Postglacial Drainage Evolution and Stream Geometry in the Ontonagon Area, Michigan

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By JOHN T. HACK

SHORTER CONTRIBUTIONS TO GENERAL GEOLOGY

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A parallel drainage network controlled by glacial grooves provides data for the study of fluvial geomorphic phenomena, especially the geometry of stream bifurcations

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POSTGLACIAL DRAINAGE EVOLUTION AND STREAM GEOMETRY IN THE ONTONAGON AREA, MICHIGAN

By John T. Hack

ABSTRACT

The clay plain on the south shore of Lake Superior in Ontonagon County, Mich., is underlain by glacial deposits of Valders and post-Valders age consisting of till interbedded with fine-grained glaciolacustrine sediments. The surface of the plain is strongly grooved by glacial flutings parallel to the direction of ice motion. The grooves have a regular wavelength that averages 380 feet. In places the grooves are buried by lacustrine sediments of glacial Lake Duluth. Seven or more glacial lake shorelines cross the plain, in places interrupting the grooves. Six shorelines between 650 and 1,200 feet above sea level are attributed to glacial Lake Duluth. One ancient shoreline at about 615 feet altitude is attributed to Lake Nipissing. Lake Duluth was short lived. It formed and withdrew completely between 10,300 and 9,500 before present. More conspicuous shore features were formed by Lake Nipissing at its stand only 15 feet above the level of present Lake Superior.

As Lake Duluth withdrew from the Ontonagon Plain, a drainage network formed. The streams followed the grooves that extended downslope toward the lake. The resulting drainage pattern is strongly attenuated downslope, but, except for the attenuation, is similar in its geometry to drainage patterns in well-graded landscapes. The evidence available supports the theory that the spacing of stream bifurcations or junctions in a typical drainage network is random and that stream lengths tend to be proportional to the six-tenths power of the drainage area because of the random arrangement of channels. Very little change has taken place in the Ontonagon drainage pattern as a result of piracies or headward migration of divides. The depth of incision of the present valleys is roughly proportional to a function of the drainage area and there is little or no evidence to support the idea of nickpoints that migrate upstream. Instead, the valleys are cut to depths proportional to the available discharge.

The longitudinal profiles of the larger streams are slightly concave. The streams are adjusted to the bed and bank materials and those in bedrock have steeper gradients than those in unconsolidated deposits. Most of the streams have intricate meander patterns, both of their channels and their valleys. Streams flowing in bedrock, however, have straighter courses. Meander wavelength increases in a regular manner as average discharge increases, and is also affected by the materials in the valley; streams in bedrock have meanders about four times the wavelength of streams in till. It is concluded that the evolution of the drainage system in the Ontonagon area has proceeded in response to upstream factors, such as available discharge and the geology of the drainage basin. Changes in base level have affected the streams mainly by changing the relief and the available potential energy in the landscape.

INTRODUCTION

The Ontonagon area includes an extensive clayey plain south of the Lake Superior shore. This plain, referred to herein as the Ontonagon Plain, is deeply trenched by an unusual drainage network that has developed since the Valders glaciation and the draining of glacial Lake Duluth about 9,500 years ago. Because of the unusual nature of the drainage network, the short time in which it has developed, and the controls available by which to date the development, the area is one where much can be learned about the evolution of a drainage system and some of the geologic factors that are involved in the equilibrium conditions in stream channels.

The plain lies on the south shore of Lake Superior between the Porcupine Mountains and the Firesteel River; it is bordered on the south by a range of bedrock hills, known as the Copper Range, that rises more than 1,000 feet above the lake. The plain is about 8 miles wide and slopes toward the lake with an average gradient of 50 feet per mile. It is underlain by fine-grained glacial-lake sediments and till. A remarkable series of long shallow grooves crosses the plain more or less parallel to the direction of slope. Because the grooves have controlled drainage development, the streams flow from the Copper Range across the plain in straight but narrow valleys and have narrow attenuated drainage basins.

Most of the study of the drainage network was done by means of aerial photographs and topographic map interpretation. Photogrammetric techniques were employed in studies of some areas. A total of 9 weeks, however, in 1958 and 1960 was spent in the field in order to examine the materials that enclose the stream channels and to make observations of the bed materials of the streams at various localities (pl. 1). Because the deposits beneath the plain have not previously been described, a reconnaissance study was made of the glacial geology. The stratigraphy and petrography of the glacial deposits beneath the Ontonagon Plain were studied by Friedrich Brandtner who was in the field with the writer in 1960.

The choice of the area as a place to study a drainage network was suggested by Walter S. White of the U.S. Geological Survey, who noted the grooves and recognized that they were produced by glacial scouring. Much of the detailed study of glacial deposits was done on the property of the White Pine Copper Co. Many courtesies and valuable help were extended by the company.

CLIMATE AND STREAMFLOW

The Ontonagon plain has a humid but cold continental climate. At Bergland, south of the Copper Range, the mean monthly temperature ranges from 11°F in January to 67.7°F in July. The average growing season is only 84 days and precipitation in the form of snow is common from November to May. Precipitation averages 32 inches and is only slightly seasonal, being heaviest in the late spring and summer and lightest in the fall. Snowfall just south of the Lake Superior shore averages 115 to 130 inches (U.S. Dept. Agriculture, 1941). Lake Superior has a marked ameliorating effect on the climate of the Ontonagon Plain, as is well shown by the vegetation. Whereas spruce and fir forests are prevalent south of the Copper Range, the area between these hills and Lake Superior is occupied by a typical northern hardwood forest. The most abundant trees are hemlock, maple, birch, and aspen; oaks are found in limited numbers on dry, generally south-facing slopes.

Within the study area, the only stream gage measuring flow uncontrolled by a dam is on the Ontonagon River west of Rockland (pl. 1). The gaged flow at this point averages 1,320 cfs (cubic feet per second) from a drainage area of 1,290 square miles (U.S. Geol. Survey, 1961). Examination of the records at other gaging stations on streams in northern Michigan tributary to Lake Superior indicates that the average discharge in cubic feet per second of each of the streams is roughly equal to its drainage area in square miles (fig. 1). Seasonal fluctuations are large. The highest average flow over an extended period generally is in April and May when mean monthly discharges more than 5 times the mean annual discharge are common. Momentary peak discharges of more than 15 times the mean annual have been recorded in almost every stream, though none of the stations have a very long period of

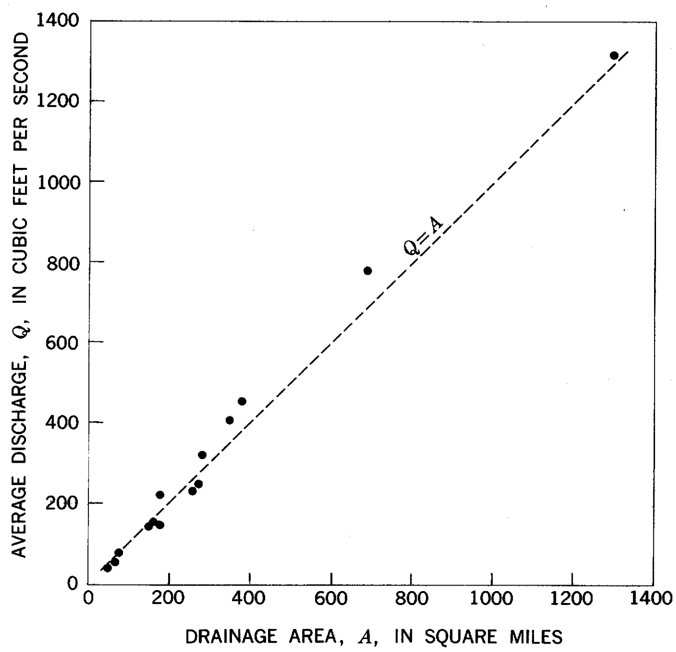


FIGURE 1.—Relation of average discharge to drainage area at gaging stations in northern Michigan west of the Sturgeon River, in streams tributary to Lake Superior.

record. The peak flows are not necessarily in the spring. The largest flood in the region during the 18 years prior to 1960 occurred on August 18, 1942, when the discharge on the Ontonagon River at Rockland was estimated at 42,000 cfs, almost 40 times the mean.

Comparison of the streamflow characteristics of the Ontonagon Plain with those of the central Appalachians, where the writer has made similar studies (Hack, 1957), indicates that except for peak flows the stream regimens are not very different. On the Ontonagon Plain the seasonal variation in flow is more pronounced and regular. The peak recorded discharge, however, is much less. In the Potomac River basin of Virginia, West Virginia, and Maryland, for example, many streams have recorded momentary discharges more than 100 times the mean annual.

SUMMARY OF BEDROCK GEOLOGY

The Ontonagon area is mostly within the outcrop area of the Keweenawan Series of Precambrian age. The character of the bedrock has a strong influence on the topography and on the material in the glacial deposits. The following description of the rocks is taken from papers by Van Hise and Leith (1911), Butler and Burbank (1929), White and Wright (1954), Hamblin (1961), and Hamblin and Horner (1961).

Portage Lake Lava Series.—These rocks of middle Keweenawan age underlie the hills known as the Copper Range (fig. 2). They are bordered on the south by the

Keweenaw fault and on the north are overlain unconformably by the Copper Harbor Conglomerate. The Portage Lake Lava Series includes more than 20,000 feet of interbedded lava flows and thin conglomerate. Most of the lavas are basaltic in composition but felsite and other silicic rocks are included in the lava series. In the Porcupine Mountains there is a large area of such rocks. The lava series contains many copper-bearing beds, particularly the interbedded conglomerate and scoriaceous tops of the lava flows. Once-famous mining districts no longer in production are at Norwich, Victoria, Rockland, Greenland, and Mass.

Copper Harbor Conglomerate.—The Copper Harbor Conglomerate of late Keweenawan age underlies the northern and eastern slopes of the Porcupine Mountains and the north side of the Copper Range. The rocks consist mostly of reddish, rudely stratified conglomerate with rhyolitic boulders predominating. Medium- to coarse-grained sandstone, some thin and persistent, some lenticular, is interbedded in the conglomerate. Thin beds of volcanics also occur. The thickness of these rocks in the Ontonagon area is 2,300 to 5,500 feet (White and Wright, 1954).

Nonesuch Shale.—The Copper Harbor Conglomerate is overlain by a band of dark-gray siltstone, shale, and sandstone, generally 600 feet thick. These rocks (the Nonesuch Shale of late Keweenawan age) contain a cupriferous zone near the base which, at White Pine, is

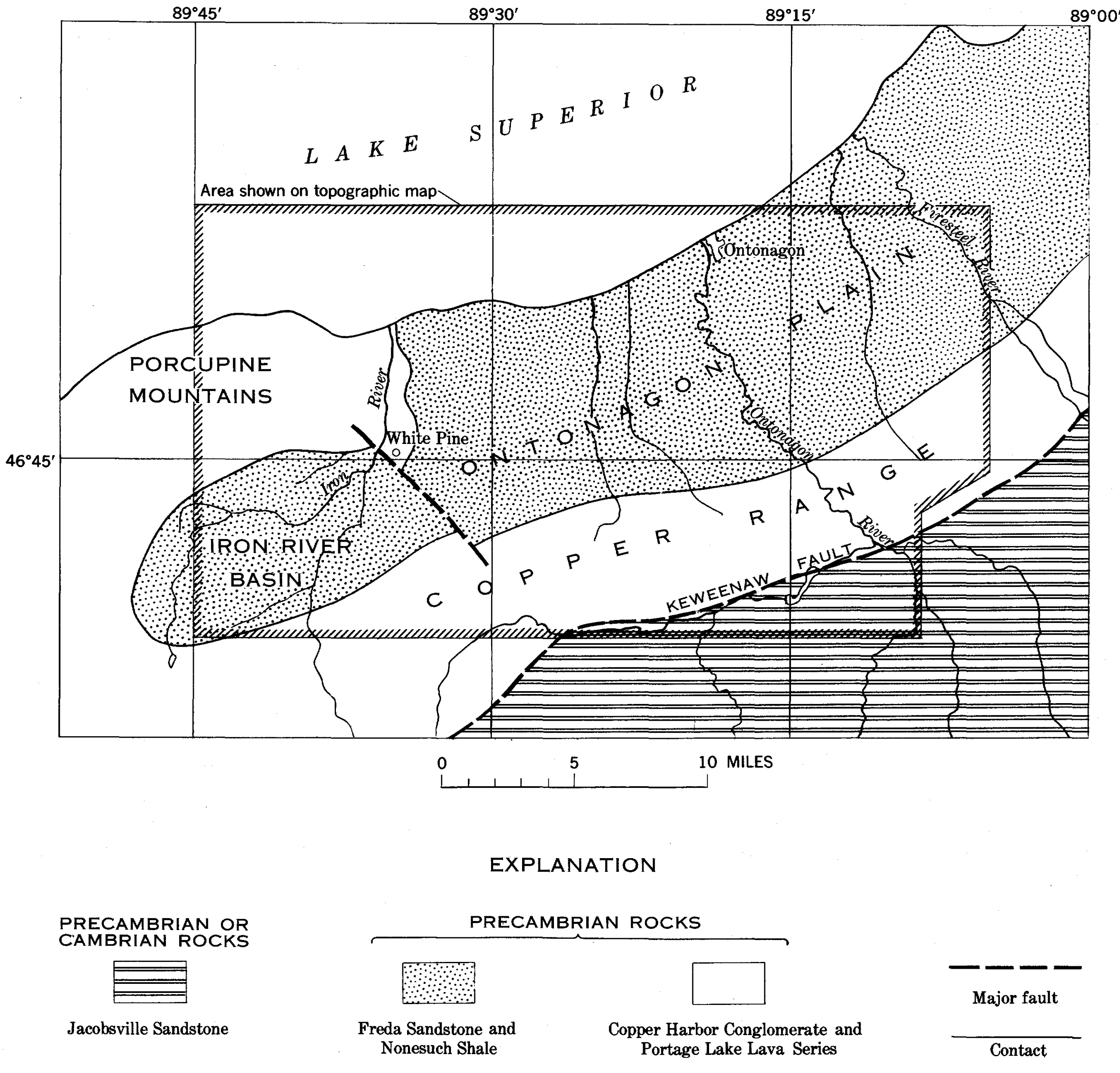


FIGURE 2.—Simplified geologic map of the Ontonagon area showing major geographic and geologic divisions.

mined for copper. The Nonesuch Shale forms a band along the south edge of the Ontonagon Plain and underlies an extensive area in the Iron River basin. It crops out only in the bottom of the deeper valleys and along the Lake Superior shore near Silver City (pl. 1).

Freda Sandstone.—This formation of late Keweenawan age consists of alternating layers of reddish-brown fine arkosic sandstone and red micaceous silty shale. Some conglomerate occurs near the base. The material is derived mostly from basaltic lavas. Quartz is not abundant. Ripple marks, raindrop imprints, and cross-stratification are common. The Freda Sandstone is more than 14,000 feet thick and crops out extensively along the Lake Superior shore and in many valley bottoms of the Ontonagon Plain. Inasmuch as this sandstone underlies glacial deposits over a wide area and also, from the shoreline to well out into the lake, forms the lake bottom, it is undoubtedly the source material of much of the till that underlies the Ontonagon Plain, and has imparted to the till its characteristic reddish-brown color.

Jacobsville Sandstone.—The area south of the Copper Range is underlain by gently dipping sandstone beds of Precambrian or Cambrian age. The Jacobsville Sandstone is separated from the Portage Lake Lava Series by the Keweenaw fault. The rocks are quartzitic rather than arkosic and, though reddish, are of considerably lighter shade than the Freda Sandstone. They are medium grained, well sorted, and crossbedded, and contain lenses of conglomerate. Outcrops in the Ontonagon area are not extensive and are confined to the valley of the Ontonagon River and its tributaries above Rockland.

Structure.—The structure of the Keweenawan rocks, except those in the Porcupine Mountains, is monoclinal, the dip being to the north. In the Copper Range, dips may exceed 50°, but they flatten to the north and, under the Ontonagon Plain, are generally less than 10°. The Porcupine Mountains are a structural dome or arch and have steep dips on both the north and south flanks.

PLEISTOCENE DEPOSITS

The glacial geology of the Lake Superior region is known principally from the reconnaissance work of Leverett (1928), though his chronology and nomenclature have been modified by more recent work in various parts of the region. This newer work has been summarized by Hough (1958, 1963). The Ontonagon area is entirely in the belt of late Wisconsin drift commonly referred to as Valders. The terminal moraine of the Valders Stade lies about 60 miles south of Ontonagon (Flint and others, 1959). A minor recessional moraine traverses the Copper Range, though till is thin on most of the hills and bedrock exposures are numerous. The Ontonagon Plain and the Iron River basin are underlain by thick layered deposits of till and lake sediments.

Till and outwash of the Copper Range and Porcupine Mountains.—In the hilly areas of the region, drift is thin and outcrops are numerous, especially on the south sides of the hills. Most of this thin till is different from the thick till of the Ontonagon Plain; it is more stony, contains volcanic rocks derived locally, and has a loamy matrix. One area of glacial outwash along Sleepy Creek in sec. 7, R. 42 W, T 49 N occupies about 1 square mile.

Glacial till and lacustrine deposits of Ontonagon Plain and Iron River Basin.—Most of the area north of the Copper Range is underlain by reddish-brown glacial till and lake deposits which are interbedded in many places. The arrangement of the deposits suggests that they were laid down by an ice sheet that retreated into an ice marginal lake and then readvanced. In the western part of the area where they were examined in greatest detail, the deposits can be subdivided into three distinct beds referred to as the lower, intermediate, and upper units. The lower unit consists of stony till containing subangular boulders and fragments derived mostly from the bedrock under the Ontonagon Plain. The bedrock over most of the area is the Freda Sandstone, which imparts a silty to fine sandy texture and reddish-brown color to the till matrix. The intermediate unit is in distinct layers or strata consisting of till and laminated silt and clay believed to be lacustrine sediments. Some of the layers contain large erratic boulders of gabbroic and basaltic rock, though in general, this unit is much less stony than the lower unit. The upper unit is a clayey till that is much less stony than either of the lower units. There are few erratics of large size, and a smaller proportion of the stones is derived from the Freda Sandstone.

Younger lacustrine deposits.—The upper clayey till is overlain in places by strongly laminated clay, silt, and sand that vary widely in thickness. These are typical lacustrine deposits presumably laid down in late Pleistocene lakes marginal to the ice. South of the Copper Range in the eastern part of the area, where the three branches of the Ontonagon River come together, the lacustrine deposits make up almost the entire section above the bedrock and are more than 200 feet thick. This area was occupied by an ice-marginal lake known as Lake Ontonagon (Leverett, 1928, p. 57). The lake formed when the glacier stood on the south side of the Copper Range and blocked the Ontonagon River Basin. When the ice retreated north of the Copper Range, a large part of the present Lake Superior basin was uncovered, and an ice-marginal lake formed that occupied half of the basin and had its outlet near Duluth, Minn. The shoreline of the lake, called Lake Duluth, is below an elevation of 1,200 feet in the Ontonagon area. The lake extended through the lowest gaps of the Copper Range and its south shore lay south of the area shown in plate 1. The north shore was presumably bounded by ice and lay somewhere north of the area. The deposits of glacial Lake Duluth are mostly a thin veneer and are somewhat patchy in distribution. East of the Ontonagon River they are, in places, 20 feet thick or more; they form a thick fill in the valleys of the Flintsteel and Firesteel Rivers where they rest on till. In most of the area west of the Ontonagon River the lacustrine deposits form a veneer less than 2 feet thick. In some places they are lacking entirely.

THICKNESS OF THE GLACIAL DRIFT

The bedrock surface beneath the drift is quite irregular and the pre-Valders topography of the Ontonagon Plain must have had valleys at least 100 feet deep. As shown in holes drilled by the White Pine Copper Co., the drift near the Iron River is in places more than 150 feet thick; in the vicinity of the tailings pond, however, it is only 20 to 30 feet thick and it continues to thin to the north to less than 10 feet. Along the Cranberry River the drift ranges in thickness from more than 80 feet to a minimum of 10 feet. Along the Ontonagon River valley the drift in places is at least 200 feet thick and the absence of bedrock exposures in the valley walls suggests that the Ontonagon River follows a preglacial course.

The present drainage system is deeply intrenched; in some places the valley walls are mostly bedrock and in others mostly drift. Unfortunately, not enough data are available to show the character of the preglacial topography under most of the Ontonagon Plain.

RECENT DEPOSITS

Shore deposits older than Lake Nipissing.—Sandy deposits, that are apparently partly dune sand, but that may in part be delta sand occur on the Ontonagon Plain near the Lake Superior shore. These deposits veneer the upland plain inland from the scarp that marks the shoreline of Lake Nipissing. They form a belt averaging about a mile wide that widens to 2 miles near the Ontonagon River. In part, they may be dunes blown up over the low escarpment from the Nipissing beach. Near the Ontonagon River, however, the sand is too thick to have had this origin, for it forms the entire escarpment as though deposited in a delta when the Ontonagon River was at a higher level.

Lake Nipissing and younger beaches.—The Ontonagon Plain ends abruptly about 1,000 feet south of the Lake Superior shore where it drops from 5 to 30 feet to a low sandy belt that borders the lake. This belt is a complex of beach and dune sand, partly shore deposits of Lake Nipissing, which stood at an altitude of about 615 feet, and partly shore deposits of modern Lake Superior. The deposits are coarse sand and in places, especially marginal to the larger streams, gravel.

Recent stream deposits.—Most of the streams longer than a mile or two are intrenched in the till or in bedrock, and are bordered by a floodplain or, in places, by terraces. These deposits are generally reddish brown, similar in color to the glacial drift. They are derived both from the drift and the Precambrian bedrock, and are typical graded-stream deposits, generally gravelly at the base and sandy or silty at the surface. The gravel has a wide range in size and commonly contains large boulders that have been eroded from till as the streams cut through it.

SURFACE FEATURES OF THE ONTONAGON PLAIN

The surface irregularities on the Ontonagon Plain and their origin are of particular importance in relation to the drainage network because they controlled, in part, the initial drainage pattern that formed as the glacial lake withdrew from the area. The most important of these features are the glacial grooves and the glacial-lake shorelines. Other features of the plain that have had less influence on the drainage are kettle holes and beaver ponds. Glacial moraines occur in the Copper Range, but none have been found on the Ontonagon Plain itself.

GLACIAL GROOVES OR FLUTES

Prominent grooves cross the Ontonagon Plain and are evident in most of the more or less level areas. The grooves have not been previously described in the Ontonagon area but they were noted by W. S. White (oral commun., 1957) who interpreted them correctly as glacial grooves. Somewhat similar grooves, or flutings, of about the same wavelength have been described in Alberta, Canada (Gravenor and Meneley, 1958). The grooves in the Ontonagon area are not developed equally but are more distinct and densely distributed in the inner part of the plain. They become indistinct downslope toward Lake Superior. Grooves are completely lacking in low swales, as for example along the Ontonagon River and the Flintsteel and Firesteel Rivers. In these areas they are presumably covered by lake sediments.

The best display of grooves is in the area to the east of White Pine on the lakeward side of the Algonquin shore of Leverett which crosses the plain at about 940 feet. This set of grooves consists of a series of long parallel ridges and troughs that are a regular distance apart, much like a series of waves. The average wavelength measured in two traverses between White Pine and the Cranberry River is 380 feet. Grooves unmodified by running water are generally less than 10 feet deep, but many grooves are deeper because they have been eroded and scoured by water.

Not all the grooves in the Ontonagon area trend in the same direction (fig. 3); rather, they reflect the direction of motion of the ice. The fact that in some places one set of grooves crosses another suggests that groups of grooves were formed at different times by ice that moved in slightly different directions and that younger sets obliterated older ones. The grooves are not continuous nor all of the same length. Some are as much as 3 miles long; others are less than a quarter of a mile long.

Similar grooves or glacial flutings in Alberta, Canada, have been studied by Gravenor and Meneley (1958). Their data show that the flutings are parallel to the direction of ice motion. The Alberta grooves resemble closely the Ontonagon grooves and their wavelengths are similar. The mode of the wavelength frequency distribution in Alberta is between 350 and 400 feet; in the Ontonagon area the mode is approximately 350 feet and the average is 380 feet. Gravenor and Meneley (1958, p. 724) made microfabric studies of the till beneath the Alberta flutings and found that at depths less than 5 feet the particles had a strong preferred orientation parallel to the flutings. At depths of 11 feet and below the orientation was not parallel to the flutings. In the Ontonagon area, F. Brandtner in 1960 made fabric analyses of the till at three localities and found that the pebbles in the upper till showed a strong preferred orientation roughly parallel to the grooves (fig. 3).

In order to further establish the direction of ice motion in relation to the grooves, the writer searched bedrock outcrops in the Copper Range and the Porcupine Mountains for glacial striations. Nine localities were found within the area shown on plate 1 that give clear evidence of the direction of ice motion (fig. 3). Such glacially polished bedrock surfaces are not common on the volcanic rocks because of the rapidity of

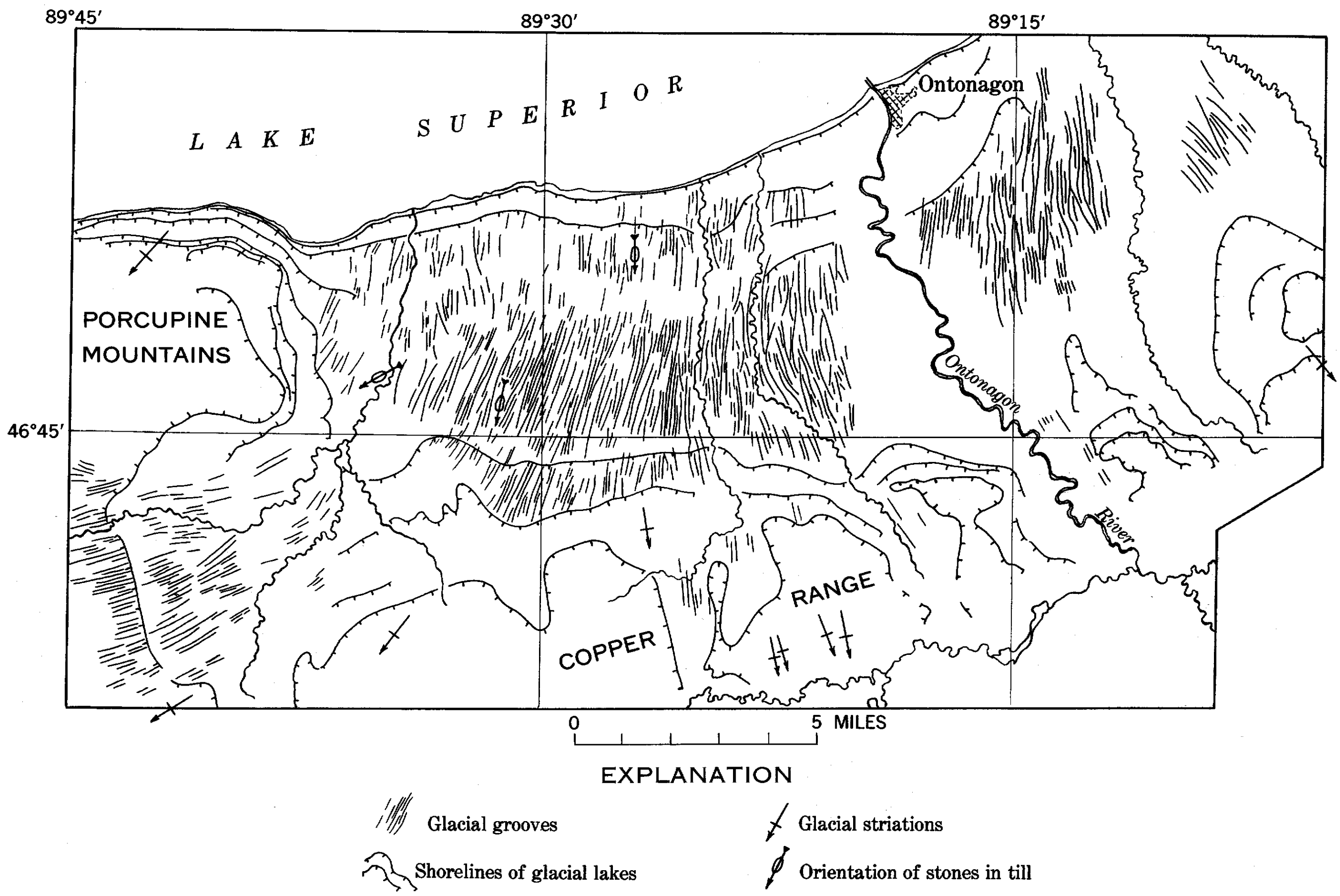


FIGURE 3.—Ontonagon area showing minor surface features including glacial grooves, glacial-lake shorelines, and localities where direction of ice movement could be determined.

weathering, but they can be observed in some places by stripping off the cover of soil and moss, if it is thin enough, and washing the rock surface with water and a scrub brush. In every case where striae or grooves on the rock surface were found, their directions corresponded to the directions of the large grooves on the surface of the till plain nearby.

As shown in figure 3, in some parts of the Ontonagon Plain the grooves are missing. In some places, as along the Flintsteel and Firesteel Rivers, this is because the surface is underlain by thick lacustrine deposits that are younger than the grooves. The grooves, if present, are buried.

The grooves were modified slightly by shore erosion and deposition after the ice withdrew. The thin cover of lake sediments that occurs on most of the surface, however, is too thin to obscure the grooves. The continuity of the ridges, however, is interrupted in various places and the grooves are weaker along the old shorelines, if not entirely obliterated.

GLACIAL LAKE SHORELINES

The traces of at least seven glacial lake shorelines were found in the Ontonagon area (fig. 3). These features are best observed on aerial photographs on which changes in texture of the vegetation or breaks in continuity of various surface features are apparent. The shorelines are most conspicuous in bedrock areas where gravelly beaches and nips cut in the rock are well preserved. In till areas, the shorelines are rather obscure and some cannot be found on the ground. Locally, gravelly spits, bars, and beach ridges are used as sources of aggregate and road metal.

Highest Lake Duluth shore.—The best most complete of the higher shore lines found by the writer in the area shown on plate 1 corresponds to what Leverett (1928) called the highest Lake Duluth shore. In most of the area it is marked by a prominent nip with sand and gravel deposits in front of it. Spits, bars, and even tombolos are well formed. The shore is readily traced on aerial photographs and also can be seen on the ground without difficulty almost anywhere along its trace. According to Leverett (1928, p. 62), the Lake Duluth shore is tilted and rises in elevation toward the east from 1,163 feet in the Porcupine Mountains to 1,192 feet near Greenland; it rises sharply near the Ontonagon River. The writer's tracing of this shore, based mostly on aerial photographs, corresponds closely to Leverett's description and the shore apparently rises from a minimum of 1,150 feet in the Porcupine Mountains to 1,190 feet north of Greenland, a difference of 40 feet.

There are at least two shorelines between the Lake Duluth and the Lake Algonquin shore of Leverett. Only the lower of these two can be traced across the Ontonagon Plain and even it is discontinuous. It is most distinct south of White Pine where gravel has been mined from it for use on the railroad. This shoreline rises from 1,020 feet to 1,060 feet from west to east.

Lake Algonquin shore of Leverett.—A fairly distinct shorelines occurs at an altitude of 920 feet in the Porcupine Mountains and at 960 feet in the Greenland-Rockland area. This shoreline corresponds to the one that Leverett (1928) referred to as the Lake Algonquin shore. It is well formed and fringed by gravelly beach deposits from the Porcupine Mountains eastward to and beyond White Pine. It is also well formed in the Greenland-Rockland area, but between these places it is indistinct and, though traceable on aerial photographs, it cannot everywhere be found on the ground in the course of a traverse. It crosses and obscures the glacial grooves, however, and this is the best means of recognizing it (fig. 3).

At least three lower shorelines cross the plain between Leverett's Algonquin shore and the Lake Nipissing shore. The highest of these is fairly well marked by its effect on the glacial grooves. It rises from 715 feet in the Porcupine Mountains to 755 feet east of Ontonagon.

Lake Nipissing shore.—A very conspicuous shoreline occurs at an elevation between 615 and 620 feet above sea level or approximately 15 feet higher than the present shore of Lake Superior. The base of the nip is covered by bars and dune sand, and the writer did not determine whether this shore is tilted within the area studied; the tilt, if present at all, must be slight. The Lake Nipissing shore is much better developed than any of the others, including the highest shoreline of Lake Duluth. The nip or scarp is very well marked and in places is 20 feet high, though generally only 5 to 10 feet high. Furthermore, a fairly wide belt south of the cliff is mantled by sand presumably blown from the shore below. On the lakeward side, the shoreline is fringed by a wide belt of beach ridges and dunes that form a sandy complex far larger than any of the older beaches.

AGE OF THE SHORELINES

In the late Pleistocene and Recent chronology of the ancient Great Lakes outlined by Hough (1963), the Valders ice retreated rapidly and glacial Lake Duluth must have been short lived. The maximum advance of the Valders ice sheet occurred at about 11,000 B.P. (before present). The ice retreated from the area and formed the ice-marginal Lakes Ontonagon and Duluth; it eventually reached a level below the present shore of Lake Superior when the water of the basin drained over a sill at Sault Ste. Marie into lower lakes to the east.

The age of these lower lakes is thought to be about 9,500 B.P. (Hough, 1963, p. 103). Thus the ice must have retreated from northern Michigan, including the Ontonagon area, and the glacial Lake Duluth drained within a period of less than 1,500 years.

Samples of wood collected by F. Brandtner in 1960 provide a local chronology in agreement with Hough's analysis, and show that within the Ontonagon area itself the retreat of the ice and draining of the lake probably occupied an even shorter time. Brandtner collected samples of wood from a kettle-hole swamp near White Pine uncovered by a power shovel during excavation for a tailings pond of the White Pine Copper Co. The kettle is near Caribou Creek in sec. 2, T. 50 N., R. 42 W., at an altitude of about 880 feet, 280 feet below the high stage of Lake Duluth. A log found in the upper clayey till beneath the kettle hole yielded a carbon-14 date of 10,230±280 years B.P., as determined by Meyer Rubin of the U.S. Geological Survey (W–964). Dates obtained from a log found within the kettle and from gyttja at the base of the kettle were 9,600±350 B.P. (W–965) and 9,500±350 B.P. (W–1150) respectively. The date of the till presumably represents a minor readvance of the Valders ice. The kettle could not have contained a fill of gyttja until the water of Lake Duluth had withdrawn from the area. These dates indicate that Lake Duluth may have formed and drained in a period less than 780 years.

The rapid withdrawal of water from the Ontonagon Plain is also suggested by the uniform tilt of the traces of the successive shorelines. All the older shore lines are tilted nearly the same amount and they rise about 40 feet between the Porcupine Mountains and the area of the Ontonagon River. By contrast, the Lake Nipissing shore has no measurable tilt within the area studied; furthermore, none of the older shorelines are prominent features as compared with the Lake Nipissing shore.

The evidence on the Ontonagon Plain seems to be in close agreement with the chronology of Hough (1958, 1963) but not with that of Leverett (1928). Leverett believed that Lake Algonquin included the area of the Superior basin and formed a prominent shoreline in the Ontonagon area at 920–960 feet. Although a shoreline does cross the plain at that level, it is tilted the same amount as the other shorelines, including the high shore of Lake Duluth. The kettle hole from which the date of 9,500 B.P. was obtained lies below Leverett's Algonquin shore. Thus, the evidence indicates that all the shorelines north of the Copper Range older than the Lake Nipissing shore should be assigned to the Duluth stage.

No evidence was found in the area by which to date the Nipissing stage, but it is well documented in other areas. The high stage presumably occurred at 4,200 B.P. (Hough, 1963).

KETTLE HOLES

Small subcircular kettle holes, which have been referred to locally as "bathtubs", are abundant on the Ontonagon Plain. They are especially numerous in places where the glacial grooves are best developed. Very few of these kettles are more than 100 feet in diameter and most are about 50 feet or less. Some are ponds or lakes, others are bogs, and still others are dry sediment-filled depressions. The larger kettles are readily visible on aerial photographs.

Their position at the top of the till sequence and the fact that they contain undisturbed lake sediments indicate that the kettles must have formed during the last advance of the ice, probably contemporaneously with the grooves; they must also have formed either under the ice or under water. They probably are holes left by ice that became entrapped in the clayey till and later melted. These features are not large enough to have influenced the development of the drainage.

BEAVER PONDS

The dense distribution of beaver dams and ponds on the Ontonagon Plain suggests that beavers may have influenced the drainage development. Residents of the region classify the beavers as bank beavers and pond beavers, depending on the environment in which they make their homes. The bank beavers inhabit the intrenched stream valleys, including streams as large as the Cranberry River. They build small dams across the channel, but these dams are low and generally do not reach from one floodplain to the other. Because beaver lodges are generally not in evidence along the large streams, the animals presumably inhabit burrows in the stream banks. Dams on the large streams are temporary and probably have no permanent influence on the channel equilibrium. The habit of some beavers of digging bank burrows along streams that are regularly subject to floods is apparently well known to biologists and has been described by Dugmore (1915, p. 28).

The pond beavers live in more permanent bodies of water on the upland area of the plain. Glacial grooves make excellent enclosing walls for lakes and beaver ponds, averaging one or two per square mile, are especially abundant in the grooved areas. Often the dams are very elaborate and long-lasting structures on which shrubs and trees grow and which may enclose bodies of water many acres in size. The pond beavers live in typical beaver lodges that are much in evidence at the upland sites.

The pond beavers may have some influence on the geomorphic development of the landscape. It is possible that some of their dams are large enough to divert small upland streams from one groove to another. Study of the aerial photographs indicates that such diversions have been rare, however, because most dams do not rise as high as the top of the ridge between the grooves.

CHRONOLOGY OF EMERGENCE OF THE ONTONAGON PLAIN

The chronology of events during the uncovering of the area and the changes in position of the lake shore have an important bearing on the development of the present subaerial drainage system, for the lake shore is the base level of erosion. The events that took place following the high stage of Lake Duluth are summarized below as they are now known from evidence found in the Ontonagon area, as well as from Hough's (1958, 1963) synthesis of the history of the Great Lakes:

11,000–10,000 B.P.—Valders ice retreated from the area and formed the ice-marginal Lake Duluth that covered the Ontonagon Plain to an altitude of 1,100–1,200 feet above present sea level. This open water body had sufficient fetch for wave energy to cause considerable shore erosion in a short period of time.

10,000–9,500 B.P.—The level of the ice-marginal lake was lowered and remained for short periods at various levels. At the end of this period the shore was probably somewhat below that of the present Lake Superior.

9,500–4,200 B.P.—The events in the Lake Superior basin during this interval are not well known, but presumably in the Ontonagon area the surface of the plain was tilted, rising to the northeast, as a consequence of the unloading of the ice. At the same time the level of the lake rose both because of the tilt and because of changes in the outlets of the lower lakes of the Great Lakes basin. At 4,200 B.P. the shoreline of Lake Nipissing in the Ontonagon area must have been at about 615 feet, 13 feet higher than the present level of Lake Superior and a strong nip, or escarpment, was cut back into the Ontonagon Plain.

4,200 B.P. to present.—The level of the lake was lowered slightly; a broad beach and series of bars were built lakeward from the Nipissing scarp. The lake shore reached its present level of 602 feet. Some tilting occurred after the Nipissing stage, but the effects were not large enough to be observed in the Ontonagon area.

EVOLUTION OF THE STREAM VALLEYS

The evidence indicates that the area must have been free of lake and glacial water and exposed to erosion by streams as long as 9,000 years ago. Except for the narrow belt below the Lake Nipissing shore, it has been subject to subaerial erosion ever since. The gyttja and buried log in the kettle hole near White Pine (p. B8) show that the plain was very soon covered by dense vegetation when the water receded. Probably the only areas free of vegetation were the larger runoff channels where floods prevented cover growth.

The lack of evidence of piracy (pl. 1) indicates that little change in the drainage pattern has taken place since the drainage system formed. Most streams follow the pattern of grooves, and captures have occurred only where the larger valleys have grown laterally and engulfed smaller ones. As an example, the Ontonagon River has widened its valley somewhat, apparently as its meander belt shifted, and has captured Mill Creek. Smaller valleys, however, especially where the grooves are most dense, show little evidence of this process. Tributaries of the Floodwood River, for example, parallel the Potato River for miles without joining, even though the Potato is a much larger stream. Stream junctions more commonly occur where initial irregularities on the surface of the plain controlled the direction of flow. Deer Creek (the upper part of the Potato River) serves as a striking example. This stream flows down a series of grooves that are oblique to the general trend. On its west bank it picks up a number of smaller tributaries that flow parallel to the more general trend of the grooves. The initial drainage pattern must have been determined in detail by the surficial irregularities of the emerged sublacustrine plain.

Valley cutting has not generally proceeded headward from the lake; instead, the drainage channels have deepened all along the course an amount roughly proportional to a function of drainage area, and hence discharge. Rudimentary channels generally begin 1 to 2 miles below the head of a groove where drainage areas are only ½ to 1 square mile in size. Because the grooves are shallow and on the average are 380 feet wide, the smallest stream channels form only shallow and narrow gullies that occupy a small part of the floor of the groove. Gradually a valley begins to form which contains a meandering channel. Valleys more than 5 miles long may have widths equal to the wavelength of the original groove in which the valley was formed. In the uppermost reaches the small channels are grassy and floored with fine-grained material. The size of the bed material gradually increases as the channel develops. Argentine Creek, for example (sec. 18, T. 50 N., R. 41 W.), 2.4 miles from the source, is intrenched in a steep-walled ravine 60 feet wide but only 6 feet deep. The stream channel within the ravine is 6 feet wide and less than a foot deep. The bed material in the channel is medium to fine sand. At 3.7 miles from the source the same stream is in a valley 120 feet wide intrenched 8 feet beneath the groove floor. The stream channel is 12 feet wide, 2 to 3 feet deep, and has coarse sand on its bed.

The increase in valley depth as the drainage area becomes larger is shown graphically in figure 4.

The inference is made from this relationship that, in general, the stream valleys have not developed by headward erosion from base level, that is, from Lake Superior, but have gradually deepened through time in proportion to a function of the runoff carried by the channel.

In the valleys of the streams that enter Lake Superior, no evidence is found of nickpoints that migrate upvalley. Instead, as shown in figure 5, the profiles of the smaller streams show scarcely any intrenchment at all, even within a few tenths of a mile of the low escarpment south of the lake. Larger streams have slightly convex profiles as they approach the lake. The largest of the streams, such as the Little Cranberry River which has a drainage area of about 6.5 square miles, have concave-upward profiles. The data suggest that only the smallest streams have profiles that might be considered immature. Theoretically, the profiles of even the smallest should be concave upward if the form is related to increasing discharge and if complete equilibrium of form has been achieved. The smaller streams, however, have not kept pace with the lowering of the lake level and the cutting back from the shore that occurred during the Nipissing stage. The Floodwood River profile, for example, has a slight convexity as it approaches the lake.

When the relation of the profile of a valley to its base level is considered, the evidence can be interpreted to mean that the effect of base level on the forms of small valleys like these is mainly to determine the energy gradient. Even small valleys less than 2 miles long far back on the plain are about as deep as valleys of similar size close by the lake. The dominant factor in the valley cutting appears to be the discharge available, and valleys with drainage areas less than a certain limit (about 6 square miles) still preserve marked convexities where they enter Lake Superior.

DRAINAGE COMPOSITION

The drainage network of the area studied is consequent, and few adjustments have taken place by piracy. The present drainage system must be similar to the initial drainage that developed as the lake withdrew

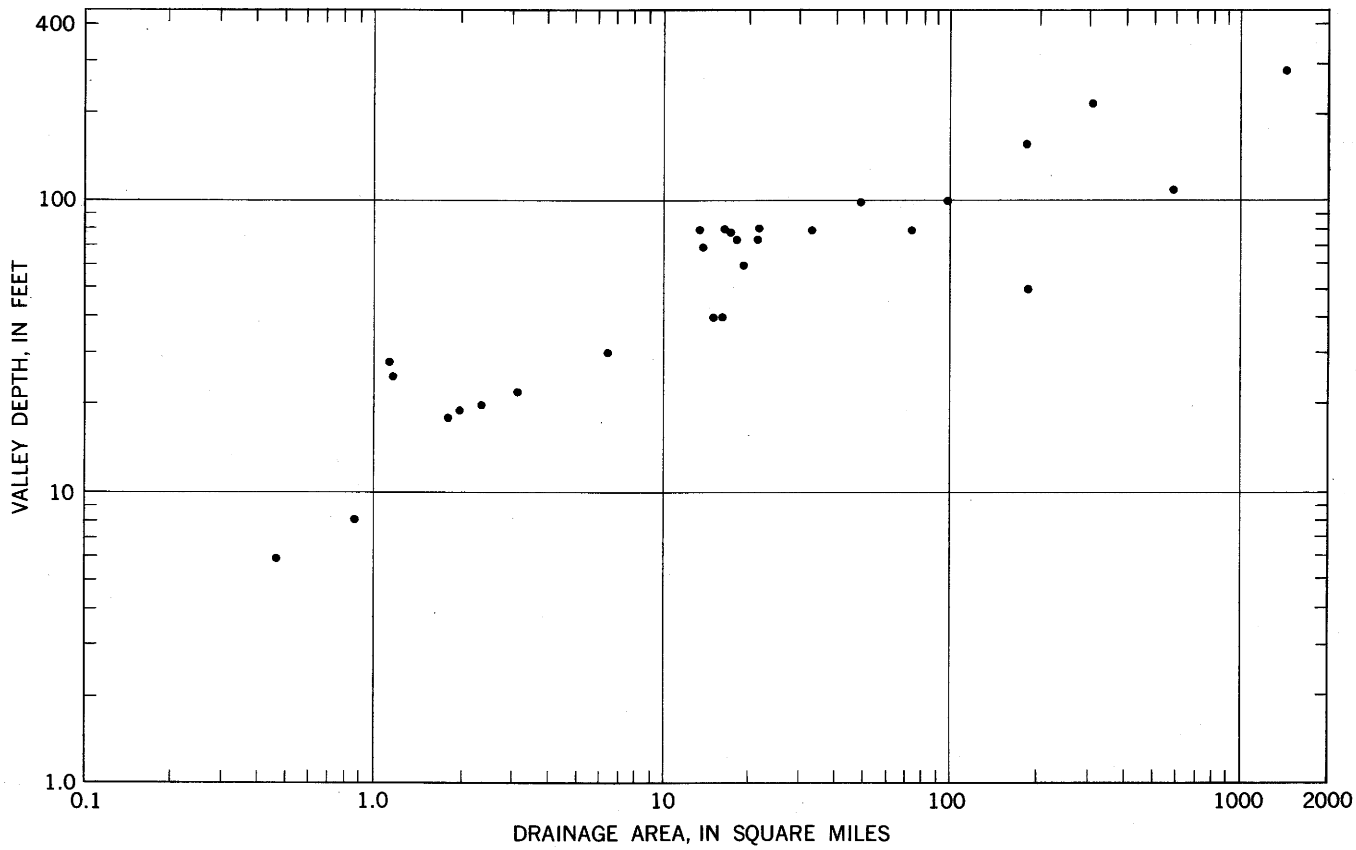


FIGURE 4.—Relation of valley depth to drainage area of valleys eroded in the Ontonagon Plain.

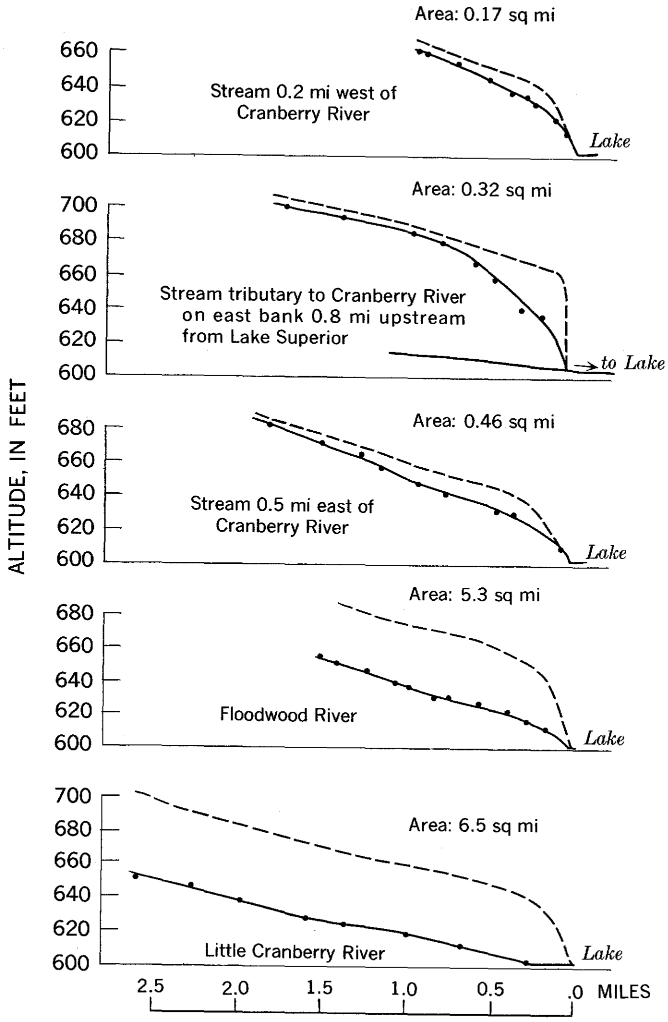


FIGURE 5.—Lower parts of profiles of several small streams showing the change in character related to the difference in size of drainage area. Solid line indicates smoothed profile of stream. Dashed line indicates upland surface. Dots on profiles are altitudes as measured by Kelsh plotter.

from the plain. The geometry of drainage patterns can be studied in various ways. The system of drainage analysis developed by R. E. Horton (1945) shows that drainage networks are organized in a regular manner and that they can be quantitatively described. In Horton's system, streams are classified according to order increasing downstream: A stream with no tributaries is a first-order stream; a stream having one or more tributaries of the first order is a second-order stream; a stream having one or more tributaries of the second order is a third-order stream, and so on. The number of streams of each order in a drainage network decreases as order number increases in a geometric series. The average length of streams of different orders increases as the order increases and other characteristics, such as average slope, also change in regular geometric series. Horton showed that various ratios, obtained by simple measurements of the drainage network, could be used to define and describe the geometry of individual drainage networks. Strahler (1957) made use of Horton's system of analysis, though he modified it slightly, and described some of the important variations in the drainage patterns of different geologic environments. Leopold and Miller (1956) described the relation between Horton's drainage composition and the hydraulic characteristics of streams. More recently Leopold and Langbein (1962) have suggested that the arrangement of stream attributes in regular geometric series is the most probable arrangement in a system in dynamic equilibrium.

The writer, interested in the shape of drainage basins, has studied the relation of Horton's geometry to stream length and drainage area in previous work (Hack, 1957). Drainage basins increase in length relative to their width as they become larger, and the rate of increase is much the same for drainage basins throughout large areas. In general, the length of a drainage basin measured along the principal stream increases in proportion to 1.4 times the six-tenths power of the drainage area. Miller (1958, p. 10) found similar values for streams in New Mexico. The relation to Horton's geometric system indicates that certain of the attributes of drainage patterns he described, such as length and bifurcation ratios, are very stable.

In the Ontonagon Plain area, because of well-developed glacial grooves, the drainage basins have elongate cigarlike shapes. The cigarlike basins, as shown by plate 1 are confined to the areas of the grooves and the drainage lines are arranged in a strongly parallel pattern. In areas without grooves, as in the Ontonagon River valley and the basins in the Copper Range, the drainage is typically dendritic in plan. Some patterns, like that of Mill Creek, are partly parallel and partly dendritic. The analysis that follows concerns principally areas of parallel drainage.

Several examples of drainage systems on the Ontonagon Plain are shown in figure 6, obtained both from topographic maps and aerial photographs. The first-order stream channels generally can be seen only on the aerial photographs. Note also that the Cranberry River basin has the characteristic elongate shape only in the lower part where it traverses the grooved area.

The relation of stream number and stream length to order are shown in figure 7. The values of the various parameters defined by Horton are shown in table 1. These parameters are defined briefly below; a complete

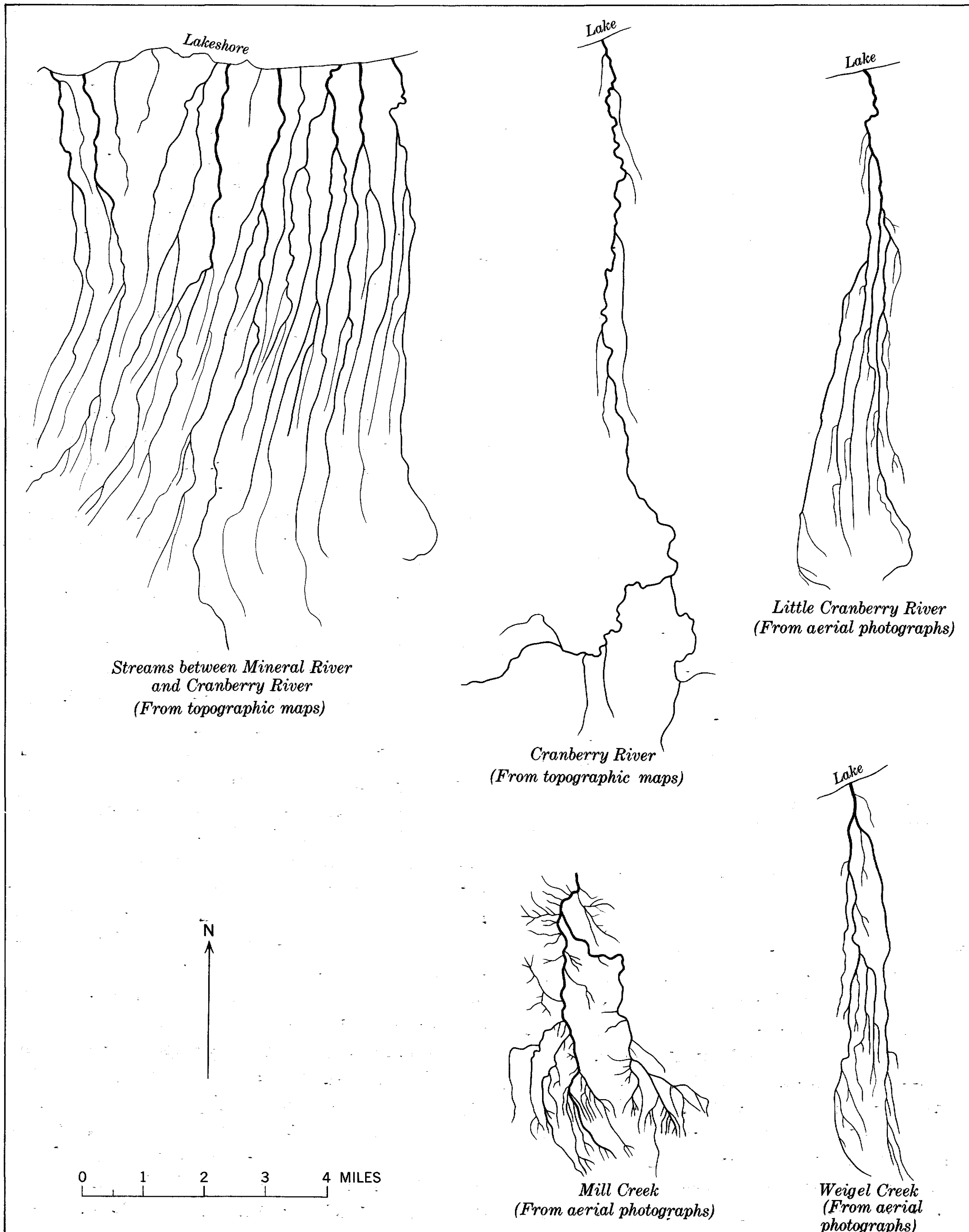


FIGURE 6.—Examples of drainage systems on the Ontonagon Plain.

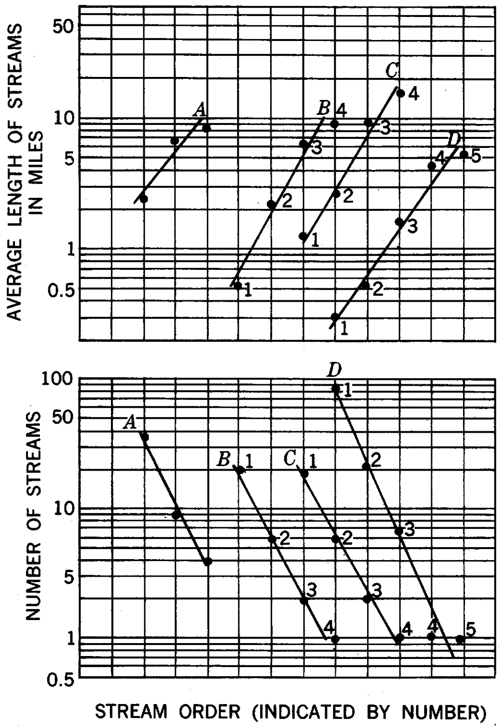


FIGURE 7.—Relation of stream order to stream length and number in drainage basins of the Ontonagon Plain. *A*, Streams between Mineral and Cranberry Rivers; data from maps; stream orders not known. *B*, Little Cranberry River. *C*, Weigel Creek. *D*, Mill Creek.

explanation of these parameters is available in Horton (1945).

A, area; area of drainage basin, generally in square miles.

L, length, in miles, of longest stream or principal stream in drainage basin.

N, number of streams.

S, stream frequency; refers to number of streams per unit area, generally per square mile.

Dd, drainage density; length of stream per unit area, generally in miles per square mile.

l_s, average length in miles of streams of order *s*.

a_s, average drainage area in square miles of streams of order *s*.

r_b, bifurcation ratio; ratio of the number of streams of one order to the number of streams of the next higher order.

r_l, length ratio; ratio of average length of streams of one order to streams of next lower order.

ρ, ratio of length ratio to bifurcation ratio.

s, order of main stream in a given drainage basin.

The drainage basins on the Ontonagon Plain are obviously well ordered (fig. 7; table 1); that is, the streams of different orders vary in number and length in the same manner as the streams described by Horton. Comparison of the bifurcation ratios and length ratios (table 1) with those given by Horton (1945, table 1) also indicates that they are similar. For example, the average of 10 of Horton's length ratios is 2.3, not far from 2.5, the average of the ratios of four segments of the Ontonagan Plain shown in table 1. The bifurcation ratios of Horton average 2.9, whereas the ratios of the four segments of the Ontonagon Plain average 3.0. Not only are these ratios similar, but the fit of the data when plotted graphically (fig. 7) appears to be about as good as the fit obtained by Horton (1945, figs. 2, 4). The factor that distinguishes the streams on the Ontonagon Plain is a high drainage density and elongation of every order of stream. Weigel Creek on the Ontonagon Plain, for example, has first order streams that average 1.2 miles in length but only 0.21 square mile in area. Calculation from Horton's data (1945, table 1) shows that in Esopus Creek in the Catskill Mountains the average drainage area of a stream 0.99 mile long is approximately 1.1 square miles. Thus, on the Ontonagon Plain, a first-order drainage basin has only one-fifth the area of a basin of equivalent length in the Catskill Mountains.

The shape of drainage basins can be studied more directly by comparing the length of the principal stream

TABLE 1.—*Characteristics of the drainage net of certain stream basins and groups of basins in the Ontonagon area*

Basin(s)	Highest order of stream	Total area (sq mi)	Number of streams	Number of first-order streams	Stream frequency	Drainage density	Average length of first-order streams	Average drainage area of first-order stream	Bifurcation ratio	Length ratio	ρ
All streams between Ontonagon and Mineral Rivers (data from topographic maps)	3	126	93	70	0.74	2.42	2.05	0.85	2.85	2.1	0.74
Same, converted to higher order [1]	4	126	293	200	2.3	3.3	.97	.30	2.85	2.1	.74
All streams between Mineral and Cranberry Rivers (data from topographic maps)	3	63	49	36	.77	2.86	2.45	.86	2.85	1.9	.61
Same, converted to higher order [1]	4	63	152	139	2.4	4.2	1.27	.30	2.85	1.9	.61
Little Cranberry River basin (data from aerial photographs)	4	6.5	29	20	3.0	7	.52	.07	2.86	3.0	1.05
Weigel Creek basin (data from aerial photographs)	4	5.7	28	19	4.9	5.6	1.2	.21	3.28	2.58	.79
Mill Creek basin (data from aerial photographs)	5	9.2	119	88	13	6.6	.31	.05	3.15	2.5	.80

[1] Inasmuch as the smallest streams are not indicated on the topographic maps, it is assumed that the smallest streams shown are on the average of the second order rather than the first, and Horton's parameters are recalculated.

with the drainage area of the basin. It has been shown that in erosionally graded landscapes in the eastern United States such as the Shenandoah Valley of Virginia, the length of the principal stream, within fairly narrow limits, is equal to 1.4 times the six-tenths power of the drainage area. In other words, the average stream with a drainage basin 1 square mile in area is 1.4 miles in length, and increases in length at a rate that is slightly greater than the square root of the drainage area and approximates the six-tenths power. This relationship is quantitatively related to the parameters in Horton's system (Hack, 1957, p. 63–67).

By using Horton's equation relating drainage density to other properties of the drainage network, the length of the principal stream of a drainage basin of order s can be expressed by the equation:

$$L_s = l_1 r_l^{s-1} \tag{1}$$

The area of a drainage basin of order s can be expressed by:

$$A_s = a_1 r_b^{s-1} \cdot \frac{\rho^s - 1}{\rho - 1}. \tag{2}$$

Thus the relation of length to drainage basin area is a complex function of the drainage density, bifurcation ratio, and length ratio (Hack, 1957, p. 66–67). Trial of different values for the variables shows that the exponent in a general equation expressing basin shape,

$$L = bA^n \tag{3}$$

is determined by the quantity ρ, the ratio of the length ratio to the bifurcation ratio. The value of the coefficient, b, is related indirectly to the drainage density, but also to ρ, for it is equal to the quantity

$$L/A^n \text{ or } l_1/a_1^n.$$

Figure 8 is a graph on which the length-area relationships for several hypothetical drainage basins have been plotted. The values of the exponents and coefficients are determined by assuming certain values of a_1, l_1, r_l, and r_b as shown in the inserted table. We can thus observe the overall change in the basin shape as the basin grows in size, as reflected in the slope and position of the curves on the graph. Curve 2 of figure 8, with a slope of 0.5, serves as a good reference line, for when the exponent n has a value of 0.5 the drainage basin does not change in overall shape as the basin becomes larger, that is, the length of the basin as measured along the principal stream is always proportional to the square root of the drainage area. If n is larger or smaller than 0.5 then the shape of the drainage basin along a given principal stream must change in a downstream direction. If n is greater than 0.5 then larger basins are narrower relative to their lengths than smaller upstream basins, whereas if n is smaller than 0.5 larger basins are wider relative to their lengths than smaller ones. The intercept at a drainage area of 1 square mile is equivalent to the coefficient and is determined by the absolute values of length and area of the average basin of a given stream order. The construction demonstrates that differences in the shape of drainage basins theoretically can be controlled by differences in bifurcation and length ratios having a wide range in value.

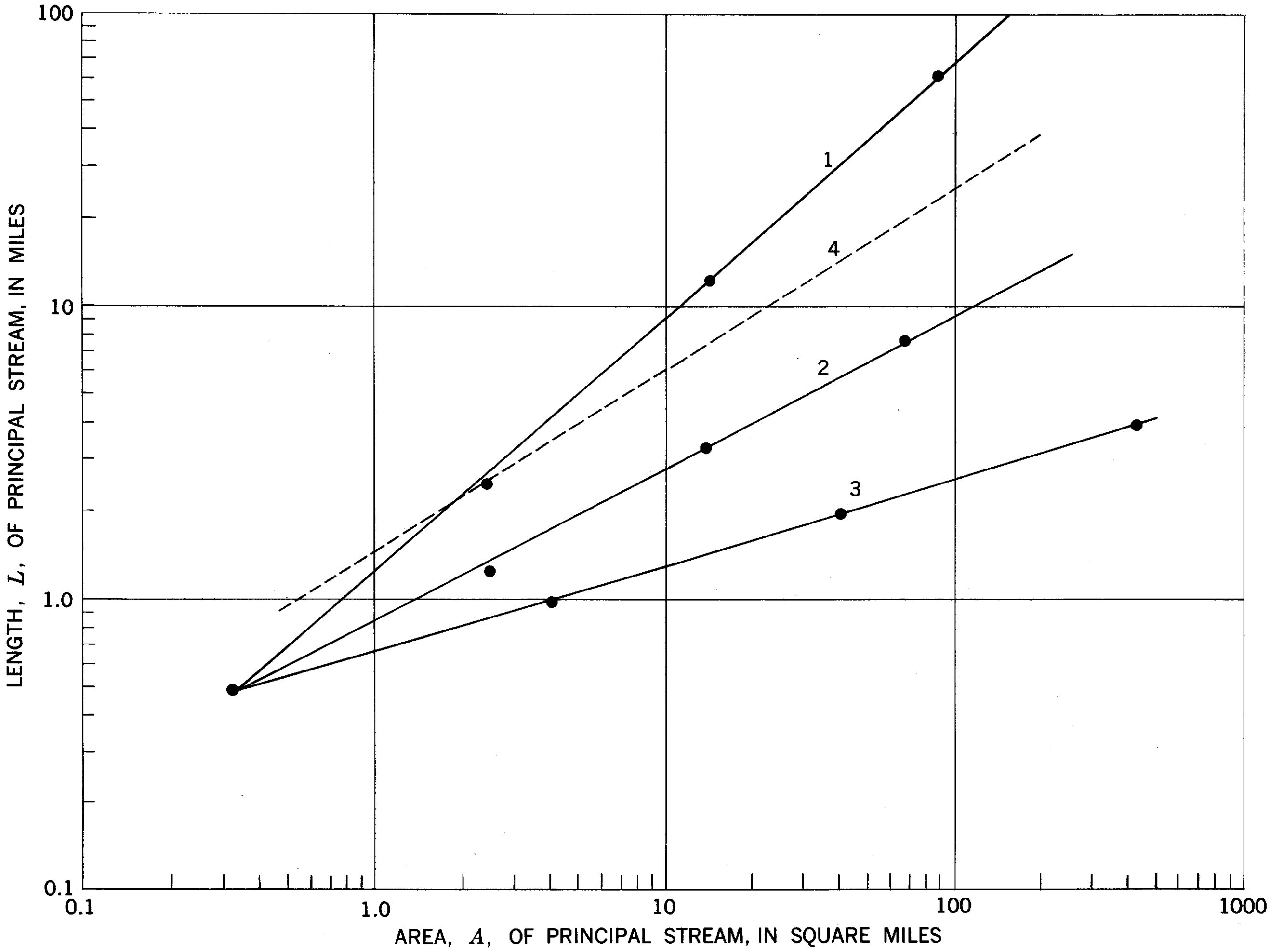
If the bifurcation ratio characteristic of a drainage network were very large, basins would tend to become more circular in shape with increasing size. If the length ratio, on the other hand, were large, drainage basins would tend to become narrower with increasing size and large basins in such a network would be very narrow and cigar-shaped.

The evidence, however, indicates that the ratios r_l and r_b are very conservative and tend to have values like those of curve 4 (fig. 8) in which ρ is 0.75, and which represents a drainage basin in the Shenandoah Valley (Hack, 1957, p. 66). Analysis of the streams on the Ontonagon Plain by Horton's method has shown that the ratio ρ is similar, in that it averages 2.5 : 3.0 or 0.83. Although the extremely elongate basins of this area do become slightly more elongate as they enlarge, the rate of change is very little greater, if any, than that in most other, more normal drainage networks, and first-order drainage basins are nearly as elongate as third or fourth-order basins.

The actual dimensions of drainage basins on typical grooved areas on the Ontonagon Plain are shown graphically in figure 9. The exponent of a curve drawn by inspection through the points is 0.6 and the coefficient is 3.6. Obviously both small and large basins tend to be very elongate and the elongation becomes slightly greater as the basin enlarges. Streams like the Cranberry and Potato Rivers, that originate south of the grooved area but cross it, follow quite a different curve, as shown in the graphs. The upper part of the Cranberry River basin with a drainage area of about 10 square miles is like any ordinary drainage basin and has a dendritic drainage pattern. In this part of its course the principal stream follows the curve

$$L = 1.4\ A^{0.6}. \tag{4}$$

Where the drainage area becomes 10 square miles, however, and the river is a fifth-order stream, it enters the grooved area and follows a narrow corridor through the grooves to the lake, picking up no tributaries any larger than second order.



Curve	l_1	a_1	r_l	r_b	ρ	b	n
1	0.5	0.33	5.0	2.5	2.0	1.25	0.85
2	.5	.33	2.5	5.0	0.5	.86	.5
3	.5	.33	2.0	10.0	.2	.78	.29
4	.35	.077	2.4	3.2	.75	1.50	.62

FIGURE 8.—Possible relations of stream length to drainage area in hypothetical drainage basins having different assumed values of length, area, length ratio, and bifurcation ratio. Dots refer to the values of L and A obtained by calculation.

Leopold and Langbein (1962, p. A14–A17) have shown that the geometry of natural drainage can be explained simply as ordered in a random manner. They show that if streams of each different order are spaced an even distance apart the streams will, on the average, join at distances that are proportional to the stream order. Leopold and Langbein further show that the length ratio must range between 2 and 4, and that natural streams do have length ratios between these limits. Furthermore, by using a random-walk technique they construct a model drainage system on the basis of this concept. The stream length of the resulting system varies proportionally to a power function of the drainage area close to six-tenths.

The drainage network on the Ontonagon Plain, in spite of its unusual plan, behaves in accordance with the requirements of the theory. The grooved area of the plain may be thought of as a homogeneous environment in which a steep slope toward Lake Superior and a strongly grooved surface texture have together imposed a constraint on the formation of the network and have tended to produce long narrow drainage basins. In other words, when the plain emerged from the lake, a stream of water flowing down the slope was required to travel an extra long distance before it by chance merged with an adjacent stream. Nevertheless, the bifurcations were arranged in a random manner as in any other drainage network.

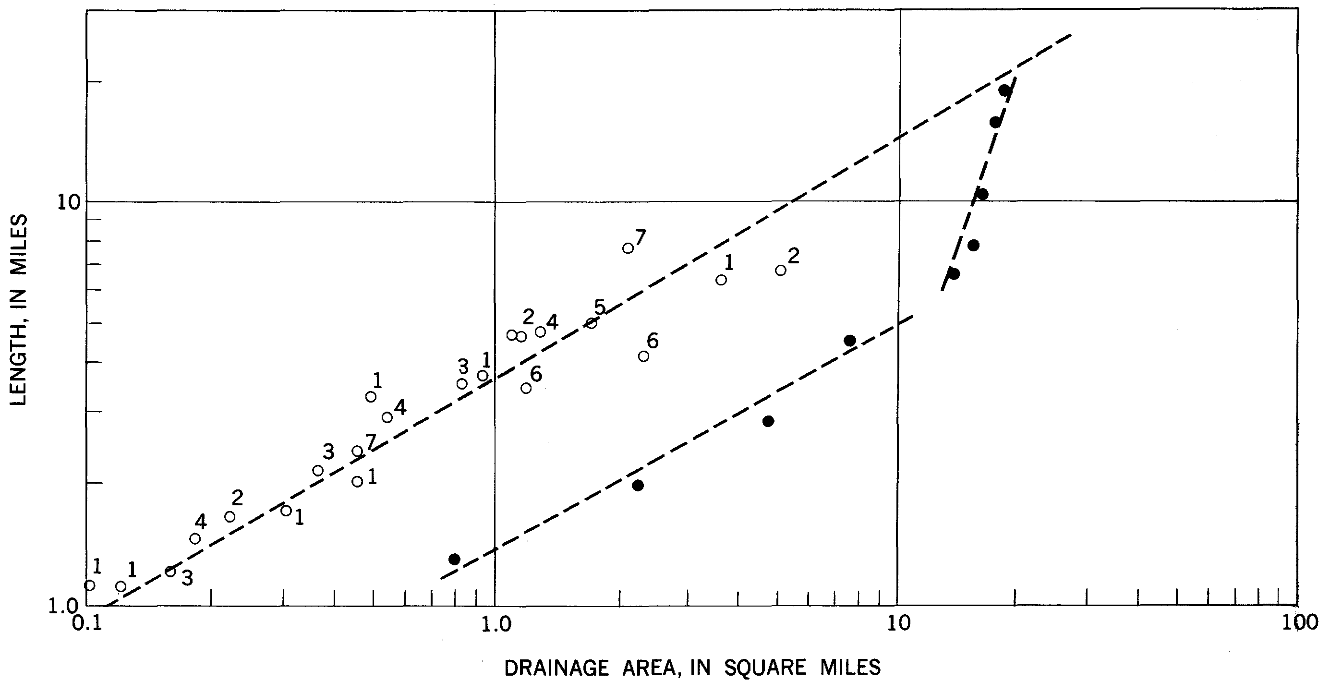


FIGURE 9.—Relation of channel length to drainage area of streams on the Ontonagon Plain. 1, Weigel Creek; 2, Bear Creek; 3, Floodwood River; 4, Miles Creek; 5, Stony Creek; 6, Duck River; 7, Argentine Creek; ○, localities on streams originating in grooved areas; ●, localities on Cranberry River.

Actually, the drainage system on the plain is homogeneous and randomly arranged only when examined as a whole. The minute details of the network were presumably determined by specific geologic causes and events. The drainage network began to form as Lake Duluth drained and the lake water receded from the grooved surface of the till and lake sediment. Because the gradient of the plain was steep, drainage could only flow down the slope and must have been confined in the grooves. Retreat of the lake was neither regular nor immediate and the shoreline stood for short or long periods at various levels on the plain. Each stand of the lake provided an opportunity for erosion at the lakeshore of part of the ridge between grooves, and thus could have formed a gap through which a stream of water running down the plain could escape into the next groove. Such gaps are in addition to gaps present where a groove ends. The drainage pattern probably was formed by the joining of streams through such gaps as the lake withdrew. Some additional junctions may have been caused by beaver dams, which are still numerous on the Ontonagon Plain.

If we knew the exact configuration of the plain as it emerged from the lake and the location of all the beaver dams in the past history of the area, we doubtless could explain the location of every bifurcation. We cannot, of course, and because the bifurcations are spaced across the plain in a complex manner without apparent order, we say they are distributed in a random manner, which is equated by Leopold and Langbein (1962) with maximum probability. By using their random-walk technique, we can set up limits or constraints and construct a model drainage pattern that resembles the actual pattern of the Ontonagon Plain. First the assumption is made that the plain is crossed by grooves 0.1 mile apart and that water must flow down every groove. In the first mile there is no chance for the water to escape, but in every succeeding mile downslope the stream has an even chance to escape to the right, to the left, or to continue down the groove. Another way of stating the conditions would be to say that the drainage density is 10 and every first-order stream has a length of 1 mile. It is not necessary to state the manner in which the stream bifurcates, for this is left to chance, the one limitation being that a stream must be 1 mile long at the first bifurcation and that it does not have another chance to bifurcate for another mile. A sample of the resulting drainage system is shown in figure 10. When the length of the principal stream is compared with the area of the drainage basin at various points in the model, the two quantities are found to be related roughly according to the equation $L=4A^{0.67}$, not unlike the actual equation for the Ontonagon Plain. A similar drainage network could have been constructed by using somewhat different limitations or constraints. For example, we might have assumed that at every 0.1 mile instead of at every mile, each stream has a chance to flow to the right or

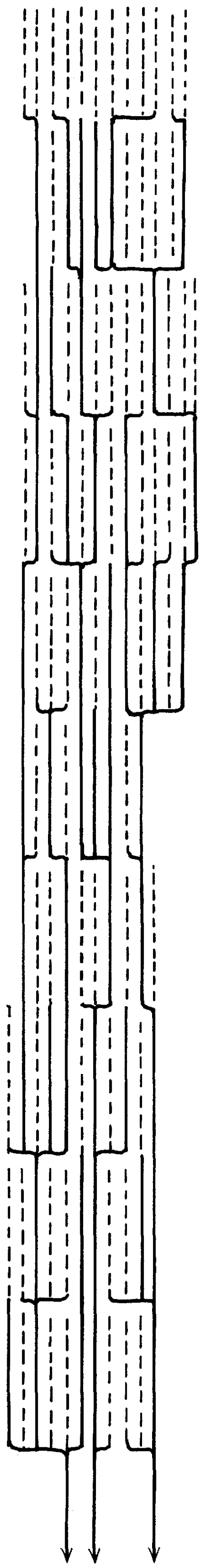


FIGURE 10.—Model simulating the drainage network of the Ontonagon Plain, constructed by the random-walk technique.

the left or continue downslope. The chances, however, would be uneven, so that the probability of the stream turning to the right would be one-twelfth, to the left one-twelfth, but to continue downslope ten-twelfths.

The conditions that determined the natural drainage network were similar to those in the model. Geologic conditions determined that a stream must flow a long distance downslope before it could join another, and the wavelength of the grooves determined that the spacing of the streams must be relatively close. Because the locations of the bifurcations are randomly arranged, as in any other drainage network, the resulting drainage pattern is attenuated in a direction parallel to the grooves, as though stretched out in one direction. Presumably, any geologic or topographic factor that restricts or impedes the development of channels in a systematic way would operate in a similar manner. Thus on a very long slope that is unusually steep, we might also expect an attenuated drainage network. Similarly, a trellised drainage pattern is in effect an attenuated drainage network caused by the presence of resistant beds of rock. If the resistant beds are far apart, the attenuation may affect only the streams of higher orders, and not the smaller streams.

In the general equation relating stream length to area,

$$L=bA^n,$$

where L is stream length, A is area, and b and n are constants, the exponent n must always have approximately the same value in any homogeneous area for its value is determined by the laws of probability and not by geologic factors. The coefficient b, however, is a measure of the attenuation of the drainage network and may be affected by any geologic or topographic characteristic that tends to direct the flow of water uniformly.

LONGITUDINAL PROFILES

Studies of the equilibrium conditions in stream channels have shown that many variables are involved in determining the form of longitudinal stream profiles. Classic geomorphic theory in America assumes that the concave upward form is a result of a graded condition that evolves as the landscape is eroded toward base level (Hack, 1960). Increasing discharge has generally been assumed to be an important factor. Leopold and Maddock (1953) in their analysis of the hydraulic geometry of streams showed that at least seven variables are involved in a solution of the problem. They are downstream changes in channel slope, depth, width, load, velocity, and roughness, and discharge. Some of them are closely related to the geology of the area through which the stream passes.

The actual forms of the profiles in small streams comparable in size with those in the Ontonagon area have been studied in the humid eastern United States (Hack, 1957). It was found, as might be expected, that the profile itself is closely dependent on the geology, probably through the interaction of some of the other variables. Most of the profiles studied could be expressed by integrals of the function

$$S \propto L^{y}$$

where S is channel slope and L is channel length or distance from the source of the stream. Because most of the profiles are concave, the constant y is generally negative and ranges in value from higher than -0.5 to lower than -2.0. In some streams, in which the size of the bed material is uniform in a downstream direction, the value of y is -1.0; the profile is then a simple logarithmic curve

$$H = C - y \log_e L,$$

where H is altitude, L is channel length, and C and y are constants. One stream studied, the North Fork of the Shenandoah River, follows such a curve for more than 100 miles (Hack and Young, 1959, fig. 3). Samples of the bed material of this stream, as of others having this type of profile were approximately the same in grain size all along the course. The beds of these streams, although smooth, are being eroded and contain many rock outcrops in the channel. The loose bed material is of the same size because it is locally derived or has been transported only a short distance and is of similar resistance. In such streams the competence must be more or less the same all along the course.

It has been shown (p. B9) that the profiles of many of the smaller streams that enter Lake Superior are affected by the escarpment at the lake edge and have convex upward profiles. Only the larger streams have absorbed the effect of the escarpment and have concave profiles. It is expectable, then, that the profiles of many streams, especially the smaller ones, reflect the initial configuration of the sublacustrine plain and have little curvature. The channel slopes of some larger streams, however, have been reduced to much below the initial slope. The Ontonagon River has been reduced to a grade of 3.2 feet per mile where it crosses the grooved plain. The grades of the Cranberry and Potato Rivers in parts of their courses are as low as 12 feet to the mile. These grades are much less than the initial slope of the lake plain, which averages 50 feet to the mile, and they suggest at least some profile evolution.

From the values for channel slope at the sample localities shown in table 2, we can construct a model stream profile that represents the average for all the streams in the area. The slopes are plotted against the stream length in figure 11. This model profile is not intended as a representation of all or any of the streams. Its purpose is to clarify the relation of length to area as it affects slope.

The field of scatter is irregular and no definite relation between the two quantities is apparent. An unbiased line through the points was drawn both by the method of least squares and by multivariate analysis (Snyder, 1962). Both methods gave the same line having the equation $S = 146\ L^{-0.73}$, where S is channel slope in feet per mile and L is length in miles. The integral of this variable is a line that represents a model average-stream profile for the grooved plain, and has the equation $H = C - 540L^{.27}$, where H is the fall in feet from the source of the stream and C is a constant. Such a profile has only a slight curvature. At 5 miles the gradient would be 45 feet per mile and at 10 miles it would be 27 feet per mile. This degree of curvature is not very high. The profiles of individual streams, of course, are all different from this average. The smaller streams, like Bear Creek, have profiles without appreciable curvature (fig. 12).

The profiles of the larger streams are also irregular and it cannot be shown that there is any systematic downstream rate of decrease in slope representative of all of them. Streams such as the Cranberry and Potato Rivers, however, are large enough to have a marked concavity (fig. 12). The irregularity of the profiles is probably due partly to the effect of the bedrock through which the stream flows and partly to irregularities in the rate at which drainage area and discharge increase downstream. The Cranberry River, for example, has a reach about three miles long in which the lower valley walls are bedrock and rugged enough in appearance to be called a gorge. The profile is so much affected by the bedrock that the gradient increases to 50 feet per mile, whereas it is less than 30 feet per mile in the till area upstream and only 16 feet per mile downstream where the valley is also in till.

The marked affect of the geology on the slope of other streams is reflected in the relation of channel slope to drainage area (fig. 13). The channel slopes at most localities in bedrock reaches are much higher than in till reaches. The fact that the gradients have become adjusted to the presence of bedrock in the valley walls is, of course, an indication that there has been considerable gradation of the river profiles.

The effect of the relation of drainage area and length on the channel slope is also shown by figure 13. Note that the average slope in a drainage area of 1 square mile is approximately 50 feet per mile. A similar coefficient was obtained for streams in the Shenandoah Valley of Virginia, in limestone and shale, and in fact

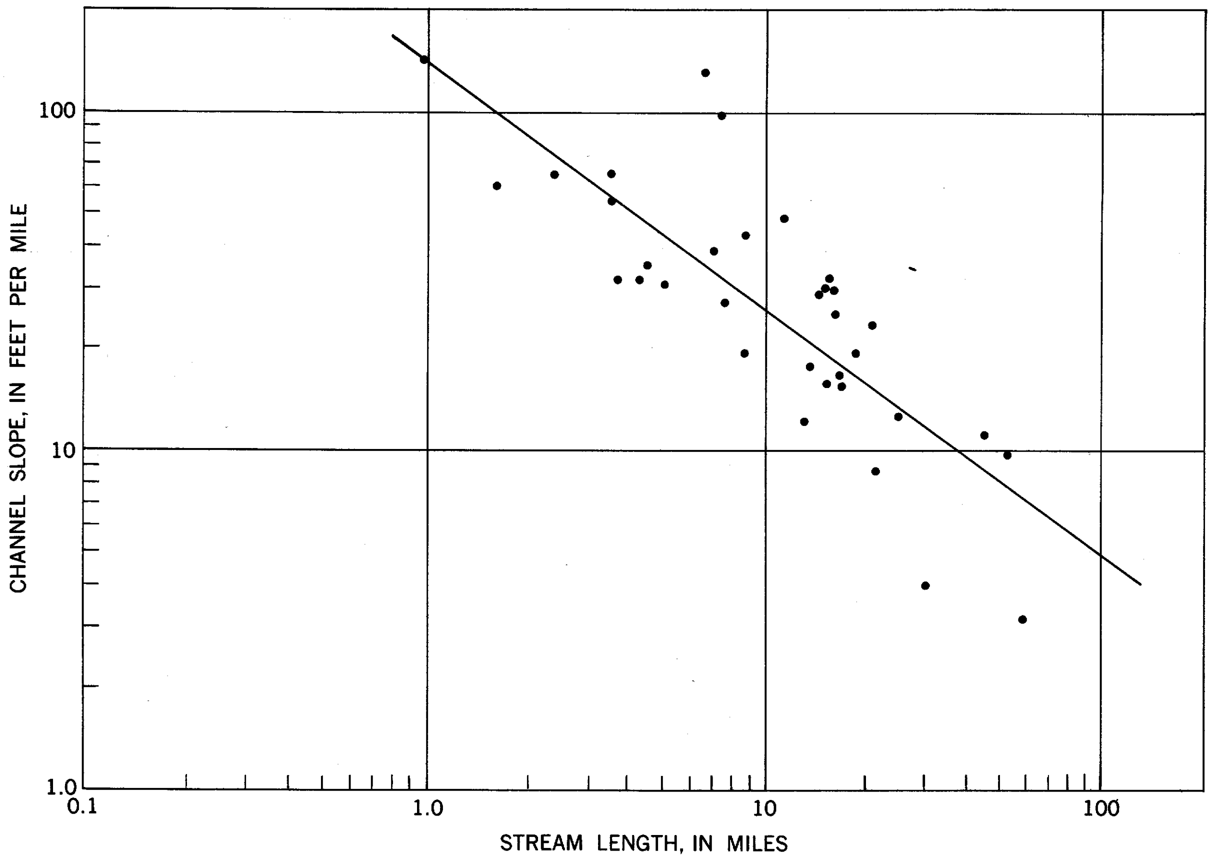


Figure 11.—Relation of channel slope to channel length at localities in the Ontonagon area (data from table 2). Equation of line: $S=146L^{-0.73}$

Table 2.—*Measurements in stream valleys at localities in the Ontonagon area*

No.	Stream	Area (sq mi)	Length (mi)	Slope (ft per mi)	Width (ft)	Depth (ft)	Estimate of bed-material size (mm)				Valley depth (ft)	Valley width (ft)
							Fragments in sample	Geometric mean	Phi mean	Phi standard deviation		
M 25	Stony Creek	1.71	5.0	32								
31	Mineral River	11.45	13.0	12.5	36	3.5	20	12	−3.4	1.2		
32	Tolfrey Creek	1.82	4.4	36	14	2.5	20	25	−4.6	1.1	18	85
33	West Branch Duck River	2.36	4.2	33	12	3.5	20	21	−4.4	1.8	20	90
64	East Branch Duck River	1.2	3.5	67	12	2.0		[1] .5	[1] 1.0		25	110
65	Argentine Creek	.47	2.4	67	6	0.8		[1] .25	[1] 2.0		6	60
66	Halfway River	1.17	1.6	62.	6	2.0		[1] 1.0	[1] 0		28	100
67	Argentine Creek	.87	3.7	33	12	2.7		[1] 1.0	[1] 0		8	120
69	Halfway River	2.00	3.5	57	15	4.5	47	40.0	−5.3	2.5	19	180
70	Iron River	97.4	20.3	24	143	14.0	56	52	−5.7	1.0	100	900
71	Little Iron River	13.3	6.6	133	98	5.2	51	80	−6.3	2.2	80	275
72	Cranberry River	16.7	11.3	50	67	6.5	52	52	−5.7	1.5	80	195
73	Argentine Creek	3.16	7.8	44	18	3.0	42	36	−5.2	1.9	22	160
75	Little Iron River	13.6	7.0	40	46	5.0	40	52	−5.7	2.0	80	160
76	West Branch Iron River	33.6	16.9	16	100	6.0	94	26	−4.7	1.6	80	500
77	do	50	16.0	26	115	8.0	45	110	−6.8	1.8	100	550
78	Mineral River	6.6	7.0	100	35	5.0	22	84	−6.4	2.5	30	
102H	Cranberry River	17.3	15.2	33	45	4.0	50	64	−6.0	1.1	80	1,000
102E	do	17.8	15.8	31	55	6.0	50	26	−4.7	.9		
102G	do	17.4	15.4	31			50	30	−4.9	1.5		
104	do	18.2	16.9	17			30	125	−7.0	1.9	75	1,200
105	do	18.1	16.6	17	45		50	26	−4.7	.8	80	1,100
111	do	19.1	18.5	20	40	5.0	33	45	−5.5	.7	60	1,400
115	Potato River	21.5	14.6	30	60	5.0	48	64	−6.0	1.4	80	700
118	do	21.7	15.4	16	80(?)	3.0	49	12	−3.5	1.5	75	1,400
127	Deer Creek	15.3	8.8	20	50	4.0		[2] 30	[2] −4.9		40	400
132	Cranberry River	14.0	7.6	28	55			[2] 80	[2] −6.3		70	1,300
144	do	16.1	11.5		38	4.0		[2] 105	[2] −6.7		40	230
166	Firesteel River	83.8	21.0	14			49	75	−6.2	2.2	80	1,400
201	West Branch Ontonagon River [3]	183	30	4.1							50	
202	do. [3]	590	52	10							110	
203	Middle Branch Ontonagon River [3]	305	45	11							220	
204	East Branch Ontonagon River [3]	183	25	13							160	
205	Ontonagon River	1,496	58	3.2							280	

[1] Average size estimated by comparison with standard reference samples.
[2] Rapid estimate by measurement of 6–10 fragments chosen at random.
[3] Locality not visited.

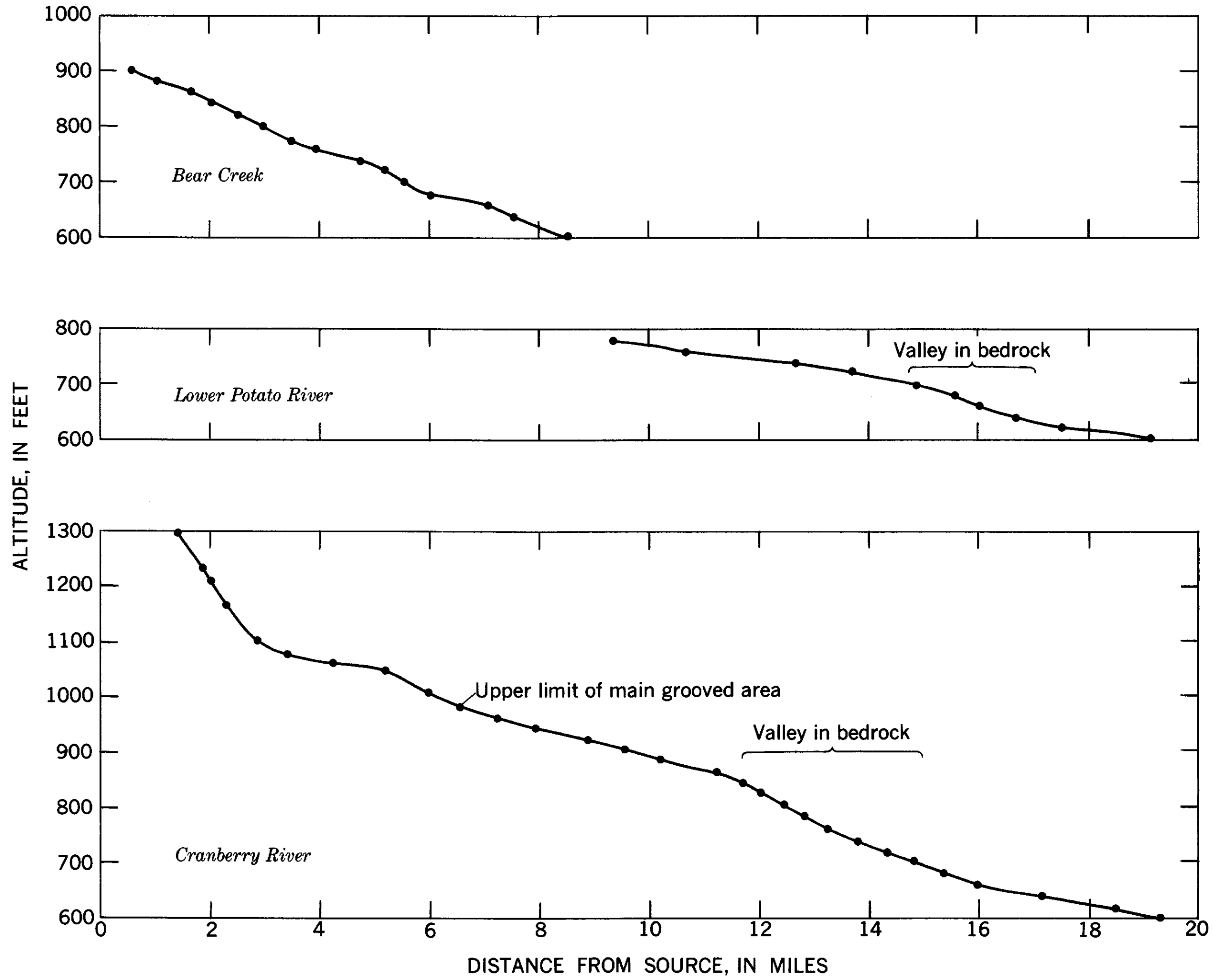


FIGURE 12.—Longitudinal profiles of parts of three streams in the Ontonagon area.

the slope-area diagram for such streams in that region is very similar (Hack, 1957, fig. 16) but the slope length relationship is quite different in the two regions. In the Shenandoah Valley the slope at a stream length of 1 mile is 80 feet per mile, about 1.4 times the value of the slope at an area of 1 square mile. On the Ontonagon Plain, however, the average slope at a stream length of 1 mile is 146 feet per mile, more than three times the slope in a 1 square mile drainage area. In the Ontonagon area the channel slopes for any given stream length are much steeper than in the limestone and shale streams of the Shenandoah Valley and the profiles must also be correspondingly steeper, even though the area-slope relations are similar. This relation must be a consequence of the elongate character of the drainage basins.

The relation of drainage area to length in an ungrooved is different from that in a grooved area. The normal relation found in the eastern United States (Hack, 1957, p. 63), and also exemplified by the upper Cranberry River (fig. 9), is $L=1.4\,A^{0.6}$. In the grooved area the relation is approximately $L=3.6A^{0.6}$. Thus, if a stream passes from an ungrooved area to a grooved area, there is a marked decrease in the rate of increase of discharge downstream, and this decrease causes the channel slope to approach zero at a decreased rate.

SIZE OF BED MATERIAL

The bed-material size was estimated by the measurement of 20 to 90 fragments selected randomly at most of the localities shown in table 4. No systematic relation of bed-material size to channel slope or other factors was found, although in the very small streams there is a very obvious increase of size with stream length. In streams having drainage areas larger than 2 square miles, no relation is apparent. The range in mean size of bed material of reaches in the larger streams is from about 12 to 125 mm.

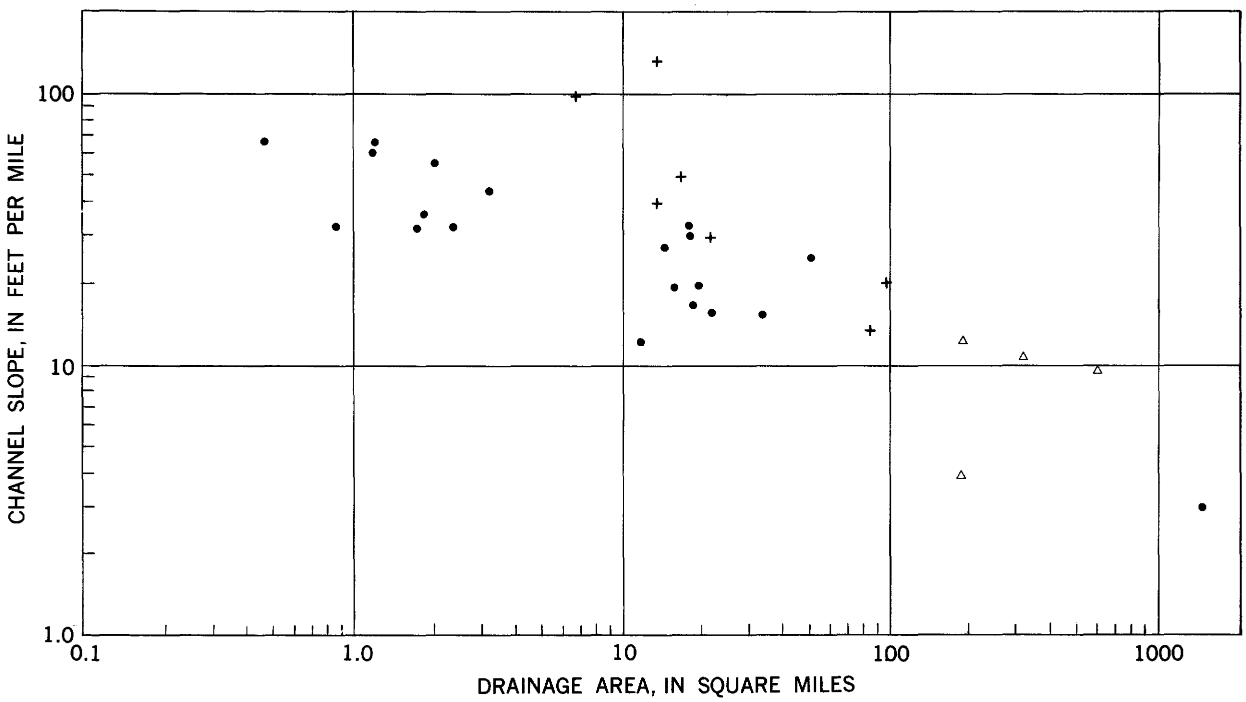


Figure 13.—Relation of channel slope to drainage area in streams on the Ontonagon Plain, ●, channel in till and lake deposits; +, channel in bedrock; △, localities in Ontonagon River basin in which character of channel is not known.

Some individual stream reaches are exceptions and show a systematic decrease in size of bed material. In both the Cranberry and Potato Rivers, directly below the bedrock reaches, the channels enter areas of till in which large boulders are uncommon. In these reaches both size and slope decrease systematically (p. B25, B29). In the area as a whole, however, the size of bed material is determined very largely by the material that encloses the channel and there is no apparent adjustment of slope in relation to size. The resistance of the bed and bank appears to be a more important factor controlling channel slope than does size of bed material.

In the central Appalachians, channel slope is related in a rough way to the ratio of bed material to drainage area. No such relation was observed in the Ontonagon area but, by comparison, the range in this ratio as well as the range in channel slopes is much smaller (fig. 14).

CHANNEL CROSS SECTION

Measurements of channel width and depth were collected at about 25 localities. The stream channels selected ranged in width from 6 to 143 feet (fig. 15). A fairly good correlation was found between channel width and drainage area and the data compares closely with similar data obtained by Miller (1958, fig. 16) in New Mexico and with data obtained by Hack (1957, fig. 22) in Virginia and Maryland. The depth-width ratio on the Ontonagon Plain declines slightly with increasing drainage area and discharge as it does in other areas.

The rate of increase of channel width is somewhat steeper in the Ontonagon area than in the central Appalachians. The curve of figure 15 has the equation $W=9A^{0.6}$, where W is channel width in feet and A is area in square miles. The exponent for the Virginia and Maryland streams is less than 0.5, as is the exponent suggested by Miller's data for New Mexico

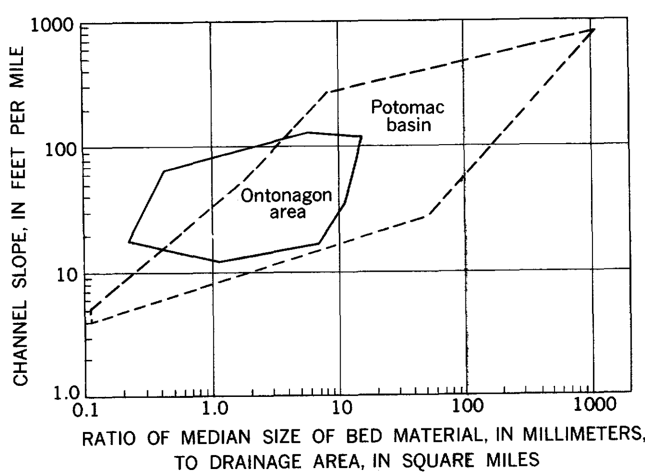


Figure 14.—Relation of channel slope to the ratio of bed-material size to drainage area. Area of scatter for localities in the Ontonagon area compared with that for localities in the Potomac River basin, Virginia and Maryland.

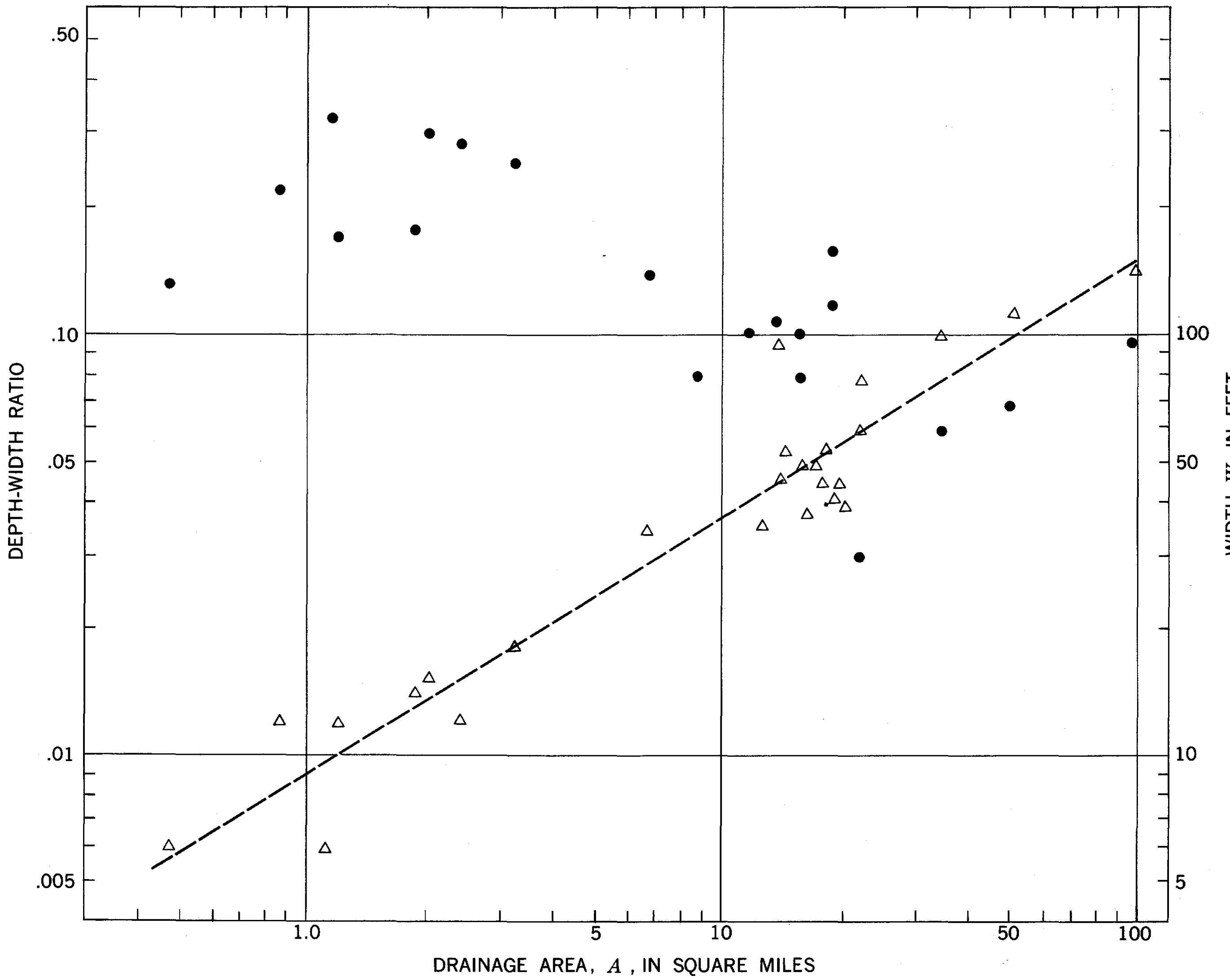


FIGURE 15.—Relation of channel width and depth-width ratio to drainage area in streams of the Ontonagon Plain. Equation of line: $W=9A^{0.6}$. ●, depth-width ratio; △, channel width.

perennial streams. The value of the exponent for the rivers of the Ontonagon area may be slightly higher because the drainage system is not in dynamic equilibrium. The initial slope of the grooves was relatively steep and, because many streams still retain a steep slope, their profiles are less concave than is normal. The lack of curvature in the longitudinal profile is compensated by a greater rate of increase of channel width.

CRANBERRY RIVER

The Cranberry River crosses the grooved Ontonagon Plain through a narrow corridor. The upper basin is nearly circular (fig. 6) and in the first 7 miles the drainage area increases to 15 square miles. Below the junction with Mason Creek, the largest tributary, the drainage area increases only 4.5 square miles in the remaining 12 miles to the lake. The lower river is of particular value as an example for study not only because of the small increase in discharge but also because, although intrenched mostly in till, it crosses several bedrock thresholds and from sec. 3, T. 50 N., R. 41 W. through sec. 29, T. 51 N., R. 40 W. (fig. 16) it is deeply intrenched in the Freda Sandstone. The bedrock has a strong influence on the behavior of the stream and changes the relative values of the variables that make up the channel equilibrium.

The river was studied in two ways. The entire valley from Mason Creek to Lake Superior was traversed in the field. In addition, the lower course from sec. 29, T. 51 N., R. 40 W. to the lake was studied on a fully corrected optical model produced by using a Kelsh plotter, part of which is sketched in figure 16. The optical model permitted observations to be made of stream terraces that are difficult to measure in the field because

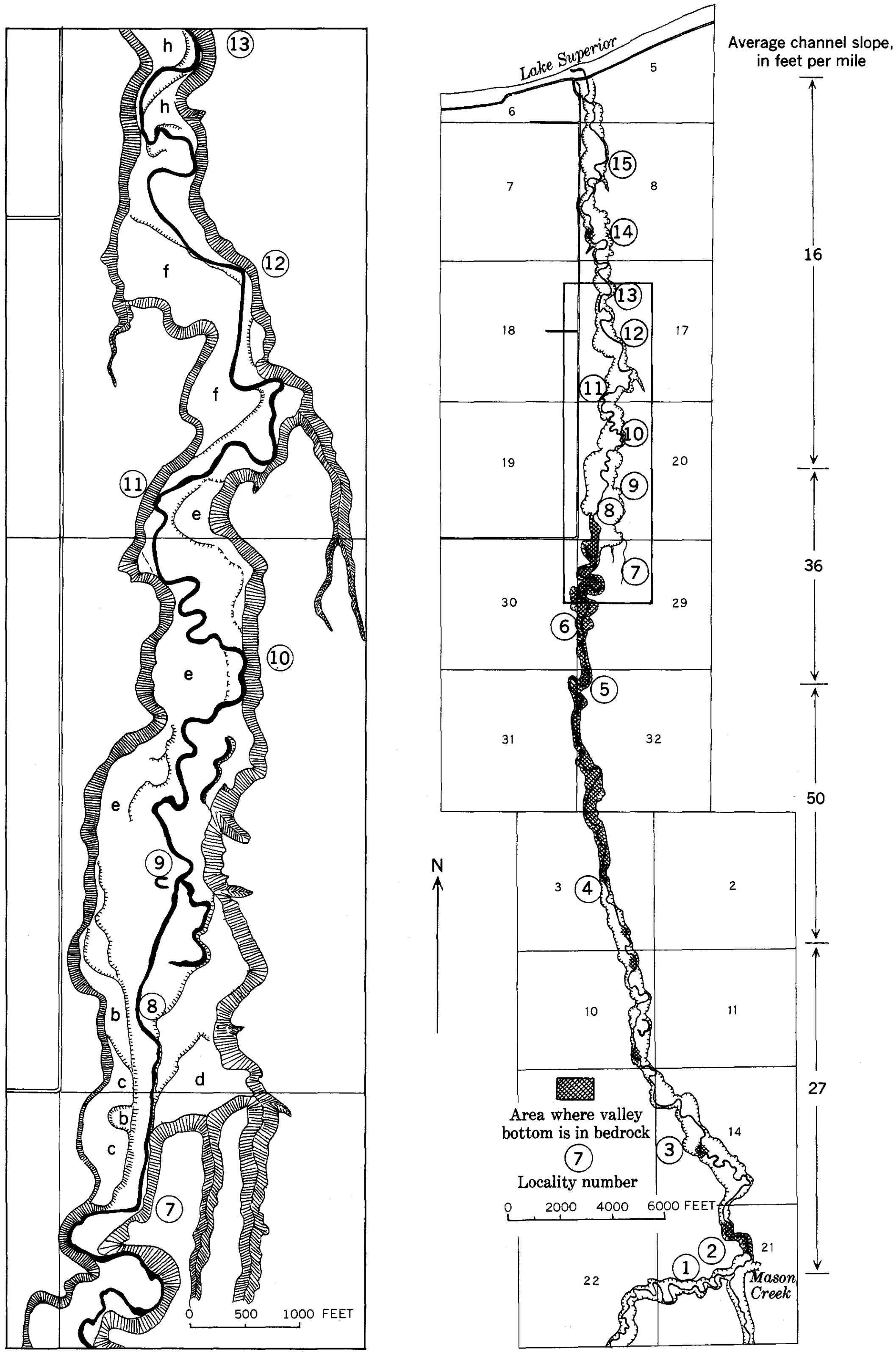


FIGURE 16.—Sketch map showing lower reaches of the Cranberry River. Numbers in circles are localities referred to in text; letters designate terraces. Sketch on left prepared by photogrammetric methods using a Kelsh plotter. Sketch on right traced from aerial photographs; no horizontal correction.

of the dense cover of tall ferns and trees. Study of the model also permitted more measurements of channel slope than could be obtained from topographic maps.

Above Mason Creek (fig. 16), the Cranberry River is intrenched about 60 feet in glacial till and lake sediments. The till is stony and contains large erratic boulders. The streambed is floored with a lag concentrate of rounded boulders washed from the till, many of which are 20 to 30 centimeters in diameter. The meander pattern consists of open smooth curves having a wavelength of about 800 feet.

Below Mason Creek (loc. 2, fig. 16), the stream passes across a resistant bed of felsite whose dip is about 30° to the north. For a distance of 2,500 feet, bedrock is almost continuously exposed in a channel only 30 feet wide within banks 3 to 4 feet high. Many large boulders are in the channel. Downstream where the felsite disappears beneath a gravelly clay till containing many stones the valley widens and a broad flood plain forms. The meander wavelength abruptly decreases to 350 feet, and the character of the streambed changes. In the bedrock reaches the bed is relatively flat and rocky and littered with angular boulders. There are few gravel bars. In the till reach downstream, the channel is floored with rounded cobbles, and there are many bars composed of fine gravel in and adjacent to the stream. Some of the bars are bare; others are covered with herbaceous plants which suggest that the bars are probably shifted by the annual spring floods and that the channel is less stable here than in the bedrock reach. In this reach the bed material is finer grained than above Mason Creek, perhaps because it does not include a layer of bouldery till.

At locality 3 (fig. 16) an outcrop of the Nonesuch Shale forms a threshold in the streambed and the channel is again littered with boulders. At this point the valley walls are in a stony clay till about 70 feet thick. The channel abruptly straightens at the shale outcrop and the meander wavelength increases.

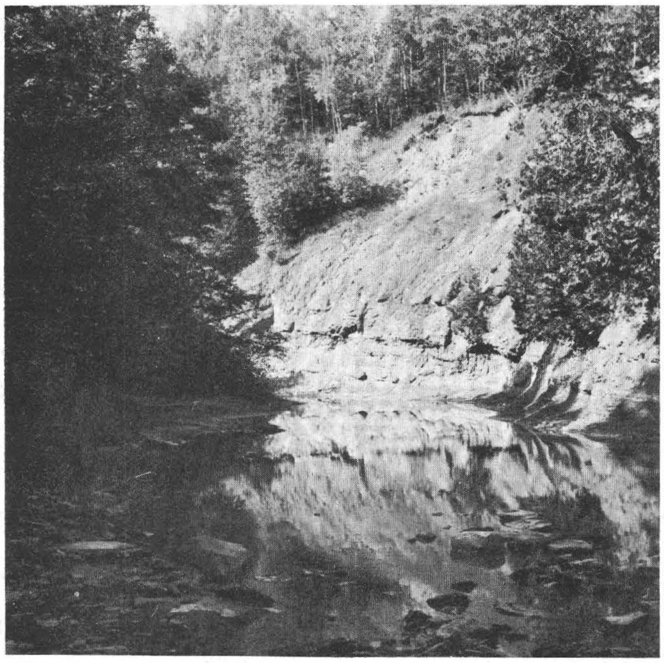
Between localities 3 and 4 the character of the stream is rather monotonous. Depth of the valley averages 40 to 60 feet, mostly in till. The till is very stony in the lower part and contains large erratic boulders of igneous rock. The upper part is less stony and the top 20 feet is generally sandy clay. Outcrops of Freda Sandstone occur at several places in the channel shown as bedrock in figure 16. The outcrops, as well as the boulders in the till, provide a source of coarse bed material and the size of the material in the bed is maintained at a fairly high value (50–150 mm) all along this reach. Channel width is 35 to 40 feet.

At locality 4, figure 16, the river becomes intrenched in Freda Sandstone and at locality 5 the top of the bedrock is 40 feet above the stream. Twelve feet of silty clay till forms the top of the valley wall. The valley gradually narrows as it passes through this bedrock reach and at the narrowest point is only 300 feet wide. Bedrock outcrops in the channel are numerous and the bed material consists mostly of cobbles of Freda Sandstone in both angular and rounded fragments. The channel slope steepens from an average of between 20 and 30 feet per mile to more than 50 feet per mile. The channel cross section is rectangular and shows few gravel or sand bars. The width averages about 50 feet. Mature trees grow on the banks. As shown in figure 16, the river has a tendency to form straight reaches between meander bends, as though the meander wavelength might be very large (as much as 1,500 ft) if the river were free to meander on an open plain.

At locality 6 the character of the bedrock changes from predominantly sandstone to shale or fine siltstone and the valley widens. The bed is mostly bedrock that forms long steplike sheets that slope gently downdip to the north. In many places cobbles, gravel, or other loose material are completely lacking on the streambed. The channel is unusually wide in this reach, in places as much as 125 feet. The meanders are again free and a few meanders of short wavelength are superimposed on the longer ones.

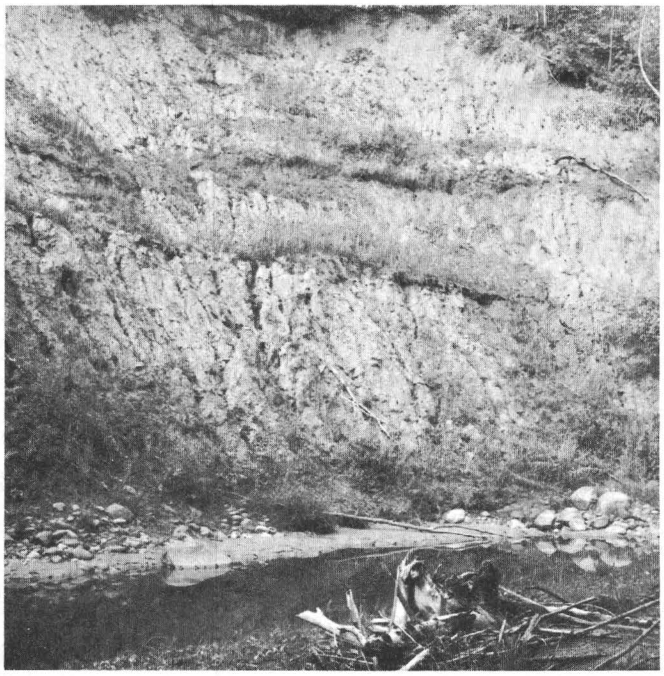
At locality 7 the river is again in sandstone; intrenchment of the stream is at a maximum, and the valley walls are about 80 feet high. Top of bedrock is 30 feet above the stream and the remainder of the valley wall consists of till and lake deposits. Average size of bed material at locality 7 is about 50 mm (fig. 17A).

Downstream from locality 7 the bedrock surface slopes so steeply to the north that within only a short distance no more outcrops of bedrock are visible in the valley walls. The last outcrop in the channel is at locality 8. The north edge of the bedrock marks a profound change in the character of the valley as well as the channel. The valley widens from 500 feet to more than 1,600 feet, and a flight of terraces borders the stream. At the point of emergence from the bedrock-walled valley or canyon (loc. 7), the meanders cease entirely but they begin again at locality 9. Between these points, a distance of about 3,500 feet, the average size of the bed material is reduced from greater than 60 mm to less than 30 mm. At locality 7 the bed contains many angular boulders and cobbles of shale and sandstone. At locality 9 the material is mostly gravelly, poorly sorted, and rounded. Shale fragments are abundant but diminish in amount further down-stream until the bed material becomes a pebble gravel of rounded sandstone fragments.



A

B



C

D

FIGURE 17.—Views of Cranberry and Potato Rivers: *A*, Cranberry River near locality 7 in canyon cut in the Freda Sandstone. *B*, Channel of Cranberry River in till near locality 12 showing bouldery bed material of glacial erratics. *C*, High bank cut by Cranberry River in the valley wall exposing the intermediate till and glacial erratics washed out of the lower layers. This locality is upstream from 17–B. *D*, Gravel bar typical of the small-scale meandering reach of the Potato River.

A marked change in the channel occurs in the vicinity of locality 9 similar to the change described between localities 2 and 3. The channel begins to meander but now has a much shorter wavelength than in the bedrock reach or in most of the upstream reaches. The average wavelength of the meanders is 450 feet, but these meanders are superimposed like a harmonic on larger meanders that have a wavelength of nearly 2,000 feet, slightly larger than the probable wavelength of the poorly developed meanders in the bedrock reach. Whereas the channel above locality 9 is rectangular in cross section and contained in almost vertical banks, the channel below this point winds between rounded gravel bars. The width averages about 50 to 60 feet but in places is much larger. Banks are about 6 feet high but bars in the channel, or adjacent to it are commonly almost as high as the banks.

The average channel slope (measured along the stream and including the bends) also changes at locality 9. Above this point the slope is about 20 feet to the mile; downstream it flattens to about 10 feet to the mile—less than half as great.

The valley walls in the reach between localities 7 and 10 are composed entirely of till and lake deposits. Although the lower layers of the till are quite stony, they contain no boulders and the river does not impinge against the valley wall. Therefore, the river does not pick up coarse rock fragments along the reach and the bed and bank materials are comminuted entirely from the material washed down from the bedrock canyon upstream. The change in character of the channel, as well as the presence of terraces, is probably associated with the change in these materials as well as with a great increase in valley width. The changes are not unlike some of those described along streams in the Appalachians that issue from the mountains onto a lowland plain underlain by relatively soft rock (Hack, 1957, p. 84; 1960, p. 91). Broad alluvial terraces form where the stream leaves the mountain. The channel slope sharply diminishes and the size of the bed material decreases.

At locality 11, although the stream is still intrenched in till and the valley is fairly wide, the wavelength of the meanders again changes and becomes much larger. The change is associated with outcrops of extremely coarse bouldery till that appear in the lower valley walls at locality 11 and downstream. The till forms a lag concentrate of coarse boulders which make the stream bed even rougher than in the bedrock reaches (fig. 17 B, C). The average size of the bed material is 125 mm, but the largest boulders are 700 to 800 mm (loc. 12).

In spite of the change in size of bed material, the channel slope does not change appreciably; downstream from locality 10 it averages 10 to 12 feet per mile. The channel narrows, however, to about 40 feet and the banks are low (4 to 5 ft.). There are no sand or gravel bars in this reach.

Another change that begins at locality 11 is a narrowing of the floodplain which at locality 13 is only 130 feet wide. This change may not be associated with the change in bed material but may be associated with a terrace formed as a result of the backwater effect of a higher stage of Lake Superior, as is discussed on page B27. The terrace may be in an early stage of dissection and the narrow floodplain just developing. The altitude of the floodplain at this point is 625 feet, 23 feet above the level of the lake and only 9 feet above the main Lake Nipissing beach.

Below locality 13 the elevation of the Ontonagon Plain declines and the valley walls become lower. At locality 14 the river crosses another buried ridge of bedrock consisting of siltstone of the Freda Sandstone. This bedrock outcrop is significant because it indicates that the valley above this point cannot be underlain by any appreciable thickness of alluvial fill. The river must be degrading. Where it crosses the bedrock, the banks are 4 feet high and the channel widens to about 60 feet. The banks here are sandy.

From locality 14 to 15 the valley walls and floor are till which, unlike the till upstream, contains no large boulders. The width of the channel again narrows to less than 50 feet, and the size of the bed material declines to an average of about 12 mm; there are occasional riffles of coarser cobbles. Large boulders of glacial erratics are scattered on the bed.

At about locality 15 the channel deepens and the river is affected by backwater from the lake. The bed is not visible below this point, and the river is bordered by a horizontal flood plain and terrace. The mouth of the Cranberry River is cut off from Lake Superior at low river stages by a gravel bar. The great power of the wave-generated currents of the lake relative to the power of the stream at ordinary flows is demonstrated by the almost straight shoreline that cuts across the narrow river estuary.

The Cranberry River suggests several conclusions relating to general problems of the morphology of river channels:

1. The channel slope is closely related to the nature of the valley walls and floor, and steep channel slopes are associated with resistant rocks in the bedrock reaches.

2. Large changes in the size of bed material are associated with changes of slope in some places, as at locality 9; in others, the two changes seem to be independent of each other, as at locality 11 where the bed material changes but the slope remains almost the same. Probably size of bed material is only one of several changes that may accompany a change in channel slope, and its effect on slope may be balanced by other factors.
3. Channel width does not vary much along the river, but it seems to be greatest in the areas that have steep channel slope, as in the bedrock reaches. Thus the channel narrows somewhat at locality 9 where the bed material becomes finer. For some reason, at locality 11 it narrows even more in spite of the coarse bed material introduced at that point.
4. The narrow floodplain below locality 11 may be causally related to the low slope. Perhaps high flood flows that are important in determining the channel conditions pass through this narrow flood channel on gentle slopes with greater depth, and hence higher velocity, in spite of the greater channel roughness.
5. The most striking lesson to be learned from the Cranberry River is perhaps the many changes that take place in the meander wavelengths, which range from 350 to 1,500 feet or more. The long wavelengths are associated with the bedrock reaches and with those that have coarse bouldery bed material. Short wavelengths occur where the bed material is fine and where the channel is bordered by bars that probably are disturbed by annual floods. The problem of the meanders is discussed on page B30.

TERRACES OF THE LOWER CRANBERRY RIVER

A vertical profile of the lower part of the Cranberry River valley was prepared by means of a Kelsh plotter from data obtained from aerial photographs (fig. 18). The area is heavily forested, but the aerial photographs used in making the map (fig. 16) and the profile (fig. 18) were taken in late April when the leaves were not on the trees. Even low terraces are generally visible on these photographs. Spot elevations, however, could be in error by several feet because of difficulties inherent in the photogrammetric technique. The terrace surfaces themselves are irregular and vary several feet in elevation away from the stream. As a result, it should be borne in mind that there may be errors in the elevation and continuity of the various surfaces shown in figure 18. The following conclusions are drawn from study of the spot elevations and the terrace profiles.

1. Rather large variations are evident in the height of the channel banks. The banks are especially high in the upper bedrock canyon and diminish at the lower end of the canyon.
2. The terraces at the mouth of the bedrock canyon between localities 6 and 9 indicate that the difference in elevation between the stream bed in the canyon reach and the stream bed in the till reach was formerly greater than now and that a rather broad transitional slope existed between the till reach and what is now the canyon reach. Presumably terraces c, d, and e are correlative, or nearly so, and are rock-defended terraces formed at a time when the river upstream flowed on the surface of the Freda Sandstone or in it at a shallow depth. At locality 11, however, the river was intrenched almost as deeply as it is now. Terraces a and b are perhaps also correlative and represent a somewhat later stage of this adjustment.
3. The terraces h and j appear to be horizontal and presumably are graded to higher levels of Lake Superior. Terrace h must be graded to the level of Lake Nipissing, for it is exactly 16 feet above the present lake level and corresponds in elevation to the level of the well-defined nip formed by Lake Nipissing. Terrace j must correspond to some lower lake stage, but, upstream, appears to merge with the present floodplain. Neither of these base-level controlled terraces can be traced more than about a mile upstream from the lake.
4. The relation of the Nipissing terrace h to the rock-defended terrace d–e is noteworthy. The rock-defended terrace descends at a steep grade and, if the correlation is correct, at locality 11 is at a lower elevation above the river than is terrace h, the Nipissing terrace. This relation suggests that perhaps most of the canyon in the Freda Sandstone has been cut since the high stage of Lake Nipissing, that is, in the last 3,600 years (Hough, 1958, p. 253), whereas the greatest amount of cutting in the till occurred prior to that time in the 4,000 years after the disappearance of Lake Duluth. The coarse bouldery bed material derived from erosion of the boulder clay at the base of the valley wall at locality 11 and below may have caused the rate of downcutting of this part of the river to be very slow. The boulders may form a limit to further flattening of the slope in this reach. Some evidence for this speculation is offered by the occurrences of bedrock above the main canyon reach (above loc. 4, fig. 16). At least 5 outcrops of bedrock are in the stream in this part of the valley but, except at locality 2 the river has not cut an appreciable depth

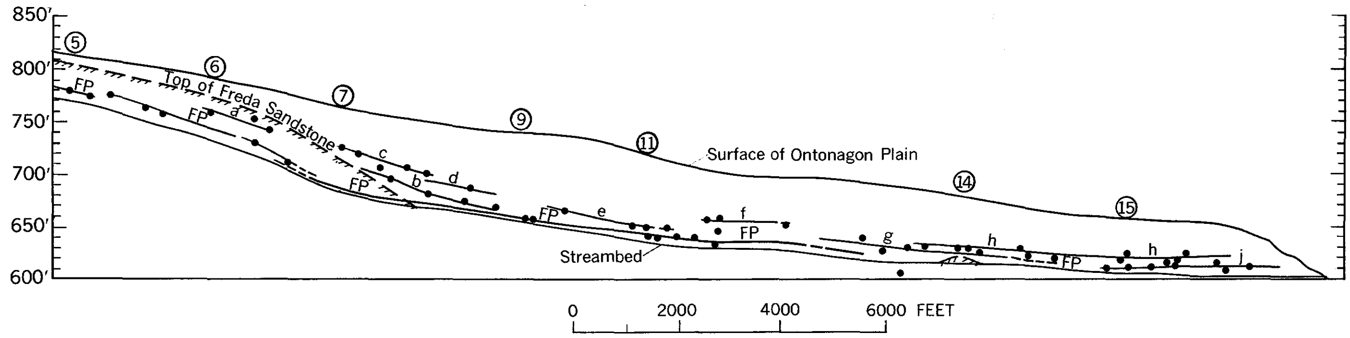


FIGURE 18.—Profile of lower valley of Cranberry River from locality 5 to Lake Superior, showing profiles of upland, valley bottom, and terraces. Numbers in circles are localities shown in figure 16; FP, flood plain; a, terrace; ●, spot elevation on optical model. Data obtained by Kelsh plotter from aerial photographs.

into them. Above locality 4 the surface of the bedrock may inhibit downcutting until the longitudinal profile is steep enough to increase the erosional energy available. A similar condition may have prevailed below locality 4 when the stream flowed at the top of the Freda Sandstone now exposed in the canyon walls (locs. 5, 6).

POTATO RIVER

The Potato River crosses a buried bedrock high area, and about 2 miles upstream from Lake Superior where it leaves the bedrock it displays a change in the character of its channel even greater than that shown by the Cranberry River. The lower course of the Potato River, shown in figure 19, is of interest particularly because of the sharp change in meander wavelength.

At locality 1 and upstream the river is deeply incised in till and bedrock. The surface of the bedrock is 36 feet above the stream and is overlain by 45 feet of interbedded till and lake sediments. The till contains no large erratic boulders at this locality but there are some large boulders on the river bed that apparently have been washed down from upstream.

As in the Cranberry River, the channel cross section is rectangular, banks are vertical, and gravel bars in the channel are small. Channel width is 60 feet and the depth 5 feet. Bed material consists of fragments of Freda Sandstone, shale, and erratic boulders. Average size is 68 mm but sorting is poor and 10 percent of the fragments are larger than 350 mm. One large boulder of diabase is 1,500 mm in diameter. The bed in most places is covered by cobbles, but in others sandstone bedrock is exposed over the entire channel.

At locality 2 the bedrock surface slopes sharply down toward the north; the last outcrop is at locality 3 where the rock surface is exposed on the west bank 7 feet above the stream. Downstream from this locality there are no rock outcrops, and the locality marks a rather sudden change in the valley.

The meander wavelength decreases at locality 3 and remains small all the way to the lake. Large gravel bars, some of which are 5 feet high, mark the channel (fig. 25, lower cross section). The bed material averages 12 mm or finer and there are no boulders. The largest fragments are about 75 mm. The low-water channel is narrow, averaging about 20 to 30 feet in width, but the entire channel cleared of trees may be as much as 125 feet wide. The channel is similar in character to that of the Cranberry River in the till reaches downstream from locality 2 and between 9 and 11, and is therefore characteristic of reaches in fine till in which the floodplain is very broad and in which all bed material must be brought from upstream. There is also a flattening of the slope, but it is not as pronounced as the change in the lower Cranberry River that takes place between localities 7 and 9 (fig. 16).

At locality 6 the river deepens and the channel narrows to 60 feet, a width that includes the bars. Bed material remains fine grained. At about this point on the river the backwater effect of Lake Superior changes the character of the channel and a short distance downstream the meandering character ceases. This point is well within the backwater zone of the lake and the channel is quite deep.

The changes that take place as the Potato River leaves the bedrock canyon and enters the till area are summarized in table 3. Ranges are estimates.

TABLE 3.—*Comparison of features of the channel of the Potato River in the bedrock canyon with those of the till area downstream*

Feature	Bedrock canyon	Till area
Average channel width .... feet	60	60–125
Average size of bed material .... mm	60–80	10–20
Average channel slope .... feet per mile	22	12
Meander wavelength .... feet	800–1000	380
Sinuosity	1.66	1.57

Terraces along the Potato River are shown in figure 19. The high terraces in the bedrock area correspond to the top of the bedrock surface and apparently are remnants of a broader valley that existed before the stream began to cut into bedrock. The terraces in the till reach downstream are apparently rock-defended terraces. No terraces border the river in the narrow valley just upstream from the lake and apparently no terraces can with certainty be said to have formed as a result of a higher base level.

MEANDERS

The example of the Ontonagon Plain should dispel any idea that meanders are phenomena of sluggish streams in areas that have been reduced to a state of low relief. The drainage here is youthful and the streams flow across the plain in almost canyonlike valleys. Alluvial meanders, intrenched meanders, and compound meanders are found in the area and meanders occur in bedrock valleys as well as in till. As demonstrated by the Cranberry and Potato Rivers, the wavelengths of the meanders change in response to changes in bed and bank materials; this change indicates that controls other than discharge affect wavelength. The extensive research that has been done on meandering streams is summarized in Leopold and Wolman (1960) and no attempt is made here to develop the theory of river meanders. Because of the complexity and variety of the meander patterns, however, the meanders of the Ontonagon Plain are worth considering because of the evidence they may add to knowledge of the problem.

CONDITIONS THAT PRODUCE MEANDERS

Leopold and Wolman (1957) have shown that meandering channels are end members of a continuum that extends from braided channels through straight channels to meandering channels. In general, the channel form that develops is related only indirectly to load and is not necessarily a function of the amount of load carried by the stream. Data from many streams show that braided streams generally have a steep slope for a given discharge whereas meandering channels have a gentle slope for a given discharge. A critical line separates observed braided channels from meandering channels on the basis of slope and discharge. The critical slope above which channels do not meander but are either straight or braided is defined by a line having the equation

$$S_c = 317\ Q^{-0.44}$$

where S_c is critical slope in feet per mile and Q is bankfull discharge in cubic feet per second (Leopold and Wolman, 1957, fig. 46).

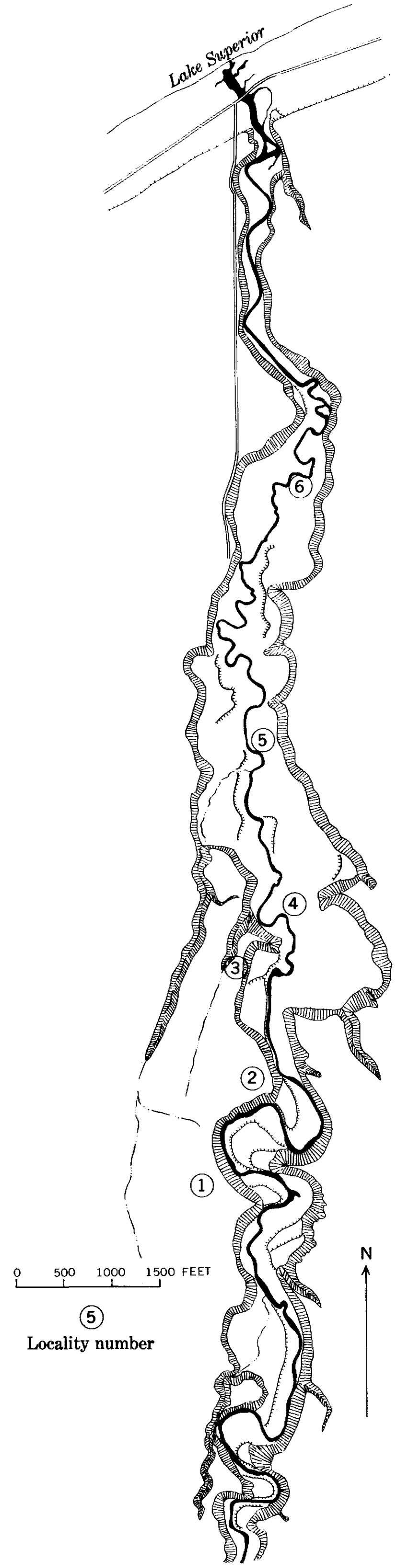


FIGURE 19.—Map of lower reaches of the Potato River. Numbers in circles are localities referred to in text. Prepared from aerial photographs using a Kelsh plotter.

Channels that have a small slope for a given discharge generally are in finer grained, less resistant materials than channels with steep slopes (Hack, 1957; Brush, 1961) and it is clear that meandering streams must be associated not only with gentle slopes but with fine-grained material and also with relatively deep channel cross sections.

Channels in the Ontonagon area meet these conditions. They are for the most part in till and lake sediments or in soft siltstone, shale, and fine sandstone of the Freda Sandstone. The steepest initial slope possible on the Ontonagon Plain itself is only 50 feet per mile, the initial slope toward the lake. For small streams, this slope is probably far below the critical value of Leopold and Wolman. Most, but not all, of the large streams have adjusted their profiles to a considerable degree and flow on much gentler slopes than the initial slope of the plain.

Unfortunately it is not possible to compare the channels of the Ontonagon area with Leopold and Wolman's meandering channels because we do not have data on bankfull discharges in this area. There are, however, a number of straight river reaches in the area and all seem to be in areas where, for one reason or another, an initial steep slope has been imposed on a fairly large stream. Several examples can be seen on plate 1; the Iron River downstream from White Pine is one of the most striking—the slope there is quite steep for so large a stream and averages 24 feet per mile. In this region the cover of glacial drift is thin and the river flows in bedrock through the reach. Furthermore, it flows along the strike for much of the course, the strata are exposed in many places in the channel and there is little fragmental material. In contrast, the Ontonagon River meanders in its lower course; it is cut in till, and its slope has been reduced to only 3.2 feet per mile, low for a stream of its size.

Several other examples of straight reaches may be seen on plate 1. Both branches of the Union River have straight reaches as they flow off the northeast slope of the Porcupine Mountains. The Little Iron River has a straight channel west of Nonesuch where it flows on a steep bedrock slope. At a locality south of White Pine in sec. 16, T. 50 N., R. 42 W. the Mineral River shows the change in pattern that accompanies a change in slope. Above the highway crossing, where the slope is 25 feet per mile, the stream meanders; below the highway, the stream crosses a buried bedrock high area, the slope increases to 100 feet per mile, and the channel straightens. The change in pattern can be seen in plate 1.

Most of the streams in the Ontonagon Plain, even the smallest, have meandering channels. Two of the larger streams in the eastern part of the area, the Flintsteel and Firesteel Rivers, are extraordinary because of the high sinuosity of their channels. They flow down low areas on the Ontonagon Plain in what are probably the axes of preglacial valleys. The initial slope before intrenchment was only about half as great as the slope west of the Ontonagon River where sinuosities are not as high.

In figure 20, the slopes of channels in the Ontonagon area are plotted against drainage area. The nonmeandering channels have high-value slopes.

MEANDER WAVELENGTH IN THE ONTONAGON AREA

In the Ontonagon area the sharp changes in the wavelength of the meanders that accompany changes in the bed and bank materials are noteworthy. Meander wavelengths are related to the size of rivers. Leopold and Wolman (1960) show that in alluvial channels the very conservative relation between channel width and wavelength is such that $\lambda=10.9W^{1.01}$, where λ is meander wavelength and W is channel width and that this relation holds true for a very wide range of river widths.

Dury (1962), in a study of misfit streams, measured the wavelengths of many meandering valleys as well as the alluvial meanders of the streams within them and shows that alluvial meanders in general have a wavelength (in feet) roughly equal to $90A^{0.44}$ where A is drainage area in square miles. Valley meanders of the streams he studied had, on the average, 10 times the wavelength of the stream meanders.

In the Ontonagon area the geology of the valley obviously is a complicating factor that affects the wavelength. A simple empirical classification is used to describe the meanders of the area (fig. 21).

Intrenched meanders.—Intrenched meanders are incised in either bedrock or unconsolidated material. The depth of incision is greater than the normal depth of the stream channel. In effect, intrenched meanders are meandering valleys in which the stream meanders more or less in the same pattern as the valley, though the stream may be bordered by an alluvial plain. There are many examples of intrenched meanders in the Ontonagon area, but perhaps the clearest are in the bedrock reaches of the Cranberry and Potato Rivers, shown in figures 16 and 19. Some short reaches of the Ontonagon River, west of Rockland are also examples (pl. 1).

Alluvial meanders.—The term "alluvial meander" is used to describe meanders of a stream developed in its own floodplain in which the depth of cutting is equal to the normal channel depth of the stream. Alluvial meanders are meanders of the stream channel rather

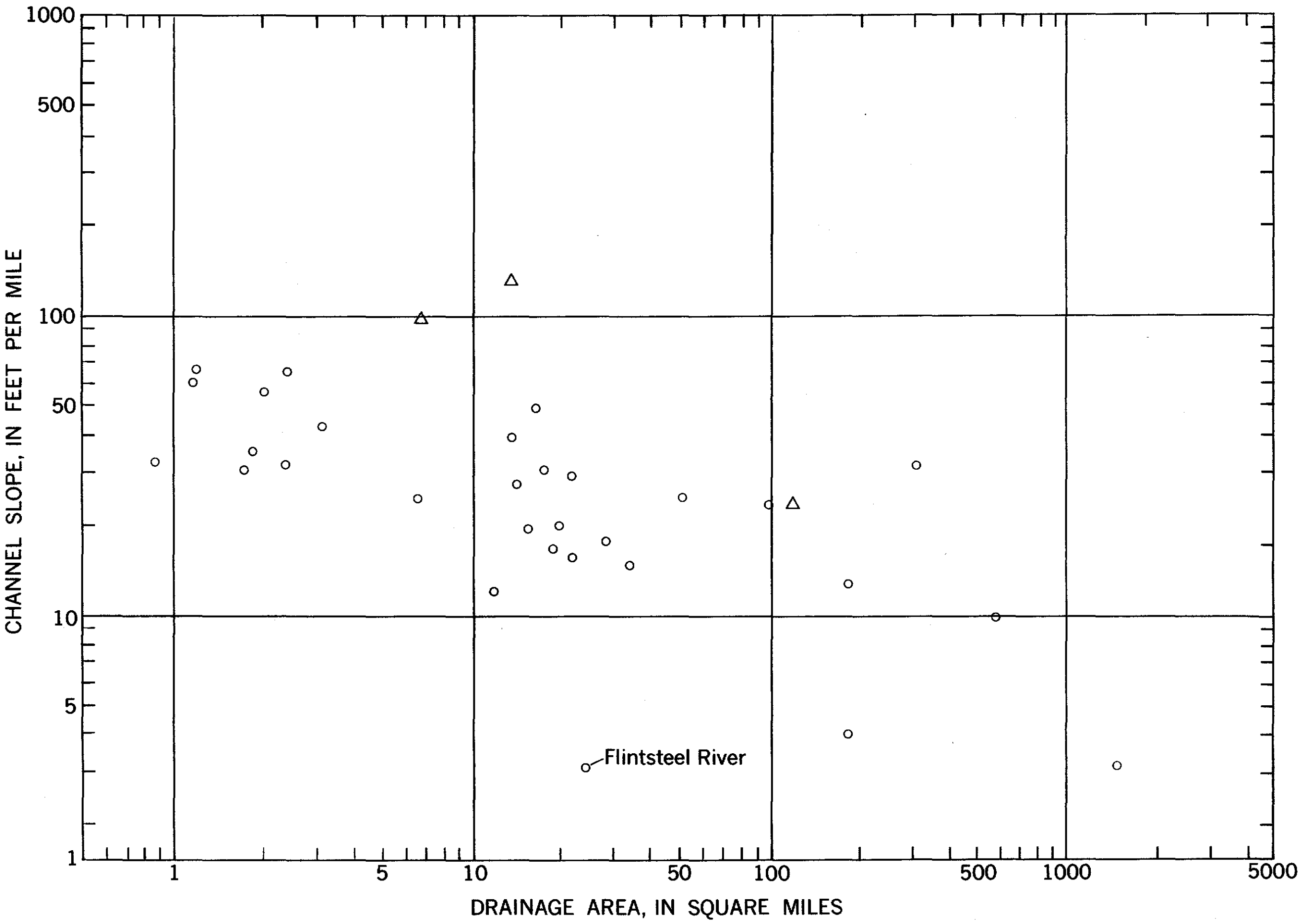


FIGURE 20.—Relation of channel slope to drainage area in meandering and nonmeandering streams in the Ontonagon area. Data are from table 2. △, straight reaches; ○, meandering reaches.

than of the valley. The valley may be many times wider than the meander belt and the pattern of the valley is not necessarily related to the pattern of the stream. This kind of meander is the most common in the Ontonagon area; three examples are shown in figure 21. The lower course of the Ontonagon River has alluvial meanders (pl. 1). Alluvial meanders may occur within intrenched meanders as shown by the meanders of the Flintsteel River (fig. 21).

Simple and compound meanders.—Either class of meanders may be simple or compound. Simple meanders have one regular wavelength like the intrenched meanders of the Cranberry River or of the tributary of the Little Cranberry (fig. 21). Compound meanders have harmonics, that is, meanders superimposed on meanders of a longer wavelength, like the alluvial meanders of the Flintsteel River (fig. 21); the most striking example is the meandering reach of the Cranberry River at locality 10 (fig. 16).

When the alluvial meanders are plotted, as on figure 22, they have a fairly conservative relation to the size of the drainage basin. A good fit is obtained with a line having the equation

$$\lambda = 60A^{1/2} \qquad (4)$$

where λ is meander wavelength in feet and A is area of the drainage basin in square miles. This line is close to, but not exactly the same as, Dury's mean line for the wavelength of alluvial meanders. Two of the reaches with alluvial meanders have wavelengths far higher than the others. These localities are on the Cranberry River where the stream flows through coarse boulder till and the bed material is composed of large erratic boulders (loc. 1, 13, fig. 16).

Intrenched meanders have higher wavelengths and a line having the equation

$$\lambda = 220A^{1/2} \qquad (5)$$

passes through the field of points; thus, the intrenched

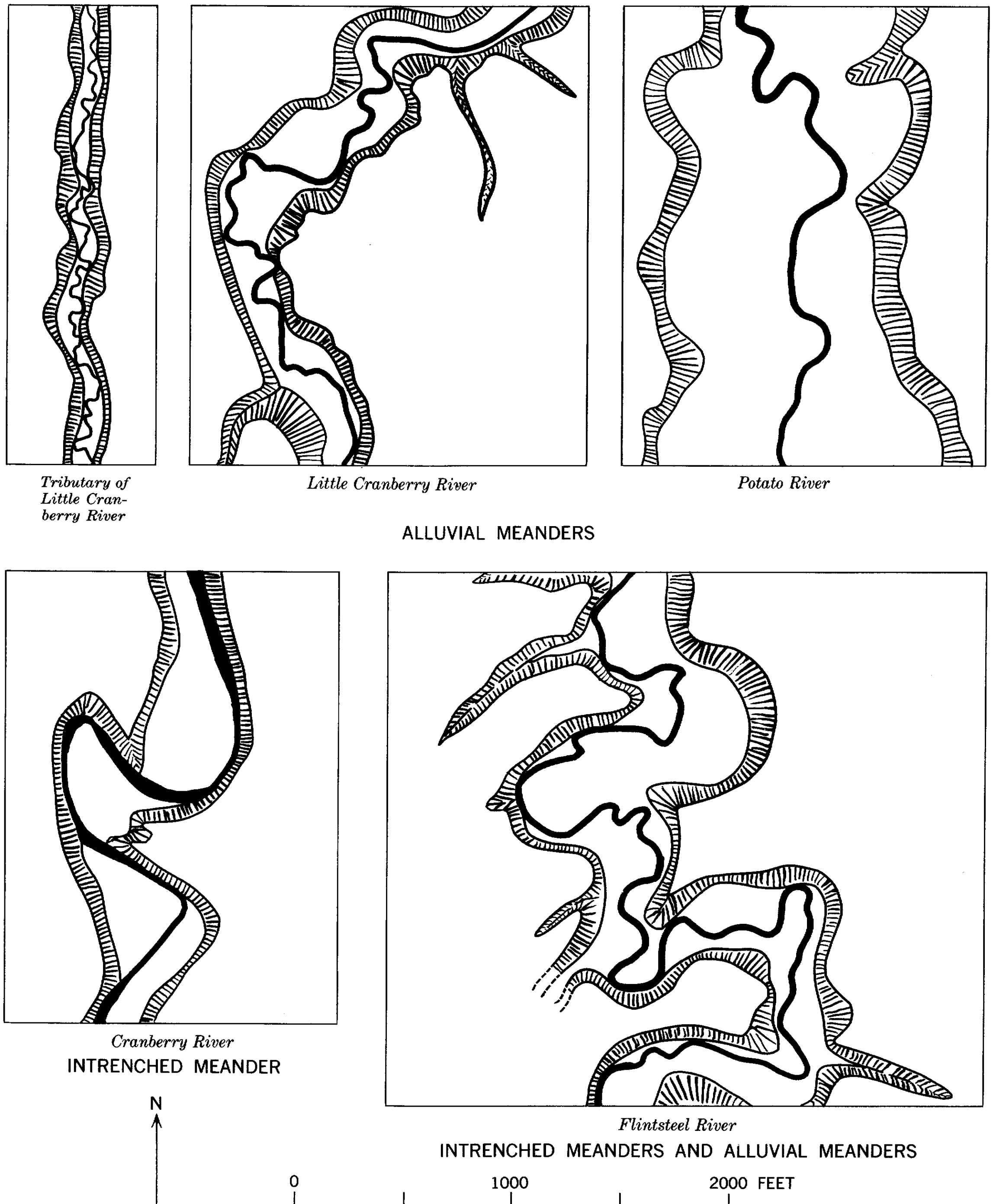


FIGURE 21.—Typical kinds of meanders of streams of the Ontonagon Plain.

meanders tend to have wavelengths about four times those of the alluvial meanders. The scatter is large, however, and the wavelengths of intrenched meanders of some smaller streams is at least five times larger than the wavelength predicted by equation 5.

The Cranberry and Potato Rivers offer an opportunity to examine changes in meander patterns in relation to geologic conditions. As shown on figure 16, at locality 1, the river has typical alluvial meanders, but they belong to the class of meanders with long wavelength ($220A^{1/2}$). In this reach the river is flowing in coarse bouldery till and the bed is completely covered with boulders, many of them more than 250 mm in diameter. The river at locality 1 resembles closely the

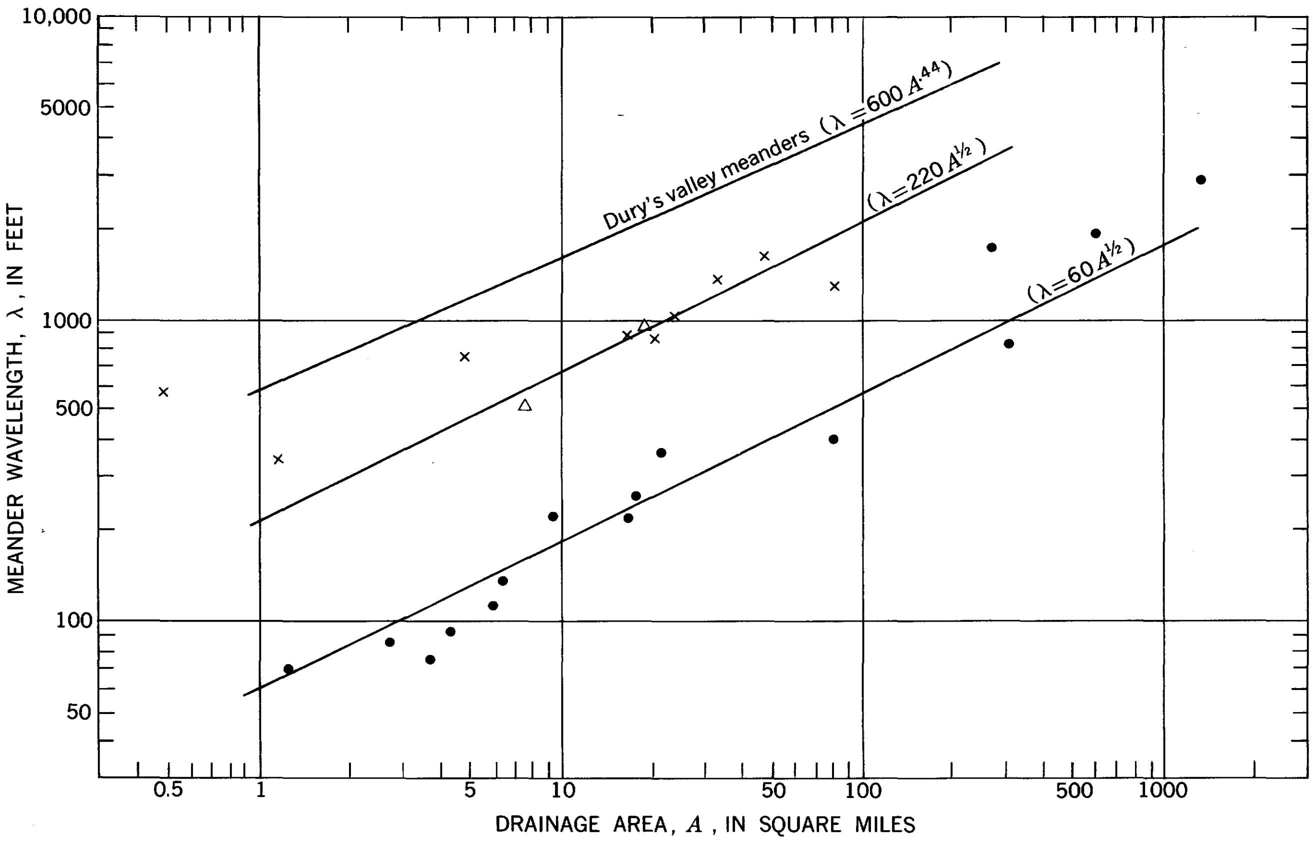


FIGURE 22.—Relation of meander wavelengths in Ontonagon area to drainage basin areas. ●, alluvial meanders; △, large alluvial meanders of Cranberry River; ×, intrenched meanders.

reach at localities 11 to 13 where it flows in similar material and also has large wavelength. In the bedrock reaches between localities 4 and 7, the wavelength is also large, but the meanders are intrenched rather than alluvial. The valley is narrow, and it meanders more or less in the same pattern as the stream. A third distinctive pattern occurs between localities 2 and 3 and between localities 9 and 11. In these reaches the valley is wide, the meanders are alluvial, and there are no bedrock outcrops; the bed material averages 25 mm in diameter or less and is brought down from the bedrock reaches above. Like the alluvial meanders of the Potato River, these meanders are bordered by high gravelly bars that are either bare or covered by grasses and herbaceous plants which suggest that the bars moved during spring flood.

Compound meanders also occur on the Cranberry River. Between localities 6 and 7 the river flows through shaly beds in the Freda Sandstone and the valley widens. Although the meanders are distinctly of the intrenched type, the stream channel itself shows a tendency to meander in a smaller wavelength between low gravel bars. At locality 10, meanders of small wavelength ($60A^{1/2}$) are clearly superimposed on larger meanders (near $220A^{1/2}$), as shown in figure 16. Close examination of the channel, however, indicates that the stream has a tendency to develop another set of meanders, or bends, of even smaller wavelength. These are shown in figure 23. Although the small meanders are not complete, they do have a regular wavelength. They appear to be formed during a lower river stage than the peak spring floods that keep the main channel open.

The Flintsteel and Firesteel Rivers have meandering patterns more complex than the others (pl. 1). They

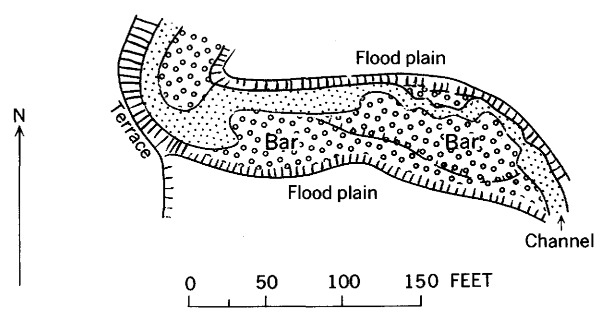


FIGURE 23.—Detailed plan of short reach of Cranberry River, showing imperfect third order meanders.

both have long reaches in which compound alluvial meanders are well developed within the valley, and the valley itself has a fairly high sinuosity. Other meandering valleys wide enough to enclose alluvial meanders either are rare or have a very low sinuosity. The Flintsteel and Firesteel Rivers are different in that they do not flow in a grooved area but occupy valleys in which till is overlain by thick lake sediments. These rivers, therefore, were not constrained by grooves during the initial stages of valley development. Presumably the streams and the valleys eroded by them were free to meander in broader belts than other streams, and therefore wide compound meanders could develop.

In summary, the data in the Ontonagon area indicate that many alluvial meanders increase in wavelength in a regular manner with increasing drainage area and discharge. The rate of increase is not far from the rate reported by Dury (1960, fig. 2). Some alluvial meanders that wander freely across a broad flood plain have abnormally large wavelengths; these meanders are in channels lined with a lag concentrate of boulders. Intrenched meanders in bedrock, like those of the Cranberry River, or intrenched meanders in till like those of the upper Iron River, also have larger wavelengths which are about four times the wavelength of the alluvial meanders. Some streams display as many as three orders of meanders of different wavelength.

MEANDER WAVELENGTHS IN THE SHENANDOAH VALLEY OF VIRGINIA

The meander wavelengths of the Ontonagon area are similar to those of the Shenandoah Valley. The relation of slope to drainage area also is similar, as are the channel cross sections. Furthermore, the river regimen is in some respects similar in that mean annual discharge increases with drainage area at the same rate in both areas. The Shenandoah Valley, like the Ontonagon Plain, contains both alluvial meanders and compound meanders, but the two do not generally occur in streams of the same size and compound meanders are not common. A typical stream in the limestone or shale lowlands of the Shenandoah Valley originates in a rather broad headwater valley with gentle slopes, and develops alluvial meanders along its channel within a distance of a mile or two from the head. Further downstream the channel becomes more rocky, the valley narrows, and at a river length of about 4 miles the enclosing valley narrows and begins to meander. In large rivers deep, well-developed intrenched meanders are the rule.

The wavelengths of the Shenandoah Valley meanders are plotted in figure 24. They clearly are related to drainage area in the same manner as the meanders of the Ontonagon Plain; hence the value $60A^{1/2}$ roughly describes the wavelength of the alluvial meanders and the value $220A^{1/2}$ fits the wavelength of the intrenched meanders.

CONDITIONS CONTROLLING MEANDER WAVELENGTH

Alluvial meanders formed inside meandering valleys, typified by the Flintsteel River (fig. 21), are common in many areas throughout the world, and a general theory has been proposed by Dury (1960) to explain them. Dury considers such streams misfit, that is, the valleys and the valley meanders are too large to have been formed by the streams that now occupy them. He believes that all such streams have valleys that were enlarged during a time when the climate was wetter and colder than now and that the dominant discharge was many times larger. The postulated discharge required is enormous and is estimated at 80 times the present (Dury, 1960, p. 236). If greater discharge is the cause of valley meandering, then it obviously occurred rather recently, and Dury thinks it may have occurred in Atlantic or sub-Atlantic time, that is within the last 6,000 years. Dury (1962) has supported his theory by making borings in many meandering valleys and his findings show that the valleys commonly contain a fill of alluvium much deeper than the stream channel itself.

This theory of misfit streams requires not only that the increase in effective discharge was very large, but also that it occurred within historic time and affected rivers in widely separated areas to a similar degree. These requirements are not easily visualized. The as-

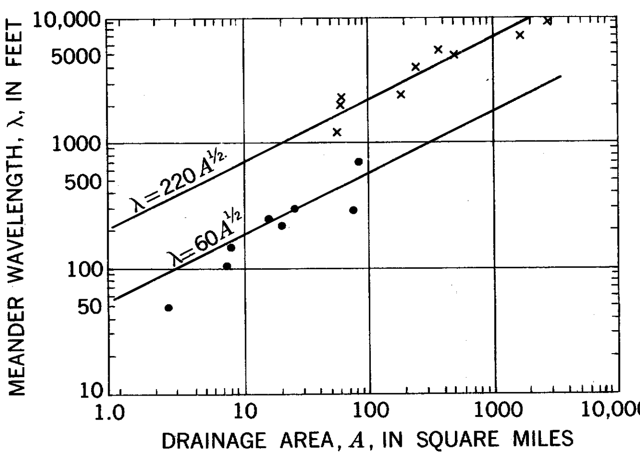


FIGURE 24.—Relation of meander wavelength to drainage area for streams in the limestone and shale lowlands of the Shenandoah Valley. Data from aerial photographs, maps, and field measurements. The lines drawn through the fields of scatter are the lines of best fit determined in the Ontonagon area. ×, intrenched meanders; ●, alluvial meanders in upstream reaches.

sociation of the meanders with a deep fill in some valleys is, of course, not proof that the streams are misfit. It means only that aggradation has occurred.

Application of the theory of misfit streams to the Ontonagon area shows that the entire Ontonagon Plain emerged less than 9,500 years ago and that only 4,200 years ago the Superior basin was occupied by Lake Nipissing (Hough, 1958, p. 253). The lower valleys of streams like the Cranberry and Potato Rivers have been cut at least 15 feet since that time. The lower Cranberry River contains good examples of large-wave-length alluvial meanders ($200A^{1/2}$) that formed below the level of Lake Nipissing (fig. 17, locs. 12–14). If these meanders formed under conditions of greater discharge, the time must have been since 4,200 B.P. (2,200 B.C.), for at locality 14, more than a mile upstream from the lake shore, the elevation of the streambed is 613 feet, only 3 feet below the level of Lake Nipissing, and the entire flood plain downstream from locality 14 is at or below the Nipissing level.

In the bedrock reach upstream from locality 7 the river is intrenched in till and bedrock in places to a depth of 80 feet. As shown by terrace profiles in figure 18, the lower 30 feet of this intrenchment was in bedrock and must have occurred late in the history of the valley, presumably mostly in post-Nipissing time. Inasmuch as the bedrock part of the valley walls have the same meander pattern as the present stream, the meanders were formed either by the present stream or by a stream that meandered in a similar pattern. The conclusion seems inescapable that the meander pattern and wavelength have not changed appreciably in the last 4,000 years and that the large-scale meanders have been formed within this period.

The increase in the dominant or average discharge that would be necessary to produce the large-scale meanders of the Ontonagon area can be deduced from the data shown in figure 22. The curves show that the ratio of the wavelengths of large- to small-scale meanders is approximately 220 : 60 or 3.7 : 1, at any given drainage area. Conversely, the drainage areas at which large- and small-scale meanders have the same wavelength are related according to the ratio 13.5 :1. If it is assumed that the average annual discharge during the period of high precipitation postulated by Dury was proportional to drainage area, as is the present average discharge, then the average discharge must have been greater than that of today by a factor of 13.5 :1 and must have involved enormously greater precipitation. The average discharge would have been comparable in magnitude to the highest momentary discharges now recorded in the region.

The large-scale alluvial meanders of the Cranberry River are especially difficult to explain as the result of increased average annual discharge, or of any change in climatic regimen. They involve the present channel and occur in a stream graded to the present Lake Superior; therefore, they are forming now or were formed very recently.

The concept of the misfit stream as a general phenomenon seems even less plausible when the meanders of the Shenandoah Valley are also considered. In that region the large-scale meanders are 3.5 to 4 times the length of the alluvial meanders and their length in feet is equal to $220A^{1/2}$ (where A is measured in square miles); in other words, the intrenched meanders there have the same relation to the alluvial meanders of the valley as in the streams in the Ontonagon area. The meanders include those of the North Fork of the Shenandoah, which are in reality a meandering valley 80 feet deep in bedrock. It is not likely that this depth of cutting could have been accomplished in the last few thousand years.

A more uniformitarian concept should be sought to explain the changes in the meander wavelengths of the Ontonagon area. That the wavelengths change where the bed and bank materials change is a coincidence that is repeated at several places in several rivers and that involves at least three kinds of material. This coincidence suggests that the cause is probably related to the change in material. No attempt will be made herein to formulate a complete explanation for the different scales of the meanders, but a partial explanation is suggested.

In the Cranberry River, the essential difference between the reaches that have large-scale meanders and those that have small-scale meanders is probably in the kind of material that the reaches are competent to handle at the dominant or effective discharge. There are three reaches in which large-scale meanders are well developed. The first is at locality 1 (fig. 16) where the bed and banks are mostly boulders, and where there are few gravel bars in the channel. Mature trees grow on the banks at the edge of the channel. In this reach, gravelly material is carried through, and out of the reach and the work done by the river is primarily that of scouring the bank and adding to the lag concentrate of boulders. Only a small proportion of the boulders are carried downstream because there are few of them in the channel below locality 2. Conditions are similar in the reach that runs from locality 11 to 14. Large scale meanders also occur in the bedrock reach between localities 4 and 7. In this reach the bed is either composed of angular sandstone cobbles and boulders or is smooth bedrock. There are few gravelly bars, and mature trees grow on the steep channel banks. Probably the channel

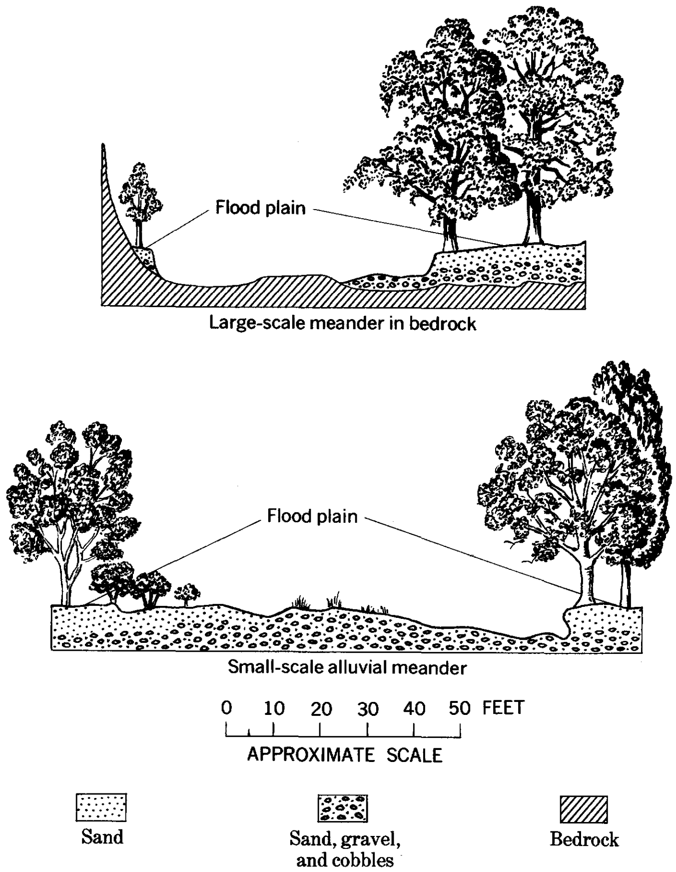


FIGURE 25.—Generalized cross section comparing a typical stream channel in an alluvial reach with a stream in a bedrock reach of the Potato River.

and bed material in these reaches are adjusted to the highest discharges that occur only very rarely, and the annual floods modify the channel only slightly.

The reaches of typical smaller scale alluvial meanders, however, as at localities 2–3 and 7–11, are quite different. In these reaches the banks are poorly defined. The river bed consists in part of a narrow low-water channel of bare rounded gravel moved from upstream. This channel is bordered by wide gravel bars commonly covered by sparse grass and in places by dense growth of annual plants or low shrubs. The outer limits of the channel are defined, in places poorly, by a forest of mature trees. It is inferred from this distribution of the vegetation that the bed material and the material in the bars is probably moved around each spring, or at least every few years. The valley also differs in these reaches; it is much wider and commonly contains terraces as well as a wide flood plain. In these reaches the channel must certainly migrate at a much faster rate than in the other reaches, for it has succeeded in shaping a much broader valley. Because the material handled by the stream is fine, more energy goes into shifting the channel and handling material eroded from the steeper reaches upstream. The differences between the two kinds of reaches are illustrated diagrammatically in figure 25 taken from measurements of the channel of the Potato River. Figure 17*D* shows a typical gravel bar in the lower meandering reach of the Potato River.

The difference in the character of the material handled by the stream probably produces a difference in the equilibrium of the channel and affects all the other variables involved, including especially depth, width, and velocity, even though the discharge may be the same. Presumably the difference in equilibrium also includes the meander wavelength. Applying this concept to other areas we can expect, as we have found in the Shenandoah Valley of Virginia, that meanders in streams that are eroding a bedrock channel will have larger meander wavelengths than streams that handle alluvial materials mostly transported from upstream. In the Shenandoah Valley, the small scale or alluvial meanders are confined to the headwater areas and the fine-grained alluvial material is derived by creep and wash from the residuum and soil formed on the valley sides (Hack, 1957, p. 85).

SUMMARY AND CONCLUSIONS

The foregoing study of the drainage system of the Ontonagon area was undertaken in order to observe a drainage network in a place where, because the character of the initial surface is known and dated, the evolution of the valleys that are now cut into it may be inferred. The conclusions arrived at are now briefly recapitulated with little reference to the supporting data.

THE DRAINAGE SYSTEM

Development of the Ontonagon Plain.—The Ontonagon Plain is a glacially grooved plain underlain mostly by clayey and silty sediments in part lacustrine, and in part glacial, of Valders and post-Valders age. The buried bedrock surface is irregular and overlain by as much as 200 feet of till, sublacustrine till, and lake sediments, predominantly fine-grained clay and silt. Of particular importance in the study of the stream valleys is the fact that the plain is cut by a set of discontinuous grooves that has exerted a strong control on the drainage pattern which formed as the plain emerged from glacial Lake Duluth. The lake withdrew from the grooved plain about 10,000 to 9,500 years ago. The evidence suggests that withdrawal of the lake was rapid, and that most of the plain was drained by 9,500 years B. P. At the time of Lake Nipissing, 4,200 years ago, the lake level was about 15 feet higher than present Lake Superior, and an escarpment was cut back into the till plain after that date. The base level of erosion in the area has been at or below 616 feet ever since. The present base level (Lake Superior) is 602 feet.

Cutting of the valleys.—Deep valleys have been cut in the Ontonagon Plain, penetrating both till and buried bedrock ridges. The valleys have been deepened by amounts roughly proportional to the quantity of water discharged through them. Small valleys have deepened little, even though they are at or close to the lake edge. The drainage network formed almost immediately as the lake withdrew. The drainage outline was determined by the pattern of the grooves and other initial surface features of the plain and has changed very little. Piracies have occurred primarily as a consequence of downcutting as larger valleys grew and engulfed smaller ones. The break in slope at the nip formed by Lake Nipissing is preserved only in the profiles of the smaller valleys and shows as a convexity that is the more pronounced the smaller the valley. The character of the valley profiles is apparently determined more by the discharge and load of the streams than by changes in base level or, as Rubey (1952, p. 134) put it, by duties imposed from upstream. The headward migration of channels and divides seems to take place very slowly and concomitantly with downcutting of the valleys.

The drainage geometry.—Although attenuated because of the grooves, the consequent drainage network, formed as glacial Lake Duluth receded, has many of the attributes of highly developed drainage systems in areas that have been erosionally graded. The characteristics of the drainage net on the Ontonagon Plain as defined by Horton (1945), including the rate at which the streams bifurcate as the drainage area increases, are similar to those of Appalachian drainage areas. The rate at which the drainage area increases with increasing stream length is also the same as in many other areas, although for any given stream length the drainage basin area is much less. The geometry of the drainage system is explained by the theory of maximum probability, that is, the system grew in a random manner as the lake receded from the plain and the runoff was discharged downslope through the grooves. The similarity to patterns in erosionally graded areas suggests that maximum probability largely controls the joining of streams in both environments. The importance of this generalization lies mainly in the conclusion that the rate of increase in discharge with stream length must, in most cases, be an independent variable in the channel equilibrium. The rate is determined partly by the climate of an area and partly by the probable six-tenths power rate of increase of length with drainage area. In a homogeneous landscape, the average rate of increase in drainage area relative to length is probably not affected by the amount of erosion in a given area, the degree of gradation, the perfection of the adjustment to geologic structure, or even by the average slope. However, local environmental factors such as structure obviously do affect the absolute value of the drainage area at a given stream length, and local changes in such factors must cause changes in the relation of area to length.

Longitudinal stream profiles.—The long profiles of the streams in the Ontonagon area are irregular as compared with streams in more completely graded landscapes. The larger streams, however, have some degree of concavity in their profiles. Because the profiles are adjusted to the nature of the materials that enclose the valley, stream reaches in bedrock are much steeper than reaches in till, other factors being equal. If the streams that are cut in till are considered by themselves, the relation of drainage area to slope is close and the gradients decrease systematically with increasing drainage area and discharge. Because the shape of the drainage basin is an important factor in determining the discharge at any given point on a stream it also has an influence on the profile. In general, the profiles of the streams in the grooved part of Ontonagon Plain do not decrease in slope at the same rate downstream as streams in areas where drainage basin shapes are more normal.

Meanders.—Meandering streams are the rule in the Ontonagon area. A few streams however, have straight courses and these streams invariably have unusually steep slopes for a given discharge. They generally have bedrock bottoms and banks. Streams such as the Flintsteel River that have very high sinuosity, on the other hand, are cut in lake sediments and have unusually low slopes. The relationship of slope to discharge that limits the meanders is similar to the relationship cited by Leopold and Wolman (1957). Meander wavelength in the Ontonagon area increases with discharge, but it also changes with the character of the bed and bank materials: in coarse materials the meander wavelengths at a given discharge are longer than in fine materials. Valley meanders also have a longer wavelength than the stream meanders. It is suggested that the meander wavelengths are determined by the discharge that is most effective in forming the channel in a given kind of material. No basis is found in this region to support the hypothesis of Dury (1960) that the large-scale meanders are inherited from an interval in late Pleistocene time when average discharges were larger than the present discharges.

EVOLUTION OF THE LANDSCAPE

The Ontonagon Plain is an erosionally graded landscape in the sense that subaerial erosion has produced some change in almost every part of it and the changes are related in an orderly way to environmental factors. On the other hand, the area as a whole cannot be considered a single large system in dynamic equilibrium. Every reach of every stream might be regarded as a small system in equilibrium; that is, the form and slope are adjusted for the transportation of a certain load through a certain geologic structure under the available regimen of discharge. A sudden change in the environmental conditions will produce a response such as a change in channel form. However, when a stream of some size such as the Little Cranberry River is considered, it is evident that its valley and channel as a whole are not in equilibrium. The valley is constantly changing as erosion continues and the changes involve the forms as well as the local relief. The Ontonagon Plain, therefore, differs from stream-eroded landscapes in areas such as the Appalachians where, presumably, all the elements of the landscape are mutually adjusted over large areas as the result of long-continued subaerial erosion.

At present the Ontonagon Plain contains many features, such as the grooves and shorelines, that are relict from past conditions, graded not by streams but by glacio-lacustrine processes. Eventually, all these features will be modified or obliterated by subaerial erosion. When all parts of the plain are mutually adjusted, equilibrium will be achieved throughout and there will be little further change in form except as the relief may slowly become lower.

By making certain assumptions, we can infer what the evolution of the Ontonagon Plain has been in the past and how it will proceed in the future. If we assume that small and large valleys evolve in the same manner, we can then infer, for example, that the Cranberry River valley once had the profile and other characteristics of adjoining smaller valleys and that the smaller valleys, as they enlarge and deepen, will evolve toward the form of the present Cranberry River valley and will develop the same slightly concave profiles. In general terms of evolutionary development, the typical valley of the Ontonagon Plain was initiated by the withdrawal of Lake Duluth. The drainage basin area and discharge were fixed and determined by the initial features of the plain and have not appreciably changed. At first the valleys were shallow but gradually deepened at rates that were greater downstream. Because valley width was determined by the width of the meander belt, largely a function of discharge, the initial width of the valleys was not very different from the present. The greatest change was in the profile. We know from study of the terraces that downcutting of the valley did not keep pace with the withdrawal of the lake. The resulting break in slope was reduced both by vertical downcutting of the stream and by a smoothing back of the break from the lake. The profile became first convex, then straight, then slightly concave. Concomitantly with the evolution of the profile, changes in channel shape took place. The initial channels, of the larger streams at least, must have been wider than they are now to compensate for the greater steepness. Initial channel slopes of some streams may have been too steep for the development of meanders, but as time went on meanders began to develop in streams, where they did not exist initially.

There are few facts to guide any inferences about the evolution of interstream areas. Nevertheless, there has been some rounding and smoothing of the valley walls. Presumably the valleys widen somewhat through time as the meander belts themselves migrate from one side to the other of the valley. If base level remains the same indefinitely this widening may be accompanied by a rounding of the upper slopes and a softening of the sharp valley crests to smoother convex forms, until ultimately the valleys will be separated by rounded hills.

If base level remains the same, the drainage pattern will retain its attenuated form for a great length of time, for the drainage basins can widen only as the topography is lowered and as larger valleys engulf smaller ones. The development of the drainage to the present gives us no reason to believe that there will be any rapid headward migration of divides away from the master streams. There is probably a limit to the distance that a divide can migrate without continued downwasting of the entire landscape. Eventually, the slopes in the area will be mutually adjusted and there will be no vestige of the grooved plain except an attenuated drainage network. This evolutionary cycle need not be postulated in great detail; it has actually progressed to only a very limited extent. The remarkable aspect of the changes that have occurred thus far is the degree of gradation of the stream valleys in spite of what classical geomorphologists might call the extreme youthfulness of the topography. The present developments includes an adjustment of the profiles both to rock type and to discharge.

The particular evolution that we envisage for the Ontonagon Plain, on the rather unlikely assumption that base level will not change, is in some respects similar to the classic geographic cycle of Davis (1899). The Ontonagon Plain is not presented, however, as a general example. The evolution that has occurred there can exemplify only areas that have had a similar geologic history. Furthermore, the evolution is only superficially like the theoretical cycle of Davis, and there are important differences. In the classic concept (Davis, 1899), the evolution of the forms is dependent on base level. Although the larger streams on the Ontonagon Plain have indeed cut deeper valleys than the smaller ones, the gradation of none of them has much dependence on base level, except as base level limits the steepness of slopes in the area as a whole. The forms of the longitudinal profiles as well as the valley cross sections are dependent primarily on discharge and on what we might refer to as upstream factors. Even larger streams, such as the Cranberry River, except for a short backwater curve, enter Lake Superior at grades that are related to their increasing discharges; the effect of an important recent change in base level is expressed only in a short stretch of terrace about a mile long.

GRADED STREAMS

The writer has suggested that the term "grade" or "graded" as applied to streams should be used simply to refer to the slope of any smooth adjusted channel (Hack, 1960, p. 83–85). The term is generally reserved for streams whose slopes are thought to be stable and determined by a load transported from upstream (Mackin, 1948). If the term is thus restricted, other channels such as those cut in hard rock are considered "ungraded," even though they may have equally smooth and regular profiles determined by a load that is locally derived, or by other geologic factors. The Cranberry River well illustrates the point. This stream has a well-developed profile that may be divided into several smooth segments. It is, however, lowering its bed along most of its course, its profile is presumably changing, and the bed load at any given place is composed partly of material transported from upstream and partly of material acquired locally from the bed or from the banks and valley walls. Nevertheless, the river is graded in the sense that its channel slope is adjusted in a manner similar to other streams in the same region for the erosion of a given type of material.

The Cranberry River does have one or two short reaches that seem to meet the conditions suggested for the more special connotation of "graded" stream. The reach, about a mile long, between localities 7 and 10 (fig. 16) is an example. In this reach the bedrock floor of the channel is covered by cobbles and gravel, the material in the banks is entirely transported from upstream, the floodplain is bordered by terraces, and the size of the bed material decreases from an average of more than 50 mm to about 25 mm. At the upstream end of this reach the bed and bank material contains many angular fragments of shale as well as sandstone derived from immediately upstream. Within a short distance, however, the shale is comminuted and absent from the bed material; only the more resistant fragments are in the deposits and they are more rounded than those upstream. This condition ends near locality 10 where the stream picks up material eroded from the till that contains coarse cobbles. Below this point there is no further decrease in the size of the bed and bank materials. At locality 11 where the river encounters a coarse layer in the till the size increases again.

The "graded" reach from localities 7 to 9 coincides with the place along the valley where the river leaves its bedrock valley and enters a valley cut exclusively in glacial till. The profile changes from a steep one that is nearly without curvature and in which the average slope is more than 30 feet per mile, to a gentler profile that averages 16 feet per mile. It is a transitional reach between two profile segments in which the equilibrium conditions differ because of a difference in the enclosing materials. Within the transitional reach the profile must be more sharply concave than either up or downstream in order to make the transition. Terraces exist because the bed material transported from upstream cannot be carried off in the gentler reach downstream and it is stored in the transitional reach until reduced by weathering and wear to smaller sizes. In all three reaches the river is presumably lowering its bed. This writer sees no advantage in calling the transitional reach "graded," as though it were more stable, and the others "ungraded."

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