other clocks and was found to be accurate; 2) the spin in air of the current meter was checked at the time of each determination; 3) the current-meter measurement reach was straight and quite uniform for about 15 ft above the measurement site and about 5 ft downstream from the site; 4) there were no important obstructions to flow on the channel bed or banks; 5) relations between elements of hydraulic geometry (discussed more fully later) were consistent and similar to those reported for other streams (see Bruun, 1966); 6) velocity and depth were measured at no fewer than 8, and generally at 10 or more, verticals in the cross section.

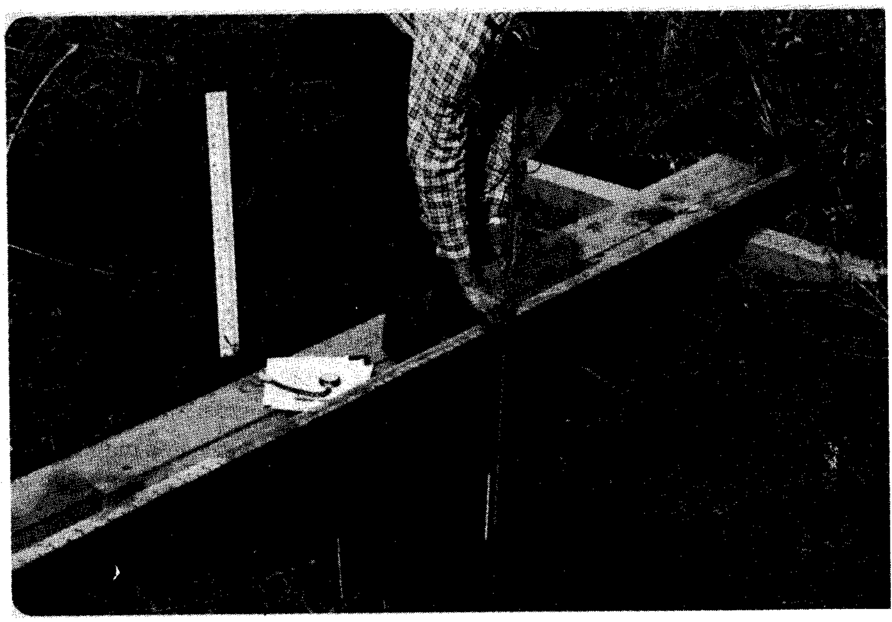

Figure 21. Method of discharge measurement, 1964.

Because the hydrographs of Glenn Creek are rather drawn out in time, the methods described above were adequate for defining the hydrographs for 1964. The short-term discharge fluctuations accompanying changes in rainfall intensity and the small rises in response to short storms noted in subsequent years were not apparent in these records.

In 1965, it was decided to record the discharge of Glenn Creek continuously. For the range of flows expected, a 3-ft, 90° V-notch weir seemed most appropriate as a control. Such a weir is highly accurate at low flows and has a capacity of about 38 $\mathrm{ft}^3/\mathrm{sec}$. It is impossible to say what flood frequency this discharge represents.

The basic equation relating discharge q ($\mathrm{ft}^3/\mathrm{sec}$) and head above the notch vertex H (ft) is

$$q = 4.277\ C_D\ H^{5/2} \tag{3a}$$

where C_D is the weir discharge coefficient. Calibration tests run at the Alden Hydraulic Laboratory of Worcester Polytechnic Institute (Alden Hydraulic Laboratory, 1965) showed that $0.6271 \geq C_D \geq 0.570$ for $H < 0.600$ ft and $C_D = 0.570$ for $H > 0.600$ ft. Thus C_D was found from a graphical relation between C_D and H for $H < 0.600$ ft ($q < 0.680\ \mathrm{ft}^3/\mathrm{sec}$), and the appropriate value used in eq 3a, while for $H > 0.600$ ft, eq. 3a becomes

$$q = 2.438\ H^{5/2}\ . \tag{3b}$$

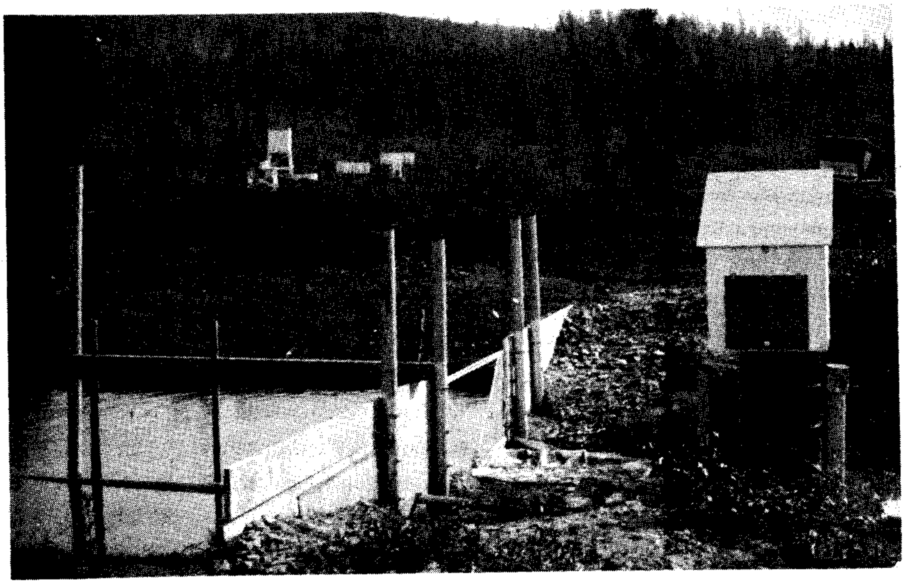

Figure 22. 90° V-notch weir in operation. Stilling well and recorder house to right, meteorological instruments in background.

A wooden dam anchored by piles set 12 ft into the ground (the permafrost table was at a depth of 2.6 ft originally) was constructed across the channel about 20 ft upstream from the discharge measuring site of 1964. A 2-ft-diam stilling well was anchored to piles downstream from the dam, with three 0.33-ft inside-diam transparent intake pipes leading from the stilling well through the dam to a pile 8 ft upstream of the dam. A Stevens A-35 water-level recorder with a 1-ft diam float and a weight-driven clock was placed in a housing above the stilling well. Figure 22 shows the weir measurement system in operation.

The gear ratios of the recorder clock and pulley assemblies were such that time could be estimated to within 2 to 3 min and head could be estimated to 0.001 ft. Head readings on the recorder could be checked by reference to a staff gage, graduated in 0.01 ft and set such that the zero point was at the same elevation as the bottom of the weir notch, and attached to the pile supporting the intake pipes. Checks of the recorder readings of time and head were made daily. The distance from the notch to the intake pipe entrances was sufficient to assure that drawdown effects were negligible (Stevens, 1968, p. 15-16). The total length of the intake pipes, the intake pipe diameter, and the area of the stilling well were such that the lag of the response of the water level in the stilling well to a change in head behind the weir was negligible.

With the entire system operating properly, discharge could be read to two significant figures, and readily estimated to three significant figures, with the head above the lower limit of calibration of the weir plate. The main problems in obtaining an accurate discharge record arose when there was leakage under the dam. Such leakage obviated the collection of reliable discharge records during the months of June, July and much of August 1965; September 1966; and May and June 1967. Leaks were repaired by adding mixtures of sand and drilling mud to the base of the dam, and by promoting freeze-back and preventing thaw of the ground at the base of the dam. Discharge measurements ceased following the flood of 12 August 1967, when extensive erosion occurred around the dam.

Air temperature and relative humidity

During 1964, air temperatures were measured by maximum-minimum thermometers in a standard U.S. Weather Bureau shelter about ⅓ mile downstream from the discharge-measurement site. Average daily temperature was computed by averaging the maximum and minimum readings for the day. However, because readings were made at about 0800 hours, the maximum and minimum readings did not correspond to those for the calendar day. To correct for this, the minimum temperature was assumed to have occurred during the current calendar day, and the maximum during the preceding day. Temperatures calculated in this way correlated well with those calculated from hourly data at the U.S. Weather Bureau station in Fairbanks (see Dingman, 1966a, p. 6). Relative humidity was not measured in 1964.

The weather station installed by the U.S. Army Meteorological Team at the weir site in August 1965 included a hygrothermograph. Air temperature measurements that year covered the period 4 August to 28 September. Similar records were obtained from 6 May to 11 October 1966 and 3 May to 3 October 1967. The hygrothermograph was installed in a standard U.S. Weather Bureau shelter, and time and temperature and relative humidity readings were checked daily, 5 days a week.

Water temperature

During 1964, water temperature was measured in Glenn Creek at the time of each discharge determination, and each time a dissolved sediment sample was taken. Measurements were made by submerging a thermometer in a vigorously flowing portion of the stream that was shaded from the sun.

On 5 August 1965, two distance-thermographs were installed: one reading water temperature in the weir pond and one in the stream about 20 ft upstream of the head of the weir pond, at the point where sediment samples were taken. The sensors of these instruments were shielded from the sun, and readings were checked against calibrated thermometers daily, 5 days a week. The instruments were removed on 29 September 1965, and reinstalled on 11 May 1966. The 1966 readings were discontinued on 14 September to prevent damage by ice formation. Water temperature measurements in 1967 covered the period 3 May to 3 October.

Evaporation

A Class-A evaporation pan was installed according to U.S. Weather Bureau specifications (USWB, 1955) at the weir site on 4 August 1965. Daily measurements of evaporation were made until 28 September that year, and again for the period 16 May to 11 October 1966 and 3 May to 3 October 1967. An accumulating anemometer located next to the pan provided readings of total wind mileage at a level about 1 ft above the bottom of the pan during these periods. In 1966 and 1967 a Six-type floating maximum-minimum thermometer provided readings of water surface temperature in the pan. Both these instruments were read at the time of evaporation measurements.

Snow surveys

In September 1965, personnel from the U.S. Soil Conservation Service established a standard snow course on the basin of Glenn Creek. The course was located in a birch forest, about 700 ft south of the weir. Measurements of snow depth and water content were made on or about the first of each month when snow was on the ground. For each determination, snow depth and water content were measured with a Mt. Rose snow sampler and a scale at five evenly spaced points along the course.

Depth of thaw

Thaw depth was simply measured by probing the ground with a graduated steel rod. The frozen ground had a high ice content, making it impossible for the probe to penetrate. Layers of coarse materials at depth in the soil also provided resistance to the probe, but these could usually be identified by the metallic sound made when the probe struck or scraped the rock particles. When doubt existed as to the nature of the resisting surface, a soil sampling spoon was used to provide samples for visual examination.

Seasonal thaw reaches its greatest vertical extent in late August. In the vicinity of Glenn Creek, the annual variation in this depth is only a small percentage of the total depth of thaw so that the depth of the permafrost table can be adequately defined by such measurements. To map the depth of the permafrost table, in the watershed, the ground was probed with a 3.35-ft rod at four points around each of the 86 grid points on the basin during the period 15-18 August 1966. The four measurements were taken at the ends of an imaginary 6-ft cross centered on the grid point, with one arm extending parallel to the grid-survey line and the other at right angles to it.

Dissolved solids

Samples for dissolved-solids determinations were taken periodically during 1964 and 1966 (see Fig. 20). The samples were taken by submerging a polyethylene bottle in the stream at the gaging site (1964) or just above the weir pond (1966). The bottle was filled with creek water, rinsed, and emptied twice immediately before each sample was collected. The bottle was then securely capped, sealed with tape, and shipped to CRREL, where the determinations of dissolved solids were made in the chemistry laboratory by the author.

In the laboratory, the residue-on-evaporation method of determining total dissolved solids was used (Rainwater and Thatcher, 1960, p. 270-271). First, each sample was filtered through no. 1 filter paper, and 100 milliliters of filtrate pipetted and transferred to a beaker. Then 10 milliliters of 30% H_20_2 was added to oxidize organic material, and the beaker was placed on a steam table and allowed to evaporate until 10 to 15 milliliters remained. This water was then transferred to a tared aluminum weighing dish, and the beaker rinsed with distilled water, which was also poured into the weighing dish. The dish was then placed in a 180°C oven and evaporated to dryness. After removal from the oven and cooling in a dessicator for about one hour, the dish and residue were weighed to 0.00002 g.

Several blank runs were carried out using distilled water to determine residue caused by pickup of ions from the filter paper and impurities in the H_20_2 and distilled water. This amount, which averaged 0.00026 g, was the correction used in calculating total dissolved solids by the formula:

$$\text{total dissolved solids (mg/liter)} = \frac{[\text{dish + residue (g)}] - [\text{dish (g)}] - [\text{correction (g)}]}{(100\ \text{ml})\ (10^{-4})} .$$

HYDROLOGY

Introduction

As stated at the outset of this report, the objective of this study was to achieve understanding of the hydrologic behavior of a small watershed typical of a significant portion of the discontinuous-permafrost zone of the subarctic. An attempt was made to accomplish this through analysis of the rainfall, streamflow, and other pertinent data, which would hopefully provide insight as to the major physical hydrologic processes acting in the watershed during the summer and a conceptual model of the ways in which they act.

Most texts on hydrology, e.g., Linsley, *et al.* (1949, p. 405-409), Wisler and Brater (1959, p. 17-18), and Ward (1967, p. 310-319), present a qualitative general discussion of the runoff process, including interception, infiltration, filling of surface depressions, and initiation of overland flow and interflow, implicitly at least, involving an entire watershed. While these general descriptions of processes may serve as a useful didactic purpose, it should not be inferred that they are strictly applicable to all, or perhaps even to most, situations, or that the runoff cycle as described is completely general and completely understood.

As evidence for this view, Betson (1964) developed a successful runoff model based on the premise that, in most storms in an area of North Carolina and Tennessee, runoff is produced only on a relatively small portion of a given watershed. This was supported by the work of Hewlett and Hibbert (1966). Detailed watershed studies by Engman (1966) and Ragan (undated) have shown that this partial-area-runoff-contribution concept is applicable in northern New England. Further, it is well established that overland flow occurs only rarely, if ever, in the forested northeastern United States (Hart, 1966; Engman, 1966; Ragan, undated).

Thus, these and other studies indicate the need for detailed investigation of hydrologic processes even in presumably well-known temperate areas, and especially in areas where precipitation, vegetation, and soil characteristics differ from those commonly encountered.

General channel description

For most of its length, the channel of Glenn Creek is well defined, but tortuous and with heavily vegetated banks (Fig. 23). Grasses and horsetails grow on the channel bottom in places. About 100 ft downstream from grid point E (see Fig. 2), the channel divides. Most of the water flows in the left channel, which has abundant evidence of recent very active erosion: trees are severely tilted toward the channel, which is generally incised 4 to 6 ft below the valley floor; the moss mat is generally breached (though in places it extends completely across the channel, with the water flowing several feet beneath); and near point D, the water flows underground for about 30 ft, well back under the left bank. The right channel shows no such evidence of active erosion, though it is incised several feet into the valley floor. The two channels join again about 100 ft above point B in an area of very diffuse flow patterns among vegetation hummocks. Below this point, flow is again in a well-defined channel incised 1-2 ft into the valley bottom. A cross section of the channel at the weir site (surveyed before weir construction began) was given in Figure 13.

Hydraulic geometry

Width, depth, velocity, and discharge. The current-meter discharge measurements of 1964 permit discussion of the hydraulic geometry (Leopold and Maddock, 1953) of Glenn Creek at the gaging site. The data are summarized in Appendix C. Figure 24 gives the stage-discharge relation at the 1964 gage site. The form of these curves is typical of those for larger streams (Linsley *et al.*, 1949, p. 213). It is immediately apparent from Figure 24 that a significant shifting of the rating curve took place in the course of the summer. Discharge measurements 1-45 (10 June - 25 July) define one curve, measurements 61-89 (17 August - 25 September) define another, and measurements 46 - 60 (27 July - 15 August) are transitional between the two.

Figure 23. Glenn Creek 100 ft above weir site (May 1965), view upstream. Creek is 1 – 3 ft wide.

For a given discharge, the curves for measurements 46 - 89 show a higher stage than does the curve for measurements 1 - 45, suggesting a silting up at the measuring section. To check this, cross sections were plotted using data taken during discharge determinations; some of these are shown in Figure 25. No significant change in the cross section occurred through 25 July, and its form for this period is represented by the section for 11 July. The section for 27 July, however, shows a silting of 0.05 to 0.1 ft across the channel. Curves of 4 August and 27 August show further silting. Calculations showed an average of 0.3 ft of sediment deposited at the cross section between 25 July and 27 August. Cross sections subsequent to 27 August show a deepening of the channel, especially near the left bank.

The at-a-station hydraulic geometry relations are shown in Figure 26; the equations relating average velocity $\bar{v}$ (ft/sec), average depth $\bar{d}$ (ft), and width w (ft) to discharge q (ft^3/sec) for the three sets of observations are:

Observations 1-45

$$\bar{v} = 0.405\ q^{0.51}$$
$$\bar{d} = 0.696\ q^{0.34}$$
$$w = 3.56\ q^{0.15}$$

Observations 46-60

$$\bar{v} = 0.434\ q^{0.48}$$
$$\bar{d} = 0.599\ q^{0.34}$$
$$w = 3.85\ q^{0.18}$$

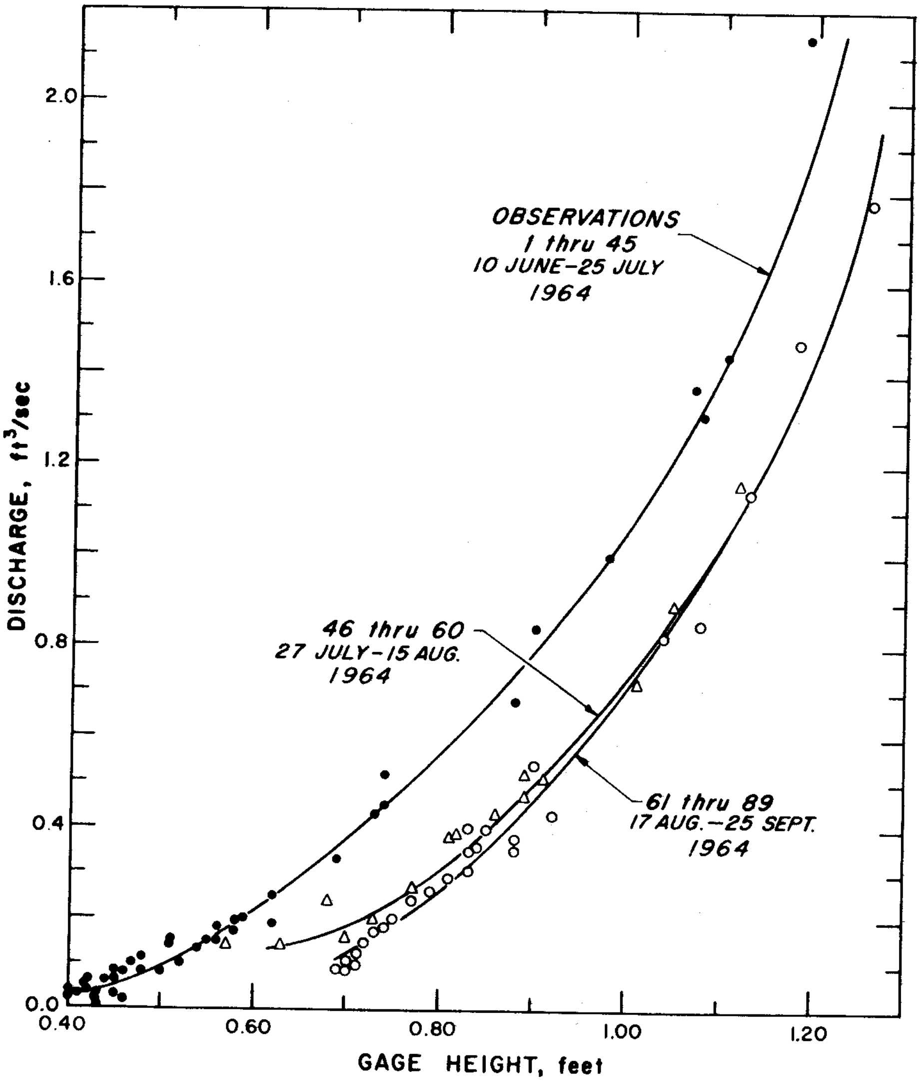

Figure 24. Stage-discharge relation at gaging site, 1964.

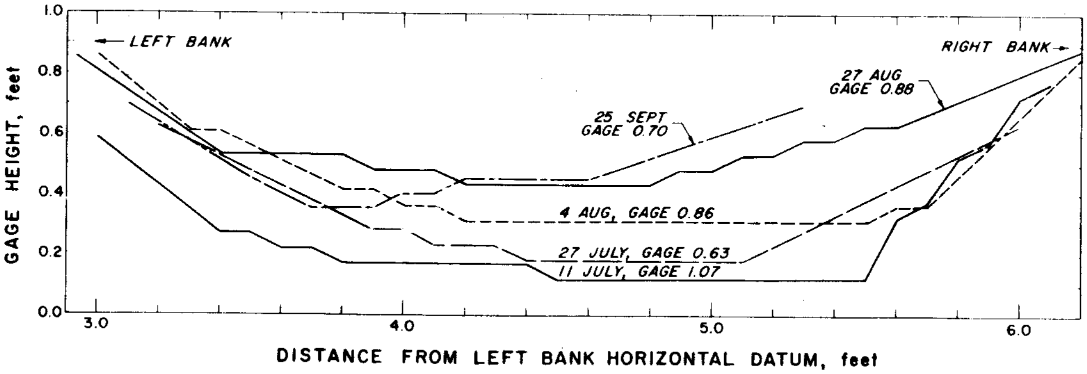

Figure 25. Successive cross sections at gaging site, 1964, showing changes due to scour and fill.

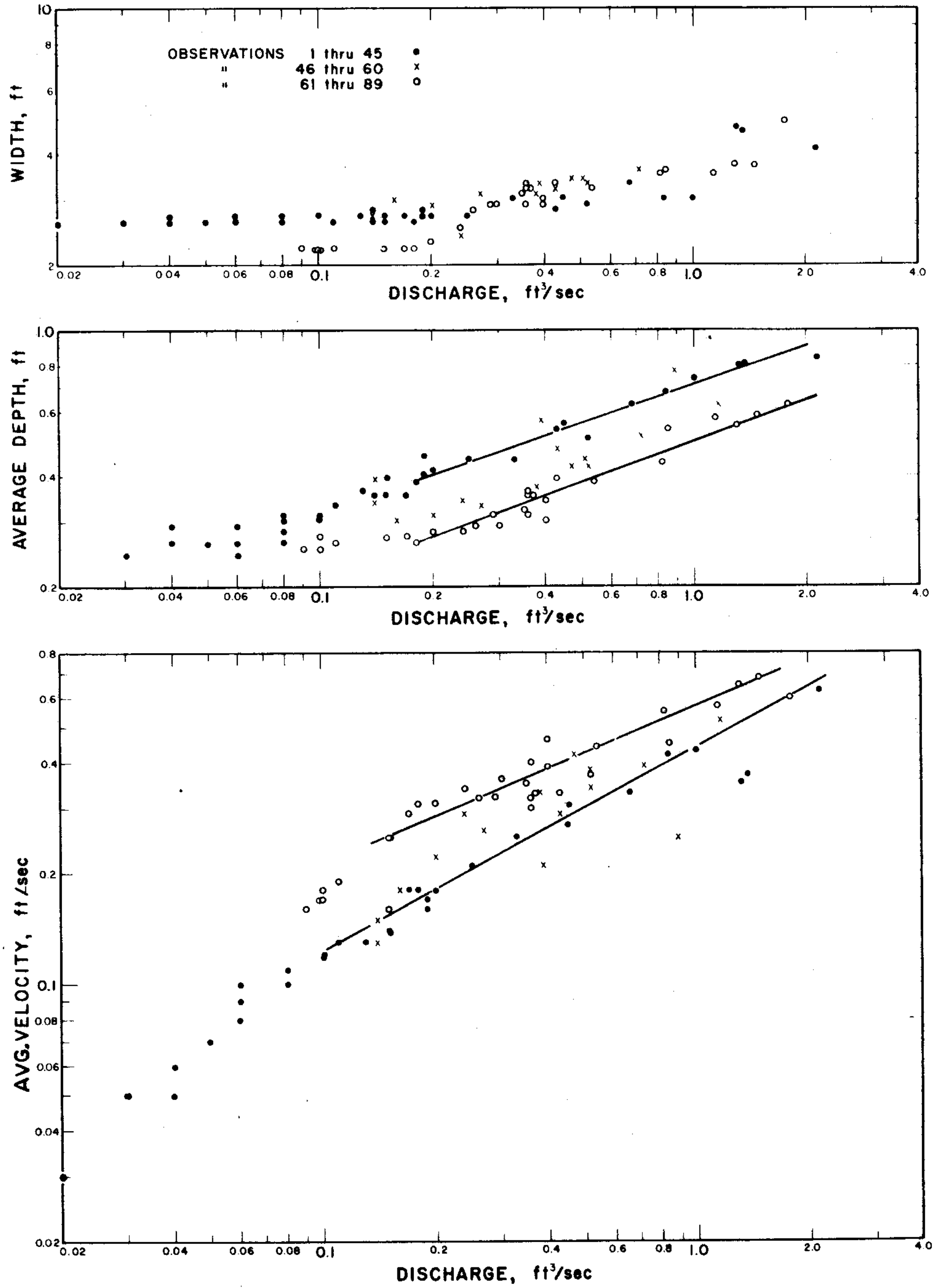

Figure 26. Hydraulic geometry at gaging site, 1964.

Observations 61-89

$$\bar{v} = 0.554\ q^{0.44}$$

$$\bar{d} = 0.481\ q^{0.32}$$

$$w = 3.75\ q^{0.24}\ .$$

The values of the exponents in the velocity relationships are slightly greater, and those for the depth relationships slightly less, than the average values for other streams summarized by Leopold *et al.* (1964, Tab. 7-5).

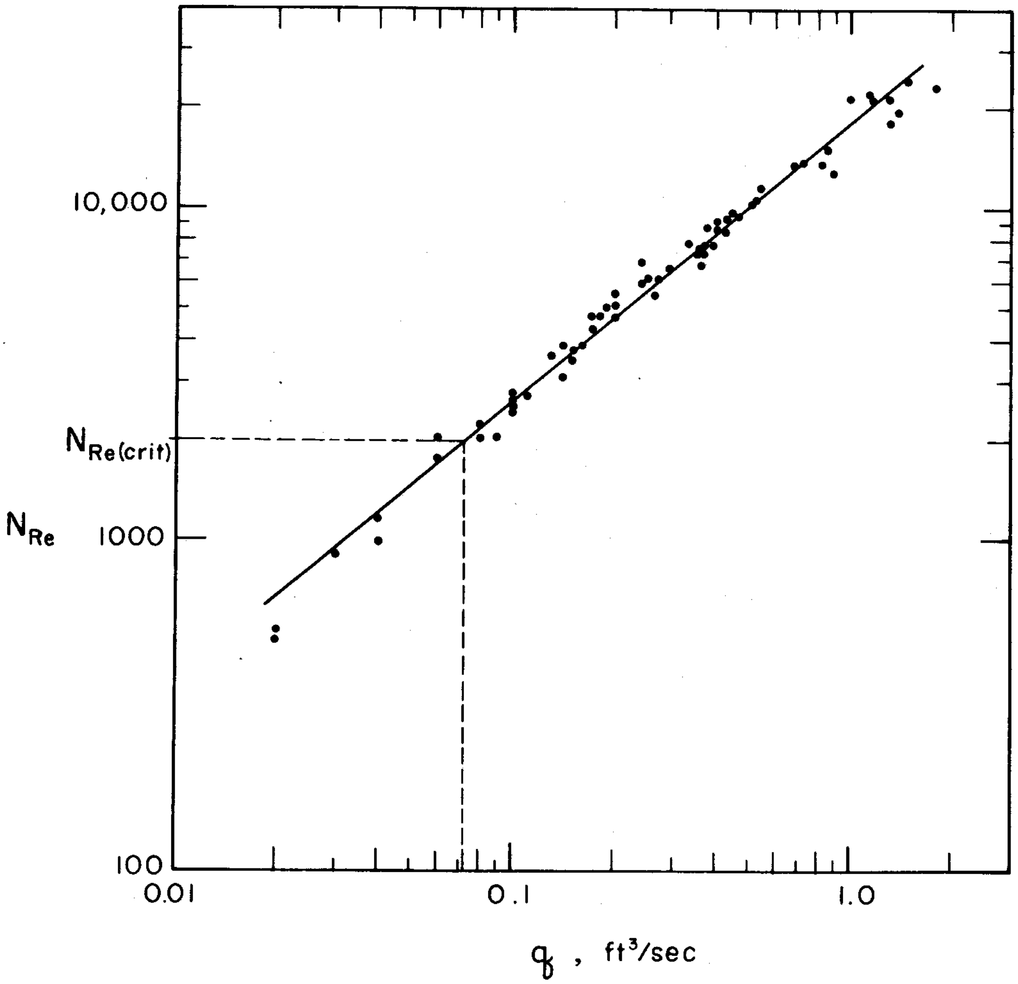

Figure 27. Relation between Reynolds number N_{Re} and discharge q at gaging site, 1964.

Flow regime. Figure 27 shows the relation between Reynolds number N_{Re} and discharge for 68 occasions in 1964. These data are also summarized in Appendix C. Reynolds number is defined as

$$N_{Re} = \frac{\bar{v}\,\bar{d}}{\nu}$$

where $\bar{v}$ is average velocity, $\bar{d}$ is average depth (essentially equal to hydraulic radius), and ν is is kinematic viscosity, using values corresponding to the water temperature at the time of measurement. The relationship is approximated by

$$N_{Re} = 17{,}920\ q^{0.84}$$

and indicates that for $q < 0.072$ ft³/sec, $N_{Re} < 2000$, the value generally given as the lower limit of fully turbulent flow. Flows less than 0.072 ft³/sec were not uncommon during the period of study: from 1 June - 25 October 1964, they occurred about 12% of the time.

The highest average flow velocity measured in 1964 was 0.68 ft/sec, and the highest average flow velocity in a vertical section of the stream was 1.28 ft/sec. Reference to Sundborg's (1956, Fig. 13) graph relating grain size and erosion and transporting velocities indicates that 0.68 ft/sec (0.21 m/sec) is just capable of eroding unconsolidated material of an average grain size of 0.02 mm, which is present along the channel of Glenn Creek. Flows of much lower velocities can readily transport material of this size (Allen, 1965, Fig. 10). These figures and the flow records for all years indicate that Glenn Creek performs very little work in scouring its channel by physical erosion, except presumably during spring runoff and rather rare summer storms. However, it does erode thermally, melting ground ice and thereby releasing fine-grained materials which it is capable of transporting.

Channel roughness. The tortuosity of the channel of Glenn Creek and its myriad vegetational obstructions would be expected to produce a high channel roughness (Chow, 1959, p. 101-104). This is borne out by calculation of the roughness factor n in the Manning equation

$$\bar{v} = \frac{1.49}{n} R^{2/3} S^{1/2}$$

where $\bar{v}$ is mean velocity in feet per second, R is hydraulic radius in feet and S is slope. To calculate n, data on $\bar{v}$ and R from 68 discharge determinations of 1964 were used (see App. C). (Measurements of cross sections showed that R was essentially equal to the mean depth, so the latter value was used in the computations.) Slope was measured on three occasions over an approximately 60-ft reach extending upstream from the discharge-measuring site, using a tape and a self-leveling level. The average of the three measurements was 0.0038 (range: 0.0036-0.0040), with no apparent relationship between slope and discharge. Since, theoretically, slope does not change with discharge at a point on a stream with cohesive banks (Leopold *et al.*, 1964, Tab. 7-8; Scheidegger and Langbein, 1966, p. 9), the average value of 0.0038 was used for all calculations. The results of these calculations are shown in Figure 28. For flow with Reynolds numbers in the turbulent regime (R > 2000), values of n range from 0.089 to 0.421, and average 0.197. This value is consistent with the method of estimation given by Chow (1959, p. 106-123). In this method, n is estimated as

$$n = (n_0 + n_1 + n_2 + n_3 + n_4)\, m_5$$

where n_0 is a basic value for a straight, uniform channel in the existing natural materials, n_1 is a correction for variations for shape and size of the channel cross section, n_3 is a correction for obstructions in the channel, n_4 is a correction for channel vegetation, and m_5 is a correction for channel meandering. Using Table 5-5 of Chow (1959), the following values are assigned:

Material:	earth	$0.020 = n_0$
Degree of irregularity:	severe	$0.020 = n_1$
Channel cross-section variability:	severe	$0.015 = n_2$
Channel obstructions:	severe	$0.060 = n_3$
Vegetation:	very high	$0.075 = n_4$
Degree of meandering:	appreciable	$\left(\frac{\text{stream length}}{\text{valley length}} = 1.3\right)$ $1.15 = m_5$

$$n = 0.020 + 0.020 + 0.015 + 0.060 + 0.075)\ 1.15 = 0.218.$$

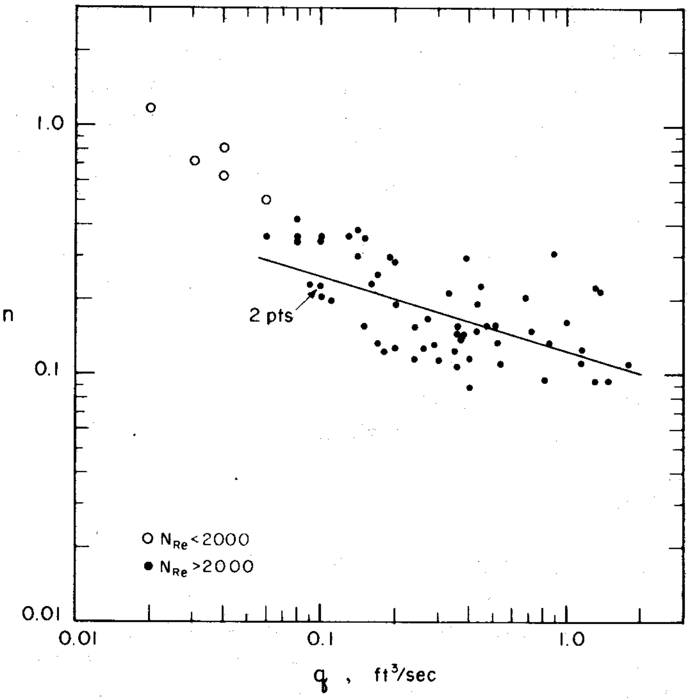

Figure 28. Relation between Manning's roughness n and discharge q at gaging site, 1964.

The relation between *n* and discharge is inverse, as is generally observed, and is approximated by

$$n = 0.127\ q^{-0.288}\ .$$

The exponent in this relation is of considerably greater absolute magnitude than the average values reported by Leopold and Maddock (1953) and Leopold *et al.* (1964, Tab. 7-8). However, Wolman (1955) and Johnson (1964) both found exponents of -0.20 for a medium-sized stream in cohesive bank materials and a small mountain stream, respectively. In Glenn Creek, the high rate of change of *n* with discharge is probably due to an increase in the ratio of the average flow depth to the height of vegetation obstructions in the channel. This reasoning is supported by Chow's (1959, p. 107-108) discussion of the effects of vegetation on channel roughness.

Time of concentration. The time it takes for a parcel of water to travel from the most distant part of the channel to the outlet is called the time of concentration for the channel. This is a significant parameter since, for a uniform rainfall intensity, if streamflow comes only from channel precipitation, it is equivalent to the time of equilibrium for the watershed, that is, the storm duration at the end of which the runoff rate at the watershed outlet is equal to the rate of runoff production ("excess rainfall") over the basin. "During a storm, once the maximum intensity has occurred for a duration equal to the time of concentration of the basin, additional rainfall will not cause a higher peak flow but will just prolong the runoff period" (Bruce and Clark, 1966, p. 245-246).*

The information on hydraulic geometry and roughness can be used to estimate the time of concentration for the channel. The length of the channel of Glenn Creek from the weir to the farthest

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point from the weir is 5340 ft measured on the 1:2400 topographic map. The average slope of this channel is 0.066. If we now select a range of discharges, the hydraulic-geometry relations established earlier can be used to find a corresponding depth at the weir site. One-half of this depth is perhaps a reasonable estimate of the average flow depth in the channel. A value of Manning's n can also be selected from its relation to discharge, so that an average velocity, and hence a time of concentration, can be calculated for the channel.*

Table X and Figure 29 show the results of these calculations, and indicate that time of concentration varies from less than 1 hour at high discharges to over 8 hours at low flows.

Table X. Estimated relation between time of concentration T_c and discharge q in the channel of Glenn Creek.

q (ft^3/sec)	$\bar{d}$ (ft)	n	$\bar{v}$ (ft/sec)	T_c (hr)
0.02	0.078	0.392	0.179	8.30
0.05	0.107	0.301	0.286	5.18
0.10	0.135	0.246	0.409	3.62
0.20	0.171	0.202	0.585	2.54
0.50	0.234	0.155	0.937	1.58
1.0	0.296	0.127	1.34	1.11

$\bar{v}$ = mean velocity
n = roughness factor
$\bar{d}$ = average depth

Kirpich (1940) presented a widely used empirical formula for calculating time of concentration for temperate-zone watersheds from watershed topographic parameters, and it is of interest to compare the values calculated for Glenn Creek with the results of that formula. Kirpich's relation is

$$T_c = \frac{L^{1.15}}{7700\,H^{0.38}}$$

where T_c is time of concentration in hours, L is the length in feet measured along and to the end of the main stream and thence in a direct line to the farthest point on the divide, and H is the difference in elevation in feet between the farthest point on the ridge line and the basin outlet. For Glenn Creek, $L = 5340 + 2600 = 7940$ ft, and $H = 775$ ft, and the calculated value of $T_c = 0.31$ hour. A similar formula is given by Bruce and Clark (1966, p. 246):

$$T_c = \left(\frac{2Ln}{3\,S}\right)^{0.467}$$

where T_c is in minutes, n is a roughness factor selected from a table, L is as in Kirpich's formula, and S is the mean slope of the main drainage channel. These writers give a maximum value of $n = 0.80$ for "coniferous timberland, or deciduous timberland with deep litter or grass." The appropriate value of S is 0.066, and the calculated value of $T_c = 93$ min or 1.55 hr. While it is not at all clear to what flow frequencies these calculated concentration times apply, they presumably are to be used in calculating storm runoff, and hence should be compared to the times corresponding to the higher flow rates of Glenn Creek. In this regard, the formula of Bruce and Clark (1966) seems

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* This procedure is used because it seems more reasonable to calculate an average depth based on the depth at the weir site and to use an average roughness found at the weir site than to calculate an average velocity based on the velocity at the weir site.

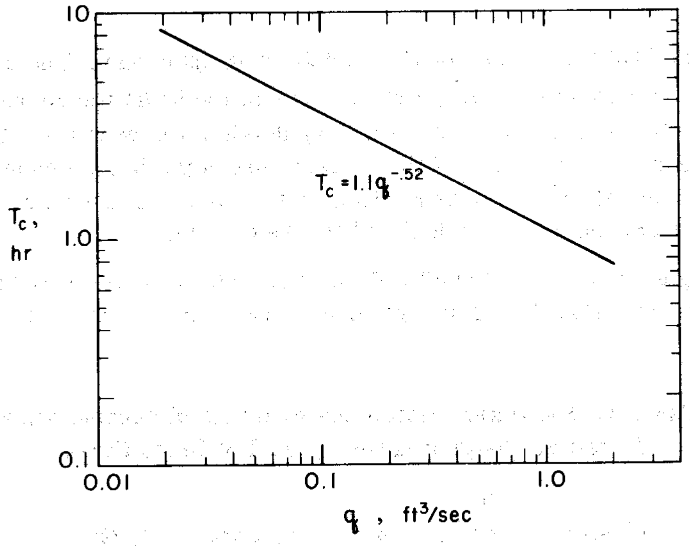


Figure 29. Estimated relation between time of concentration T_c and discharge q in the channel of Glenn Creek.

to provide a reasonable estimate when applied to Glenn Creek. However, both equations are for watershed concentration time, which supposedly includes time of travel before runoff reaches the channel.

We have seen that the time of concentration for Glenn Creek channel is a function of streamflow rate at the beginning of a storm, and hence varies with the degree of wetness (antecedent conditions) of the watershed. This must be true for all watersheds to the degree that average flow depth and roughness (or average velocity) are functions of discharge. Hewlett and Hibbert (1966) have also suggested that the stream channel network grows and shrinks in response to antecedent moisture conditions. Thus, calculation of a single time of concentration for a watershed, as by the formulae of Kirpich (1940) and Bruce and Clark (1966), appears unrealistic. This would seem to be particularly true when considering time of concentration for an entire drainage basin, since rates of runoff to the channel are probably highly dependent on antecedent conditions.

Characteristics of hydrographs

Rainfall-runoff volumes. Table XI gives the total runoff, total precipitation, and antecedent discharge (discharge at the time storm rainfall began) for 16 storms in the basin. Total runoff was found by plotting discharge (at times of measurement for 1964 and every 4 hours for 1965-1967) vs time on semilogarithmic paper. In all but a few cases, the discharge following a peak approximated a straight-line trend (exponential decay) on such paper; this line was fitted by eye and extended to a discharge of 0.01 ft³/sec. Following 8 of the 25 storms originally plotted, estimated flow from the previous storm or storms made up a large part (about 25 to 75%) of the total flow. Because of the difficulty in accurately separating flow due to each storm, the figures for these are eliminated from Table XI. Thus, the 16 periods listed are those during which only a minor portion of the total flow was estimated to be due to previous storms.

Actual points and the points on the extended line were then replotted on arithmetic paper, and the runoff due to a given storm was estimated by measuring the appropriate area with a planimeter. In the few cases where discharge showed a horizontal trend immediately preceding the rise of the subsequent storm, the slope of the recession of the subsequent or preceding storm was used to estimate the recession. The ratio runoff/precipitation ranged from 0.030 to 0.42, and the average for all storms was 0.18.

Table XI. Precipitation, runoff, and antecedent discharge for 16 storms on Glenn Creek watershed.

Runoff period	*Runoff (ft^3)*	*Runoff (in.)*	*Precipitation (in.)*	*Runoff pptn.*	*Antecedent discharge (ft^3/sec)*
9 June-28 June 64	422,000	0.26	1.69	0.15	*
22 June- 6 July 64	188,000	0.12	0.86	0.14	0.05
10 July-21 July 64	323,000	0.20	0.58	0.34	0.15
28 July-10 Aug 64	292,000	0.18	0.88	0.20	0.14
3 Aug-23 Aug 64	468,000	0.29	1.41	0.21	0.24
15 Aug-25 Aug 64	238,000	0.15	0.36	0.42	0.47
30 Aug-25 Oct 64	940,000	0.58	1.49	0.39	0.30
5 Sept-4 Oct 65	583,000	0.36	1.00	0.36	0.13
17 June-26 June 66	441,000	0.27	0.85	0.32	†
6 July-13 July 66	12,400	0.0076	0.22	0.035	0.01
9 July-16 July 66	20,700	0.013	0.39	0.033	0.03
7 July-14 July 67	34,600	0.021	0.69	0.03	0.02
19 July-24 July 67	40,100	0.025	0.76	0.033	0
21 July-27 July 67	165,000	0.10	0.70	0.14	0.19
23 July-30 July 67	515,000	0.32	1.19	0.27	0.39
8 Aug-17 Aug 67	333,000	0.20	1.05	0.19	0.06

* Discharge measurements began 10 June 64.

† Water-level recorder not operating, 16-17 June 66.

Figure 30 is a plot of runoff/precipitation as a function of antecedent discharge q_i. Antecedent discharge is taken as the discharge immediately before the stream begins to rise in response to a storm, and is a measure of the wetness of the watershed at the commencement of a storm. Although there is a fair degree of scatter, the regression relation

$$\frac{RO}{P} = 0.085 + 0.734\ q_i$$

has a correlation coefficient of 0.776, significantly different from zero at the 0.01 level. This indication that the proportion of rainfall running off is related to antecedent moisture conditions has important implications for a runoff model of the watershed, and will be discussed later in this report.

Hydrograph timing. Table XII presents data on the duration of storm T_s, the duration of the hydrograph rise (time to peak) T_p, and the apparent response (lag) time T_L for 28 storms. (Data for 1964 are eliminated because streamflow was not recorded that year.)

The duration of precipitation was determined by examination of rainfall and streamflow-recorder charts as the total length of time from the beginning to the end of rain responsible for a streamflow rise, measured to the nearest hour. It thus includes, in many cases, some periods when no measurable rainfall was falling. Its determination is somewhat subjective, but in most cases the beginning of a "block" of rainfall responsible for a rise is obvious. No rain which fell after the peak flow is included.

The apparent duration of rise is simply the total period from the initial flow increase until the occurrence of the peak flow rate, measured to the nearest hour. When the peak flow was maintained for several hours, as was the case for some storms, the initial hour of the peak flow occurrence was taken as the end of the rise time. The beginning of an actual streamflow increase is

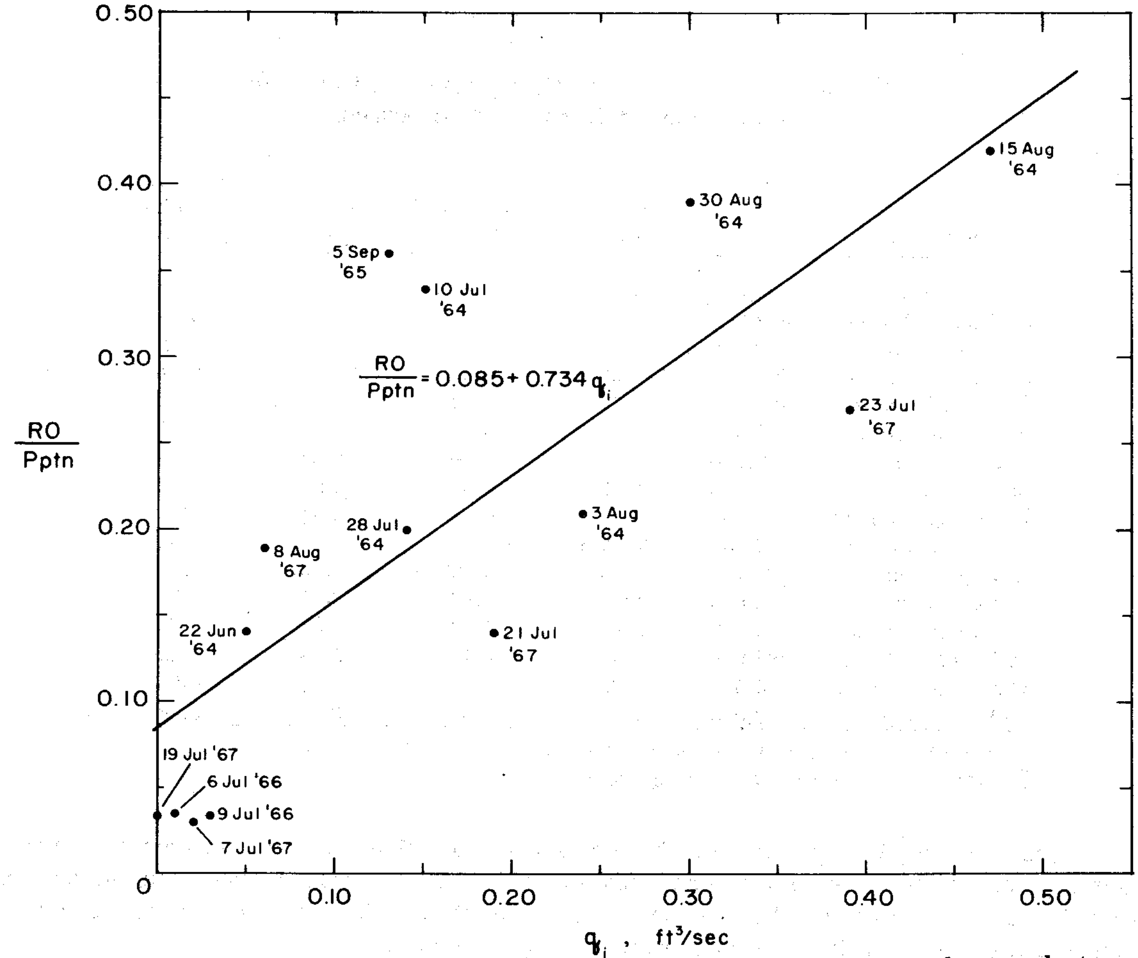


Figure 30. Relation between runoff/precipitation fraction RO/Pptn, and antecedent discharge q_i for 14 storms in Glenn Creek Basin.

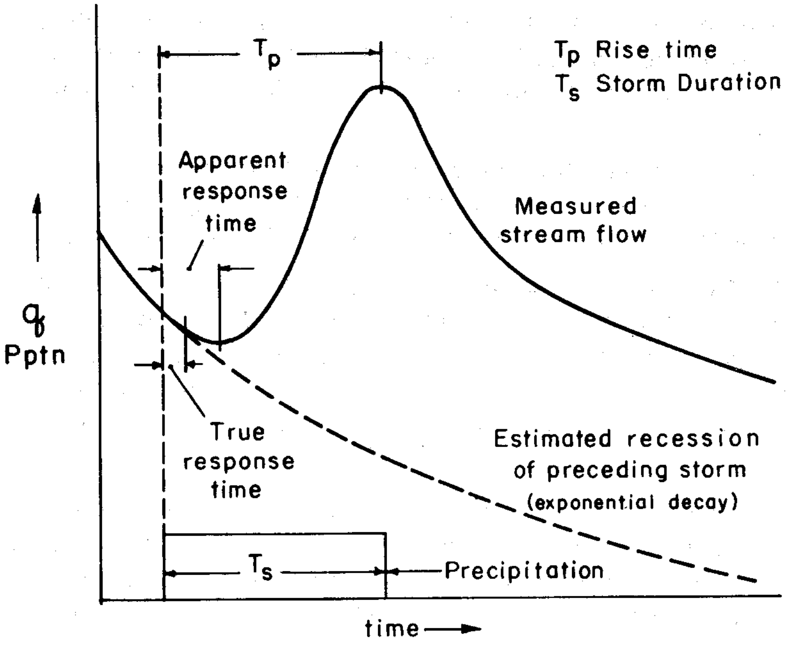


Figure 31. Diagram illustrating true and apparent response times, rise time, and storm duration.

Table XII. Storm duration T_s, duration of rise T_p, and apparent response time T_L for 28 storms on Glenn Creek watershed.

Storm	T_s *(hr)*	T_p *(hr)*	T_L *(hr)*
24 Aug 65	1	3.5	4
27 Aug 65	3	10	1
5 Sept 65	24	29	1
3 June 66	4	8.5	0
6 June 66	2	2	0
11 June 66	14	9	6
17 June 66	15	20	*
20 June 66	4	5.5	0
21 June 66	18	15	4
22 June 66	11	14	1
26 June 66	15	9	3
26 June 66	1.5	1.5	0
30 June 66	1	2	1
3 July 66	6	10	0
6 July 66	13	19	2
9 July 66	17	17.5	2
16 July 66	1	6.5	5
17 July 66	6	3	3
18 July 66	1	2.5	1
7 July 67	45	44	1
8 July 67	2	3	0
9 July 67	2	3	0
10 July 67	1	2	0
19 July 67	36	35	9†
21 July 67	16	14.5	11 (1) **
23 July 67	31	29	6
26 July 67	18	12	16 (7) **
8 Aug 67	32	36	0

* Streamflow recorder not operating at beginning of rise T_p estimated from beginning of rainfall to occurrence of peak.

† Streamflow zero at beginning of rain, weir pond level below notch.

** Numbers in parentheses "true" response times (Table XIII).

sometimes later than the true beginning of stream response. This is illustrated in Figure 31, where, when a storm occurs during a recession from a previous peak (the recession being assumed to follow an exponential decay), the streamflow continues to decrease, but at a slower rate than the exponential. However, Table XIII, which compares *true* response time as measured from the beginning of deviation from the previous exponential recession for storms where this could be determined with *apparent* response time, shows that for the most part there is a close correspondence between the two.

The data of Table XIII show that, except in some cases where circumstances were somewhat unusual, response times range from zero to 4 or 5 hours. This rather fast response is also illustrated by Figure 32, which gives plots of hourly discharge and hourly precipitation for 10 storms and associated hydrograph rises. In some cases a rather close relationship exists between rainfall intensity in a given hour and rate of increase in flow rate. This is particularly well illustrated for the storms of 5 September 1965 (Fig. 32a), 9-10 July 1966 (Fig. 32c), and 8-10 August 1967

Table XIII. True and apparent response times for 16 storms, Glenn Creek watershed (measured to the nearest hour).

Date	*True response time (hr)*	*Apparent response time (hr)*
5 Sept 65	1	1
3 June 66	1	0
20 June 66	0	0
21 June 66	0	4
26 June 66	3	3
6 June 66	2	2
9 July 66	2	2
7 July 67	1	1
19 July 67	13	14*
21 July 67	1	11†
23 July 67	6	6
26 July 67	7	16**
28 July 67	1	††
30 July 67	0	††
8 Aug 67	0	0
11 Aug 67	0	1

* Zero or very low streamflow for 7 days preceding this storm.

† Discharge constant for 11 hours following beginning of rain.

** Rain intensity very low (drizzle) for 5 hours after initial rain.

†† No rise; runoff continued to decrease following these storms.

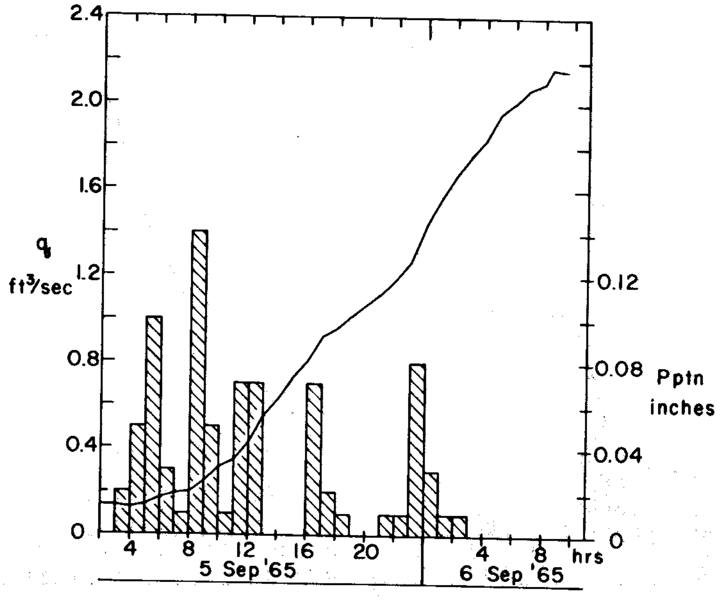
(Fig. 32h); for most rises, there is a less specific, but still apparent, correspondence. On 19 July 1967 (Fig. 32d), when the stream had zero flow at the onset of rain, some 14 hours of generally low intensity rain (total = 0.17 in.) fell before streamflow commenced. It can also be seen from Figure 32 that the peak flow rate was reached within a few hours of the end of rain.

Table XII gives data on the duration of storm T_s and duration of the hydrograph rise T_p for 28 storms on Glenn Creek Basin; the data are plotted in Figure 33. A simple regression between the two parameters gave the equation

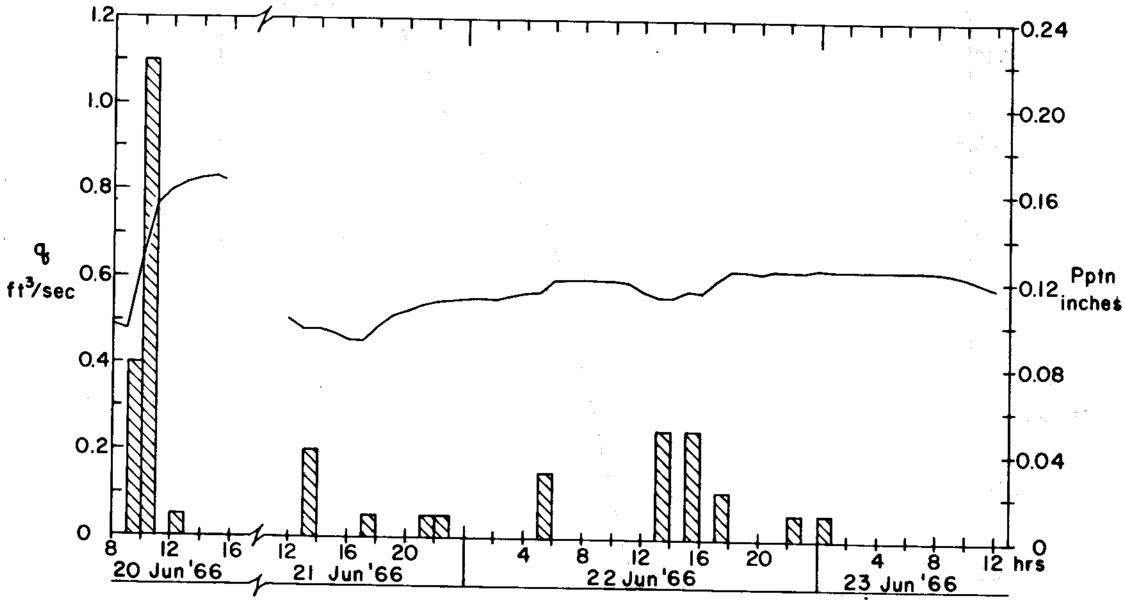
$$T_p = 1.79 + 0.93\ T_s$$

with a correlation coefficient of 0.96. Although there is considerable scatter, the correlation is significant at the 0.01 level of significance. Further tests reveal that the slope of this relation is not significantly different from unity and the intercept is not significantly different from zero at the 0.05 level. Thus, T_p is essentially equal to T_s.

A comparison between the stream discharge rate at the beginning of rainfall q_i, and the response time T_L, sheds further light on the runoff process. As noted earlier, the value of q_i can be taken as an index of the *wetness* of the watershed at the beginning of rainfall. If runoff during a rise comes to a significant extent from interflow or "Hortonian" overland flow (i.e., overland flow which begins after infiltration capacity has decreased to a value less than rainfall intensity), one would expect a faster response (shorter T_L) from a wetter land surface (higher value of q_i). Table XIV gives data to evaluate this relation for 27 storms. A simple correlation analysis between

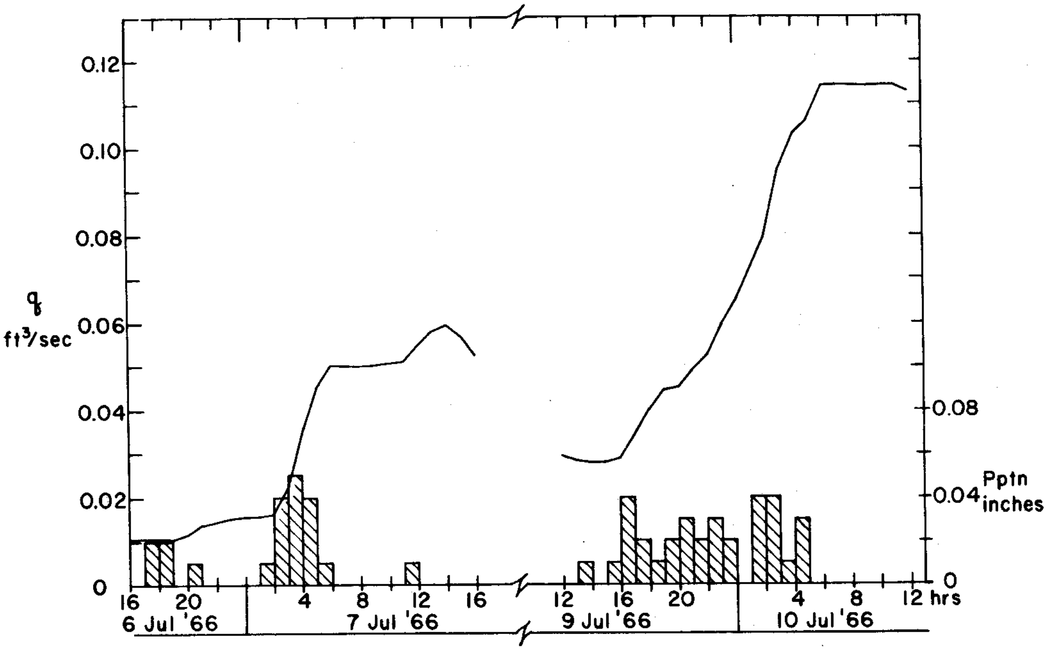


a. 5-6 September 1965.

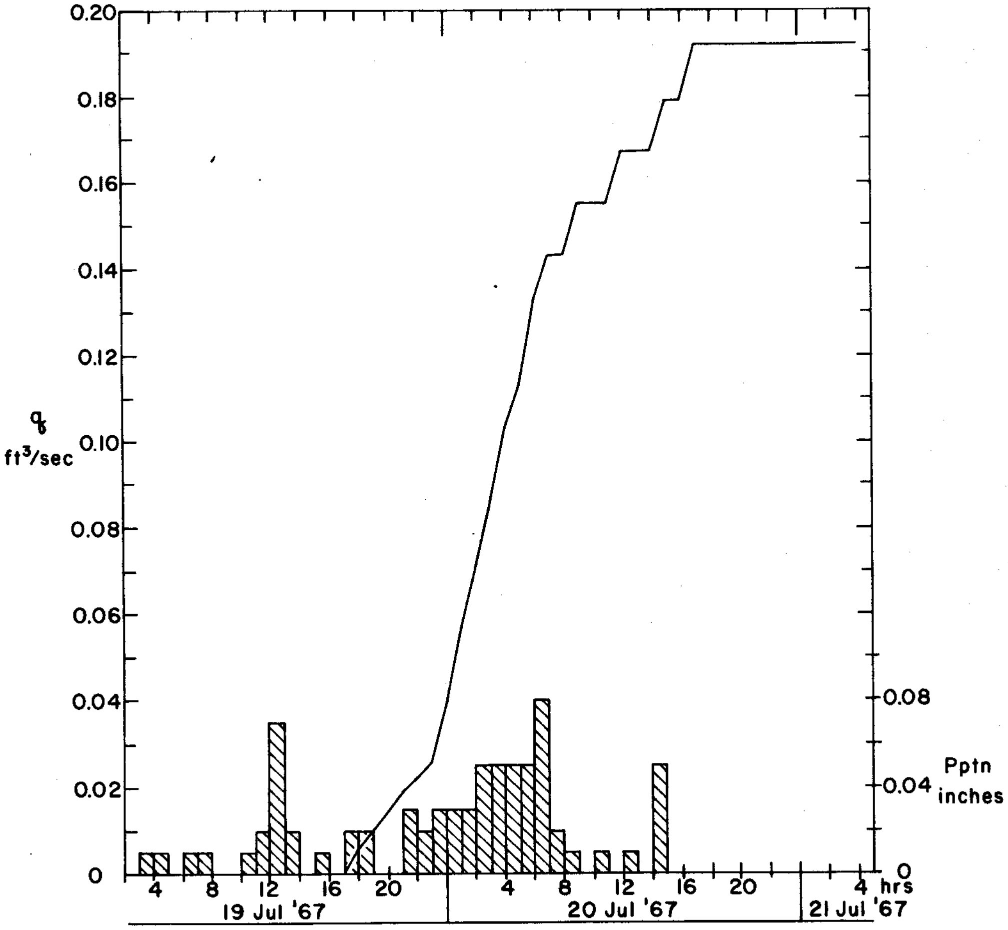


b. 20 June 1966 and 21-23 June 1966.

Figure 32. Precipitation and streamflow rise, Glenn Creek.

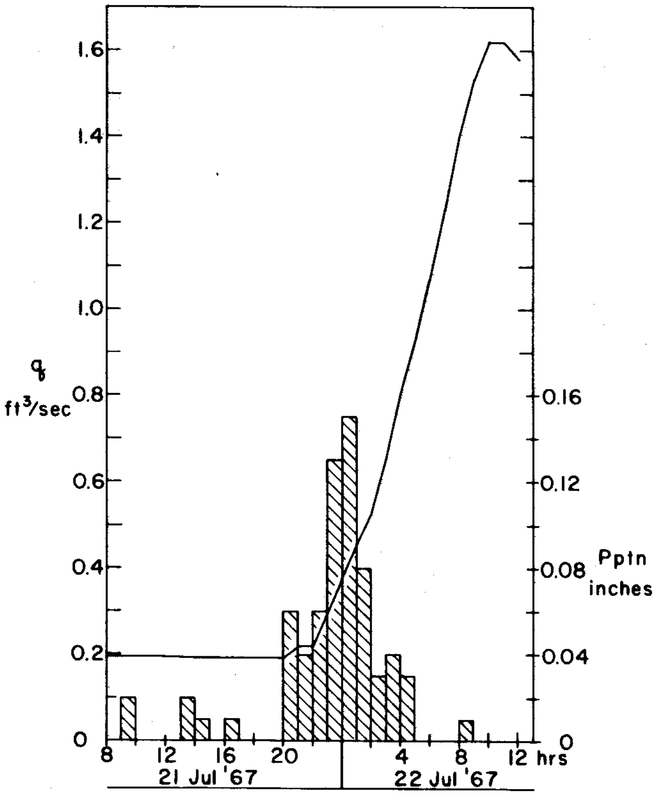


c. 6-7 July 1966 and 9-10 July 1966.

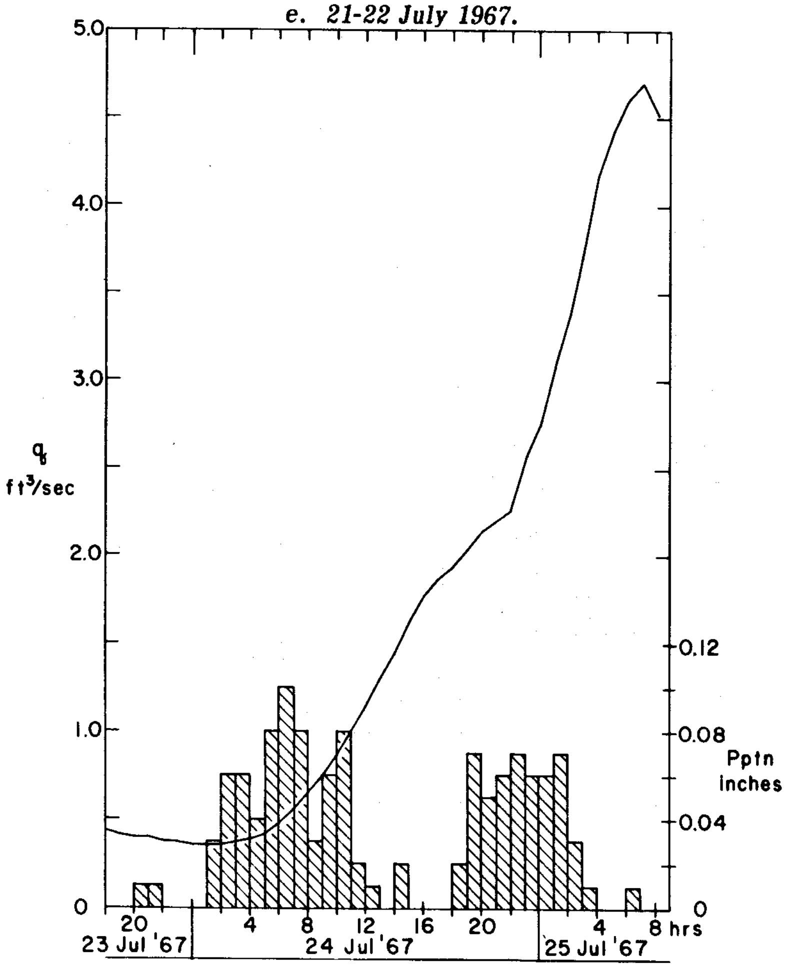


d. 19-21 July 1967.

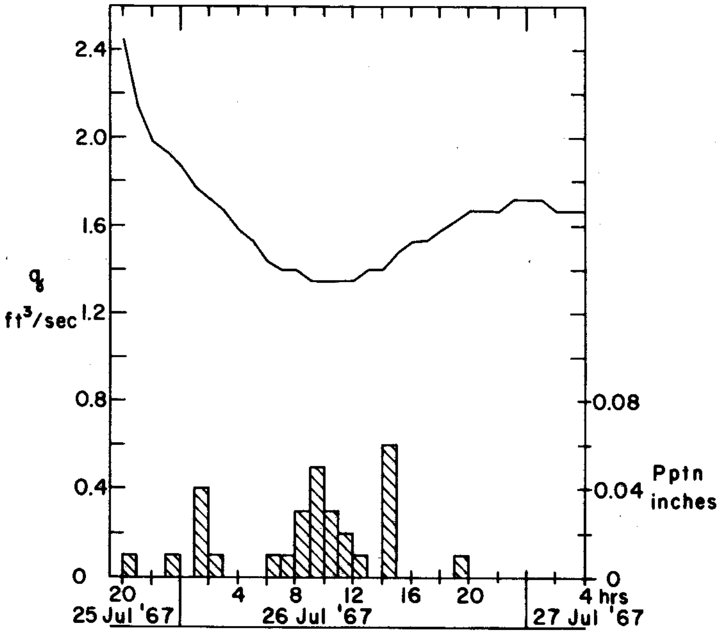
Figure 32 (Cont'd). Precipitation and streamflow rise, Glenn Creek.



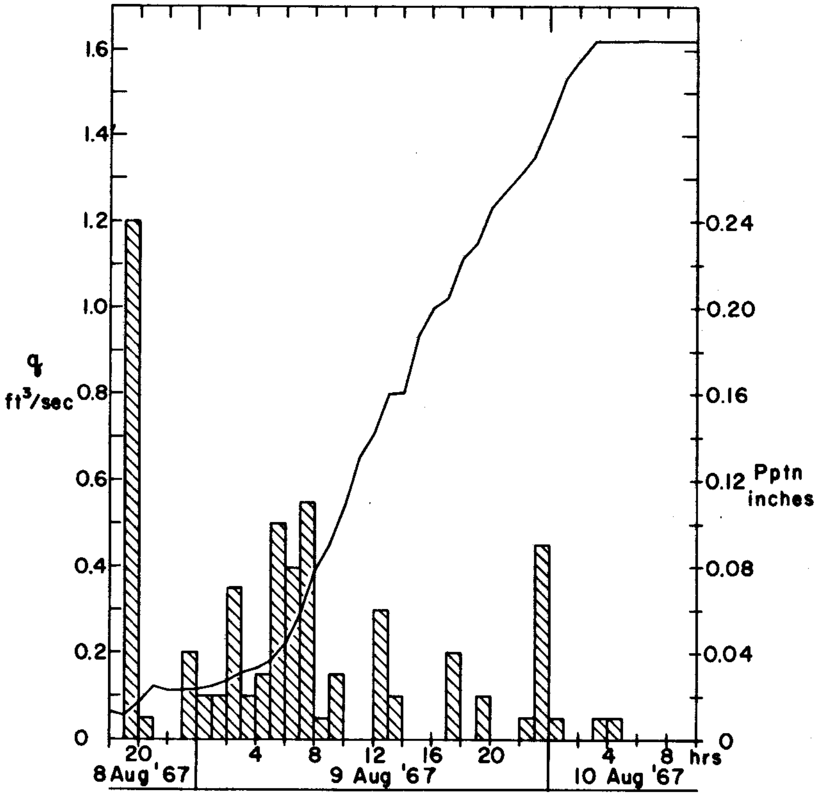
e. 21-22 July 1967.



f. 23-25 July 1967.



g. 25-27 July 1967.



h. 8-10 August 1967.

Figure 32 (Cont'd). Precipitation and streamflow rise, Glenn Creek.

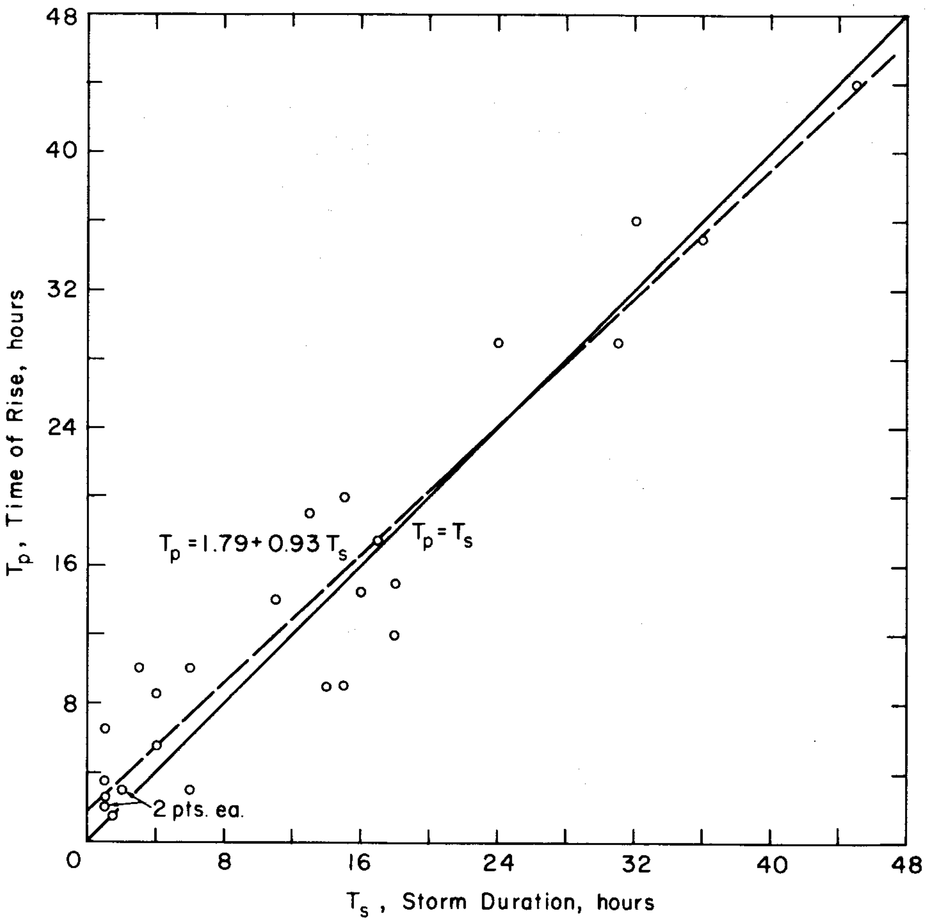


Figure 33. Relation between duration of rise and storm duration for 28 storms on Glenn Creek watershed.

the two variables shows a correlation coefficient of 0.225, which is not significant at the 0.05 level (and is, in any case, positive, which is opposite to the relation expected if watershed wetness affects T_L).

Recession characteristics. The general equation for base flow or recession flow is derived from the basin storage equation when there is no recharge to the basin:

$$\frac{dV}{dt} + q = 0 \tag{4}$$

where V is the volume of water in storage in the basin at time t and q is discharge at time t. A general relation between q and V is stated as

$$q = kV^n \tag{5}$$

where n and k are constants (see Hall, 1968a). Substitution of eq 5 into eq 4 and integration lead to the recession equations

$$q = q_0\,(1 + t/t^*)^{\frac{n}{1-n}} \qquad n \neq 1 \tag{6}$$

$$q = q_0\, e^{-t/t^*} \qquad n = 1 \tag{7}$$

Table XIV. Response time T_L and antecedent discharge q_i, for 27 storms, Glenn Creek watershed.

Storm	T_L *(hr)*	q_i *(ft³/sec)*
24 Aug 65	4	0.21
27 Aug 65	1	0.15
5 Sept 65	1	0.13
3 June 66	0	0.39
6 June 66	0	0.25
11 June 66	6	0.08
20 June 66	0	0.48
21 June 66	4	0.48
22 June 66	1	0.56
26 June 66	3	0.16
26 June 66	0	0.16
30 June 66	1	0.06
3 July 66	0	0.02
6 July 66	2	0.01
9 July 66	2	0.03
16 July 66	5	0.01
17 July 66	3	0
18 July 66	1	0
7 July 67	1	0.02
8 July 67	0	0.07
9 July 67	0	0.11
10 July 67	0	0.09
19 July 67	14	0
21 July 67	11 (1)*	0.19
23 July 67	6	0.39
28 July 67	16 (7)*	2.14
8 Aug 67	0	0.06

* Values in parentheses, the *true* response times, used in correlation.

where q_0 is discharge at time $t = 0$ and t^* is a recession constant. Equation 7 is a simple exponential decay, which is most commonly used as an approximation of streamflow recessions. In eq 6, t^* is a function of n and k, and in eq 7, $t^* = 1/k$.

As pointed out by Hall (1968a), there is no analytical way of determining the value of n in eq 5 by examining actual recession curves. He has therefore suggested (Hall, 1968b) a matching-curve method for comparing the actual curves with those that would result from various values of n.

To determine the parameters of recession curves, one selects time periods during which there is no precipitation and during which flow from only one storm is occurring. This latter condition can almost never be fulfilled with certainty, but in the present case it was approximated by selecting such recessions from the hydrograph separations previously made for the measurement of total storm runoff. To provide a larger sample, periods when part of the total discharge included runoff from a previous storm were included when this part was less than 5% of the total flow.

When Hall's (1968b) matching-curve method was used (which involves log-log plots of q/q_0 vs t) for the 12 recessions selected, the actual curves fit the exponential curve ($n = 1$) at least as well as any other, justifying the use of this simple relation. To determine t^* for each period, a simple correlation of $\ln q$ vs t was carried out; with the regression coefficient (slope) taken as equal to k (= $1/t^*$). These results are summarized in Table XV and the data used in the computations are listed in Appendix D. Four-hourly values of discharge were used except for the 1964 data, when approximately daily values were all that were available. The values of t^* for the 12 recessions range from 19.6 hours to 76.9 hours, and average 39.2 hours.

Table XV. Recession constants for 12 recessions, Glenn Creek watershed.

Recession period	k (hr^{-1})	t^* (hr)	Correlation coefficient*
1964			
11 June 1015 - 16 June 0945	0.022	45.5	0.971
25 June 1010 - 28 June 1425	0.028	35.7	0.991
12 July 1630 - 17 July 1400	0.027	37.0	0.992
16 Aug 2000 - 19 Aug 1115	0.023	43.5	0.993
1966			
18 June 1200 - 20 June 0800	0.046	21.7	0.987
3 July 1200 - 5 July 2400	0.018	55.6	0.878
7 July 1400 - 9 July 1400	0.013	76.9	0.949
13 July 1100 - 16 July 1900	0.023	43.5	0.939
1967			
10 July 1300 - 14 July 1700	0.035	28.6	0.978
22 July 1100 - 23 July 2300	0.043	23.3	0.996
25 July 0700 - 26 July 0700	0.051	19.6	0.997
10 Aug 1000 - 11 Aug 1400	0.025	40.0	0.999

* For linear relationship between the natural logarithm of discharge and time.

It is of interest to compare these values with the values given by Holtan and Overton (1963, Tab 1) for 40 streams in the conterminuous United States. Their data are plotted in Figure 34 to show t^* as a function of drainage area. Figure 35 is also taken from Holtan and Overton (1963, Fig. 11), and shows t^* as a function of drainage area within 4 drainage basins. Compared with these data, it is clear that data from Glenn Creek show an exceedingly large recession constant for a basin of its size; basins of 0.7-square mile area shown in Figure 35 have a constant of less than 1 hour. Stated another way, the drainage of water from Glenn Creek following a rain occurs much more slowly than in the basins studied by Holtan and Overton (1963).

This characteristic of the recessions of Glenn Creek, along with the wide range of recession constant values (19.6 to 76.9 hours), invites further inquiry. Equation 4 assumes that water leaves the basin only as stream flow q. In reality, there is generally also a loss of water by evapotranspiration; this loss has apparently been universally ignored in the derivation of recession equations. However, if a constant term for evapotranspiration rate E is included, eq 4 becomes

$$\frac{dV}{dt} + q + E = 0 . \tag{8}$$

Assuming that eq 5 holds with $n = 1$, eq 8 leads upon integration to

$$q = (q_0 + E)\, e^{-t/t^*} - E. \tag{9}$$

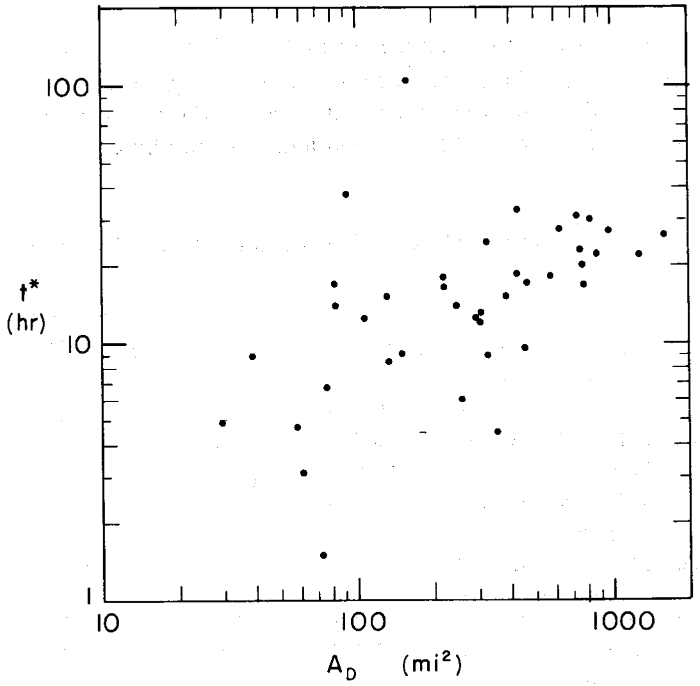


Figure 34. Recession constant t* vs drainage area A_D for 40 streams in the conterminous United States (data from Holtan and Overton, 1963).

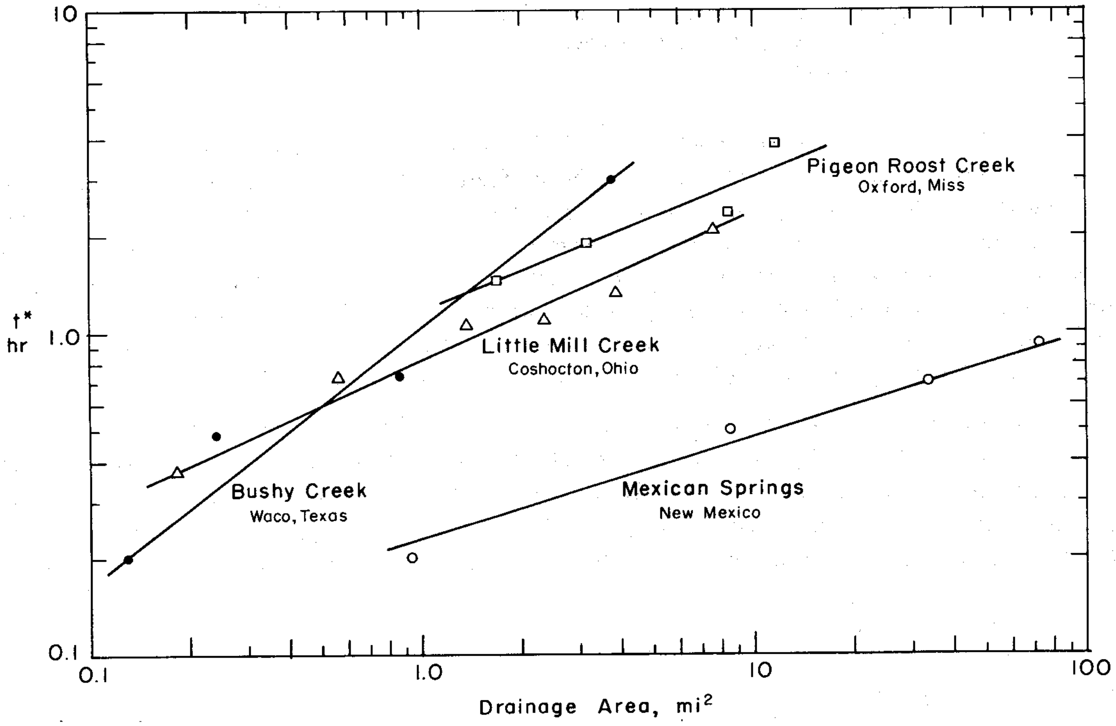


Figure 35. Recession constant t* vs drainage area A_D, within 4 drainage basins (from Holtan and Overton, 1963, Fig. 11).

To determine the effect of the evapotranspiration term on the recession, some typical values of q_0, E, and t^* can be used in eq 9. With $q_0 = 2$ ft³/sec, and t^* taken as the largest value observed at Glenn Creek (77 hours), values of E ranging from 0 to 0.01 in./hr (the largest average daily pan evaporation rate measured during a recession) were tried; the results are shown in Figure 36.

It is clear from eq 9 and Figure 36 that the effect of evapotranspiration is to steepen the recessions and to cause them to deviate from a simple exponential decay. However, in their early stages, each of the recessions shown might reasonably be modeled as simple exponentials. It is of interest to get an indication of how the apparent simple exponential decay constant (the value of t^* that most closely fits the early stages of the recessions) varies with evapotranspiration rate. This is done by comparing the time required for the discharge to decrease to 1/e of its original value for each of the curves of Figure 36. (This time equals t^* for a simple exponential decay.) The results are given in Table XVI and Figure 37.

The model data suggest that: 1) there is a strong effect of evapotranspiration rate on apparent recession constant; 2) a fourfold variation in the recession constant of Glenn Creek could be accounted for by differences in evapotranspiration rate; and 3) the recession constant is approximately a negative exponential function of evapotranspiration rate.

It is possible to compare actual recession constants with an evapotranspiration index for Glenn Creek. The index having the closest correspondence to actual evapotranspiration rate is undoubtedly the rate of evaporation from the Class-A pan at the weir site. Unfortunately, these data are available for only 7 of the 12 recessions listed in Table XV. Table XVII lists the recession constants, average pan evaporation rates, and periods of time for which they apply for the 7 recessions; the relation is presented graphically in Figure 38.

With the exception of one point, there is a distinct decrease of t^* as E increases, as predicted theoretically. If all 7 points are included, the linear correlation coefficient is -0.660 (significant between the 0.10 and 0.20 levels); without the point for 22-23 July 1967, the correlation coefficient becomes -0.959 (significant at less than the 0.01 level).

There does seem to be adequate justification for thinking that the pan evaporation rate during the actual recession period on the 22nd and 23rd was considerably higher than the average rate measured between the 22nd and 24th and presented in Table XVII. The hygrothermograph charts indicate very low humidities on the afternoon of the 23rd, following that day's evaporation measurement, rising abruptly to 100% at about 2100 hours. This condition of saturation persisted all day on the 24th, suggesting that essentially no evaporation took place on the morning of the 24th prior to the evaporation measurement. Thus, presumably, almost all the evaporation recorded between measurements on the 23rd and 24th took place on the 23rd during the recession period, so that the true rate was considerably greater than calculated from the daily pan measurements between the 22nd and 24th.

There thus seems to be good reason to believe that most of the variation in recession constants at Glenn Creek is due to variation in evapotranspiration rates during the same period. If the linear trend determined by the six points (not including the one for 22-23 July 1967) in Figure 38 is extrapolated to $E = 0$, $t^* \simeq 110$ hours. However, as noted in Figure 37, the theoretical curve is not linear, and presumably the true value of t^* at $E = 0$ is much higher than this.

The apparent fact that the variation in recession constants at Glenn Creek is largely due to variation in evapotranspiration suggests that the very long recessions there as compared with recessions in temperate regions may be at least in part due to the considerably lower evapotranspiration rates in central Alaska.

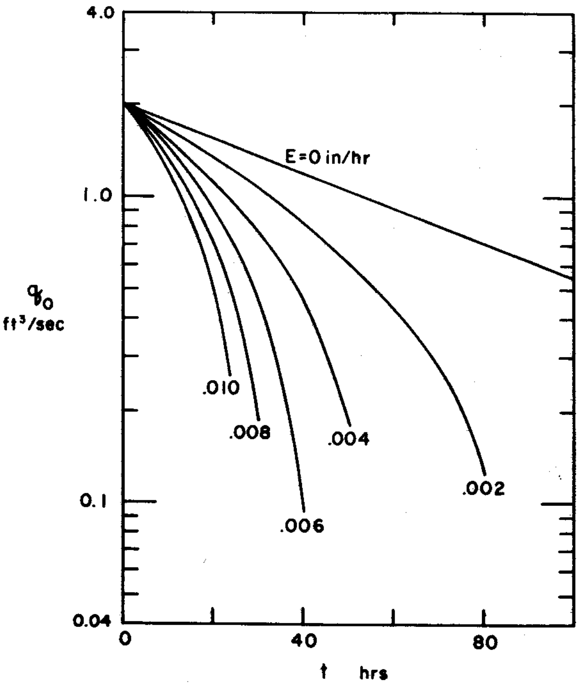


Figure 36. Theoretical effects of evapotranspiration rate E on recession $t^* = 77$ hr, $q_0 = 2$ ft³/sec.

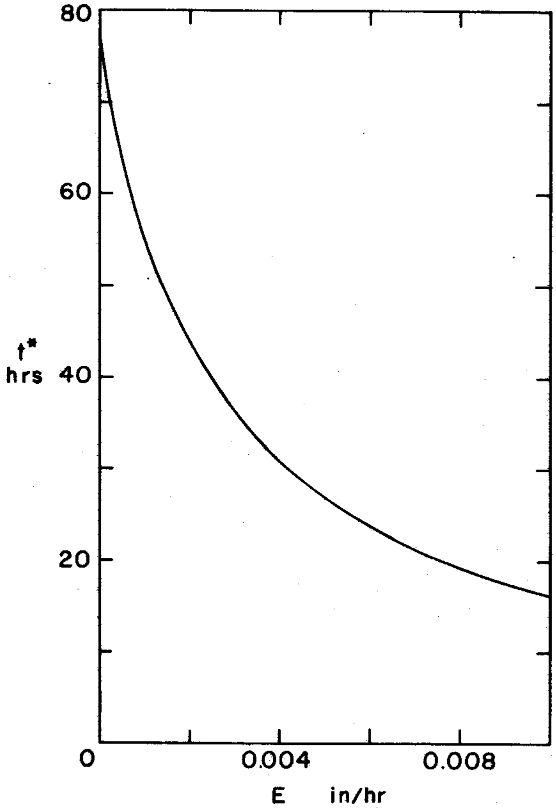


Figure 37. Theoretical effects of evapotranspiration rate E on apparent recession constant t^*, $q_0 = 2$ ft³/sec.

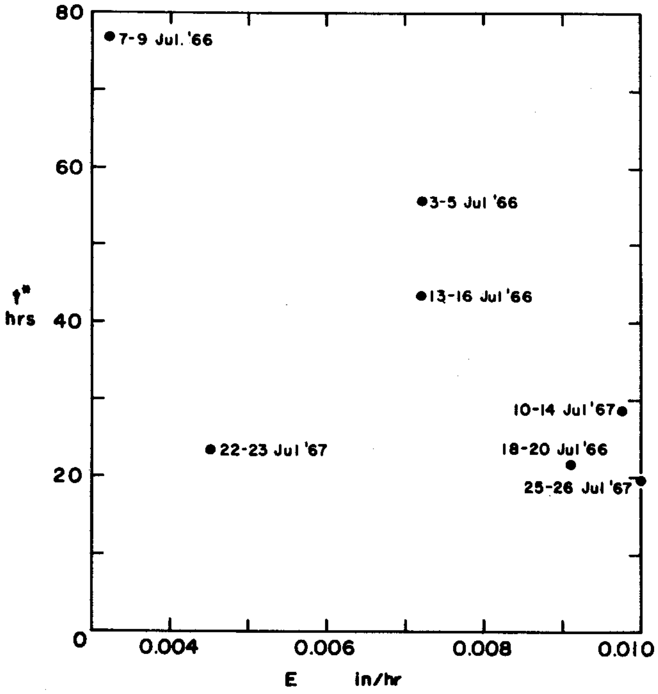


Figure 38. Measured recession constants t^* for 7 storms vs average class-A pan evaporation rate E at Glenn Creek.

Table XVI. Relation between evaporation rate *E* and apparent recession constant *t for model recessions.**

E (in./hr)	*t* (hr)*
0	77
0.002	44
0.004	31
0.006	24
0.008	19
0.01	16

Table XVII. Recession constants *t and average pan evaporation rates *E* for 7 recessions of Glenn Creek.**

Recession period	*t* (hr)*	*Evaporation period*	*E (in./hr)*
		1966	
18 June 1200 - 20 June 0800	21.7	18 June 1125 - 20 June 1600	0.0091
3 July 1200 - 5 July 2400	55.6	3 July 1130 - 6 July 1000	0.0072
7 July 1400 - 9 July 1400	76.9	7 July 0930 - 10 July 1240	0.0032
13 July 1100 - 16 July 1900	43.5	13 July 1250 - 16 July 1245	0.0072
		1967	
10 July 1300 - 14 July 1700	28.6	10 July 1435 - 14 July 1000	0.0098
22 July 1100 - 23 July 2300	23.3	22 July 1000 - 24 July 1030	0.0045
25 July 0700 - 26 July 0700	19.6	25 July 0910 - 26 July 1010	0.0100

Sources of streamflow

Introduction. Figure 39 indicates the possible routes by which water can reach a stream channel in a typical temperate watershed in the absence of snow: overland flow, interflow (unsaturated flow), groundwater flow, flow from bank storage, and direct channel precipitation. Figure 40 is a diagram of a typical midbasin cross section of Glenn Creek watershed, showing relations among ground cover, ground surface, water table, permafrost, and bedrock (compare profile F, Fig. 18). Consideration of the previous discussions of watershed geology, soils, permafrost, climate, and vegetation allows a preliminary evaluation of the routes by which water can travel to Glenn Creek.

As noted earlier, the possibility of overland flow from the moss-covered north-facing side of the watershed can be immediately eliminated on the basis of the high permeability of the moss (App. B). Similarly, overland flow from the duff-covered slopes seems at best very rare when the low rainfall intensities and high permeabilities of such areas are considered. However, the area of tussocks and bare ground with a high water table, which is widespread in the valley bottom, must be considered a potential source of overland flow.

Groundwater flow must be considered a possible source of streamflow in the basin. There were many periods during this study when streamflow was sustained for many days following a rainstorm, clearly indicating a delayed source. The most striking such period occurred in September and October 1964 (Fig. 41); other long periods of flow unsustained by rainfall occurred from 1-17 June, 23 June - 6 July, and 18 July - 26 August 1966.

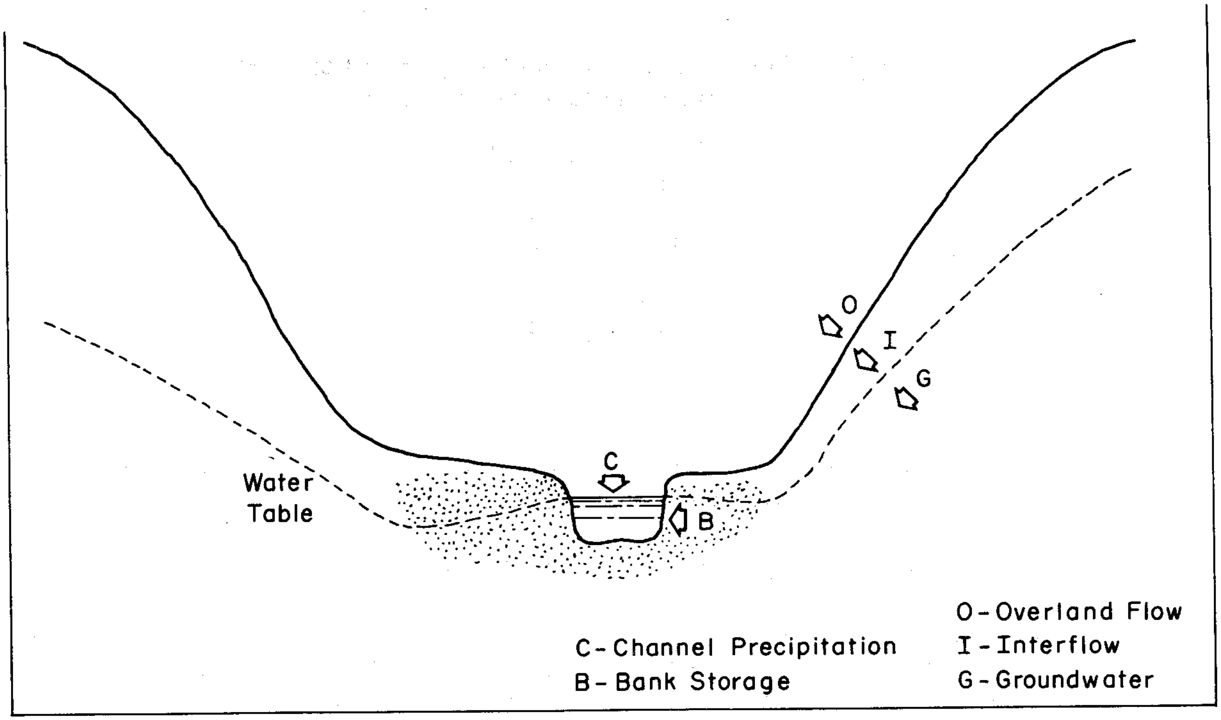


Figure 39. Schematic diagram showing routes of water to a stream in a typical temperate-zone watershed.

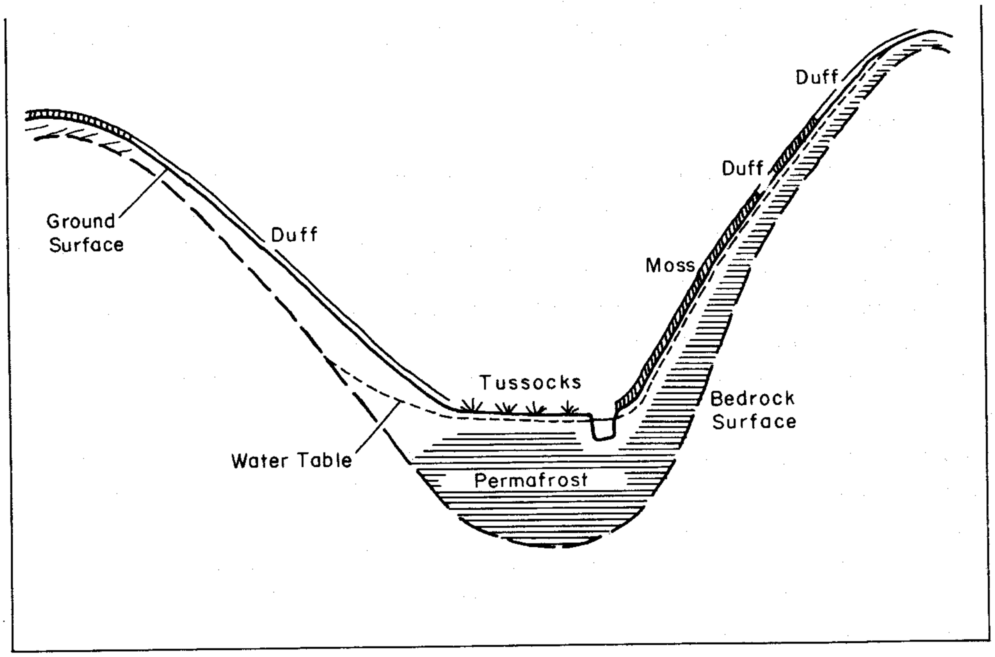
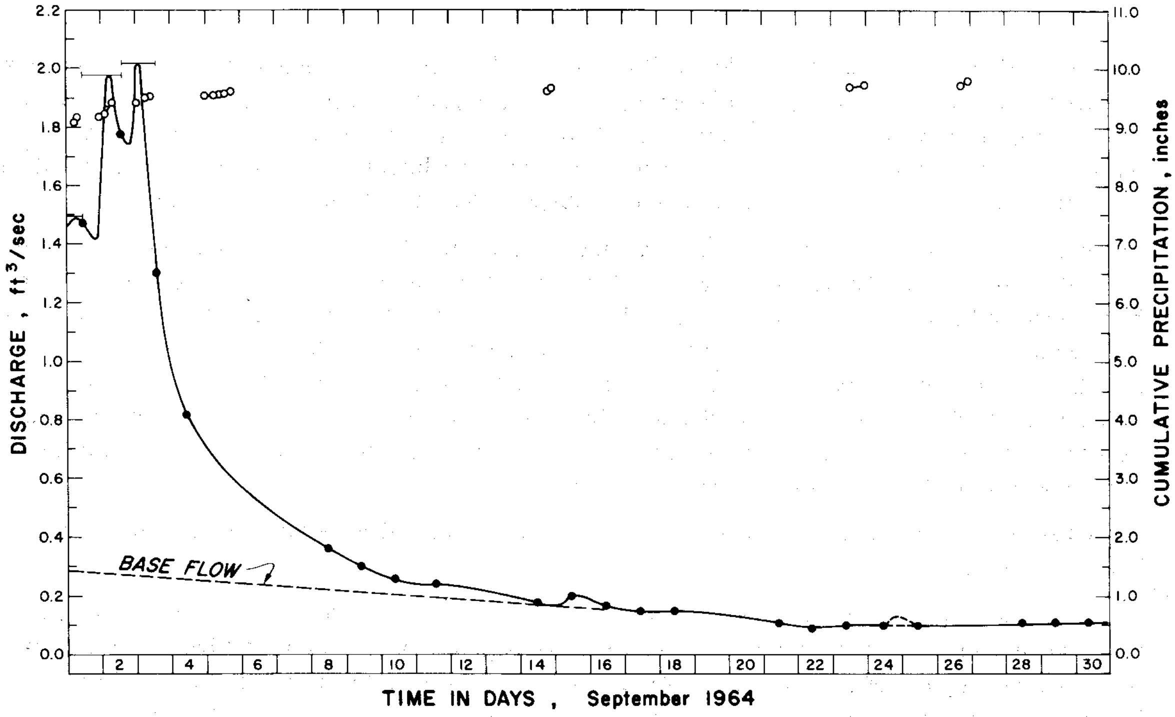


Figure 40. Schematic diagram showing ground cover, water table, permafrost, and bedrock in a typical cross section of Glenn Creek watershed.



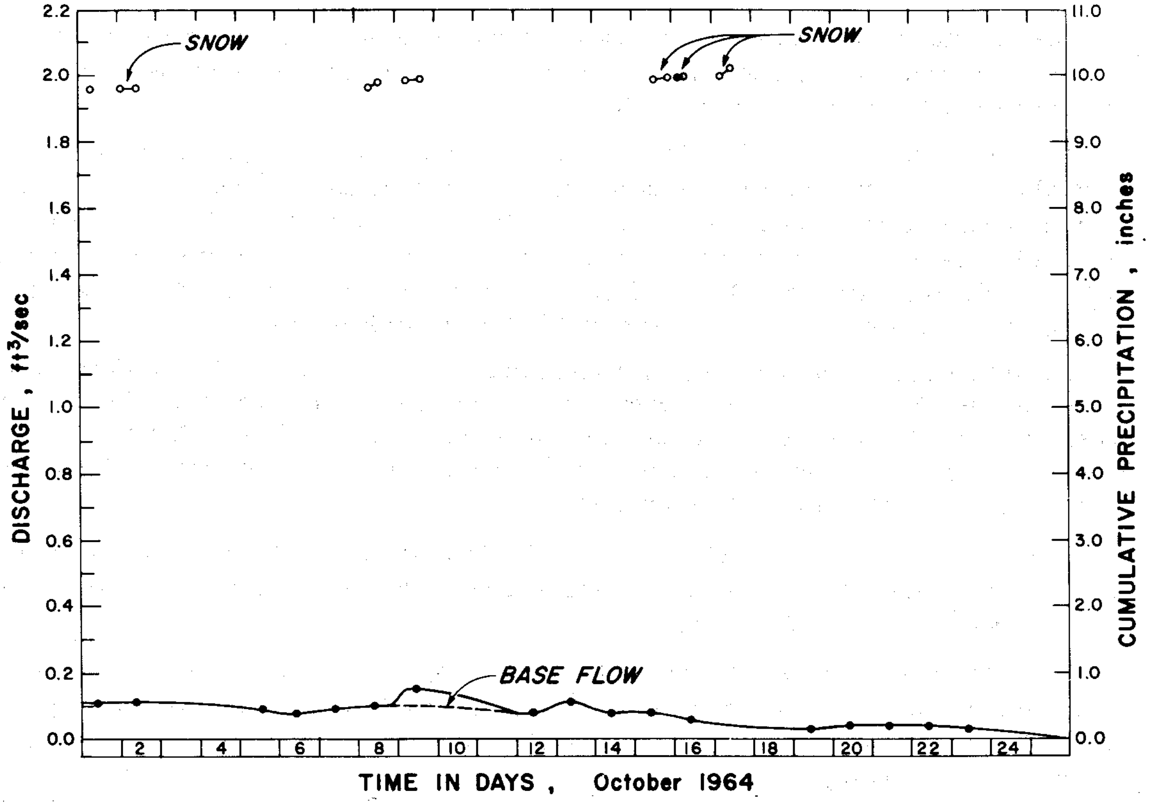


Figure 41. Glenn Creek hydrograph and cumulative precipitation, September-October 1964.

Bank storage is streamflow which infiltrates into channel banks when stream level rises above the local groundwater table; this water returns to the stream by seepage when stream level falls. This phenomenon is generally characteristic of streams with significant deposits of relatively coarse-grained alluvium in their banks. Since the banks of Glenn Creek consist of silts and clays (Goldstream silt loam), there is little reason to suspect that bank storage is significant.

It is conceivable without more detailed analysis that interflow is at least part of the source of the delayed flow in the basin. However, if present, this must arise in the mosses on the north-facing side of the basin. This statement is based on the reasoning that, since in the valley bottom area, which lies to the north of the stream, the top of permafrost is within a few feet of the ground surface, and the water table is at or very close to the surface, there is little opportunity for unsaturated flow to occur from that side. Similarly, there is little opportunity for unsaturated flow to occur in the wet, largely frozen soils on the south flanks of the valley. The possibility of interflow through the mosses is examined later.

Channel precipitation is a definite source of streamflow, and its significance is considered in the following section.

Thus, because of the conditions imposed by the physical characteristics of the watershed, overland flow outside the valley bottom, interflow through the basin soils, and bank storage can be considered highly improbable as significant sources of streamflow to Glenn Creek. Delayed flow from swamps and ponds is also eliminated, since there are no swamps in the watershed, and the only pond noted is a small one, about 20 ft in diameter near the basin outlet.

Overland flow from the valley bottom, interflow through the mosses, and groundwater flow are possible streamflow sources, and channel precipitation a definite source. These possibilities are examined in detail in subsequent sections.

Channel precipitation. If channel precipitation is the dominant source of streamflow, several streamflow characteristics should be evident: 1) the ratio runoff/precipitation should not exceed the ratio stream surface area/drainage area; 2) the duration of streamflow following a storm should not greatly exceed the length of time it takes a parcel of water to travel from the most distant part of the channel to the gage; 3) streamflow should begin to rise almost immediately after rain begins; and 4) rates of rise should rather closely reflect storm intensities.

As noted earlier, field examination and air-photo study reveal that the total length of channels in the basin is 9760 ft. The hydraulic-geometry measurements show that channel width at the outlet ranges from 2.5 to 4 ft over a discharge range of from 0.02 to 2.14 ft^3/sec. Multiplying the maximum width observed times channel length gives a stream surface area of 39,000 ft^2 or 0.0014 square mile. This area is 0.2% of the total basin area.

It is readily apparent from Table XI that runoff/rainfall ratios greatly exceeded this amount for all storms. Table XVIII shows that the volume of runoff due to channel precipitation does not exceed a small percentage of total runoff.

The data on the response of Glenn Creek to rainfall inputs show that the duration and rate of the hydrograph rise are closely controlled by the duration and intensity of rainfall. Although this might suggest the dominance of channel precipitation during a rise, some further calculations serve to obviate this possibility.

Assuming that all the rain that falls directly on the channel runs off, we can, as before, calculate the total volume of channel precipitation as the product of the total rainfall P and the stream surface area A_s. Following Linsley *et al.* (1949, p. 390-392), we can assume that the duration of rise for the runoff due to channel precipitation is equal to the storm duration T_s, and that the total *time base* of the hydrograph due to channel precipitation is equal to T_s plus the time of concentration T_c. Approximating the hydrograph due to channel precipitation by a triangle, and referring to Figure 42, we have

$$V = P\ A_s \tag{10}$$

$$V = \tfrac{1}{2}\ q_{pc}\ (T_s + T_c) \tag{11}$$

where V is volume of channel precipitation and q_{pc} is peak discharge from channel precipitation. Equating eq 10 and eq 11

$$P\ A_s = \tfrac{1}{2}\ q_{pc}\ (T_s + T_c)$$

$$\frac{2\ P\ A_s}{(T_s + T_c)} = q_{pc}\ . \tag{12}$$

We now can use values of P and T_s from some actual storms on Glenn Creek, and compare the estimated peak discharge from channel precipitation with the actual peak and the estimated runoff of channel precipitation with the total runoff. The calculations are shown in Table XVIII, where q_{pc} was computed by eq 12, taking $T_c = 2$ hr (see Fig. 29) and $A_s = 39{,}000$ ft^2 as previously determined. The total volume of channel precipitation runoff was found by eq 10. The data show that the peak flow due to channel precipitation is considerably less than the actual peak for a given storm, and suggest that the major portion of the runoff during the rise has a source other than channel precipitation. The typical situation, idealized from data of Table XVIII, is shown in Figure 43.

Thus, although it has been demonstrated that streamflow begins to rise shortly after rain begins (see Table XIII) and that streamflow rises reflect variations in precipitation intensity (see Fig. 32), the foregoing analysis has indicated that runoff due to channel precipitation makes up only a few percent of total runoff. Further, streamflow rises are apparently not dominated by channel precipitation. If estimates of times of concentration are at all accurate, then the long recessions of Glenn Creek must arise from some other source (see Fig. 43).

Overland flow. As noted earlier, the physical characteristics of Glenn Creek watershed obviate overland flow from all portions of the basin outside the valley bottom. However, a source of runoff which is quickly responsive to rainfall is indicated by the hydrograph characteristics. It has also been noted that although the proportion of rainfall running off is related to the antecedent discharge of the watershed (see Fig. 30), the speed of response of the stream to rainfall is not related to antecedent discharge.

These characteristics would seem to be explained by invoking the concept of a variable source area, as suggested by Betson (1964) and Hewlett and Hibbard (1966). Figure 44a shows that when the watershed is relatively dry the water table in the valley bottom is relatively low and intersects the ground surface in only a few places; these places are more frequent nearer the stream channel. Under wet conditions (Fig. 44b), the water table intersects the ground over a wider portion of the valley bottom. When rain begins, flow to the channel begins almost immediately under both conditions, if there is some degree of connection among the water-filled depressions and between the depressions and the channel. However, under wet conditions, more of the rain falls on water standing or flowing at the surface, and less is lost in wetting the ground and filling "detention storage."

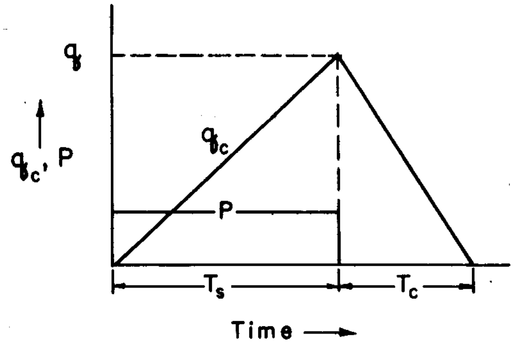


Figure 42. Diagram for calculation of peak discharge due to channel precipitation q_{pc}.

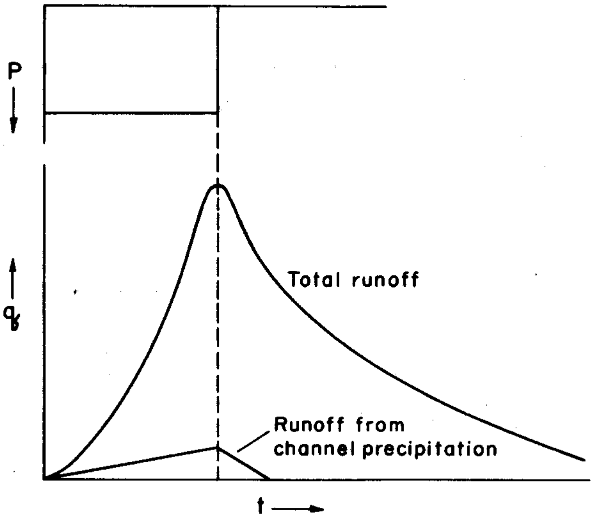
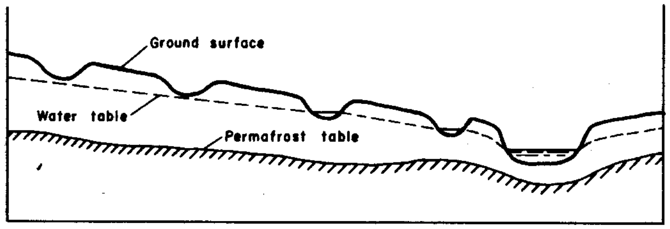
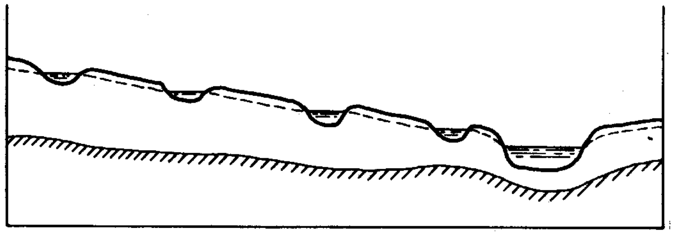


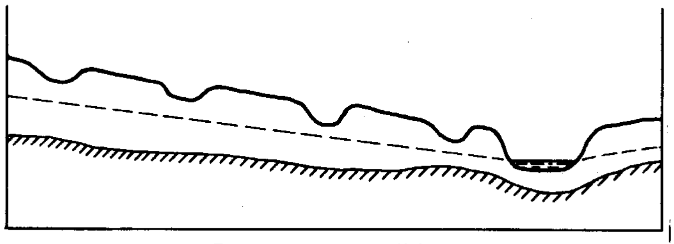
Figure 43. Typical relation between runoff due to channel precipitation and total runoff, Glenn Creek watershed.



a. Dry conditions



b. Wet conditions



c. Extremely dry conditions

Figure 44. Diagram illustrating the concept of variable source area in the valley bottom of Glenn Creek watershed.

Table XVIII. Comparison of streamflow caused by channel precipitation and total streamflow for 16 storms on Glenn Creek watershed.

Storm	P *(in.)**	P *(ft)**	T_s *(hr)*	T_s *(sec)*	q_{pc} *(ft³/sec)†*	q_p *(ft³/sec)***	q_{pc}/q_p *(%)*	V_c *(ft)††*	V_c/V *(%)****
5 Sept 65	0.80	0.067	24	86,400	0.055	1.98	2.8	2600	0.45
2 June 66	0.03	0.003	4	14,400	0.010	0.16	6.3	120	0.16
17 June 66	0.76	0.063	15	54,000	0.080	3.45	2.3	2460	0.56
20 June 66	0.32	0.027	4	14,400	0.100	0.41	24.4	1050	1.60
21 June 66	0.10	0.008	18	64,800	0.009	0.48	1.9	310	0.25
26 June 66	0.07	0.006	15	54,000	0.008	0.08	10.0	230	0.77
6 July 66	0.20	0.017	13	46,800	0.024	0.06	40.0	660	5.30
9 July 66	0.29	0.024	17	61,200	0.027	0.09	30.0	940	4.50
7 July 67	0.45	0.038	45	162,000	0.018	0.12	15.0	1480	4.30
19 July 67	0.22	0.018	36	129,600	0.010	0.19	5.3	700	1.70
21 July 67	0.65	0.054	16	57,600	0.065	1.47	4.4	2100	1.30
23 July 67	1.18	0.098	31	111,600	0.064	4.44	1.4	3820	0.74
26 July 67	0.21	0.017	18	64,800	0.018	1.23	1.5	660	0.26
28 July 67	0.04	0.003	2	7,200	0.016	0.15	10.7	120	0.24
30 July 67	0.22	0.018	50	180,000	0.007	0.20	3.5	700	0.59
8 Aug 67	1.01	0.084	32	115,200	0.054	1.57	3.4	3280	0.98

* Before peak.

† $q_{pc} = (2\,PA_s) / (T_s + T_c)$; $A_s = 39{,}000$ ft²; $T_c = 2$ hr = 7200 sec.

** Largest difference between measured (total) flow and flow estimated from previous storms.

†† $V_c = PA_s$.

*** Total volume of runoff from Table XI.

Thus, a higher percentage of rainfall runs off under wet conditions than under dry conditions, but streamflow responds rapidly to rainfall under both conditions, as observed in the data from Glenn Creek watershed.

Under extremely dry conditions (Fig. 44c) the groundwater table may be well below the ground surface virtually everywhere, and a long period of rain is required before the stream responds. Such a condition explains the long response time noted for the storm of 19 July 1967 (see Table XIII and Fig. 32d), and for other storms of that dry period.

Although this mode of runoff supply is referred to as *overland flow,* it is not infiltration-limited overland flow as originally described by Horton (1933) and discussed in most hydrology texts.

The volumetric importance of this source of runoff varies depending on antecedent conditions. However, the maximum amount of runoff that can be supplied in this way is equal to the rainfall over the valley-bottom area, i.e. that area covered by vegetation units 5 and 6 (see p. 30). Table VIII shows that together these make up 0.054 square mile, or 7.7% of the total basin area. Table XIX compares maximum possible valley-bottom runoff (overland flow) to total runoff for the 16 storms for which the latter could be accurately determined. For four of these storms, which occurred when the watershed was very dry, the maximum possible valley-bottom runoff exceeded the total runoff; this is consistent with the proposed valley-bottom runoff model. For the other 12 storms, the ratio varies from 0.189 to 0.574.

The ratio maximum possible overland flow/total runoff is not significantly related to antecedent discharge. However, this fact is not damaging to the proposed model, since 1) the values shown are for maximum possible overland flow, not actual overland flow; and 2) in any case, the amount of runoff from other sources may also vary with antecedent wetness, so that the ratio of overland flow to total runoff may not change markedly as antecedent conditions change.

Table XIX. Relation of maximum possible overland flow to total runoff for 16 storms, Glenn Creek watershed.

Runoff period	*Precipitation (in.)*	*Total runoff (ft^3)*	*Maximum possible overland flow (ft^3)*	*Overland flow total runoff*
9 June 64 - 28 June 64	1.69	422,000	212,000	0.502
22 July 64 - 6 July 64	0.86	188,000	108,000	0.574
10 July 64 - 21 July 64	0.58	323,000	72,700	0.225
28 July 64 - 10 Aug 64	0.88	292,000	110,000	0.377
3 Aug 64 - 23 Aug 64	1.41	468,000	177,000	0.378
15 Aug 64 - 25 Aug 64	0.36	238,000	45,100	0.189
30 Aug 64 - 25 Oct 64	1.49	940,000	187,000	0.199
5 Sept 65 - 4 Oct 65	1.00	583,000	125,000	0.214
17 June 66 - 26 June 66	0.85	441,000	107,000	0.243
6 July 66 - 13 July 66	0.22	12,400	*	
9 July 66 - 16 July 66	0.39	20,700	*	
7 July 67 - 14 July 67	0.69	34,600	*	
19 July 67 - 24 July 67	0.76	40,100	*	
21 July 67 - 27 July 67	0.70	165,000	87,800	0.532
23 July 67 - 30 July 67	1.19	515,000	149,000	0.289
8 Aug 67 - 17 Aug 67	1.05	333,000	132,000	0.396

* Exceeds total runoff.

The values in Table XIX indicate that valley-bottom runoff can be a significant portion of total runoff. Runoff from this area would be expected to quickly respond to rains, and to dominate the rising portion of the hydrograph, causing the close relationship between rise rate and intensity observed for many of the storms in the basin. Later in this report, an attempt is made to estimate the time distribution of overland flow.

Interflow and groundwater flow. Preceding discussions have shown that the recession portions of the hydrographs of Glenn Creek are greatly extended in time over those reported for even much larger watersheds in temperate areas, and that overland flow apparently does not account for much more than half the total runoff, except perhaps when conditions are extremely dry. Clearly, a significant delayed runoff source is indicated in the watershed. Melting snow, ponds, and swamps can be immediately eliminated as sources, since none of these features (except for one very small pond, which is not visibly connected to the stream network) are present in the watershed during the summer months.

One or more subsurface flow routes, saturated and/or unsaturated, would initially seem the most likely source of this delayed flow. The possibilities would seem to be: 1) the moss on the north-facing slopes; 2) the thin layer of unfrozen soils beneath the moss; and 3) the south-facing slopes via the valley bottom.

It is possible to construct a simple model of the drainage of a partially saturated sloping porous slab, and to compare its drainage characteristics with those observed in Glenn Creek. Referring to Figure 45, the permeable layer is initially saturated to a thickness h_0, and the assumption is made that the slope of the water surface S is constant in time and space and equal to the general ground slope. (Although not strictly true, this assumption appears reasonable in the present case; deviations of the water surface slope from the ground slope should be minor except locally near the downstream end.) Considering a recession period, when there is no recharge to the permeable layer from precipitation, and neglecting evapotranspiration, the continuity equation for a unit width of the aquifer is

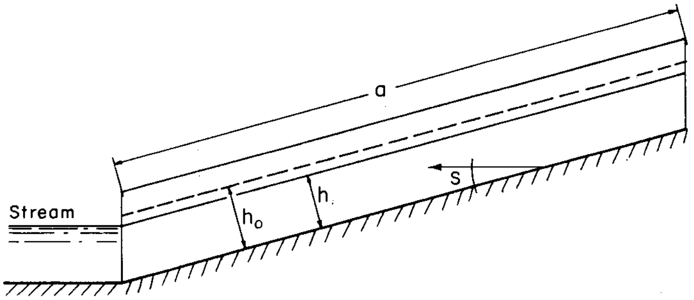


Figure 45. Sketch to define variables in analysis of drainage of a sloping porous slab. a = length of permeable layer.

$$q = -\frac{dV}{dt} \tag{13}$$

where q is flow rate per unit width, V is volume of water in storage in the aquifer, and t is time. Since the layer is assumed to be saturated, Darcy's law describes the flow rate:

$$q = khs \tag{14}$$

where k is permeability and h is depth above the impermeable layer. The volume of water in storage is

$$V = nha \tag{15}$$

where n is aquifer porosity and a is the length of the permeable layer (slope length). Substituting eq 15 in eq 13

$$q = -na\frac{dh}{dt} \tag{16}$$

and from eq 14

$$dh = \frac{1}{ks}\,dq\,. \tag{17}$$

Now 17 can be substituted into eq 16

$$q = -\frac{na}{ks}\,\frac{dq}{dt}\,. \tag{18}$$

Separating variables and integrating, and noting that $q = q_0$ when $t = 0$, gives

$$q = q_0\,e^{-\frac{ks}{na}t}\,. \tag{19}$$

Thus, if streamflow during a recession is maintained by saturated flow through the mosses

$$t^* = \frac{na}{ks}\,. \tag{20}$$

Appendix B presents results of a study of the water-holding and water-transmitting properties of the moss cover which is present over about 51% of the watershed. The permeability of the moss decreases with head to a constant value of about 0.22 cm/sec at heads below about 0.04. However, the average slope of the basin is 0.184, and to be consistent with the simplified model just described, the permeability at this gradient, 0.14 cm/sec (16.5 ft/hr), should be used. The porosity of the moss was measured at about 0.95, and the average slope length is 2400 ft. Thus,

$a = 2400$ ft

$k = 16.5$ ft/hr

$n = 0.95$

$s = 0.184$

$$t^* = \frac{(0.95)\,(2400)}{(16.5)\,(0.184)} = 750 \text{ hr.}$$

This value is about 10 times larger than the largest observed recession constant, and about 20 times greater than the average value of t^*. It is possible that this approximately represents the *true* recession constant for the basin when evapotranspiration is zero, as discussed earlier. Figure 46 is a plot of three pairs of recessions, each pair representing a different value of q_0. For each q_0, two recessions are shown, both using $t^* = 750$, with one for conditions of no evapotranspiration and the other for an evapotranspiration rate equal to the highest observed pan evaporation rate, 0.01 in./hr. It can be seen that the *apparent* recession constant may be reduced to approximately one-twentieth of the true value by evapotranspiration, at least when initial discharges are low.

Thus, saturated Darcy-type flow through the mosses may be at least part of the source of the delayed flow to Glenn Creek. Flows through much less permeable and vertically thin (due to permafrost) silts beneath the moss, and through the slowly permeable and gently sloping valley-bottom silts, would be much slower than in the moss, and are probably insignificant contributors of streamflow. Unsaturated flow in all mediums would, of course, be even slower.

A faster mode of flow which may contribute significantly to delayed flow involves channels beneath the moss. There is some field evidence indicating that such flow routes exist, at least after heavy rains. Figures 47 and 48 show a silt deposit in a moose trail, observed about 2 weeks after the extremely heavy rains of 8-12 August 1967. This silt deposit extended for some 50 ft, and, at its head, water was running out of the moss. A hole punched through the silt with a steel probe brought water under artesian pressure bubbling to the surface. Clearly water had been flowing sufficiently fast beneath the moss to erode the underlying silt, which was deposited after the water broke out of the moss and flowed along the moose trail. If such submoss channels are active during most rains, they may be responsible for the recession characteristics of Glenn Creek. In addition, some groundwater must enter the stream through the soils on the south-facing slopes and the valley bottom, and through the thin seasonally-thawed zone on the north-facing slopes.

In regard to this latter source, the possibility that the thawing of the seasonally frozen soil releases water which contributes to streamflow was raised on the basis of 1964 streamflow data (Dingman, 1966b). This process has also been suggested as being responsible for the streamflow patterns of two rivers in the discontinuous permafrost zone of Canada (Sommer and Spence, 1968, p. 63-64). These writers suggested that

> "During the spring snowmelt period the temperatures in the active layer of the ground remain below freezing. This layer is also relatively dry and receptive to moisture as a result of water deficiency in the late summer and fall of the preceding year. Therefore, the snowmelt waters are able to pass into the ground where they become refrozen in the active layer. This refrozen moisture is then slowly released in summer as the active layer thaws. Thus, the maximum streamflow would be expected during the summer corresponding to the warmest period." *

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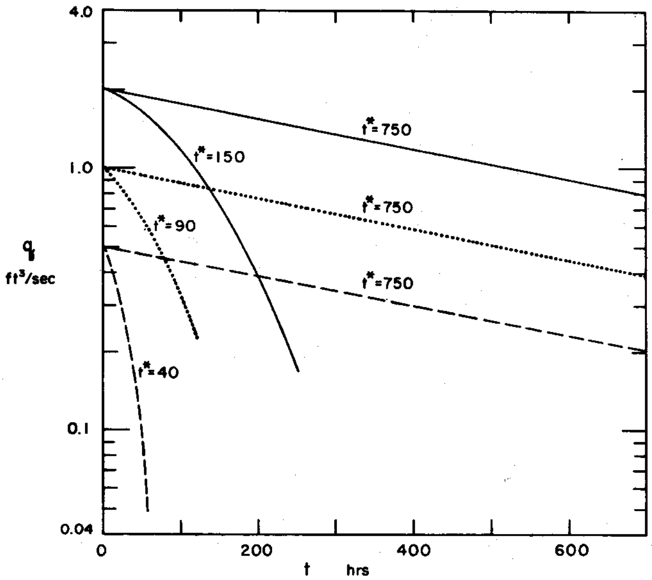


Figure 46. Theoretical effects of evaporation rate of 0.01 in./hr on apparent recession constant when the true recession constant is 750 hr, $q_0 = 2, 1, 0.5$ *ft*^{3}*/sec.*

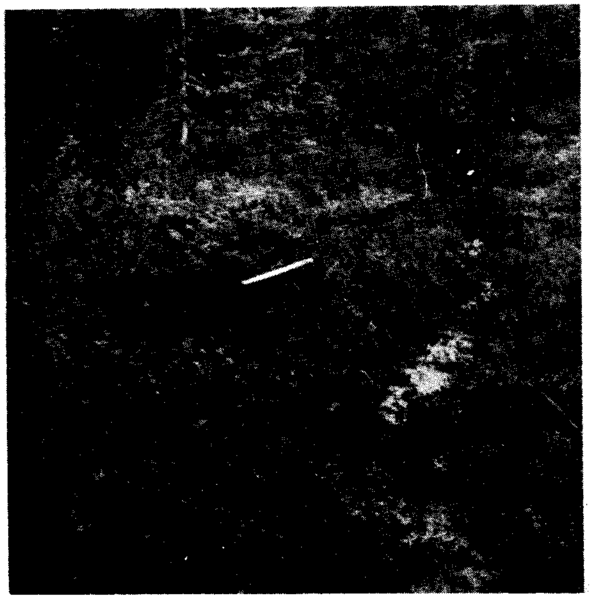


Figure 47. Upslope view of silt-covered moose trail on north-facing slope of Glenn Creek watershed, 27 August 1968.

The correspondence of maximum monthly streamflow with maximum monthly temperatures, rather than with maximum monthly precipitation or spring snowmelt, was noted for the Yellow Knife and Snare Rivers, whose basins contain widespread permafrost. The Hay River, which drains an area with only scattered permafrost, did not display this correspondence.

The explanation offered by Sommer and Spence does not seem compelling. In subarctic areas with a similar precipitation regime, but where permafrost is absent, extensive seasonal freezing occurs. It is not clear why the same phenomenon of infiltration and freezing of snowmelt waters,

Figure 48. Close-up view of cross section of silt deposit in Moose Trail, north-facing slope of Glenn Creek watershed, 27 August 1968; graduations on probe are 1 cm apart.

with later release as thawing progresses, would not occur in such areas. If transpiration were higher in nonpermafrost areas, which is presumably the case, it is possible that this difference in water loss could account for the difference in streamflow regimes, as observed by Sommer and Spence.

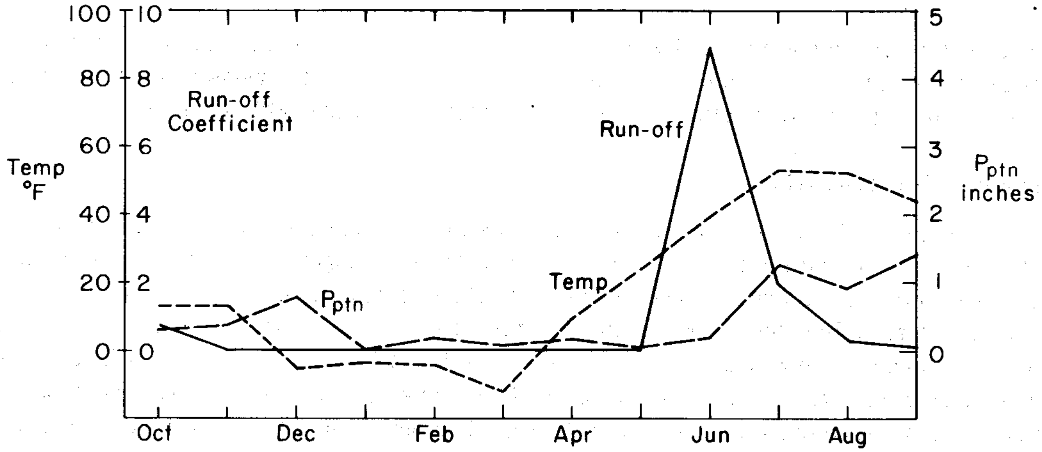
The flow of Glenn Creek, with about 60% of its drainage area underlain by permafrost, should reflect the water contribution of thawing soil if it exists. The very low flows of several mid-summer periods, particularly July and August 1966, and July 1967, strongly suggest that the process is not operative.

Plots of monthly runoff coefficient (mean monthly runoff divided by mean annual runoff), mean monthly temperature, and mean monthly precipitation for three rivers in the permafrost region of Alaska were made for comparison with the results of Sommer and Spence (1968). Figure 49 shows these data for the rivers listed in Table XX.

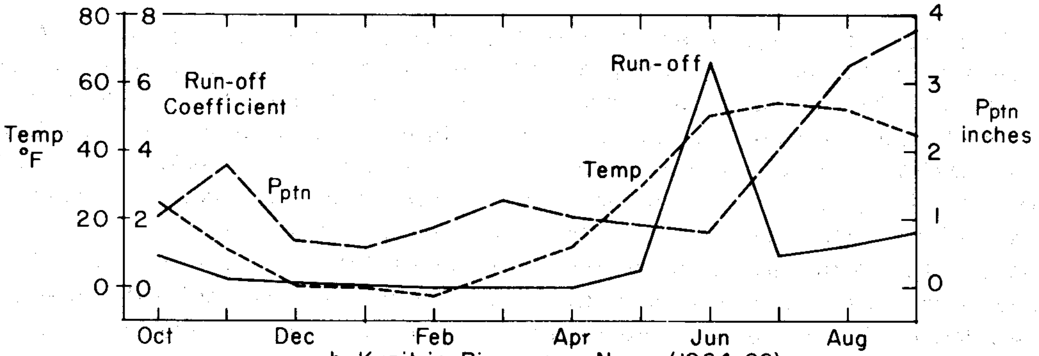
Table XX. Streamflow and precipitation records examined for possible thaw-water contributions.

River	*Drainage area (mi^2)*	*Weather station*	*Water years of record*
June Creek near Kotzebue	10.9	Kotzebue	1966
Kuzitrin River near Nome	1,720	Moses Point	1964-1966
Koyukuk River near Hughes	18,700	Hughes	1961-1966

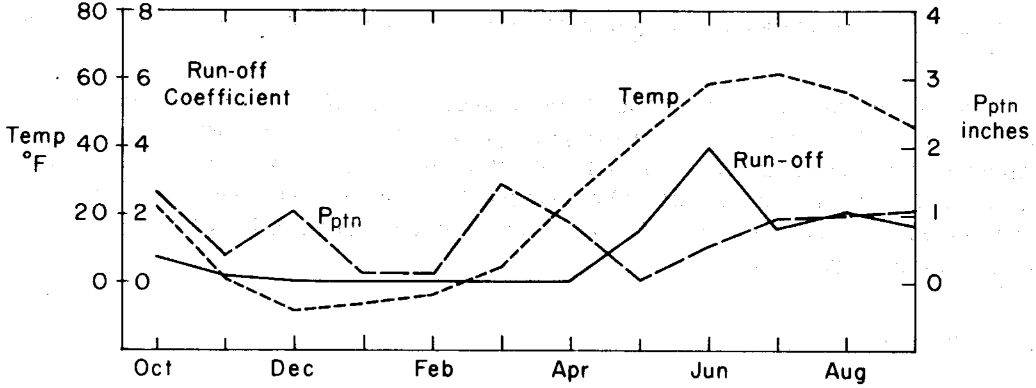
For June Creek, the peak runoff in June is clearly due to snowmelt; precipitation and temperature are both high in July, the second highest runoff month, and runoff drops sharply in August, as precipitation decreases but temperature remains high. Thus high temperatures are associated with decreasing discharge, reflecting the effect of evapotranspiration rather than flow contribution from melting.



a. June Creek near Kotzebue (1966)



b. Kuzitrin River near Nome (1964-66)



c. Koyukuk River at Hughes (1961-66)

Figure 49. ***Precipitation, temperature, and monthly runoff coefficient for 3 streams draining permafrost areas in Alaska.***

In the Kuzitrin River, the peak runoff again occurs in June, due to snowmelt. Streamflow increases from July to September, while precipitation is increasing and temperature is decreasing. Again, no effect of thaw release of water is evident. Similarly, the data for the Koyukuk River show no correspondence between high temperature and high runoff in mid- and late-summer.

Summary. Consideration of the physical characteristics of Glenn Creek watershed and its streamflow has led to the conclusion that there are three sources of runoff: 1) channel precipitation, which is a very minor component of flow at the basin outlet during and immediately after a storm; 2) overland flow that rises from those areas of the valley bottom where the groundwater table is at the ground surface, and which makes up a maximum of about 50% of total runoff; and 3) delayed subsurface flow, which may be from channels beneath the moss and from seepage through the moss and soils of the watershed, and which makes up at least 50% of runoff.

Attempts at flow separation

Introduction. The characteristics of two major and one minor source of runoff have been identified, and some information on their relative volumetric importance gained in preceding sections. Some conclusions on the time distribution of these components of runoff have also been drawn: 1) the time distribution of channel precipitation has been specified (see Fig. 43); 2) overland flow dominates the rising hydrograph; and 3) delayed flow dominates the hydrograph recession. In this section, an attempt will be made to elucidate more specifically the time distributions of the two major runoff sources, i.e. to separate the total hydrograph into hydrographs of the two major components.

The problem of hydrograph separation has been approached from many directions. Arbitrary separations (see Linsley *et al.*, 1949, p. 399-400) are commonly used, and in some cases assumptions are made that each runoff component has its own recession characteristics, which can be identified by examination of hydrographs. A third method, which has been recently employed, uses stream chemistry as a basis for separation (Rainwater and Guy, 1961; Voronkov, 1963; Kunkle, 1965; Toler, 1965; Pinder and Jones, 1969). Arbitrary separations will provide no insight to the behavior of the runoff process in Glenn Creek. Hall (1968a) has pointed out the difficulties of using recession hydrographs to identify runoff components; breaks in slope on a plot of log discharge vs time may arise because of drainage of one nonlinear reservoir rather than from changes in the relative importance of two or more linear reservoirs during the recession, as is commonly assumed. In any case, the data for Glenn Creek indicate that where recession trends can be identified with some confidence, they can be well-represented by a single recession constant (see Table XV).

Chemical separation. Since the analysis of vegetation, soils, and hydrologic data of Glenn Creek has pointed to the existence of two principal runoff sources, a simple chemical model was examined to see if it could provide insight into flow separation. Total runoff q_t was assumed to be made up of two components, valley-bottom runoff q_v and delayed groundwater runoff q_g. Each component was assumed to have a characteristic constant concentration of dissolved solids, c_v and c_g, respectively. Thus

$$q_t = q_v + q_g \tag{21}$$

$$c_t q_t = q_v c_v + q_g c_g . \tag{22}$$

Equations 21 and 22 can be combined to give

$$q_g = \frac{c_t - c_v}{c_q - c_v} q_t . \tag{23}$$

The results of 35 determinations of total dissolved solids for Glenn Creek (Fig. 50) show the empirical relationship

$$c_t = 54 \; q_t^{-0.15} \tag{24}$$

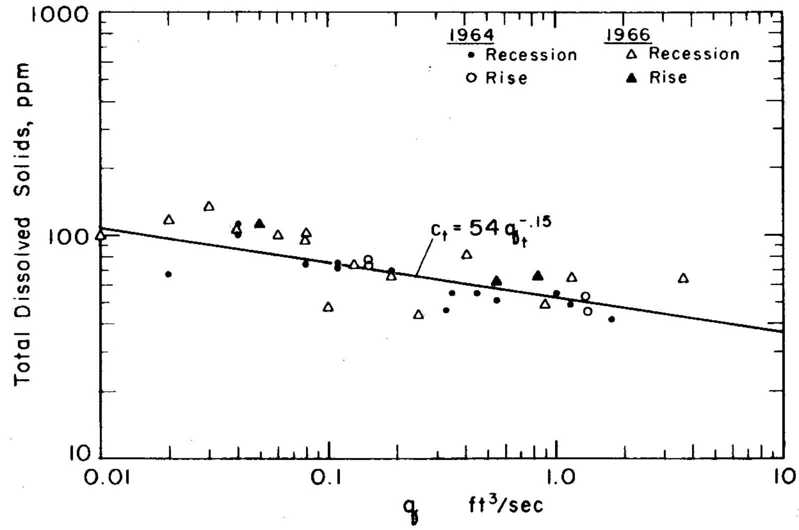


Figure 50. Relation between total dissolved solids and streamflow, Glenn Creek.

where c_t is in parts per million and q_t is in ft³/sec. Substituting eq 24 into eq 23, we have

$$q_g = \frac{54\ q_t^{-0.15} - c_v}{c_q - c_v}\ q_t\ . \tag{25}$$

Following Pinder and Jones (1969, p. 441), the highest dissolved-solids concentration measured in streamflow (135 ppm, 17 and 24 July 1966) was used as an estimate of c_g. (This value corresponded closely to concentrations measured in stagnant pools on the creek bed on 13 and 27 August: 132 and 133 ppm, respectively.)

For a value of c_v, concentrations at the higher flows were examined. The lowest concentration measured was 42 ppm. Extrapolation of the curve relating q_t and c_t (Fig. 50) indicates that $c_t = 40$ ppm when $q_t \simeq 6$ ft³/sec. Thus, c_v was estimated at 40 ppm, and eq 25 becomes

$$q_g = \frac{54\ q_t^{-0.15} - 40}{90}\ q_t\ . \tag{26}$$

Using this, q_g can be found as a function of q_t.

The results are shown in Table XXI and Figure 51; they indicate that delayed runoff makes up more than 50% of total runoff only at very low discharges. Since we have seen that the minimum possible delayed runoff is almost always more than 50% of total runoff (Table XIX), the simple two-component chemical model does not seem to adequately separate the flow components. Using a smaller value for c_v improves the situation somewhat, but even with $c_v = 10$, the proportion of valley-bottom runoff is too high.

This discrepancy suggests either that 1) the two-component hypothesis is faulty; or 2) there are two components of runoff, but one or both do not have a constant concentration of dissolved solids. Components other than the valley bottom and delayed (and very minor) channel precipitation have been eliminated in the previous discussion. Since the two together would seem to account, at least qualitatively, for the dominant features of the hydrograph, fast response and long recessions, it seems most reasonable to assume that the chemical composition of one or both major components is not constant; one or both may increase in concentration as functions of time while moving toward the stream.

Table XXI. Groundwater flow as a function of total flow, as calculated from eq 26.

q_t (ft^3/sec)	q_g (ft^3/sec)	q_g/q_t
0.001	0.0012	1.2
0.002	0.0020	1.0
0.005	0.0042	0.84
0.01	0.0071	0.71
0.02	0.012	0.60
0.05	0.023	0.47
0.1	0.038	0.38
0.2	0.061	0.30
0.5	0.10	0.20
1	0.15	0.15
2	0.18	0.090
5	0.13	0.025
10	−0.19	

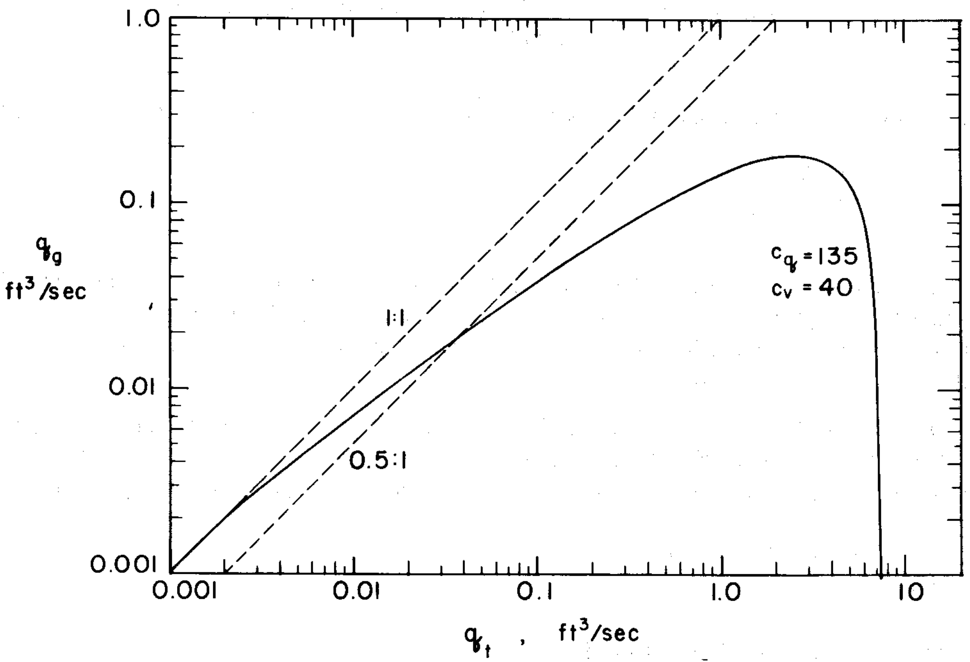


Figure 51. Relation between groundwater flow q_g and total flow q_t, as calculated from eq 26, for $c_v = 40$.

Volumetric separation. Unfortunately, aside from chemical separation, there are no other means of determining the time distributions of the two runoff components that are not completely arbitrary. However, it is possible to estimate the time distribution of the two components if it is assumed that 1) the volume of total runoff from a storm is known (based on the method used previously, where the exponential recession trend of one storm is extended beyond where runoff from the succeeding storm begins); 2) the volume of valley-bottom runoff is known (calculated as 100% runoff of rain falling on the valley bottom); 3) surface runoff is dominant from the beginning of runoff until sometime after the peak; and 4) groundwater runoff begins essentially when runoff begins. If there is some consistency in these time distributions from storm to storm, some inferential support for the conceptual model proposed can be gained.

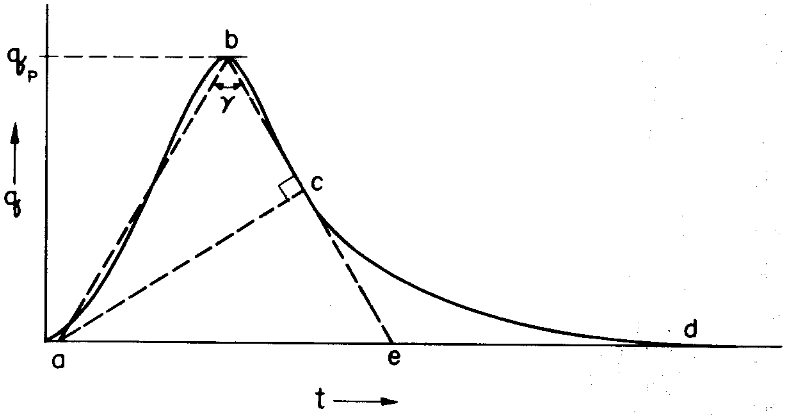


Figure 52. Diagram showing computation of volumetric hydrograph separation.

Referring to Figure 52, the estimation is done as follows: the delayed-flow hydrograph is assumed to have the shape defined by the lines connecting points a-c-d, with the position of point c to be determined for each storm. This general shape is one of the common forms assumed in arbitrary hydrograph separations (Linsley *et al.*, 1949, p. 399-400). The area of the triangle a-b-c is the volume of valley-bottom runoff V_v, which here is assumed known as 100% runoff from that area. Points a and b are fixed at zero and the peak flow q_p respectively, and line b-e is drawn to match as closely as possible the initial portions of the recession. Line a-c is perpendicular to b-e. If the angle between a-b and b-e, γ, is measured, then

$$\overline{bc} = \frac{2\ V_v}{\overline{ab}\ \sin \gamma}$$

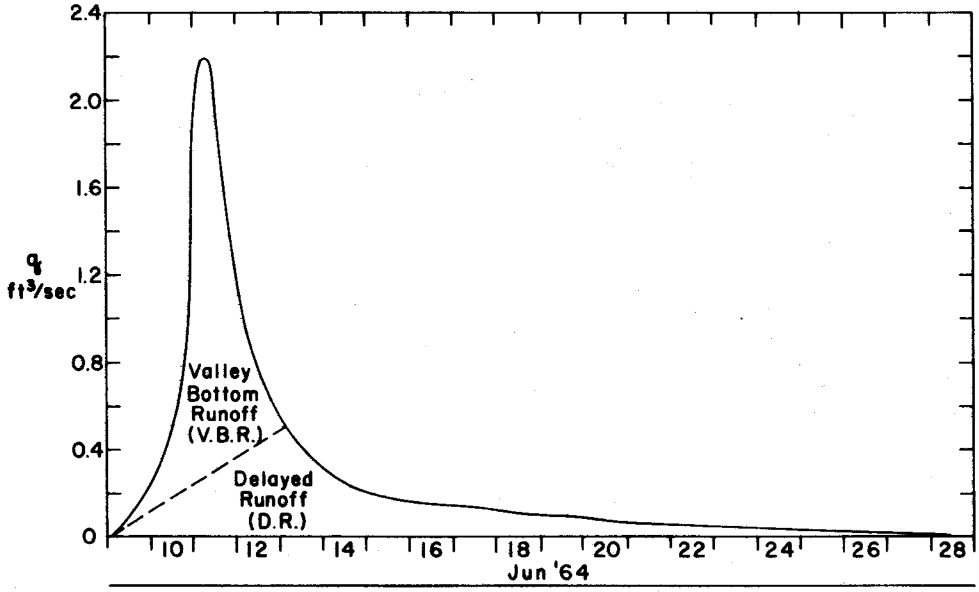
where $\overline{bc}$ and $\overline{ab}$ are the distances from b to c and a to b, respectively. Only 12 storms are sufficiently close to have simple one-peaked hydrographs to allow use of this method, and the separated hydrographs for these are shown in Figure 53. The volume of runoff represented by the area between the measured hydrograph and the constructed separation line was measured by planimeter, and adjustments made as necessary until this volume was within 10% of the calculated valley-bottom runoff.

In summary, the hydrograph separations shown in Figure 53 may be considered suggestions which are consistent with the two-component hypothesis of runoff from Glenn Creek. The volumes of runoff from each source have been forced to fit this hypothesis, and the time distribution appears qualitatively reasonable in that the rise, which has been shown to be quite responsive to rainfall intensity, is dominated by the valley-bottom runoff (overland flow) and the exceptionally long recessions are due to the delayed source or sources.

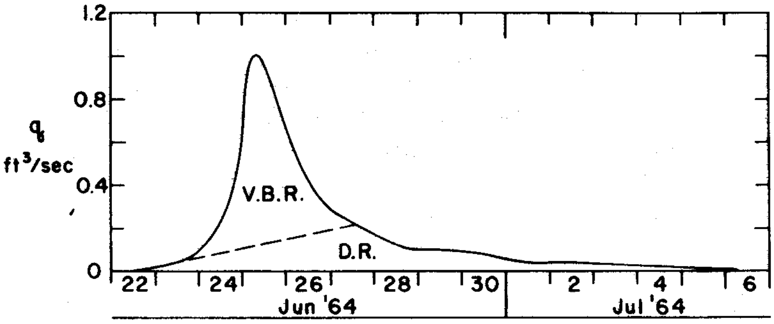
Hydrograph modeling

Enough regularities of the hydrographs of Glenn Creek have been identified to suggest a method of synthetic hydrograph construction. Consider a storm of total precipitation P and duration T_s falling on the watershed, with the stream discharge equal to q_i at the beginning of the storm. The regression relation established earlier

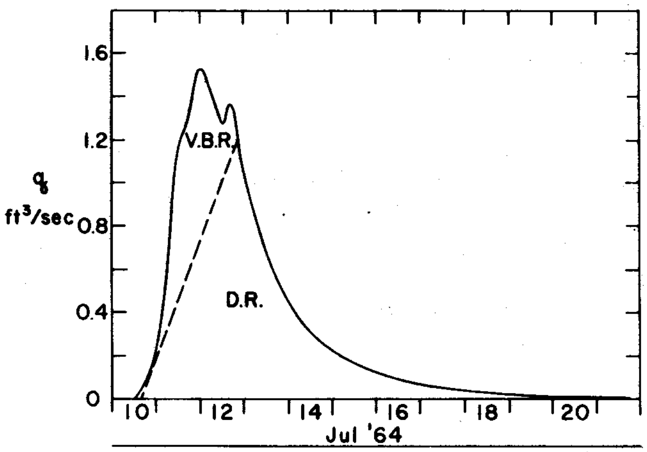
$$\frac{RO}{P} = 0.084 + 0.734\ q_i \qquad (27)$$



a. 9-28 June 1964.

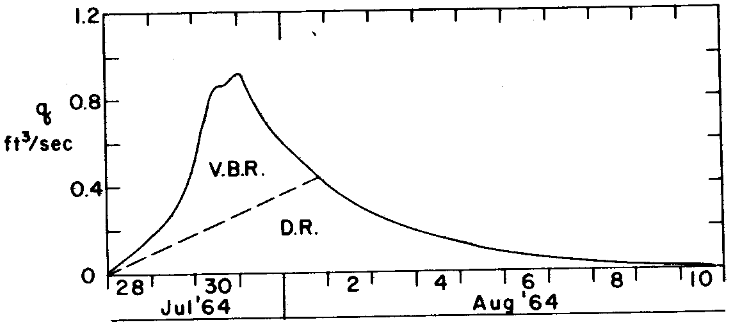


b. 22 June - 6 July 1964.

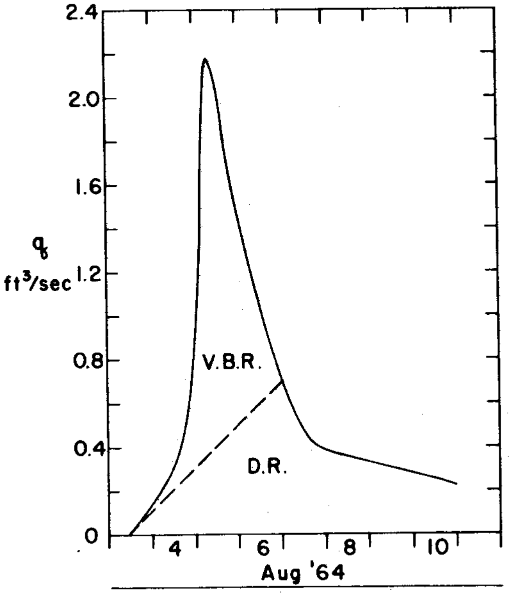


c. 10-21 July 1964.

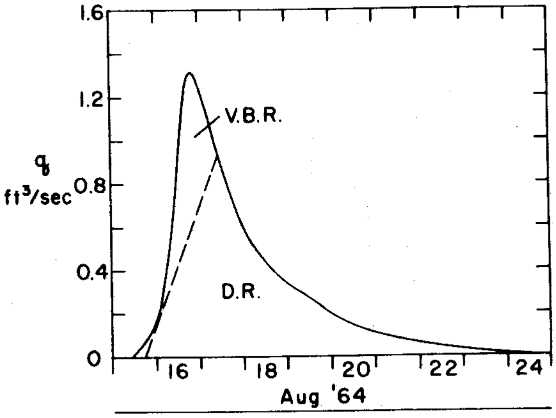
Figure 53. Volumetric separation of hydrographs.



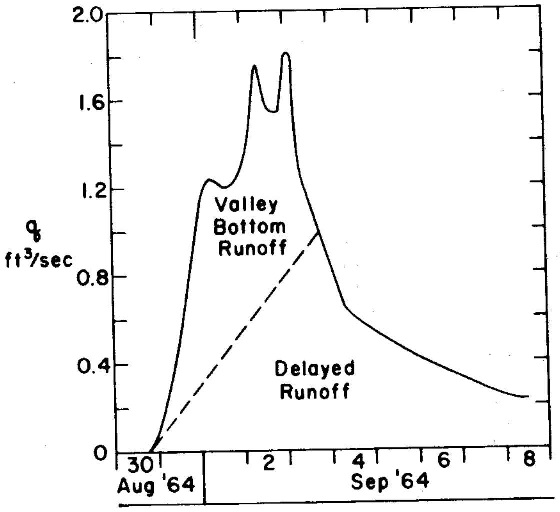
d. 28 July - 10 August 1964.



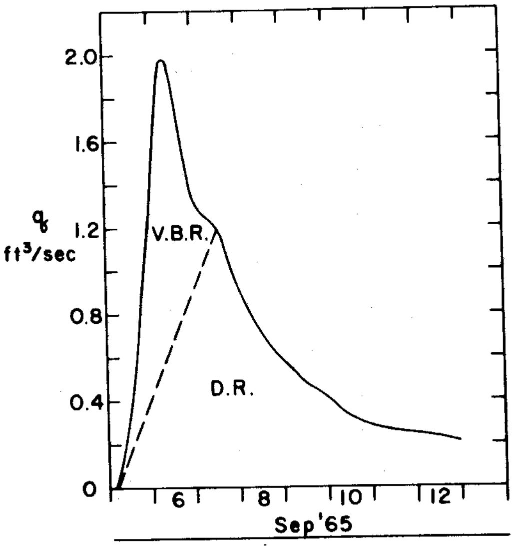
e. 3-11 August 1964.



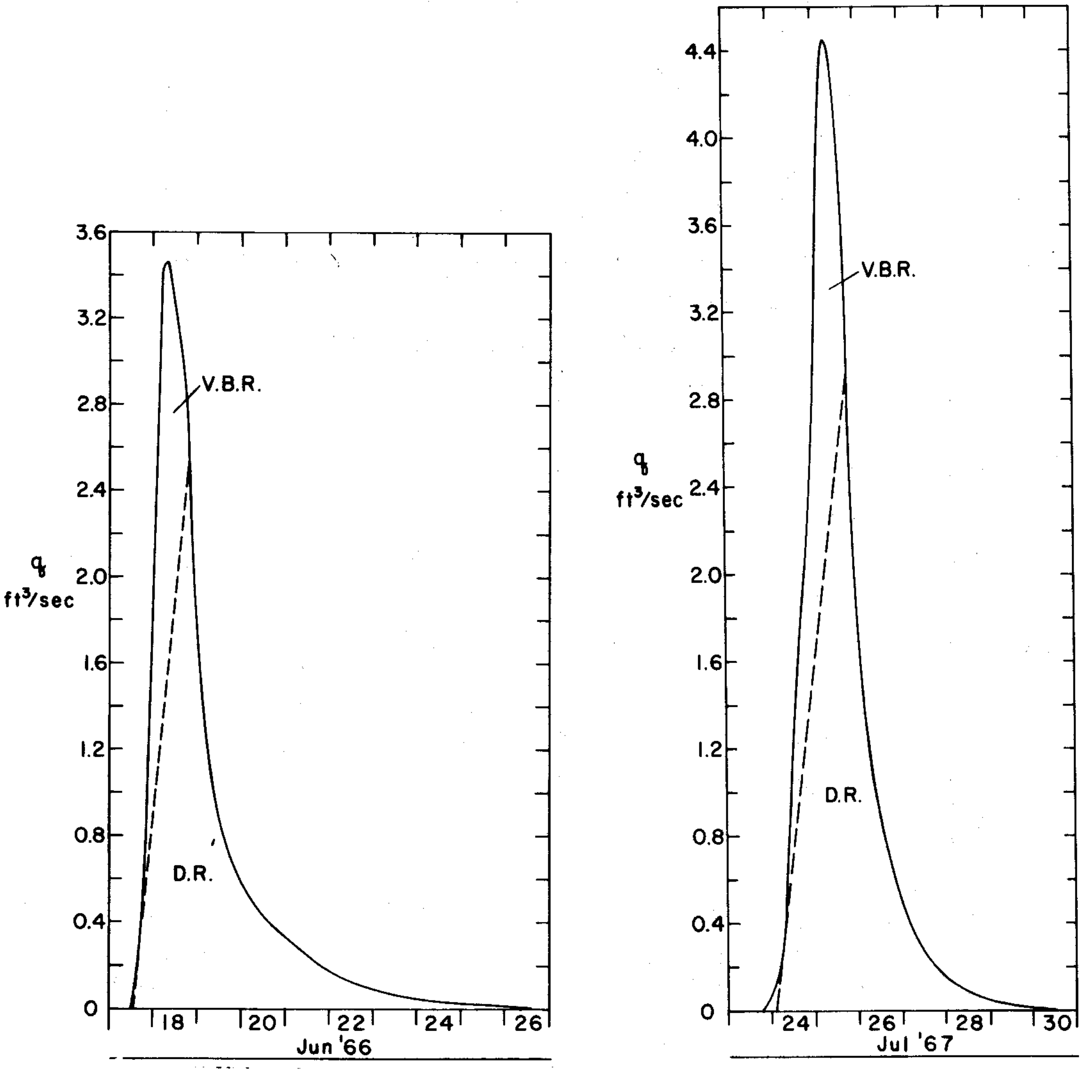
f. 15-25 August 1964.



g. 30 August - 8 September 1964.

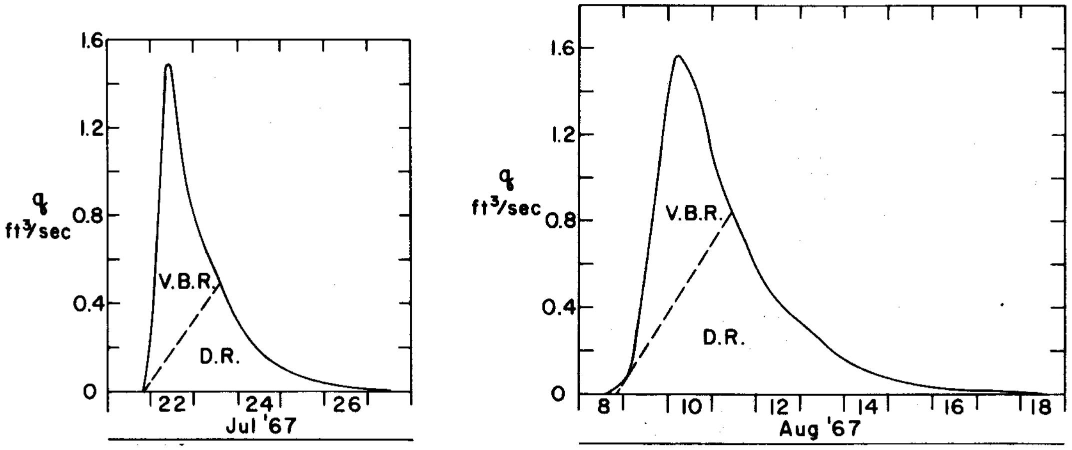


h. 5-12 September 1965.



i. 17-26 June 1966.

k. 23-30 July 1967.



j. 21-27 July 1967.

l. 8-17 August 1967.

Figure 53 (Cont'd). Volumetric separation of hydrographs.

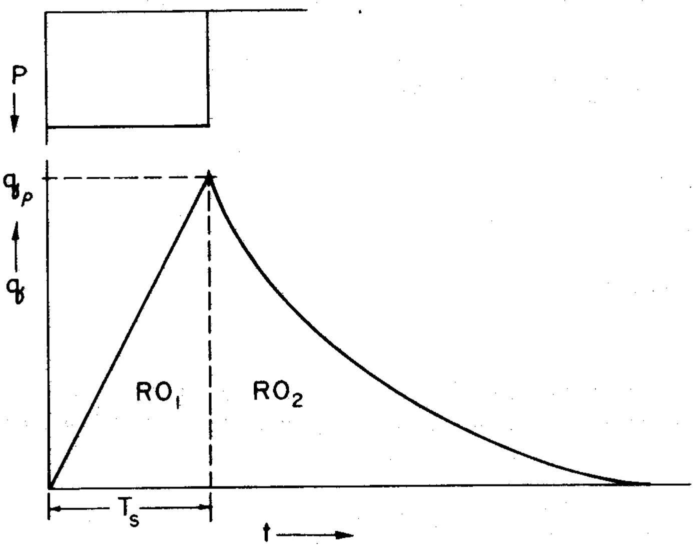


Figure 54. Diagram illustrating hydrograph synthesis.

where q_i is in ft³/sec, can be used to estimate the percentage of rainfall running off, and hence the total runoff volume *RO*. Then, referring to Figure 54, we consider the hydrograph in two segments, one before the peak and one after. If the volume of runoff occurring before the peak is designated RO_1, and that occurring afterwards as RO_2

$$RO = RO_1 + RO_2 \,. \tag{28}$$

If the rise portion is approximated by a triangle

$$RO_1 = (½)\, q_p T_p \tag{29a}$$

where q_p is peak discharge and T_p is time to peak. But, as found earlier, it is essentially true that $T_p = T_s$. Since the T_s is given, we have

$$RO_1 = (½)\, q_p T_s \,. \tag{29b}$$

The recession portion is well represented by an exponential decay

$$q = q_p\, e^{-t/t*} \tag{30}$$

where $t*$ is initially considered to be a reasonably constant characteristic of the watershed. Thus

$$RO_2 = q_p \int_0^\infty e^{-t/t*}\, dt$$

$$RO_2 = q_p\, t* \,. \tag{31}$$

Substituting eq 31 and eq 29b into eq 28

$$RO = (½)\, q_p T_s + q_p\, t* \tag{32}$$

in which the only unknown is q_p. Solving for q_p

$$q_p = \frac{RO}{(T_s/2 + t^*)} . \tag{33}$$

If eq 27 is substituted into eq 33

$$q_p = \frac{(0.085 + 0.734\ q_i)\ P}{(T_s/2 + t^*)} \tag{34}$$

which expresses peak discharge as a function of antecedent conditions q_i, total precipitation P, storm duration T_s, and recession constant t^*. Since q_p in eq 34 is peak discharge due to the storm, the actual discharge at peak can be estimated as

$$q_p = \frac{(0.085 + 0.734\ q_i)\ P}{(T_s/2 + t^*)} + q_i\ e^{-T_s/t^*} . \tag{35}$$

Peaks estimated from eq 35, assuming t^* is constant at the average value of 39.2 hours are compared with actual peaks for 14 storms in Table XXII and Figure 55. The correlation coefficient between actual and observed values is 0.925, which is significant below the 0.01 level.

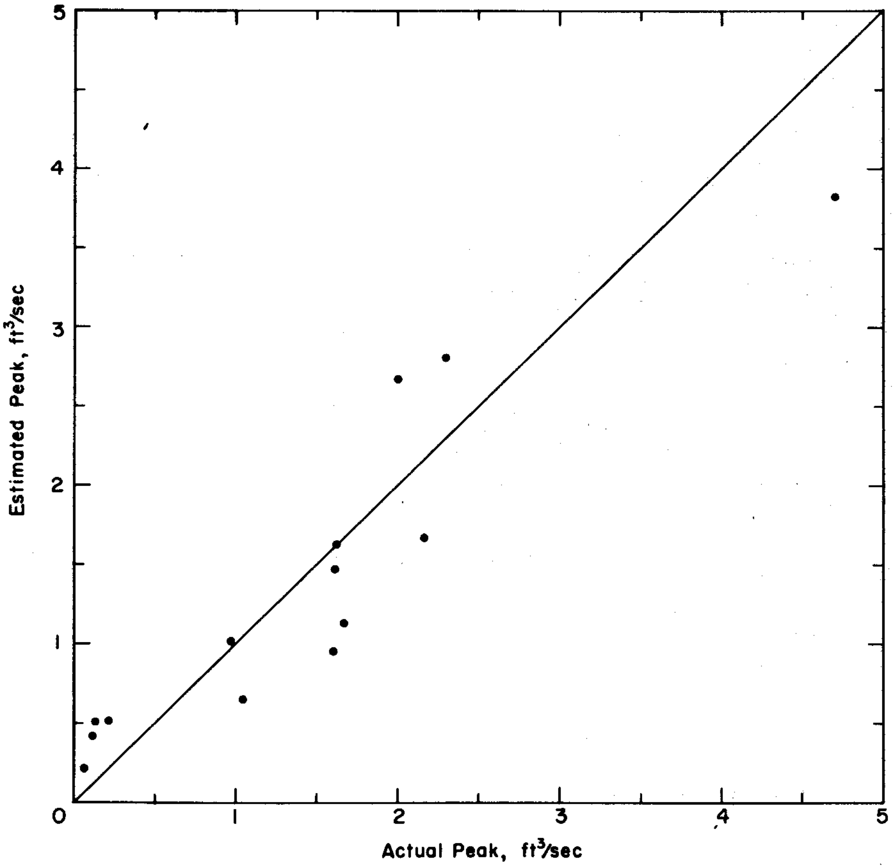


Figure 55. Estimated and observed peak discharges, 14 storms, Glenn Creek watershed.

Table XXII. Data and results of comparison of estimated and observed peak discharge, 14 storms, Glenn Creek watershed.

Runoff period	*P (in.)*	q_i *(ft³/sec)*	T_s*(hr)*	*Calculated* q_p *(ft³/sec)*	$q_i e^{-T_s/t^*}$ *(ft³/sec)*	*Calc peak*	*Observed peak*
22 June - 6 July 64	0.86	0.05	68	0.646	0.009	0.655	1.04
10 July - 21 July 64	0.58	0.15	36	0.894	0.059	0.953	1.60
28 July - 10 Aug 64	0.88	0.14	72	0.992	0.023	1.015	0.97
3 Aug - 23 Aug 64	1.41	0.24	44	2.72	0.08	2.80	2.30
15 Aug - 25 Aug 64	0.36	0.47	32	1.27	0.20	1.47	1.61
30 Aug - 25 Oct 64	1.49	0.30	78	2.63	0.04	2.67	2.00
5 Sept - 4 Oct 65	1.00	0.13	24	1.59	0.07	1.66	2.16
6 July - 13 July 66	0.22	0.01	13	0.201	0.007	0.208	0.059
9 July - 16 July 66	0.39	0.03	17	0.395	0.020	0.415	0.114
7 July - 14 July 67	0.69	0.02	45	0.503	0.007	0.510	0.133
19 July - 24 July 67	0.76	0	36	0.510	0	0.510	0.206
21 July - 27 July 67	0.70	0.19	16	1.50	0.13	1.63	1.62
23 July - 30 July 67	1.19	0.39	31	3.65	0.17	3.82	4.70
8 Aug - 17 Aug 67	1.05	0.06	32	1.11	0.03	1.14	1.37

Presumably, predictions by this model could be improved by relating t^* to evapotranspiration during the recessions. If one could relate evapotranspiration to weather conditions (air temperature, relative humidity, wind speed), the relation given in eq 35 could form the basis for a rather complete model relating streamflow to meteorologic conditions. Given an initial discharge and a sequence of weather conditions (precipitation duration and depth and the factors affecting evapotranspiration), a continuous hydrograph for a summer could be modeled.

In the present study, there are insufficient data to fully investigate the relationship between t^* and evapotranspiration. However, reasonable success in estimating peaks was achieved. In part this can be explained by noting that in eq 35, an overestimate of t^* tends to make the first term smaller and the second term larger. Thus, the effects of incorrectly estimating t^* tend to be buffered, and eq 35 in its present form, with a constant t^*, could also be used to model a continuous summer hydrograph.

While the form of eq 35 may be generally applicable to low-order watersheds in the Yukon-Tanana uplands, the constants in the relation between antecedent discharge and proportion of rainfall running off and the recession constant vary depending on geological-permafrost conditions and watershed size.

SUMMARY AND CONCLUSIONS

Glenn Creek watershed seems to be representative of the lower elevations of the Yukon-Tanana uplands physiographic province of central Alaska in regard to its topography, geology, soils, permafrost, vegetation, climate and, presumably, hydrologic characteristics. It covers an area of 0.70 square mile, and ranges in elevation from 842 ft to 1618 ft above sea level. A wedge (0 to over 30 ft thick) of eolian, residual, and colluvial silts, with widespread gravel layers, overlies the bedrock, which is Birch Creek schist. Poorly-drained soils have formed above a shallow permafrost table that is present beneath some 50% of the watershed. The remainder of the area is permafrost-free, and is covered by well-drained soils. A close correspondence between permafrost conditions and vegetation was revealed by detailed mapping of both. The climate of the basin is strongly continental; the summers are short and warm and total precipitation and precipitation intensities are low.

The four summers included in this study, 1964-1967, included the driest (1966), the coolest (1965), and the second-wettest (1967) of the last 30 years. The summer of 1964 was near normal in regard to precipitation and temperature. Analysis of rainfall-runoff data for 1964, for which records are complete from June to 25 October, indicates that about 6 in. of the 12.3-in. normal annual precipitation is returned to the atmosphere as evapotranspiration. This is considerably less than has been calculated by others (Patric and Black, 1968), and only 31% of potential evapotranspiration. Some 4 in. of the mean annual runoff derives from spring snowmelt, the remainder being from summer and fall rainfall.

Equivalent latitude θ', which depends on the slope and aspect of the ground surface, was a useful quantitative measure of the relative insolation within the watershed. Relative insolation is the major factor controlling the presence or absence of permafrost: where $\theta' > 65°$, permafrost is present; and where $\theta' < 60°$, permafrost is absent. Further, depth to permafrost was inversely correlated to θ'.

For individual storms during the study period, runoff/rainfall proportions ranged from 0.03 to 0.42, and were positively correlated with the antecedent discharge of the stream, which is a measure of the wetness of the watershed when rain begins. The stream responds rapidly to the onset of rainfall, except when the basin is very dry, and the duration of the hydrograph rise is essentially equal to the duration of the storm. Times from the beginning of rainfall to the beginning of runoff were not significantly related to antecedent discharge. Direct channel precipitation makes up a negligible portion of streamflow. These facts, and consideration of the soils and geology of the watershed, lead to the conclusion that surface runoff from the valley bottom is the dominant source of streamflow during a hydrograph rise. The maximum possible fraction of total runoff arising from this source ranged from 0.19 to 0.57 for 12 storms.

The hydrograph recessions could be well modeled by a simple exponential decay. The decay constant varied from 19.6 hours to 76.9 hours for 12 storms, and averaged 39.2 hours. There is a strong indication that the variation in recession constants is controlled by evapotranspiration rates during the recession. The low evapotranspiration rates of central Alaska may explain, at least in part, why the recession constants for Glenn Creek are extremely large when compared with those of temperate-region watersheds. In any case, a delayed source of runoff to Glenn Creek is present, and is probably a combination of *tunnel* flow on the mineral soil beneath the moss and typical groundwater flow through the moss and soils.

Enough regularities in the streamflow characteristics of Glenn Creek could be identified to propose a method of constructing synthetic hydrographs for the stream when antecedent discharge and rainfall depth and duration are specified. The fraction of rainfall running off is estimated from an empirical relation with antecedent discharge. The time of rise is simply equal to the storm duration, and the recession is portrayed by an exponential decay (using the average value of the decay constant for the basin) beginning at the peak discharge. This latter quantity is estimated from

$$q_p = \frac{(0.085 + 0.734\, q_i)\, P}{(T_s/2 + t^*)} + q_i\, e^{-T_s/t^*} \tag{35}$$

where q_p is peak discharge (ft^3/sec), q_i is antecedent discharge (ft^3/sec), P is total precipitation (ft^3), T_s is storm duration (sec), and t^* is the recession constant (sec). Peak discharges calculated by this formula were found to correspond well with actual peaks for 14 storms.

The form of eq 35 is probably general for watersheds approximately equal in size to the watershed of Glenn Creek, although the recession constant and the relation between antecedent discharge and fraction of rainfall running off vary depending on watershed size and distribution of soils and permafrost. On considerably larger watersheds, the close relation between storm duration and rise time would begin to break down as channel characteristics began to dominate the hydrograph.

This report has provided the first estimates of the water balance in central Alaska based on actual measurements of rainfall and runoff. While it is known that precipitation increases and evapotranspiration decreases with altitude in the area, these data should provide information which is useful in evaluating water resources of the region. Recently, interest in environmental modeling of the tundra and taiga, as part of the International Biological Program (IBP), has burgeoned. Two outcomes of the present study, the water balance data and the small-watershed runoff model, should be of special interest to the IBP, as well as to the International Hydrological Decade Program. In addition, the floods of August 1967 have increased interest in predicting rainfall-caused floods in central Alaska. The method of quantifying the role of antecedent moisture conditions in determining flood peaks, which resulted from this study, suggests approaches to this problem.

It is hoped that more detailed watershed studies in central Alaska, currently contemplated jointly by the U.S. Army Corps of Engineers, the U.S. Weather Bureau, and the University of Alaska, will probe further into the mechanisms of runoff in the region. In particular, the hypothesis of runoff contributions from the valley-bottom areas should be confirmed, and detailed observations on the source of delayed flow should be made.

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APPENDIX A: SUMMARY OF PERMAFROST PROBE OBSERVATIONS, GLENN CREEK WATERSHED

Grid point	*Depth to refusal at 3 ft in indicated directions from grid point (ft)* *N*	*E*	*S*	*W*	*Avg depth*
AN2	*	*	*	*	*
AN1	1.64	1.87	2.36	1.64	1.88
A	2.03	1.54	2.03	1.57	1.79
BN3	*	*	*	*	*
BN2	*	*	*	*	*
BN1	1.64	1.74	1.54	1.57	1.62
B	1.51	1.57	1.80	1.94	1.71
BS1	1.21	0.95	1.05	1.05	1.07
CN3	*	*	*	*	*
CN2	*	*	*	*	*
CN1	*	*	*	*	*
C	1.41	1.61	2.00	1.57	1.65
CS1	1.31	1.44	1.25	1.08	1.27
CS2	*	*	*	*	*
DN3		gravel layer at 1.90			
DN2		gravel layer at 1.90			
DN1		gravel layer at 1.71			
D	†	1.44	1.12	†	1.28
DS1		gravel layer at 1.31			
DS2	*	*	*	*	*
EN4	1.87	2.07	1.97	2.03	1.99
EN3		gravel layer at 1.87			
EN2		gravel layer at 1.71			
EN1	*	*	*	*	*
E	1.08	0.82	1.34	1.48	1.18
ES1	1.44	1.38	1.38	1.31	1.38
ES2	*	*	*	*	*
FN4	1.64	2.23	2.00	2.03	1.98
FN3	*	*	*	*	*
FN2		gravel layer at 1.77			
FN1		gravel layer at 0.98			
F	1.61	1.51	2.16	1.87	1.79
FS1	0.85G	1.25G	1.31G	2.10	
FS2	1.31	1.67	2.13	1.67G	1.70
FS3	1.77	1.61	1.87	1.90	1.79
GN4		gravel layer at 1.34			
GN3		gravel layer at 0.75			
GN2		gravel layer at 1.71			
GN1		gravel layer at 0.75			
G	1.31	1.25	1.48	1.51	1.39
GS1	1.34G	1.74G	1.54G	1.38	
GS2	1.48	0.32	1.15	1.02	1.12
GS3	1.12	1.12G	1.12G	1.54	1.33
GS4	1.21	1.12	0.92G	1.28	1.20

Grid point	*Depth to refusal at 3 ft in indicated directions from grid point (ft)* *N*	*E*	*S*	*W*	*Avg depth*
HN4		gravel layer at 1.18			
HN3		gravel layer at 1.44			
HN2		gravel layer at 1.67			
HN1	0.98	1.41	1.44	1.18	1.25
H	1.28	1.64	1.41	1.21	1.39
HS1	2.23	2.59	2.07	1.94	2.21
HS2	3.02	2.23G	2.20G	1.71G	
HS3	1.84G	1.41G	1.74G	1.74G	
HS4	1.44	1.48G	1.31G	1.31G	
HS5	1.74G	1.77G	1.71G	1.94G	
IN4		gravel layer at 0.92			
IN3		gravel layer at 2.26			
IN2		gravel layer at 1.12			
IN1	1.31G	1.02G	1.54G	1.67G	
I	1.02	1.31	1.02	0.98	1.08
IS1	1.34	1.77	1.57	1.44	1.53
IS2	1.38	1.34	1.38	1.31	1.35
IS3	1.38	1.31	1.25	1.61	1.39
IS4	1.77	1.38	1.12	1.51	1.45
IS5	1.80G	1.44G	2.23G	2.39G	
JN3		gravel layer at 1.48			
JN2		gravel layer at 1.94			
JN1		gravel layer at 1.51			
J		gravel layer at 1.84			
JS1	1.44	1.25	1.57	1.64	1.47
JS2	1.84G	1.97G	2.59G	1.48G	
JS3	1.38G	2.26	1.94G	1.94G	
JS4	1.87	2.13	1.74	1.87	1.90
KN3	1.61G	1.67G	1.48G	1.02G	
KN2	1.48G	1.84G	1.57G	1.71G	
KN1	1.51G	1.31G	2.72	2.13	2.43
K	1.48	1.77G	1.34	1.48G	1.41
KS1	1.28	1.44	1.48	1.25	1.36
KS2	1.34	0.98	1.38	1.12	1.21
KS3	*	*	*	*	*
LN2	1.67G	2.76G	2.26G	1.57G	
LN1	1.64G	1.57G	2.03G	1.97G	
L	1.41G	1.87G	1.12G	1.64G	
LS1	1.15	1.31	1.21G	1.61	1.36
MN2	1.77G	1.77G	1.87G	2.46G	
MN1	1.61G	1.54G	2.49G	1.38G	
M	1.61G	1.44G	2.20	1.67G	

* Greater than 3.35 ft.

† Points not present due to presence of stream.

G Indicated gravel layer.

APPENDIX B: WATER-HOLDING AND -TRANSMITTING PROPERTIES OF MOSS

Introduction

A series of laboratory tests was undertaken to determine the porosity, permeability, and water-holding properties of the moss which covers much of the watershed of Glenn Creek.

Sampling

A plot of moss (*Sphagnum* spp. and *Pleurozium schreberi*), free of all but a few woody plants (labrador tea and blueberry), was located near the outlet of Glenn Creek Basin near Fairbanks, Alaska. A sample measuring 2.5 ft by 2.5 ft was cut out of the plot with a small saw, and gently separated and lifted from the mineral soil below. The thickness of the sample was about 0.5 ft. This was placed in a plastic-lined wooden box, covered with plastic, and shipped to CRREL in Hanover, New Hampshire. The moss and associated plants were living and appeared healthy after one-week transit time. The box was opened and placed outside, and was kept watered up to and during the testing periods. The vegetation continued to flourish during the testing periods.

Samples for testing were cut from this larger sample with a sharp-rimmed brass cylinder which could sever the moss and roots with little difficulty or disturbance. The test samples measured 0.3 ft in diameter and were 0.4 to 0.5 ft long.

Permeability

Laboratory procedure. Horizontal and vertical samples were transferred directly from the sampling tube to the permeability-testing apparatus (Fig. B1). Vertical samples were inverted so that the flow of water was from the top to the bottom of the moss. A large-mesh wire screen held the sample several cm from the bottom of the cylinder and above the cylinder inlet. The upper outlet of the testing cylinder was sealed, and a vacuum applied to the top of the cylinder while water was allowed to rise slowly into the cylinder from the bottom. When the water had risen over the sample, the vacuum and seal were removed. A wire-mesh screen was placed on top of the sample, and the sample length was measured.

Heads were varied from 10 cm to 1 mm by adjusting the elevation of the cylinder relative to the water reservoir. At each head, the volume of water collected in a graduated cylinder in a given time was measured. Three volume and time readings were taken at each head, and their average is recorded in Table BI. The permeability of four samples was measured, each over a range of heads, in this way. Samples 1 and 2 were each tested twice after an interim of several hours during which they remained saturated; samples 3 and 4 were tested once only. Water temperature was recorded for each run.

Results. Since measurements in the laboratory were made using cm-g-°C units, results are reported here in those units.

The permeability, or hydraulic conductivity k (cm/sec), was determined for each run by the formula

$$k = \frac{QLR_T}{hAt}$$

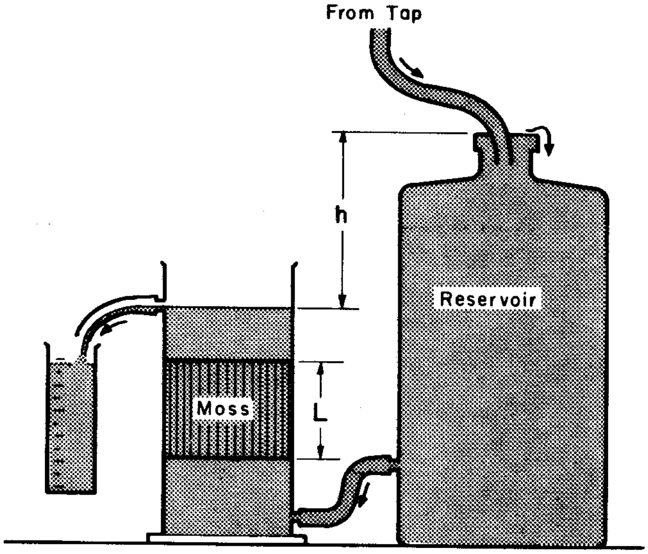

Figure B1. Permeability-measuring apparatus.

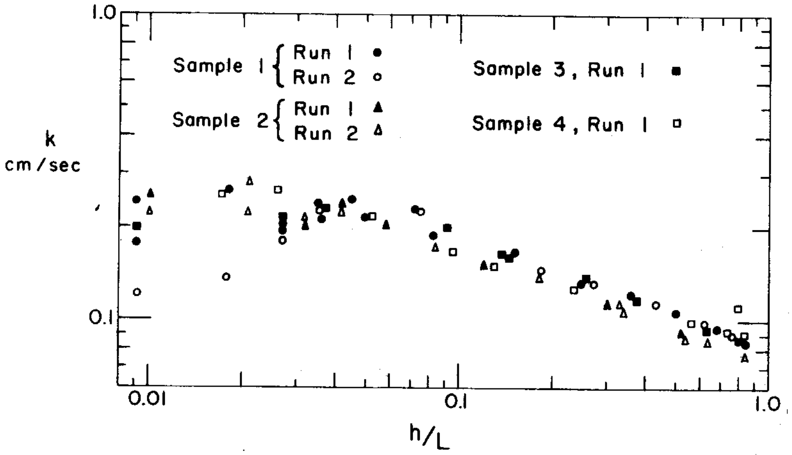

Figure B2. Relation between permeability k and hydraulic gradient h/L for moss samples. Samples 1-3 vertical, sample 4 horizontal.

where:

Q ≐ the quantity of water (cm^3) passing through the sample in time

t = time (sec)

h = the difference in elevation between the reservoir overflow and the cylinder outlet (cm)

L = the sample length (cm)

A = the cross-sectional area of the sample (cm^2)

R_T= the correction factor for viscosity of water.

The results are given in Table BI and plotted in Figure B2.

Samples 1-3 are vertical, and sample 4 is horizontal. Since the results for sample 4 fell so close to those of samples 1-3, further measurement of horizontal permeability was considered unnecessary. Above a gradient of 0.04, permeability decreases rapidly as head increases, while at

APPENDIX B

Table BI. Results of permeability determinations.

Sample 1, run 1 (19 June 1968) $L = 11$ cm, $A = 61.5$ cm^2, $T = 24°$C (Temperature correction = 0.91).

Q (cm^3)	t (sec)	h (cm)	h/L	k (cm/sec)
31.6	60	0.4	0.036	0.215
22.0	200	0.1	0.0091	0.179
22.5	60	0.3	0.027	0.204
62.5	60	0.9	0.082	0.188
100.	59.3	1.65	0.15	0.167
100.	43.5	2.8	0.25	0.134
100.	34	3.95	0.36	0.121
100.	27.5	5.5	0.50	0.107
100.	23	7.45	0.68	0.095
100.	20.2	9.3	0.84	0.086
66.5	60	0.8	0.073	0.226
43.7	60	0.55	0.050	0.216
45.7	60	0.5	0.045	0.248
19.6	60	0.2	0.018	0.266
34.7	60	0.4	0.036	0.236
21.7	60	0.3	0.027	0.197
29.7	200	0.1	0.0091	0.242

Sample 1, run 2 (20 June 1968) $L = 11$ cm, $A = 61.5$ cm^2, $T = 24°$C (Temperature correction = 0.91).

Q (cm^3)	t (sec)	h (cm)	h/L	k (cm/sec)
67.	60	0.8	0.073	0.227
34.2	60	0.4	0.036	0.232
20.	60	0.3	0.027	0.181
17.	100	0.2	0.018	0.138
4.5	60	0.1	0.0091	0.122
100.	55.3	2.	0.18	0.147
100.	40	3.1	0.28	0.131
100.	30.3	4.7	0.43	0.115
100.	24.6	6.8	0.62	0.097
100.	21.5	8.4	0.76	0.090

Sample 2, run 1 (24 June 1968) $L = 9.5$ cm, $A = 61.5$ cm^2, $T = 26°$C (Temperature correction = 0.87).

Q (cm^3)	t (sec)	h (cm)	h/L	k (cm/sec)
19.2	100.	0.1	0.01	0.258
27.5	60.	0.3	0.032	0.205
43.4	60.	0.4	0.042	0.243
74.5	60.	1.1	0.12	0.151
50.5	60.	0.55	0.058	0.205
100.	40.1	2.9	0.30	0.116
100.	29.5	4.9	0.52	0.093

Sample 2, run 2 (24 June 1968) $L = 9.5$ cm, $A = 61.5$ cm^2, $T = 26°$C (Temperature correction = 0.87).

Q (cm^3)	t (sec)	h (cm)	h/L	k (cm/sec)
100.	21.9	7.95	0.84	0.077
100.	25.2	6.1	0.64	0.087
100.	29.6	5.0	0.53	0.090
100.	38.4	3.1	0.33	0.113
100.	39.1	3.0	0.34	0.107
100.	56.9	1.7	0.18	0.139
100.	97.4	0.8	0.084	0.172
70.	100.	0.4	0.042	0.235
42.3	100.	0.2	0.021	0.284
33.5	100.	0.2	0.021	0.225
56.8	120.	0.3	0.032	0.212
16.8	100.	0.1	0.010	0.226

Sample 3, run 1 (24 June 1968) $L = 11$ cm, $A = 61.5$ cm^2, $T = 24°$C (Temperature correction = 0.91).

Q (cm^3)	t (sec)	h (cm)	h/L	k (cm/sec)
95.5	60	1.6	0.145	0.162
100.	33.5	4.1	0.373	0.118
100.	24.9	6.9	0.627	0.095
100.	20.3	8.9	0.809	0.088
100.	42.0	2.8	0.254	0.138
100.	80.3	1.0	0.091	0.201
39.4	100.	0.3	0.027	0.214
57.6	100.	0.4	0.036	0.235
48.	400.	0.1	0.0091	0.196
100.	66.7	1.5	0.136	0.163

Sample 4, run 1 (16 April 1969) $L = 11.5$ cm, $A = 61.5$ cm^2, $T = 27°$C (Temperature correction = 0.85).

Q (cm^3)	t (sec)	h (cm)	h/L	k (cm/sec)
100.	18.4	9.5	0.826	0.091
100.	20.3	8.5	0.739	0.092
100.	24.7	6.5	0.565	0.099
100.	30.9	4.6	0.800	0.112
100.	46.5	2.7	0.235	0.127
100.	69.8	1.5	0.130	0.151
83.25	100.	0.6	0.052	0.218
50.	100.	0.3	0.026	0.265
32.3	100.	0.2	0.017	0.257
100.	86.5	1.1	0.096	0.167

lower gradients the values seem to scatter (rather widely) about a constant value. The average value of k for all measurements where $h/L < 0.04$ is 0.218 cm/sec.

The break in slope in the relation between k and h/L can be shown to be due to the breakdown of Darcy's law when inertial forces become significant relative to viscous forces. Following Todd (1959, p. 47-48), a plot of Reynolds number N_{Re} vs a dimensionless friction factor f where

$$N_{\mathrm{Re}} = \frac{Vdp}{\nu}$$

$$f = \frac{dhg}{2LV^2}$$

and

V = Darcy flow velocity
d = average grain diameter
g = acceleration due to gravity
p = fluid density
ν = fluid viscosity
h/L = the energy gradient

illustrates this for granular materials. There is a break in slope in the inverse relation between these variables at the value of N_{Re} where inertial forces become significant. For the moss, it is not clear what the value of d should be, so a plot of N_{Re}/d vs f/d was made, using the data of Table BI and adjusting to a water temperature of 20°C for all runs (Fig. B3). This plot shows a clear break in slope at $f/d \simeq 2 \times 10^5$ cm^{-1}, $N_{\mathrm{Re}}/d \simeq 1$ cm^{-1}. In the graph by Todd (1959, Fig. 3.2), the slope break is gradual, with deviations from the line $f = 1000/N_{\mathrm{Re}}$ beginning at about $N_{\mathrm{Re}} = 1$ and becoming pronounced at $N_{\mathrm{Re}} = 10$. If we take $N_{\mathrm{Re\ c}}$(critical Reynolds number) = 1 as the Reynolds number at which Darcy's law no longer applies, the moss data give

$$\frac{N_{\mathrm{Re\ c}}}{d} = 1\ \mathrm{cm}^{-1},\ d = 1\ \mathrm{cm}.$$

This suggests that the moss has a pore opening equivalent to that of a granular material with a grain diameter of about 1 cm, a "pea gravel." In agreement with this, the average permeability, 0.218 cm/sec, falls in the range for gravels as given by Todd (1959, Fig. 3.4). However, the value of f/d at $N_{\mathrm{Re\ c}}$ for the moss is about 2×10^5 cm^{-1}, so that $f = 2 \times 10^5$ at $N_{\mathrm{Re\ c}}$. The data plotted by Todd give $f = 10^3$ at $N_{\mathrm{Re\ c}}$, obviously a much lower value than for the moss. Since, for the definition of f,

$$V = \left(\frac{1}{f}\,\frac{dg}{2}\,\frac{h}{L}\right)^{1/2}$$

a higher value of f for a given grain size means a lower velocity for a given gradient – the higher the friction factor, the higher the flow resistance. Thus moss would appear to have a considerably higher resistance to flow than mineral grains. This may result from the shape of the pore spaces in the moss, since for laminar open-channel flow, channel shape determines the relation between f and N_{Re} (Chow, 1959, p. 10).

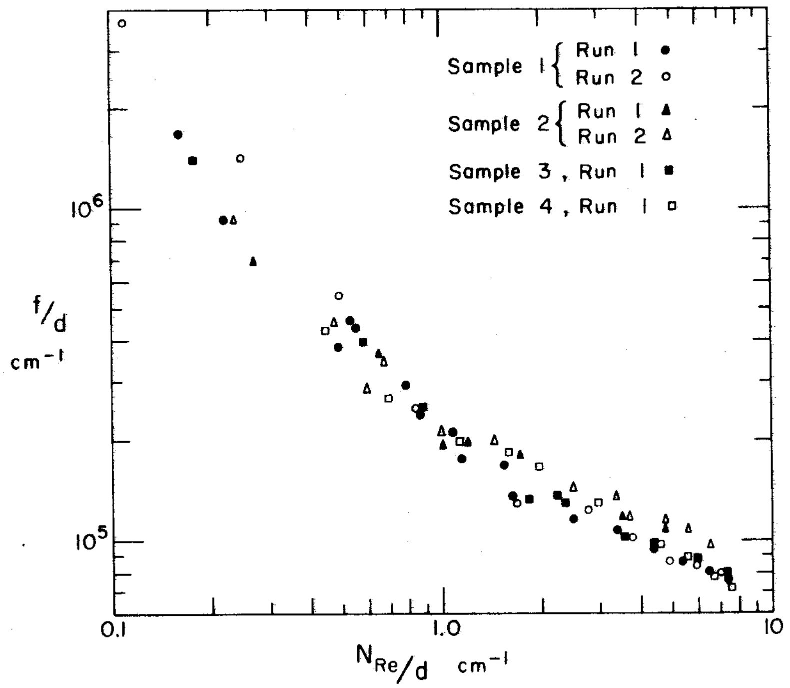

Figure B3. Relation between friction factor/depth ratio f/d and Reynolds number/depth ratio N_{Re}/d for moss samples.

Water-holding properties

Laboratory procedure. A separate sample was cut from the larger moss body in the same manner as for the permeability tests and placed in the bottom of the cylinder used for the permeability tests. The cylinder was sealed and a vacuum applied to the cylinder. Water was then introduced to the sample and the vacuum seal removed. Length and cross-sectional area of the sample were measured to determine volume. Weight was determined by subtracting the weight of the cylinder from the total weight to give the weight of the sample plus water.

Next, the cylinder was inverted and the sample was allowed to drip-drain for about four hours, when gravity drainage ceased. The weight of the sample plus water at this time was used to estimate field capacity. No change in sample volume occurred during gravity drainage. The confinement of the sample in the testing cylinder may have restricted sample shrinkage.

Further drying was accomplished by placing the sample in an evaporating dish, with a fan nearby to increase air circulation. After some 600 hours, no further weight reduction was observed, and the air-dry weight (38 g) was used as the basis for calculation of water-holding properties. During the air-drying, the sample volume decreased from 566 cm³ to 449 cm³ (21%). However, the initial volume, 566 cm³, was used as the basis for calculation of water-holding properties. After air-drying, the sample was oven-dried at 105°C for 24 hours, then weighed and measured again. Although only 3 g of water were lost in this process, volume decreased by another 77 cm³.

Results. The results of the measurements of water-holding properties are given in Table BII.

Table BII. Results of water-holding capacity measurements.

	Saturation	*Drip-dried*	*Air-dried*	*Oven-dried*
Sample volume V (cm³)	566	566	449	372
	.98 g/cm³	.46 g/cm³	.085 g/cm³	.09 g/cm³
Sample weight W (g)	575	258	38	35
	1542 %	637 %	8 %	

$V_{TOTAL} = 566 \text{cm}^3$

$V_{water} = V_{voids} = \frac{540 \text{g}}{1 \text{g/cm}^3} = 540 \text{cm}^3$

$V_{organic} = 566 - 540 = 26 \text{cm}^3$

The porosity of the moss n is

$$n = \frac{V_v}{V}$$

where V_v is volume of voids. At saturation,

$$\begin{aligned} W_w &= W_s - W_m \\ &= 575 \text{ g} - 38 \text{ g} \\ W_w &= 537 \text{ g} \end{aligned}$$

where W_w is weight of water, W_s is weight of sample at saturation, and W_m is weight of moss (air-dry). Since

$$V_v = V_w = \frac{W_w}{w}$$

where V_w is volume of water, W_w is weight of water, and w is the unit weight of water:

$$V_v = \frac{537 \text{ g}}{1 \text{ g/cm}^3} = 537 \text{ cm}^3$$

$$n = \frac{537 \text{ cm}^3}{566 \text{ cm}^3} = 0.949 \ .$$

To find the equivalent depth of water d held per unit depth of moss at saturation and at "field capacity" (drip-drained): Saturation:

$$d = \frac{V_w}{AL}$$

where A is cross-sectional area of the sample, and L is sample length

$$d = \frac{V_w}{AL} = \frac{537 \text{ cm}^3}{(61.5 \text{ cm}^2)\ (9.2 \text{ cm})} = 0.949 \text{ cm/cm} \ .$$

Thus, depth of water per unit depth of moss is equivalent to porosity. At "field capacity":

$$W_w = W_f - W_m$$

where W_f is weight of drip-drained sample;

$$W_w = 258 \text{ g} - 38 \text{ g}$$

$$W_w = 220 \text{ g}$$

$$V_w = \frac{220 \text{ g}}{1 \text{ g/cm}^3} = 220 \text{ cm}^3$$

$$d = \frac{V_w}{AL} = \frac{220 \text{ cm}^3}{(61.5 \text{ cm}^2)\ (9.2 \text{ cm})} = 0.391 \text{ cm/cm}$$

APPENDIX C: SUMMARY OF HYDRAULIC GEOMETRY DATA FROM 1964 DISCHARGE MEASUREMENTS

Date	Time	Gage ht (ft)	q (ft^3/sec)	w (ft)	$\bar{d}$ (ft)	$\bar{v}$ (ft/sec)	ν (10^{-5} ft^2/sec)	N_{Re}	n
10 June	1355	0.74	0.52	2.9	0.50	0.37			
11	1015	1.19	2.14	4.1	0.83	0.63			
12	0940	0.90	0.84	3.0	0.67	0.42			
13	1030	0.73	0.43	2.8	0.53	0.29			
14	1205	0.62	0.25	2.7	0.44	0.21			
15	0950	0.56	0.18	2.6	0.38	0.18			
16	9045	0.51	0.15	2.6	0.35	0.16			
17	1025	0.51	0.14	2.6	0.35	0.15			
18	1000	0.48	0.11	2.6	0.33	0.13			
19	1400	0.47	0.10	2.7	0.31	0.12			
20	1000	0.46	0.08	2.7	0.30	0.10			
21	1435	0.42	0.06	2.6	0.26	0.09			
22	1020	0.42	0.05	2.6	0.26	0.07			
23	0945	0.45	0.08	2.6	0.30	0.10			
24	0945	0.55	0.15	2.7	0.39	0.14			
24	1535	0.62	0.19	2.7	0.45	0.16			
25	1010	0.98	1.00	3.0	0.73	0.46			
25	1115	0.97	0.97E						
25	1520	0.94	0.89E						
26	1020	0.74	0.45	3.0	0.55	0.27	1.59	21,119	0.162
27	1140	0.59	0.20	2.7	0.41	0.18	1.53	9,706	0.228
28	1425	0.54	0.13	2.7	0.36	0.13	1.36	5,426	0.282
29	1200	0.47	0.10	2.7	0.31	0.12	1.30	3,600	0.358
30	1115	0.45	0.08	2.7	0.31	0.10	1.41	2,638	0.351
1 July	1145	0.42	0.04	2.6	0.26	0.06	1.36	2,279	0.421
2	1110	0.43	0.04	2.7	0.29	0.05	1.36	1,147	
3	1240	0.40	0.03	2.6	0.24	0.05	1.46	993	
4	1350	0.40	0.02	2.6	0.23	0.03	1.32	909	
6	1215	0.41	0.03	2.6	0.24	0.05	1.32	523	
		0.47P	0.08E						
7	1155	0.44	0.06	2.7	0.29	0.08	1.30	1,784	
9	1345	0.58	0.19	2.8	0.40	0.17	1.36	5,000	0.293
10	1105	0.56	0.15	2.7	0.39	0.14	1.46	3,739	0.350
11	1420	1.07	1.37	3.6	0.80	0.37	1.53	19,346	0.214
11	1620	1.10	1.42E						
		1.14P	1.60E						
12	1235	1.08	1.31	4.7	0.79	0.35	1.53	18,072	0.224
12	1630	1.10	1.42E						
13	1050	0.88	0.68	3.3	0.62	0.33	1.51	13,550	0.202
14	1025	0.69	0.33	3.0	0.44	0.25	1.41	7,801	0.212
15	1130	0.58	0.17	2.7	0.35	0.18	1.32	4,773	0.253
16	1030	0.52	0.10	2.7	0.30	0.12	1.32	2,727	0.343
17	1400	0.45	0.06	2.6	0.24	0.10	1.19	2,017	0.355
18	1200	0.50	0.08	2.7	0.28	0.11	1.36	2,264	0.357
20	1110	0.48	0.08	2.7	0.26	0.11	1.41	2,028	0.340
21	1115	0.46	0.02	2.6	0.24	0.03	1.36	529	
21	1630	0.45	0.03V						
22	1055	0.43	0.03V						
23	0950	0.43	0.02V						
24	1045	0.43	0.02V						
25	1020	0.43	0.01V						
		0.82P	0.39E						
27	1045	0.63	0.14	2.8	0.39	0.13	1.64	3,091	0.377
28	1115	0.57	0.14	2.7	0.34	0.15	1.32	3,864	0.298
28	1530	0.68	0.16E						

Date	Time	Gage ht (ft)	q (ft³/sec)	w (ft)	$\bar{d}$ (ft)	$\bar{v}$ (ft/sec)	ν (10^{-5} ft²/sec)	N_{Re}	n
29 July	1000	0.82	0.39	3.3	0.56	0.21	1.51	7,788	0.297
30	1050	1.05	0.89	4.6	0.77	0.25	1.51	12,748	0.309
30	1530	1.06	0.91E						
		1.08P	0.97E						
31	1115	1.01	0.76E						
3 Aug	1120	0.68	0.24	2.4	0.34	0.29	1.41	6,993	0.154
4	1050	0.86	0.43	3.2	0.47	0.29	1.46	9,336	0.191
		1.32P	2.30E						
6	1100	1.12	1.16	3.6	0.62	0.52	1.51	21,351	0.128
7	1025	0.89	0.52	3.3	0.42	0.38	1.51	10,570	0.136
8	1135	0.81	0.38	3.1	0.37	0.33	1.41	8,660	0.143
10	1050	0.77	0.27	3.1	0.33	0.26	1.41	6,085	0.169
11	0915	0.73	0.20	2.9	0.31	0.22	1.46	4,671	0.191
12	1020	0.70	0.16	3.0	0.30	0.18	1.41	3,830	0.229
13	1000	1.01	0.72	3.6	0.51	0.39	1.46	13,623	0.150
		1.02	0.79E						
14	1010	0.91	0.51	3.4	0.44	0.34	1.46	10,247	0.156
15	1150	0.89	0.47	3.4	0.42	0.33	1.46	9,493	0.156
		1.24P	1.31E						
17	1110	1.13	1.14	3.5	0.57	0.57	1.51	21,517	0.111
18	1430	0.90	0.54	3.2	0.38	0.44	1.46	11,452	0.110
19	1115	0.83	0.40	2.9	0.30	0.46	1.51	9,139	0.089
		0.86P	0.47E						
20	1005	0.85	0.40	3.0	0.34	0.39	1.53	8,667	0.115
21	1025	0.83	0.35	3.1	0.32	0.35	1.53	7,320	0.123
22	1155	0.81	0.29	2.9	0.31	0.32	1.51	6,570	0.131
23	1205	0.77	0.26E						
23	1540	0.76	0.25E						
24	1025	0.88	0.37	3.2	0.35	0.33	1.51	7,649	0.138
24	1530	0.87	0.35E						
25	1115	0.88	0.36	3.2	0.35	0.32	1.51	7,417	0.143
25	1550	0.90	0.40E						
26	1010	0.93	0.46E						
26	1455	0.92	0.43	3.3	0.39	0.33	1.53	8,412	0.149
27	0955	0.88	0.36	3.3	0.36	0.30	1.59	6,792	0.155
27	1420	0.87	0.35E						
28	1045	0.84	0.33E						
31	1435	1.08	0.84	3.6	0.53	0.45	1.59	15,000	0.134
		1.19P	1.49E						
1 Sept	1110	1.18	1.47	3.7	0.58	0.68	1.64	24,049	0.094
		1.29P	1.98E						
2	1350	1.26	1.78	4.9	0.61	0.60	1.59	23,019	0.110
		1.30P	2.02E						
3	1355	1.15	1.30	3.7	0.54	0.65	1.64	21,402	0.094
4	1000	1.04	0.82	3.5	0.43	0.55	1.80	13,139	0.095
8	1130	0.84	0.36	2.9	0.31	0.40	1.70	7,294	0.105
9	1030	0.83	0.30	2.9	0.29	0.36	1.70	6,141	0.112
10	0950	0.79	0.26	2.8	0.29	0.32	1.70	5,459	0.126
11	1355	0.77	0.24	2.5	0.28	0.34	1.59	5,987	0.116
14	1110	0.74	0.18	2.2	0.26	0.31	1.70	4,741	0.121
15	1020	0.75	0.20	2.3	0.28	0.31	1.70	5,106	0.127
16	1040	0.73	0.17	2.2	0.27	0.29	1.80	4,350	0.132
17	1035	0.72	0.15	2.2	0.27	0.25	1.93	3,497	0.153
18	1045	0.72	0.15	2.2	0.27	0.25	1.80	3,750	0.153
21	1020	0.71	0.11	2.2	0.26	0.19	1.80	2,744	0.197
22	0920	0.69	0.09	2.2	0.25	0.16	1.93	2,073	0.228
23	0920	0.70	0.10	2.2	0.25	0.18	1.86	2,419	0.203

Date	Time	Gage ht (ft)	q (ft^3/sec)	w (ft)	$\bar{d}$ (ft)	$\bar{v}$ (ft/sec)	ν (10^{-5} ft^2/sec)	N_{Re}	n
24 Sept	1100	0.71	0.10	2.2	0.27	0.17	1.80	2,550	0.226
25	1030	0.70	0.10	2.2	0.27	0.17	1.80	2,550	0.226
28	1030	0.72	0.11V						
29	0950	0.72	0.11V						
30	0925	0.72	0.11V						
1 Oct	0940	0.73	0.11V						
2	0935	0.73	0.11V						
5	1350	ICE	0.09V						
6	1030	ICE	0.08V						
7	1000	0.73	0.09V						
8	1010	0.73	0.10V						
9	1110	0.77	0.15V						
12	0945	ICE	0.08V						
13	0920	0.80	0.11V						
14	1000	ICE	0.08V						
15	0940	ICE	0.08V						
16	1045	ICE	0.06V						
19	1030	ICE	0.03V						
20	1030	ICE	0.04V						
21	1030	ICE	0.04V						
22	1030	ICE	0.04V						
23	1030	ICE	0.03V						

q = discharge; w = width; $\bar{d}$ = average depth; $\bar{v}$ = average velocity; ν = kinematic viscosity; N_{Re} = Reynolds number; n = roughness factor. P = peak stage; E = discharge estimated from stage-discharge relationship (Fig. 24); V = volumetric measurement.

APPENDIX D: SUMMARY OF DATA USED IN COMPUTING RECESSION CONSTANTS

Recession period	t (hr)	q (ft^3/sec)
1964 11 June 1015 - 16 June 0945	0	2.14
	23.50	0.84
	48.25	0.43
	73.75	0.25
	95.50	0.18
	119.50	0.15
1964 25 June 1010 - 28 June 1425	0	1.00
	1.00	0.97
	5.25	0.89
	24.25	0.45
	49.50	0.20
	76.25	0.13
1964 12 July 1630 - 17 July 1400	0	1.42
	18.25	0.68
	41.75	0.33
	66.75	0.17
	89.75	0.10
	117.50	0.06
1964 16 Aug 2000 - 19 Aug 1115	0	1.61
	15.25	1.14
	42.50	0.54
	63.25	0.40
1966 18 June 1200 - 20 June 0800	0	3.46
	4.0	3.01
	8.0	2.20
	12.0	1.66
	16.0	1.33
	20.0	1.10
	24.0	0.92
	28.0	0.77
	32.0	0.67
	36.0	0.60
	40.0	0.53
	44.0	0.49
1966 3 July 1200 - 5 July 2400	0	0.026
	4	0.019
	8	0.018
	12	0.018
	16	0.018
	20	0.019
	24	0.015
	28	0.0092
	32	0.0085
	36	0.010
	40	0.011
	44	0.012
	48	0.013
	52	0.0096
	56	0.0066
	60	0.0066
	64	0.0080
1966 7 July 1400 - 9 July 1400	0	0.059
	4	0.048
	8	0.044
	12	0.040
	16	0.040
	20	0.039
	24	0.037
	28	0.032
	32	0.031
	36	0.031
	40	0.031
	44	0.031
	48	0.028
1966 13 July 1100 - 16 July 1900	0	0.035
	4	0.025
	8	0.020
	12	0.020
	16	0.020
	20	0.020
	24	0.019
	28	0.015
	32	0.012
	36	0.012
	40	0.012
	44	0.013
	48	0.013
	52	0.0080
	56	0.0051
	60	0.0050
	64	0.0066
	68	0.0080
	72	0.0079
	76	0.0051
	80	0.0035
1967 10 July 1300 - 14 July 1700	0	0.12
	4	0.10
	8	0.094
	12	0.086
	16	0.086
	20	0.078
	24	0.064
	28	0.051
	32	0.045
	36	0.045
	40	0.045
	44	0.040
	48	0.030
	52	0.022
	56	0.019
	60	0.019
	64	0.019
	68	0.019
	72	0.013
	76	0.008
	80	0.008
	84	0.008
	88	0.008
	92	0.008
	96	0.005
	100	0.002

Recession period	*t (hr)*	*q (ft^3/sec)*	*Recession period*	*t (hr)*	*q (ft^3/sec)*
1967 22 July 1100 - 23 July 2300	0	1.62	1967 25 July 0700 - 26 July 0700	0	4.70
	4	1.53		4	3.85
	8	1.19		8	3.02
	12	0.93		12	2.38
	16	0.80		16	1.93
	20	0.68		20	1.67
	24	0.57		24	1.40
	28	0.48	1967 10 Aug 1000 - 11 Aug 1400	0	1.62
	32	0.41		4	1.48
	36	0.37		8	1.35
				12	1.19
				16	1.07
				20	0.96
				24	0.90
				28	0.80

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13. ABSTRACT

The results of a four-summer (1964-1967) hydrologic study of the watershed of Glenn Creek, about 8 miles north of Fairbanks, Alaska, in the Yukon-Tanana uplands physiographic province, are presented. This work was initiated to provide initial base line hydrologic data for a small subarctic watershed, the first of its kind in North America. Standard hydrologic and meteorologic instrumentation was used, and streamflow characteristics were analyzed by standard hydrograph-analysis techniques. The stream is second-order, and drains an area of 0.70 square mile. Basin elevations are from 842 ft to 1618 ft. In regard to topography, geology, soils, permafrost, vegetation, and climate, the watershed seems to be representative of low-order, low-elevation drainage basins in the province. Analysis of rainfall-runoff data indicates that about half the 12.3-in. normal annual precipitation is runoff. The remainder is the actual evapotranspiration, which equals only about 30% of estimated potential evapotranspiration. For individual storms, runoff/rainfall proportions were from 0.03 to 0.42, and were positively correlated with antecedent discharge of the stream, which is a measure of watershed wetness. The stream responds rapidly to rainstorms except when the basin is very dry, and has markedly slow recessions compared with temperate-region streams of similar size. Rate of recessions is apparently controlled by concurrent evapotranspiration rates. Analysis of hydrographs and knowledge of the physical characteristics of the basin indicate that storm runoff occurs initially as surface runoff from bare soil areas adjacent to the stream, while recessions are dominated by a combination of tunnel flow beneath moss-covered parts of the basins and typical groundwater flow through the moss and soils. Peak discharges for individual

14. KEY WORDS	LINK A		LINK B		LINK C	
	ROLE	WT	ROLE	WT	ROLE	WT
Central Alaska - climatology						
Central Alaska - hydrology						
Evapotranspiration						
Geology						
Hydrology						
Hydrometeorology						
Permafrost						
Precipitation						
Soil properties						
Stream flow						
River basins						
Runoff						
Vegetation						
Watersheds						

13. Abstract (Cont'd)

storms could be well estimated by an equation including antecedent discharge, total precipitation and storm duration, and average recession constant. These results represent the first detailed hydrologic data from the discontinuous permafrost zone of the North American taiga and should be of significance to the International Hydrological Decade and International Biological Program.