General Theory of Meandering Valleys

By G. H. DURY

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UNITED STATES DEPARTMENT OF THE INTERIOR

STEWART L. UDALL, *Secretary*

GEOLOGICAL SURVEY

Thomas B. Nolan, *Director*

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PRINCIPLES OF UNDERFIT STREAMS

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ABSTRACT

Meandering valleys with systematic windings are defined as subclass of winding valleys, the bends of which can be inifferently systematic or irregular. The terms "valley meander" nd "valley bend" are correspondingly applied to single features, lthough they are conveniently interchangeable in practice. 'ree meanders on a flood plain are distinguished from incised ıeanders, which are subclassified into intrenched and ingrown ɔrms. Misfit streams, either too large or too small for the alleys in which they flow, are grouped as overfit and underfit. Vavelength of meanders is used as a criterion of size and is :atistically referable to discharge at the bankfull stage. The ɜrm "underfit stream" also is applied to streams that either fail ɔ describe meanders in their present channels or that lack alley meanders. Meandering streams in more amply meanderıg valleys are called manifestly underfit.

Stream capture or other forms of derangement of drainage re incapable of supplying the general hypothesis of the origin f underfitness that is required by the facts of distribution. A ritical reexamination of W. M. Davis' attempts to explain nderfitness by capture reveals grave weaknesses in his arguıent. Authentic diversions from the French Rivers Meuse and ·ar were less effective in reducing wavelength of meanders ıan was the change which made all the neighboring rivers lso underfit.

The Wabash and Souris Rivers, the drainage on the emerged oors of glacial Lakes Agassiz and Harrison, and the back-slope treams of the English Cotswolds provide data on spillways. 'hese data show that overspill from proglacial lakes, or direct ischarge of melt water from ice fronts, is irrelevant to the eneral problem of underfitness as it is separable from reducion of stream volume in either space or time, or both.

Regional development of manifestly underfit streams requires . climatic hypothesis of origin. Examples of such streams, from he till plains of Iowa, from the Humboldt, Shenandoah, Salt, .nd Cuivre Rivers, and from the Ozarks, show that lack of ıanifest underfitness can be due either to the destruction of ·elevant valley forms by erosion or to the failure of streams o develop meanders in their existing channels. Alternatively, his lack may be merely apparent, owing to deficiencies of map ·vidence. Where stream meanders are actually lacking, poolınd-riffle sequences can occur with spacing that is appropriate o stream meanders and far closer than that appropriate to 'alley meanders.

Davis' claim that the terraces of the Westfield River indi:ate no significant change in volume during terracing is directed ıt the views of Emerson. But the claim seems not to bear on he general problem of underfitness, especially as manifestly ınderfit streams occur on the emerged floor of glacial Lake Hitchcock. In Arizona, Padre and Diablo Canyons and Oraibi Wash are manifestly underfit streams, providing evidence in arid regions of reduced channel-forming discharge comparable to that noted for humid regions. In association with observations for the Humboldt River, this evidence extends the general hypothesis of underfitness to regions that are currently semiarid or arid.

Contact of streams with bedrock is provisionally rejected as a direct cause of the absence of meanders from present unbraided channels, and variation in cohesiveness of alluvium is suggested as an alternative. The view that incised meanders are necessarily inherited from free meanders is controverted. Numerous minor canyons are thought to act as flumes after heavy rain, but valley nets in dry regions are inferred to have been intitiated and developed under more humid conditions than now exist.

INTRODUCTION

The object of this paper and and subsequent papers is to present a general theory of underfit streams. Incorporated in these papers are the results of fieldwork in the United States and in England, analytic techniques applied to hydrologic and dimensional data, and inferred hydrologic changes in Quaternary chronology. Because the necessary discussion involves reference to studies of rivers in many aspects, to meteorology and climatology, and to the whole corpus of sciences that deal with Pleistocene events, it is impracticable at this point to review previous work. As far as is possible, references are concentrated in particular sections of the text, and cross references are kept to a minimum.

In the customary but too restricted sense, an underfit stream is a stream that meanders on a flood plain in a meandering valley. Figures 1 through 3 illustrate the relevant landforms in an oblique aerial view, a topographic map, and a stereoscopic vertical photograph. The writer's planned work on streams of this type began in 1946, although scattered observations had been collected for some time previously. Results obtained before 1950 were included in a thesis [1] which, relating to parts of the English Midlands, noted the widespread contrast between the surface forms of mean-

[1] Dury, G. H., 1951, Some aspects of the geomorphology of part of the Midland Jurassic belt: London Univ. Ph. D. thesis.

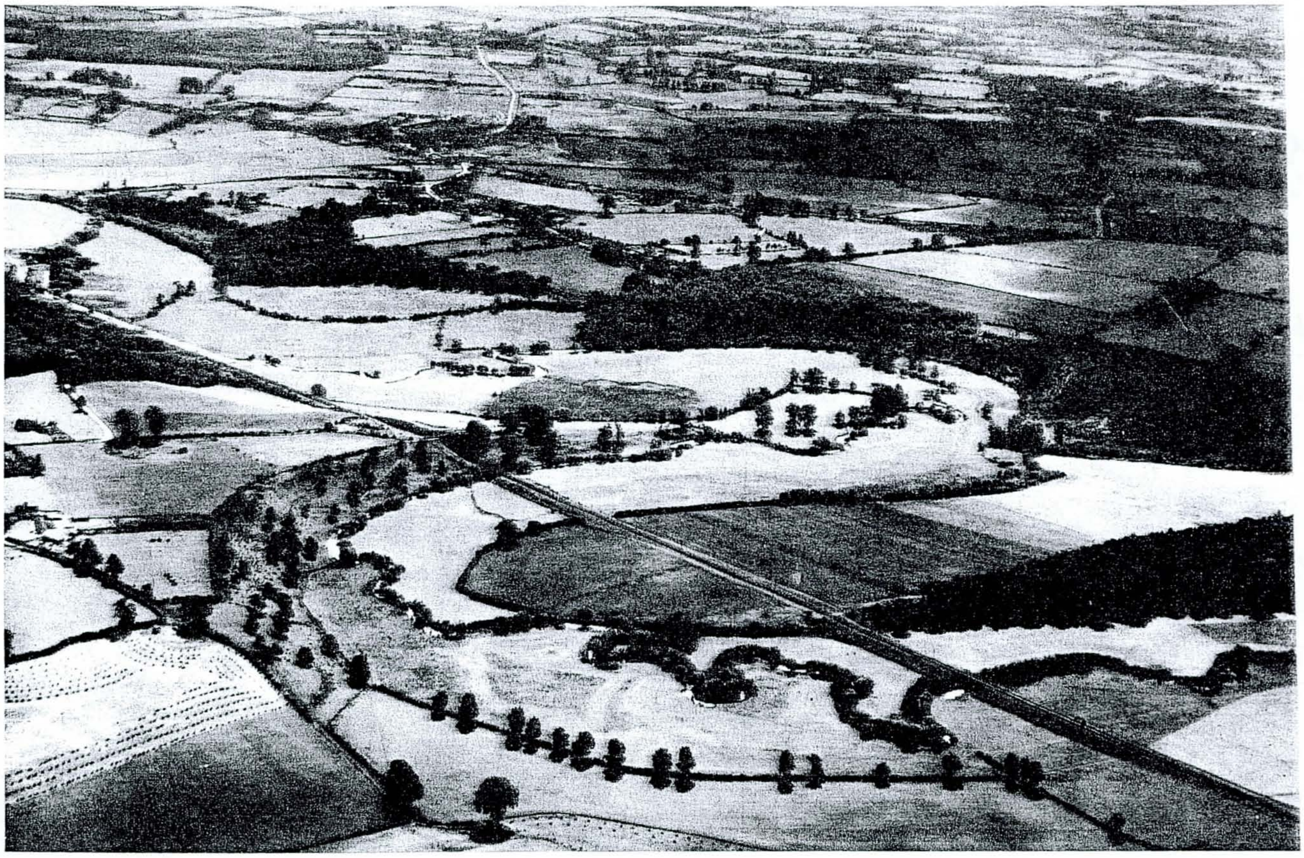

FIGURE 1.—Oblique aerial view of valley bends and stream meanders on the Evenlode River, Oxfordshire, England. The Evenlode i underfit stream. View is toward the southeast; length of the railroad shown is about 1 mile. Photograph by Photoflight, Ltd., El Aerodome, Elstree, Herts, England.

dering valleys and those of meandering stream channels. Subsurface exploration (Dury, 1952; 1953a,b,c) proved certain flood plains to be underlain by large meandering channels, which were taken to be the channels of the former rivers that carved the valley meanders. Investigation was later extended to parts of southeast England, which, unlike the Midland sites, was not ice covered during the Quaternary, and to the borderland of Wales. At each site explored, a large filled channel was located. These various observations were brought together in an account (Dury, 1954) that rejected all general hypotheses of the origin of underfit streams with the exception of some climatic hypotheses, that sketched an initial chronology, and that attempted to determine the discharges required by the large channels and by the valley meanders.

In the time between the writing of the 1954 paper and its publication, Leopold and Maddock (1953) published their analysis of hydraulic geometry. Dr. Leopold drew this account to the writer's attention. Both he and Mr. Maddock visited the University of London, to which the writer was then attached, to discuss implications of work then in progress; they also visited field areas and recommended lines of attack. Statistical techniques of the kind used by Leopold and M dock were found to support the view that the la filled channels are associated with valley mean (Dury, 1955), and these techniques were applied (D 1956) to a reexamination of the diversion of the up Moselle from the Meuse. Later (Dury, 1958), t also were applied to an amplified set of data that cluded the results of subsurface exploration in the C wold Hills (Dury, 1953a). At the suggestion of L pold, the writer made a general review of prog (Dury, 1960) which included comments on the w distribution of underfit streams in meandering vall By this time it was becoming clear that one of main episodes of hydrologic change postulated to count for underfit streams is referable to events t occurred at end-glacial times, in the loose general s nificance of that expression.

Meanwhile, Leopold had arranged for the writer be attached to the U.S. Geological Survey as a divisi staff scientist in the Water Resources Division, bas in the Washington, D.C., office. The project, supervis by Leopold, was to last for 1 year, July 1960 throu June 1961. Working with the broadest terms of ref ence to channel habit, valley form, and chronolog he writer executed fieldwork in the environs of Washngton, D.C. (Coastal Plain, Piedmont, Appalachians); outhern New England; the neighborhood of Evanston, ll.; the Driftless Area of Wisconsin and part of its nargin; parts of the till plains of Iowa; the northern)zarks; the vicinity of Denver, Colo.; the Wasatch Aountains of Utah; parts of the Humboldt River basin n Nevada; and Arizona, mainly on tributaries of the Little Colorado River. Other visits were made to West Virginia; the vicinity of Fort Collins, Colo.; northern Aichigan; and the Lake Bonneville country around Salt Lake City. Four months of the year were spent n fieldwork; the remaining 8 months were used in the ollection and analysis of material, parts of which are not required by the present study and will be treated eparately on another occasion.

The plan of this and the sequential papers is the ollowing: (a) The reexamination of the connotation, usage, and overtones of the terms "underfit stream" and "meandering valley," with examples to illustrate and ustify the conclusions reached, and to demonstrate the very wide occurrence of underfit streams of one type or other (it is in this part that previous work is prinipally cited); (b) the introduction of new field evidence, part of which supplies critical dates, and a liscussion of dating in general; and (c) the discussion n some detail of the hydrologic and climatic implicaions of the general theory. The final part includes an

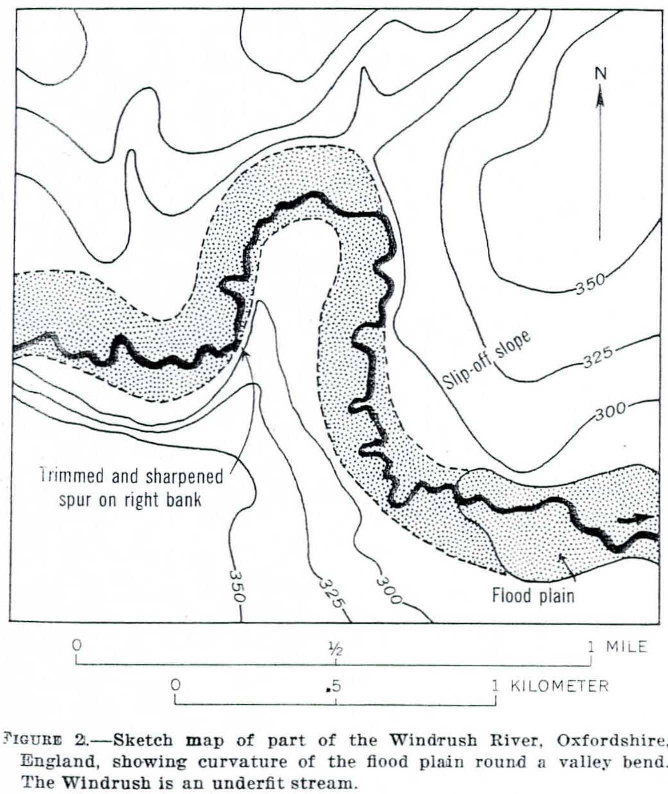

FIGURE 2.—Sketch map of part of the Windrush River, Oxfordshire, England, showing curvature of the flood plain round a valley bend. The Windrush is an underfit stream.

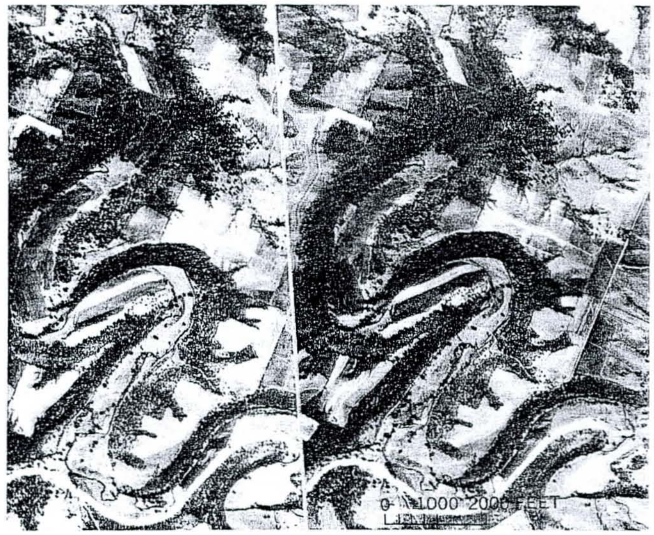

FIGURE 3.—Stereoscopic photograph of part of Bois Creek, Grant County, Wis., showing incised valley meanders, flood plain, and stream meanders. Bois Creek is an underfit stream.

amplification of existing data on the wavelength and discharge relation, which is crucial to the whole discussion, and considerably revises and extends views previously expressed on the former discharge of rivers that are now underfit.

ACKNOWLEDGMENTS

Everyone who has savored the essays of W. M. Davis will appreciate the debt of stimulation owed to that author. This debt, first incurred by the present writer in 1935, should perhaps be all the more freely admitted, for the following text includes reexamination of some of Davis' evidence, criticism of some of his hypotheses, and firm rejection of some of his conclusions.

Unreserved gratitude is due to members of the U.S. Geological Survey and of State geological surveys, to members of universities, and to private individuals for a whole range of help. The work of 1960–61 was done under the authority of Thomas B. Nolan, Director, U.S. Geological Survey; it was supervised by Luna B. Leopold, Chief Hydrologist, U.S. Geological Survey, whose encouragement, counsel, and guidance were invaluable.

Members of the U.S. Geological Survey who assisted with fieldwork and with protracted discussion in the field include John H. Adamson, Friedrich Brandtner, Charles F. Berkstresser, Jr., Keith S. Brigham, Denzel R. Cline, Theron R. Dosch, Fred C. Dreher, Donald W. Ericson, John T. Hack, R. F. Hadley, Joseph H. Hartshorn, Charles L. H. Holt, Jr., Anthony E. Homyk, Jr., Burns L. Kaupanger, Charles E. Knox, Doyle B. Knowles, Lawrence E. Lopp, Lawrence E. Martens, Robert W. Myrick, James G. Rickher, Roy W. Ryling, Francis T. Schaefer, John P. Schafer, S. A. Schumm, Daryl B. Simons, William K. Summers, Joseph E. Upson 2d, and Milton T. Wilson. Assistance, guidance, and lengthy discussion in the field were provided also by Charles L. R. Holt 3d, Madison, Wis.; Wallace B. Howe, Missouri Geological Survey; George F. Hanson and Meredith E. Ostrom, Wisconsin Geological and Natural History Survey; Arthur E. Moodie and William E. Powers, Northwestern University; David Harris, University of Colorado; Neil E. Salisbury, Iowa State University; and H. T. U. Smith, University of Massachusetts.

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To all those named, and to many others in addition, the writer is deeply grateful for unstinted help, stimulus, encouragement, friendliness, and—by no means seldom—generous hospitality.

PERSPECTIVE ON TERMINOLOGY

This investigation touches on many points of doubt and controversy. Because the landforms in question can be discussed in more than one geomorphic context, a wide range of suggested interpretations is not prising. Writers have varied in their approach to is here the central problem, according to their env ment, experience, interests, and intentions. sequently the study of meandering valleys interloc a highly complex fashion with cognate topics, an tention can be limited to a desired range of mat only with great difficulty. Particularly is this s the technical terms available for use frequently (an unwanted charge of genetic significance, whethe cause they are by nature definitive, or because they become definitive through association. Attempte implied classification, where classification is impos or purposeless, increase the confusion of language. rors of fact, resulting from hasty observation or efforts to make observation fit hypothesis, embed t selves in the literature and react on terminology.

In the broadest sense, where it implies nothing a the size of a contained stream, the term "meande valley" is already definitive. It connotes systen qualities that are expressed in alternate steep cresc slopes on the outside curves and gentler lobate sl on the opposing spurs. Even if alternate steep and tle slopes are not obvious, roughly similar dimens throughout a train of bends can indicate systen qualities. As the necessary arrays of landforms recognizable on sight in many places, the term is m ingful; but the certain presence of meanders in s valleys does not imply that their absence elsewhere invariably be proved.

Three continuous series probably connect the t extremes of straight valleys, meandering valleys, valleys that are not straight but yet do not mean No convenient term exists for the third group bec the term "winding valley" is preempted for val other than straight valleys, whether they meande whether they are merely irregular in plan. Simila "valley meander" is definitive and "valley bend" me indicative, although these terms are often interchar for the sake of euphony. Many winding valleys in actuality, regular enough to be classed as mean ing. Obscure instances may be illuminated by m urement—that is, by the comparison of dimension meanders with length of stream or with the drair basin. Systematic variation is strong presumptive dence of meandering as opposed to a merely irreg habit. But if continuous transition is possible f meandering valleys to straight valleys and to val which wind but do not meander, then criteria whether or not a doubtful case is or is not a mean ing valley become wholly arbitrary. The meander habit remains subject to demonstration but, in a w zone of uncertainty, not to disproof.

Words connoting the origin, form, and mode of development of meandering valleys vary in import according to the views of particular writers on tectonic and cyclic history and on the necessity for incised meanders to be inherited from a flood plain. The choice of available words is liable to vary with powers of observation. Two main distinctions, however, are usual: that between free meanders and incised meanders, and, among incised meanders, that between intrenched and ingrown forms. Free meanders are meanders on a flood plain. Incised meanders have cut down, so that projecting spurs—commonly, spurs of bedrock—obstruct their downstream sweep. It is, of course, possible for the spurs to be cut away and a continuous flood plain, wide enough to permit unobstructed sweep, to be formed; the meanders then pass from the incised to the free condition. As will be shown, however, the development of numerous trains of incised meanders has been arrested. Intrenched meanders are contained by walls that differ little or none in slope on the two sides of the valley, whereas ingrown meanders require steep slopes on the outside curves and gentle slopes on the inside curves. Intrenched meanders, by definition, result from vertical downcutting, whereas ingrown meanders result from lateral movement during incision. However, the terms cannot be allowed to retain the implication attached to them by Rich (1914), who claimed that the difference between ingrowth and intrenchment corresponds to a difference in rate of uplift. Indeed, insofar as uplift means differential tectonic movement, its relevance to the present inquiry is denied. Intrenched meanders command little attention in the literature, perhaps because erosion tends to obscure their nature, but more probably because they are rare. Field observation convinces the writer that incised meanders are normally ingrown.

The manner in which very simple descriptions of meandering valleys become entangled with cyclic history, hypotheses of planation and inheritance, and with varying nomenclature is well illustrated by one of the regions treated in this essay—the northern Ozarks. Davis (1893), discussing the valley meanders of the Osage River, postulated inheritance of the meander train from a planed-down surface and stated that the slope of the Osage in a former cycle "* * * had become very gentle, and [the river's] current had taken to a deviating path, peculiar to old streams, which so generally meander on their flat flood plains." This erroneous claim to relate a meandering habit to some condition of slope seems to be one of the earliest, equally erroneous, efforts to associate meandering with some stage in the Davisian cycle. Davis was challenged in an exchange of correspondence by Winslow (1894), who maintained that meanders can develop during incision; ironically enough, Davis himself was later criticized for incorporating simultaneous meander growth and incision in his own scheme of river development (Lehmann, 1915; Flohn, 1935; Hol, 1938, 1939). The debate concerning the Osage River resolved itself into questions of local planation and uplift. Whereas Davis undoubtedly was justified in positing one or more episodes of planation—whatever may be thought of the peneplain concept—Winslow was equally correct in his general claim that meanders can form, and grow, during the incision of an originally straight river—that is, that incised meanders are not necessarily inherited from a flood plain. Tarr (1924), writing of the Ozark rivers Gasconade and Meramec, proposed to substitute "incised" for the "ingrown" of Rich; fortunately, Tarr has not influenced usage. In opposition to Winslow, Tarr called the Gasconade and the Meramec typically intrenched. In actuality they are manifestly ingrown, as can be seen readily on the ground, from the air, and on all relevant maps so far published on the scales of 1:62,500 and 1:24,000. Already, therefore, in advance of reference to lithology and structure or to meanders of streams as distinct from meanders of valleys, errors in observation, discord in nomenclature, and complexly ramifying views on the development of meanders, valleys, and landscape have all been exemplified.

On streams, as in valleys, meander trains can often be recognized on sight. Continuous ranges of intermediate pattern, however, seem to link boldly meandering channels with braided channels on the one hand and with straight or slightly irregular single channels on the other. An equally continuous transition seems to lead from unique straight channels to highly anastomosed braided channels. Nothing said here is meant to deny that significant changes in behavior accompany changes in channel pattern, that certain patterns represent steady states (Leopold and Wolman, 1957), or that conversion from one pattern to another can be rapid. The point at issue is that any attempt to define meandering habit in terms of sinuosity must rely on arbitrary criteria. It may be possible to affirm, but not to deny, that a stream is a meandering stream.

When the terms "meandering valley" and "meandering stream" are used contradistinctively, complications multiply. Contradistinction is required by the observed fact that some meandering valleys contain flood plains, whereon the rivers trace meanders far less ample than those of the valleys. The term "meandering valley" now acquires additional significance: it implies that the valley meanders are homologues of stream

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meanders, but that conditions have greatly changed since they were cut.

Examples of the identity of the surface forms of meandering valleys with those associated with ingrown meandering streams are provided by Davis (1896, 1899, 1906). Reasoning from this identity and from the general circumstance that size of meanders varies with size of stream, Davis postulated a reduction in volume to explain the inferred reduction in size of meanders. He estimated size mainly by radius of curvature, which is by no means the most suitable property, and his statements about volume are so cast as to be meaningless. Furthermore, he tried on at least one occasion to hold accidental obstructions responsible for small meanders (Davis, 1913, p. 14). Nevertheless, he was correct in perceiving a disparity between the meanders of certain valleys and those of the contained streams, and justified in applying the term "underfit" to rivers that are too small for the valleys in which they flow. But if streams can be too small for their valleys, they can also be too large, hence the term "overfit." The term "misfit" includes both the overfit and the underfit classes. Overfit streams are rather difficult to imagine and would probably be unrecognizable by the criterion of radius of curvature which Johnson (1932) sought to apply. Small meander scars cut by a small ancestral river could scarcely remain long intact before the onset of large meanders developing in the stream. Observed results of dam bursts suggest that sudden natural increases in discharge—due, for example, to river capture or to the overspill of proglacial lakes—are likely to cause rapid enlargement of channels; although in some contexts the size of channel must be distinguished from the size of valley, it is improbable that overfit streams would remain identifiable for long. Streams recognized as misfit are so usually underfit that the two names are frequently interchanged.

Davis' exclusive reference to underfit streams that now meander may be responsible for Johnson's attempt to define the misfit condition in terms of meander size. In any event, Davis' examples are so well known and have been so repeatedly presented or matched that the word "underfit" has come by association to imply a meandering trace, both of valley and of stream. As a wide range of channel pattern occurs in nature, however, it seems possible that the changes which make streams underfit need not invariably preserve the meandering habit. This amounts to saying that every stream in a meandering valley is not a meandering stream, a proposition which, far less paradoxical than it may sound, will be substantiated. If it can be accepted for the time being, pending substantiation later, then the sense of meandering valley must be extended to valleys with meanders too large for the pre streams, irrespective of present channel pattern, wh as underfit must be extensible to present streams, s of which fail to meander.

Criteria of size are now urgently needed. Siz meanders is best expressed as wavelength. W length of meanders is known to bear a close statist relation to bed width, which in turn is causally rel to discharge (Leopold and Wolman, 1957). La existing evidence for a close empirical connection tween wavelength and discharge will be amplified siderably. Whereas amplitude of meander belt radius of curvature of meanders may be signific when measured on flood plains, they change du ingrowth. Ingrowing loops commonly increase b their amplitude and their radius to the limit of cu In special circumstances, radius can stay sensibly stant, but amplitude then greatly increases as ingro continues (Strahler, 1946). Wavelength, by contr is suddenly fixed when incision starts, and its aver value for meander trains is not greatly affected distortions of single loops. Only where loops are c pletely obliterated by cutoff—an infrequent hap ing—does apparent wavelength change significantl

Concentration on radius and amplitude, in prefere to wavelength, has allowed confusion to enter dis sions of structural, lithologic, and tectonic influence the dimensions and forms of valley meanders. reconcilable views expressed by Vacher (1909), M set (1928), Cole (1930), Masuch (1935), Flohn (19 Blache (1939, 1940), and Wright (1942) probably a fair sample of the work of their period. These vi were previously reviewed (Dury, 1954, p. 196–197) will not be dealt with here. It should be made cl however, that accommodation of ingrowing mean to structures and qualities of bedrock is not denied.

Where a meandering stream is contained in a m amply meandering valley, the disparity between two sets of wavelengths is determinable by a plot eit of one series against the other or of both aga drainage area. Where the valley has meanders but stream can be shown or suspected not to have meand properties additional to wavelength must be used. ditional properties are always required to test hypothesis that underfit meandering streams have b reduced in volume and that nonmeandering unde streams can exist in nature. Size of channel can expressed by a range of hydraulic dimensions, of wh bed width is usually the most available. In one se size of stream is identical with size of channel; another sense, it is expressed as discharge, in quan per unit time. Neither dimensions of channel nor charge, however, possess meaning unless they eferred to a particular stage, or frequency, of flow. The appropriate stage seems to be that of discharge at bankfull; alternatively, perhaps, a rather small range of stages in which the bankfull stage is included. The reduction in volume required to make a stream underfit must therefore be taken to mean reduction in volume at bankfull. Reductions in dimensions of channel similarly relate to bankfull conditions.

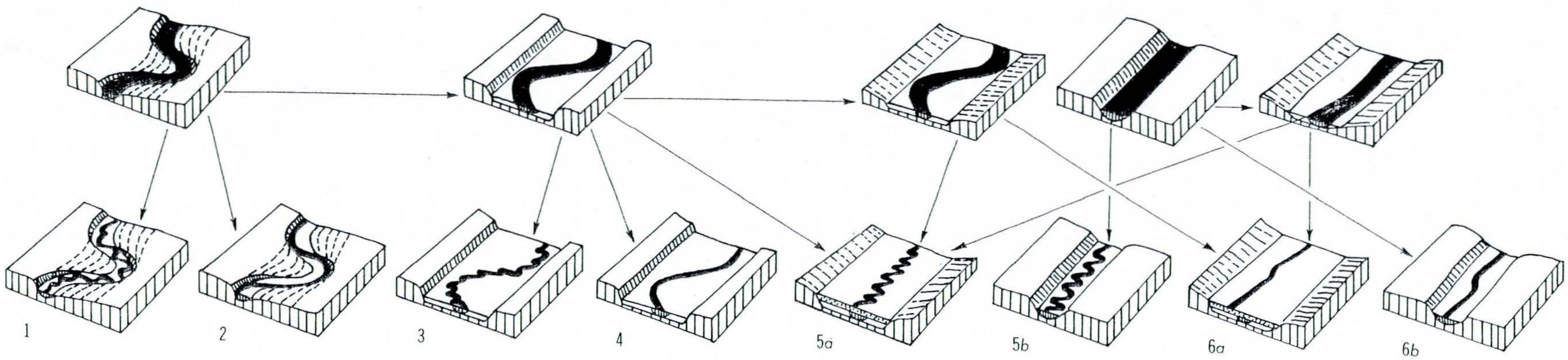
The above-mentioned interrelations of discharge, bed width, and wavelength relate specifically to the bankfull stage. They enable appropriate wavelengths to be calculated for given values of discharge and bed width and make it possible to demonstrate that certain trains of incised meanders are too large for the present streams. Conversely, that is, some existing streams are inappropriately small for the incised meanders which they occupy. In this way, the claim that a stream in a meandering valley need not be a meandering stream, and that underfit streams need not invariably meander, gains support. Confirmation will be provided later, with the aid of data on channel form and of free meanders in reaches of open valley.

The assumption that ingrown valley meanders were cut by streams of the present size—that is, that the present streams are not underfit—occurs in the work of Jefferson (1902) and Bates (1939). These writers claim that the ratio between width of meander belt and width of stream is normally greater where the rivers are incised than where meanders are free. A corresponding disparity of wavelength follows. But the general theory of meandering valleys presented in the present essay, combined with the hypothesis that underfit streams do not always meander, suggests that Jefferson and Bates were measuring wavelengths not on incised streams but on incised valleys. The relevant sites need investigation in the field.

From the inference that valley meanders were cut by former large rivers, the possibility follows that these rivers may, in places, have cut large meander troughs. If so, not all reaches of valley which contain underfit streams need themselves meander. Underfit streams in reaches of nonmeandering valley do in fact occur in nature. A leading example is the upper Evenlode in the Cotswold Hills of England (Dury, 1958; 1960, fig. 1). On the Evenlode, the large meanders had passed from the incised to the free condition before the stream became underfit; the trace of the meanders is, however, preserved, with the present meanders superimposed on it. Because the large meanders were free, they cannot properly be called valley meanders; some broader term is required. As the present essay will justify separation of the two series of meanders not only by wavelength but also by age, the contrasting terms "former meanders" and "present meanders" eventually will be adopted.

As noted previously, a meandering habit, which is expressed in pattern of channel, may be possible to affirm but not to deny. It will be shown below that a meandering tendency, recorded in the form of the bed, may operate even though it is not reflected in the channel pattern. A meandering tendency cannot be denied unless bed form is investigated and may not be open to disproof even then. Nevertheless, it remains true that some reaches of some natural single channels manifestly fail to display a meandering habit. But if nonmeandering reaches be conceded to present streams, general reasoning suggests that streams responsible for valley meanders also may have failed to meander in some places. If so, a meandering valley can in practice include reaches devoid of meanders not because intervening spurs have been destroyed but because meanders were never present to begin with. Accordingly, the possible meaning of the term "underfit stream" must be extended still further to cover at least six combinations of form of valley with trace of channel (fig. 4).

The first of the numbered combinations refers to underfit streams as described by Davis and as usually imagined. It is recognizable by surface form, as is combination 3 which has just been illustrated by the Evenlode. Combination 2, presented above as logically possible, will be exemplified from the Ozarks, from Iowa, and from the Great Basin. Comparison of size of channel at bankfull, or of rate of bankfull discharge, on the one hand to wavelength of meanders on the other has already been offered as a means of identification. In practice, observations of channel form also have been employed. These observations could be supplemented by exploration of the subsurface, as undertaken at sites representing combinations 1 and 3. Combination 4, although possible, is rare; no certain instances have yet been detected. Applicable tests would be identical with those for combination 2. Combinations 5 and 6 are required both by logic and by the general theory of meandering valleys as here presented; however, they cannot be detected by surface form nor by quantitative treatment of comparative wavelengths, for, by hypothesis, large meanders either failed to develop or have been completely destroyed. Subsurface exploration could imaginably reveal former stream channels large enough to show that existing streams are underfit, but no relevant or possibly relevant sites have yet been investigated. This study is confined to examples of the first three combinations. Nevertheless, the logically possible range of combinations shown in figure 4 requires that the possible significance of the term "underfit stream" be extended far beyond its usual limits.



<table>
<tr><td colspan="2" rowspan="3"></td><td colspan="3">Pattern of valley</td></tr>
<tr><td rowspan="2">Meandering</td><td colspan="2">Nonmeandering</td></tr>
<tr><td>Trace of large free meanders preserved</td><td>Trace of large meanders not preserved or never present</td></tr>
<tr><td rowspan="4">Pattern of stream channel</td><td rowspan="2">Meandering</td><td>1. Meandering stream in meandering valley</td><td>3. Meandering stream in former meander trough; large meanders identifiable on surface</td><td rowspan="2">5. Meandering stream in (a) open valley or (b) approximately straight narrow valley</td></tr>
<tr><td colspan="2">Two series of meanders combined</td></tr>
<tr><td rowspan="2">Nonmeandering</td><td>2. Nonmeandering stream in meandering valley</td><td>4. Apparently very large meanders on floor of open trough</td><td rowspan="2">6. Nonmeandering stream in (a) open valley or (b) approximately straight narrow valley</td></tr>
<tr><td colspan="2">Only one series of meanders present</td></tr>
</table>

FIGURE 4.—Block diagrams of underfit streams, showing character and possible origin of some combinations of stream-channel and valley patterns.

PRINCIPLES OF UNDERFIT STREAMS

DIVERSION UNNECESSARY

At the same time that the sense of underfit is thus widened to admit a greater range of landform combinations than is usually envisaged, it should be narrowed in another direction. Stream capture or other types of diversion should not be regarded as a necessary, or even usual, cause of underfitness. Among writers in English, the idea that underfit streams owe their condition to beheading is due to W. M. Davis, whose comments on the matter have been so influential that, with some writers, "underfit stream" and "beheaded stream" are almost synonymous. Hypotheses involving diversions other than capture—such as derangements of drainage associated with glacial advance—are, in effect, extensions of the capture hypothesis. This hypothesis will, therefore, be challenged first. The facts of capture and glacial derangement and the changes in stream volume which they produce are freely admitted. Indeed, examples of beheaded streams and of valleys which at one time were spillways will be presented. But the tacit implication, often tacitly accepted, that underfit streams generally result from diversion of some kind is firmly rejected.

Davis based his reasoning about underfit streams on examples that represent the first of the numbered combinations in figure 4—that is, on meandering streams in more amply meandering valleys. To obviate circumlocution, underfit streams in this combination will be called manifestly underfit. So long as attention was confined to such rivers and so long as capture was regarded as the sole, or at least main, cause of underfitness, the regional distribution of underfit streams could be denied (Davis, 1913; Baulig, 1948). To counter the general hypothesis of capture, it suffices, therefore, to demonstrate that all the streams in a given region are manifestly underfit, or, alternatively, to prove that they represent some other of the combinations in figure 4. In nature, irregularities in form of valley and in pattern of channel make it unlikely that all reaches of all streams in the region will be manifestly underfit, but if the relevant landforms can be identified widely and on numerous reaches of competing streams, then underfitness at once becomes a regional problem. Capture, however, is not finally disposed of until a single degree of underfitness is proved for the whole region. Proof that a stream actually is underfit, although not manifestly so, may not be possible unless data are available on discharge, dimensions, bed form of channel, and the subsurface. Nevertheless, if manifest underfitness be accepted as resulting from a change in volume, then abrupt departures from manifest underfitness suffice to show that the change is not always manifestly expressed at the surface. Special conditions can be imagined for limestone country, in which the volume of a particular stream abruptly decreases at one point and as abruptly increases at another point farther downstream; but limestone hydrology generally is not reliable. In practice, abrupt downstream changes from or to manifest underfitness do occur, and on rocks other than limestone. This fact greatly weakens the statement that underfit streams are not developed regionally. Merely to offer regional examples, however—even those of manifestly underfit streams—would leave intact the claims of Davis that certain rivers are underfit because of capture. Such examples will therefore be deferred until the growth of the Davisian thesis has been traced and until his evidence is reexamined.

When he applied his concept of sequential erosion to the English Plain, Davis (1895) was much concerned with the growth of subsequent streams and with piracy. The apparently shrunken condition of certain rivers appeared to him a natural result of capture. Thus, dealing with the apparently contrasted habits of the Seine and the Moselle on the one hand and of the Meuse on the other, he (Davis, 1896), ascribed the so-called staggering trace of the Meuse to capture by the Seine and the Moselle, whose meanders he styled as vigorous and robust. Three years later, he claimed to identify in the Swabian Alp and in the Cotswold Hills of England consequent streams which, having been beheaded, are underfit (Davis, 1899). He named the Schmeie and the Lauchert as having lost territory—and discharge—to the Eyach [2] and the Starzel, and the Stratford Avon as having gained at the expense of the Cotswold streams Cherwell, Coln, Windrush, and Evenlode. All underfit streams were recognized by the disparity between valley meanders and meanders of the stream.

All three regions produced anomalies. In his paper of 1896, Davis noted that the deprived Meuse is out of proportion to its valley not only downstream from the point where a main headstream was lost to the Moselle but also upstream. His second French example, involving the Bar, which flows to the Meuse, and the Aisne, which belongs to the Seine system, relied on the inferred capture by the Aisne of the former head of the Bar—that is, the river now called the Aire. As he recorded at the time, the present meanders of the supposedly diverted Aire are much smaller than the valley meanders of the Bar, which are farther along the reconstructed course of the Aire-Bar as it existed before capture. Davis offered the guess that the Aire had once received the water of the Ornain (fig. 11) and that the

[2] The spellings used are those on the German 1 : 25,000 map; Davis gives Schmeicha and Eilach.

upper Meuse had experienced additional, but undetected, captures.

In his 1899 paper, Davis referred in passing to an underfit but apparently not beheaded stream in Swabia and expressed surprise that the Stratford Avon and its feeder, the Stour, which should have gained what the Cotswold streams lost, are themselves misfit (underfit). He added to the recorded anomalies in France the observation that the Aisne, a supposedly captor stream, is discordant with its valley—admittedly, above the confluence of the Aire—so that the increment of water captured from the restored Bar does not come in question. Optimistically, perhaps, he suggested that the anomalous stream in Swabia might possess a wind gap which the available maps failed to show; but, taking a more general view, he concluded that a general change in volume had superimposed its effects on those of capture. This change he thought might result either from deforestation or from some climatic change of external and obscure origin. But to retain the hypothesis of capture, he found it necessary to claim that streams of the Avon system are less underfit than their competitors on the Cotswold back-slope.

A later suggestion (Davis, 1909) that ice-dammed lakes might formerly have discharged into the Cotswold valleys will be disposed of presently. Davis himself appeared to abandon the hypothesis of overspill when, a few years later (1913), he suggested that water could be lost to underflow through alluvium. Alluviation and underflow he saw as a normal consequence of cyclic development, saying that progressive grading of hillside slopes, continuing after rivers were already graded, would ensure increasing delivery of rock waste to streams and thus cause aggradation. This is not the place to discuss the Davisian concept of grade. Suffice it to say that the postulate of automatic aggradation by mature streams has not been seriously considered by Davis' followers and that underflow cannot possibly account for the losses of water which have occurred, even when no allowance is made for the shallowness of alluvium in some valleys and for its impermeability in others. To advocate underflow as a possible cause of underfitness is, however, to concede by implication that all the streams in a given region can be simultaneously underfit—a possibility which Davis rejected in the same paper of 1913.

Finally, Davis (1923) reverted to the hypothesis of capture, naming underflow and climatic change as possible further mechanisms but advancing no fresh evidence. He wrote,

> The reduced volume of a beheaded stream cannot develop meanders of the same size as those which it followed, with larger volume before its beheading * * * hence, as the reduced meanders are too small to fit their valley curves, they may be called underfit * * *. It is believed that underfit may also result from climatic change, as well as from loss of surface water to underflow in the flood-plain d of mature valleys.

When the sites named by Davis are reexamined, of the evidence on which he based the capture hy] sis immediately disappears. The following crit are largely independent of any local studies o] morphic history made since Davis wrote, althou; will appear, these studies are themselves adve] some part or other of the Davisian thesis. Davis' that the members of the Avon system are less un than the streams of the Cotswold back-slope c stand. In figure 5, concurrent readings of wavel are plotted for valley and stream meanders. As ı ders of neither series increase progressively in ' length downstream, a plot in this form is less ı in some contexts than a plot of wavelengths ag drainage areas, but the comparison made here is cisely that attempted by Davis—namely, a com son of size of meanders. Values plotted are ave for trains. While the samples are small, the} thought capable of showing that, in fact, the stı of the Avon system are no less underfit than tho the Cotswold back-slope. At once, therefore, the

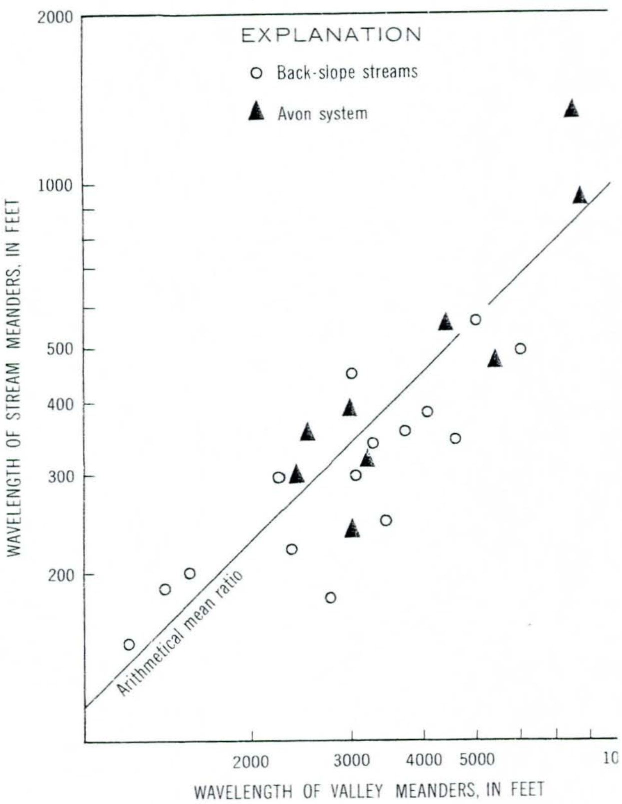


FIGURE 5.—Graph showing comparative wavelengths of valley stream meanders of members of the Warwickshire Avon system of streams of the Cotswold back-slope, England.

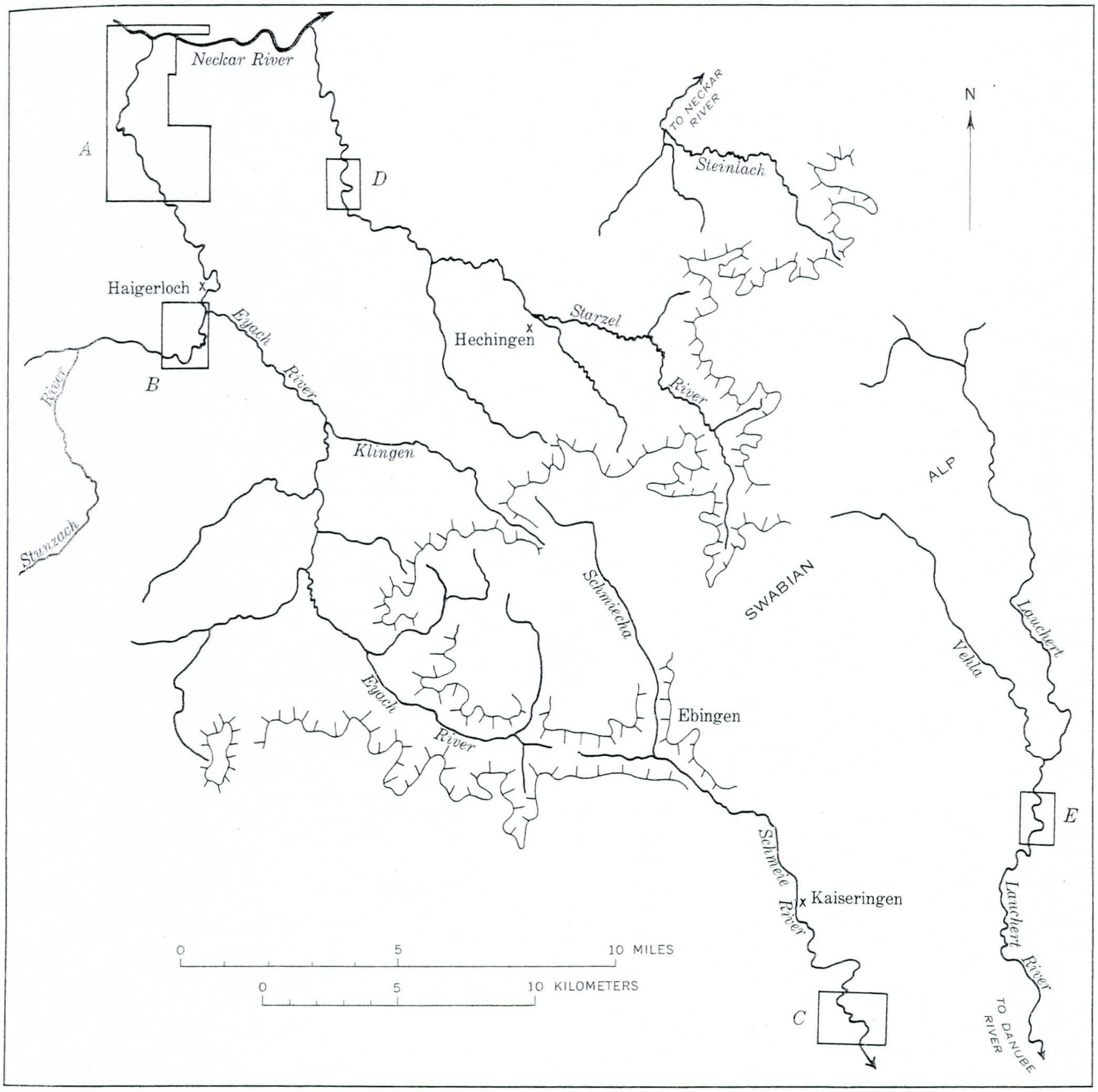


FIGURE 6.—Location map of streams in Swabia, south Germany. Outlined areas are shown in figures 7 and 8.

to appeal to capture of the Cotswold streams is removed. Dealing specifically with the Cotswold river Coln, Davis perceived signs of capture in the wind gap at each of the two heads. He maintained that the valley of the Coln displays three sets of meanders—large valley meanders, lesser scars cut into the valley walls, and the small meanders of the present stream (Davis, 1899, fig. 16). These features he supposed to have been produced, in order, by the ancestral Coln before beheading, by a river reduced by the loss of one headstream, and by the existing stream after the second feeder had also been lost. However, the alleged lesser scars do not exist (Dury, 1953a), and the main support vanishes from the postulate of successive capture. The two wind gaps at the heads of the Coln remain; but, whether or not they indicate capture, they provide no help in explaining the underfit condition of the river. The Coln is no more underfit than is the competing Avon, or than are its Cotswold neighbors. In addition to the Cherwell, Windrush, and Evenlode, the

Cotswold rivers Dorn, Glyme, Dikler, Leach, and Churn also are underfit. Some of them head either by an unbroken crest where no wind gaps can be thought to indicate capture, or midway down the smooth notch-free back-slope, where signs of capture are equally lacking. Furthermore, a single degree of underfitness is common to the whole region (Dury, 1958, fig. 8) and, as previously shown, both to the Cotswolds and to the Avon basin. None of the Cotswold evidence, therefore, sustains any part of the capture hypothesis.

The Swabian Alp is no more helpful. In point of fact, the main headstream of the Lauchert is opposed not by the Starzel but by the Steinlach, which flows to the Neckar independently of the Starzel (fig. 6, shown on page A11).

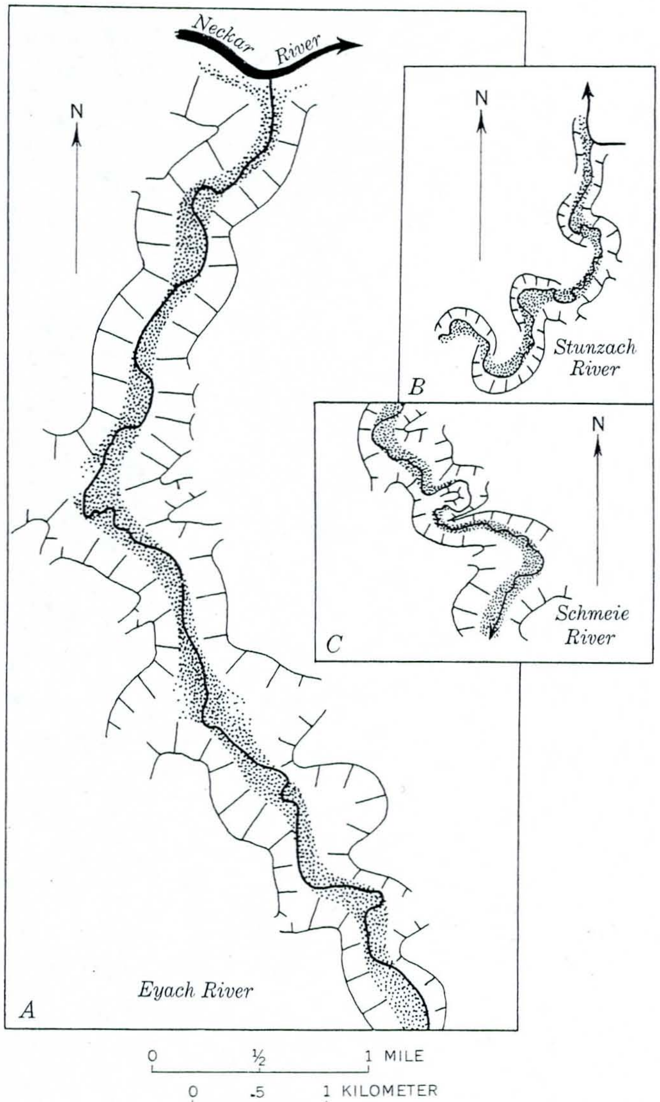


FIGURE 7.—Sketches of the Eyach, Stunzach, and Schmeie Rivers showing comparative stream-channel and valley patterns. See figure 6 for location of areas.

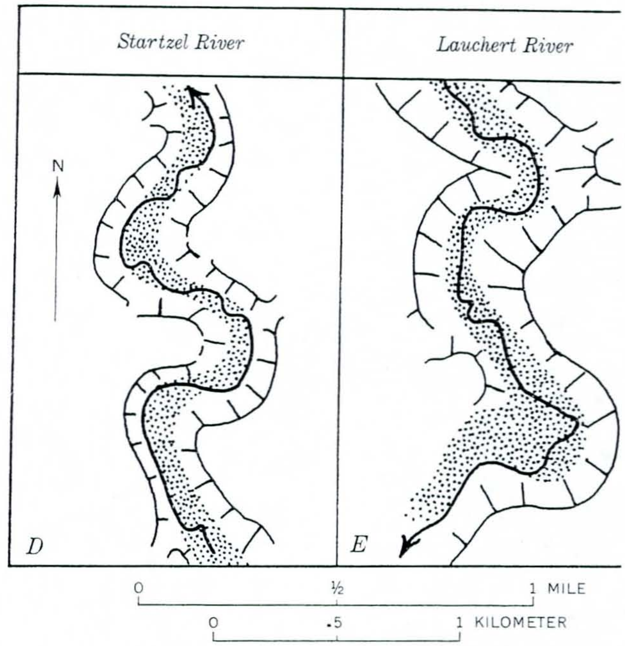


FIGURE 8.—Sketches of the Lauchert and Startzel Rivers showing parative stream-channel and valley patterns. See figure 6 location of areas.

The Starzel competes with the Vehla, which is a la stream but not the chief headstream of the Lauch The Schmiecha is the principal feeder of the Schn above Ebingen and contests territory not with the tr Eyach but with the tributary Klingen. Davis' w known block diagram can scarcely represent anyth but the trunk Schmeie, as the settlement Kaiserin is named in the caption. In all likelihood, the diagı is based on that reach of the Schmeie which is outli as C in figure 6. The attempted comparison with Eyach may therefore stand, as the general claim t back-slope feeders of the Danube have lost grounc the Neckar system is not affected by revisions of ident The view that capture along the crestal divide is sponsible for the observed underfitness is, howe false. Not only the Lauchert and the Schmeie but a the Eyach and the Starzel are underfit. Unde streams are indeed typical of the region (Dury, 196

There is obviously little to choose between the cor tion of the Lauchert and the Schmeie on the one ha and of the Starzel and the Eyach on the other (figs 8). Stream meanders are admittedly few on the tr Eyach, but they do occur; the tributary Stunzach far more obviously underfit than is the Schmeie in reach which Davis probably sketched. The Lauch seems no more underfit than the Starzel. Althou these statements are qualitative, regional analysis meander wavelengths does not seem necessary; evide of the kind used by Davis himself is enough to conf his views. Changes affecting the feeders of the Danı ıave also affected those of the Neckar. For Swabia, as :or the Cotswolds, the claim that back-slope streams ıave become underfit through capture is not supported ɔy fact.

Only the Meuse and the Bar now remain as possible ɪxamples of the effect of capture on stream volume. Let t be conceded at once that the upper Moselle has been liverted from the Meuse and that the Bar has lost the \ire. The anomalies of underfit captor streams have ;till to be explained. As will presently be shown, it is ɔossible to define the relative influence of diversion and ɪther change and to demonstrate that diversion was by :ar the less effective.

The term "diversion" is here preferred to the more ;pecific term "capture," because Tricart (1952) has ɔroved the derangement of the Meuse not to have been ;apture in the strict sense. The valley of the present ɪpper Moselle was thickly alluviated during a cold period,[3] when thaw-freeze brought unusually large quantities of rock waste down the hillsides. Rising on its own deposits at the entrance to the Toul gap, the river spilled eastward into the adjoining valley. It became, as it has since remained, the largest member of the Moselle system. To this kind of diversion Tricart applies the name "déversement," for which spilling is probably the best English rendering. Change of mechanism, of course, does not affect the general argument from diversion; diversion, however caused, must influence discharge.

The means by which Davis sought to prove the effects of diversion on discharge are, however, mainly unreliable. His example of so-called robust meanders on the Moselle can be almost duplicated—not omitting a cutoff loop—from the Meuse near Mézières (fig. 9).

[3] During the Saale Glacial of the European sequence, corresponding to the Illinoian of North America. The full significance of the date will appear later.

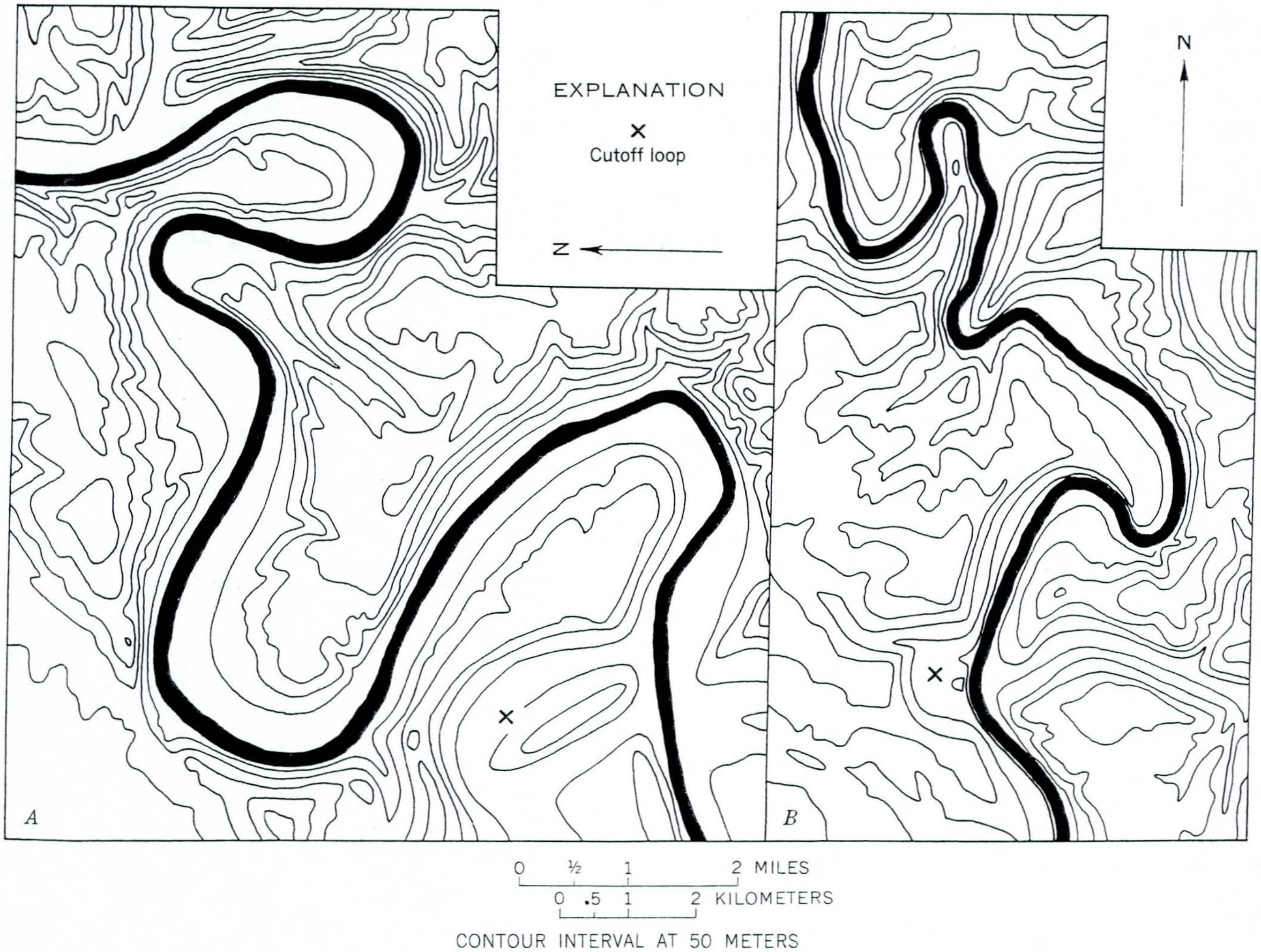


FIGURE 9.—Maps of Moselle and Meuse Rivers showing incised bends. *A*, the Moselle River near Berncastle; *B*, the Meuse River near Mézières.

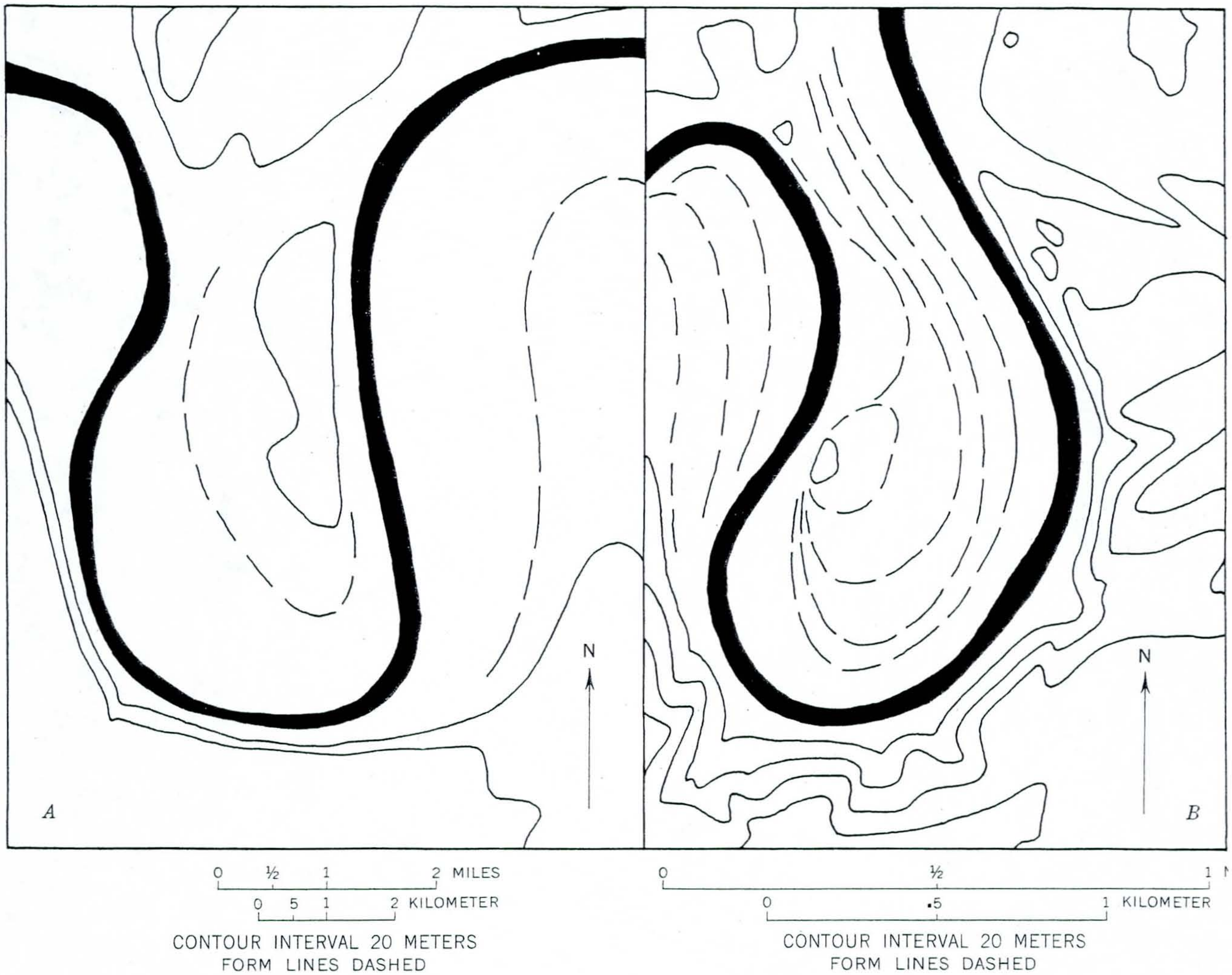


FIGURE 10.—Maps of the Seine and Meuse Rivers showing incised bends. *A*, the Seine River near Duclair; *B*, the Meuse River near Gi

His specimen "vigorous meander" on the Seine differs little from a curve of the Meuse near Givet (fig. 10). In admitting some change of habit for the Meuse near Mézières, Davis contended that the loss of the small Aire was far less serious than the loss of the upper Moselle. This may be so, but the various derangements involve a 60-percent reduction of drainage area for the Meuse opposite the Toul gap and a 40-percent reduction near Mézières. This second reduction—surely great enough to be significant—is, nevertheless, not expressed by manifest underfitness. As has been seen, the Meuse in the relevant stretch of valley is indistinguishable from the so-called robust Moselle. Although admitting that the diversion of the Aire was of no great moment to the Seine, Davis was still able to contend that it confirmed the Seine in its boldly swinging habit. In actuality, the addition of the Aire basin could not have extended the drainage basin of the Seine by more than 1 percent. This value relates to the first point where water delivered by the Aire can enter the Seine—th at the confluence of the Seine and the Oise.] Duclair, at the specimen loop, the percentage gai drainage area would be still less.

Thus far, then, the arguments of Davis fail. though diversion of the upper Moselle from the M is authentically recorded in the Val de l'Asne, the of evidence which Davis used does not validate a sential contrast in habit between the Meuse on the hand and the Seine and the Moselle on the other. logic, therefore, his contingent inferences lose all f He was right in contending that the Meuse has reduced in drainage area by loss of tributaries, bu did not succeed in proving reduction of disch (Dury, 1963).

The Bar produces difficulties of a more complex k That part of the dismembered stream which was versed after capture is now represented by the Mo Briquenay-Agron (fig. 63), which is itself manif nderfit. Although capture can be held partly respon- ible for the underfit state of the remaining Bar, it annot explain that of the Agron, the captor Aisne, the iverted Aire, or the Ornain, by which Davis sought to xtend the Aire headwards. As will presently be shown, he valley meanders of the Agron belong to a series which, widely displayed in this region, bears a quantitative relation to existing drainage areas. The matter of capture does not arise. At any rate, possible capture is not relevant to the Agron. As the Agron did not exist until the Bar had been dismembered or partly reversed, the local sequence runs: diversion of the Aire, reversal of part of the beheaded Bar to form the Agron, incision of valley meanders along the Agron, and,

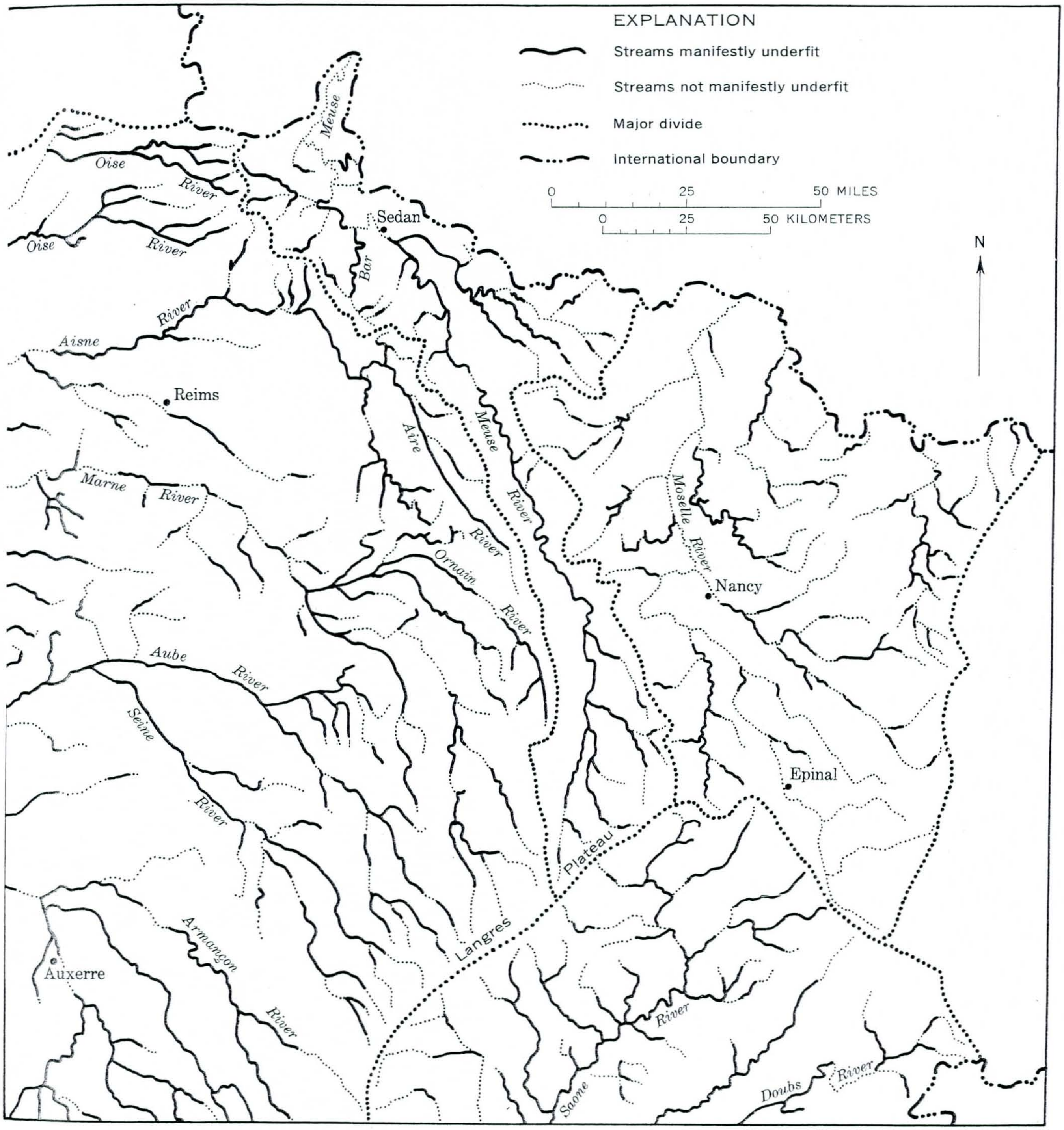


FIGURE 11.—Map of eastern France showing distribution of manifestly underfit streams.

finally, development of stream meanders. The Agron thus became underfit some time after the diversion of the Aire, thus allowing a sufficient length of time for its valley meanders to be incised 100 feet or more. Dating of the change as significantly later than the capture will shortly prove useful.

Despite what Davis wrote to the contrary, underfit streams can be regionally distributed. They are so distributed in eastern France and are well shown on the French 1 : 80,000 map from which he chose illustrations. This map is the basis of figure 11, in which manifestly underfit streams are the only type plotted. As shown, the underfit Meuse is opposed on one side by the Moselle system, parts of which are manifestly underfit, and on the other side by long manifestly underfit reaches of the Aire, Aisne, Ornain, and Marne. The underfit condition of the Saône prevents appeal to capture along the crest of the Langres Plateau. Here, as in the Cotswolds, underfitness is regional. The effects of diversion serve merely to complicate the regional pattern of underfitness. One can but regret that Davis gave first importance to diversion. His statement that some general change has superimposed its effects on those of diversion, which is incompatible in any instance with his views on distribution, is valid in a sense which he did not intend. It holds good only if it is taken to mean that diversion came first in time, and regional change second.

The shrinkage of the Agron has already been seen to postdate the diversion of the Aire. The diversion of the upper Moselle dates back to the Penultimate Glacial, whereas there is reason to place the last major regional

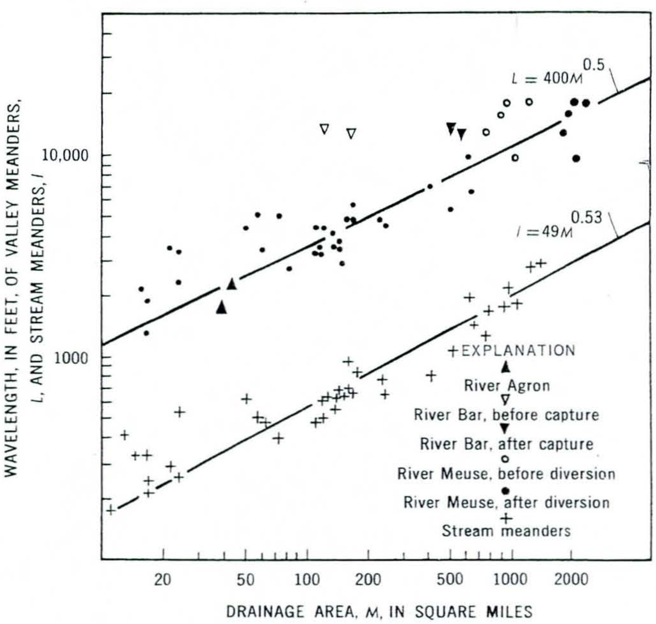


FIGURE 12.—Graph showing relation of wavelength to drainage area of selected French rivers.

shrinkage at the end of the last glacial. Althou dates are yet available from this area, it seems inev that the general change in eastern France was ap mately simultaneous with that elsewhere: dates of 10,000 years B.P. will be applied in due course English Cotswolds, the Great Basin, and Wisc and will be associated with less precise but wholly patible evidence from other regions. Accordingl relative effects of diversion and of regional shri will be calculated according to the view that ever curred in this order.

Regional graphs of wavelength, both for meanders and for stream meanders, have been mined by the usual method of least squares from for average wavelength in trains or groups and areas determined by planimetry. The results a in figure 12, where the effect of diversion on the lo of points on the area scale also is shown. Betw and 1,000 square miles, the ratio of wavelength be valley meanders and stream meanders falls from 7.5 to about 5.5; the disproportion resembles th served in a number of other regions. Eastern F is closely similar to parts of the United States in r to its degree of underfitness.

The regional graphs indicate the wavelengt valley meanders appropriate to the Meuse and th for the basins drained immediately before and im ately after diversion. It does not follow that meanders of the reduced size were actually devel Even if a meandering habit was retained by the M the shortened valley meanders might have been a modated with no great difficulty in the existin The relation among the predicted wavelengths is n theless instructive. Diversion could have re wavelength on the Meuse by about one-third and th the Bar by about one-half; the regional shrinkage have imposed a further reduction of five-sixths o wavelength of both rivers. The fractional loss o Bar was, therefore, 1½ times as great by reg shrinkage as by diversion, whereas the proportion was 3 times as great; corresponding values fo Meuse are 2½ and 4 times. Quite clearly, even diversion involved loss of more than one-half the d age area, it was potentially less effective than reg change. The hypothesis of capture, advanced in eral explanation of underfit streams, shoul discarded.

DIVERSIONS OTHER THAN CAPTURE—THE QUESTION OF SPILLWAYS

Regional distribution of underfit streams, once e lished, is in part as adverse to the hypothesis of gl derangement as to that of capture. However, str in general in extraglacial regions need to be proved as commonly and as markedly underfit as those in formerly glaciated areas before glacial derangement can, like capture, be reduced to a mere complicating factor. Alternatively, if underfit streams in formerly glaciated regions can be proved not to have become underfit until long after the ice had gone, then underfitness can be separated from glaciation in time. Demonstrations of the required sort will be forthcoming. Nevertheless, well-authenticated instances of glacial derangement demand something more than a general denial of the hypothesis as stated by Thornbury (1954, p. 156–157).

Country formerly invaded by ice sheets usually exhibits spillways of various kinds—in particular, melt-water channels that lead along or away from the lines of the ice front, or the outlets of proglacial lakes. Where such channels are now occupied by streams, such streams are underfit, for they are far less voluminous than were the former streams of melt water. As, however, many spillways fail to meander and as many are occupied largely by swamp, they frequently do not show the combination of valley meanders and stream meanders which characterizes manifestly underfit streams. In any event, streams flowing along former spillways represent a special type of underfitness with which the discussion in hand is not primarily concerned. The examples now to be examined have been selected to demonstrate the independence from the outpouring of melt water of those changes which reduce the drainage of an entire region to an underfit condition.

WABASH RIVER, IND., AND GLACIAL LAKE WHITTLESEY

The Wabash valley is well known to have functioned as a major sluiceway for melt water and outwash during part of the Wisconsin Glacial. In particular, it provided an outlet, in order, for the water of highest Lake Maumee (800 ft), possibly for lowest Lake Maumee (760 ft), and certainly for middle Lake Maumee (790 ft) (Hough, 1953, 1958). That is, water overflowed through the Fort Wayne gap in the Fort Wayne–Wabash complex of moraines at various times between the recession of the Erie ice lobe from the Fort Wayne moraine and the recession from the Lake Border moraines, an event dated at about 14,000 years B.P. (Flint, 1957, p. 347). As Thornbury (1958) observed, the present Wabash valley contains numerous abandoned braids above the present valley floor and minor scablands formed on buried uplands of bedrock. But, in common with some other rivers that occupy former outlets of melt water—for example, the upper Mississippi and the lower Wisconsin—the Wabash is not a finely meandering stream, nor does its valley possess well-developed valley meanders. The disparity between former and present discharge is established by evidence of a kind not to be expected from meandering valleys that were not spillways.

Streams that enter the Wabash from both left and right banks do, however, occupy meandering valleys of the usual sort; these streams have been reduced in volume independently of any cessation of overspill. But as they also lie beyond the Fort Wayne moraine, within conceivable range of the former discharge of melt water, attention may suitably be turned to the area within the moraine where streams exist which cannot possibly have carried water from any ice front.

The Maumee River, which drains the floor of glacial Lake Maumee and its successors toward the present Lake Erie, inherits the channel of no spillway. Still more certainly—if additional certainty be possible—the valleys of tributaries to the Maumee, which is well within the Fort Wayne moraine, can by no means have carried melt water. These valleys ramify across till and lake sediments, as they did not exist before the lake bottoms were exposed by receding waters. Nevertheless, numbers of them are manifestly underfit. On early topographic maps, their windings are often so generalized that the typical combination of valley meanders and stream meanders fails to appear, although the dimensions of the recorded windings are themselves great enough to suggest meanders not of streams but of valleys. When aerial photographs are consulted, the patterns of stream channels emerge in full. Figures 13 through 15 contrast the patterns of a sample area, some 15 to 20 miles south of Defiance, Ohio, according to evidence from topographic mapping and mapping from photographs, respectively. Figure 14*A* shows the two sets of meanders as well as the remains of scars cut by, and point bars deposited in, the large meanders despite the irregular development of both sets of meanders and despite also the extensive scalloping effected by meanders of the present stream. Former traces, including cutoffs, are reconstructed in figure 14*B*.

The flat interfluves of the sample area rise to slightly more than 700 feet above sea level and were certainly inundated not only by the three Lakes Maumee but also by glacial Lake Whittlesey (738 ft). Between the stands of middle Lake Maumee and Lake Whittlesey there intervened the stand of Lake Arkona (710–695 ft), but this episode was brief. Consequently, the valley meanders south of Defiance are considered to postdate the recession of the lake from the Whittlesey shore. As wood from beach sediments of Lake Whittlesey has been radiocarbon dated to 12,800±250 years (Barendsen, Deevey, and Gralenski, 1957) and as highest Lake Warren (690 ft) succeeded Lake Whittlesey

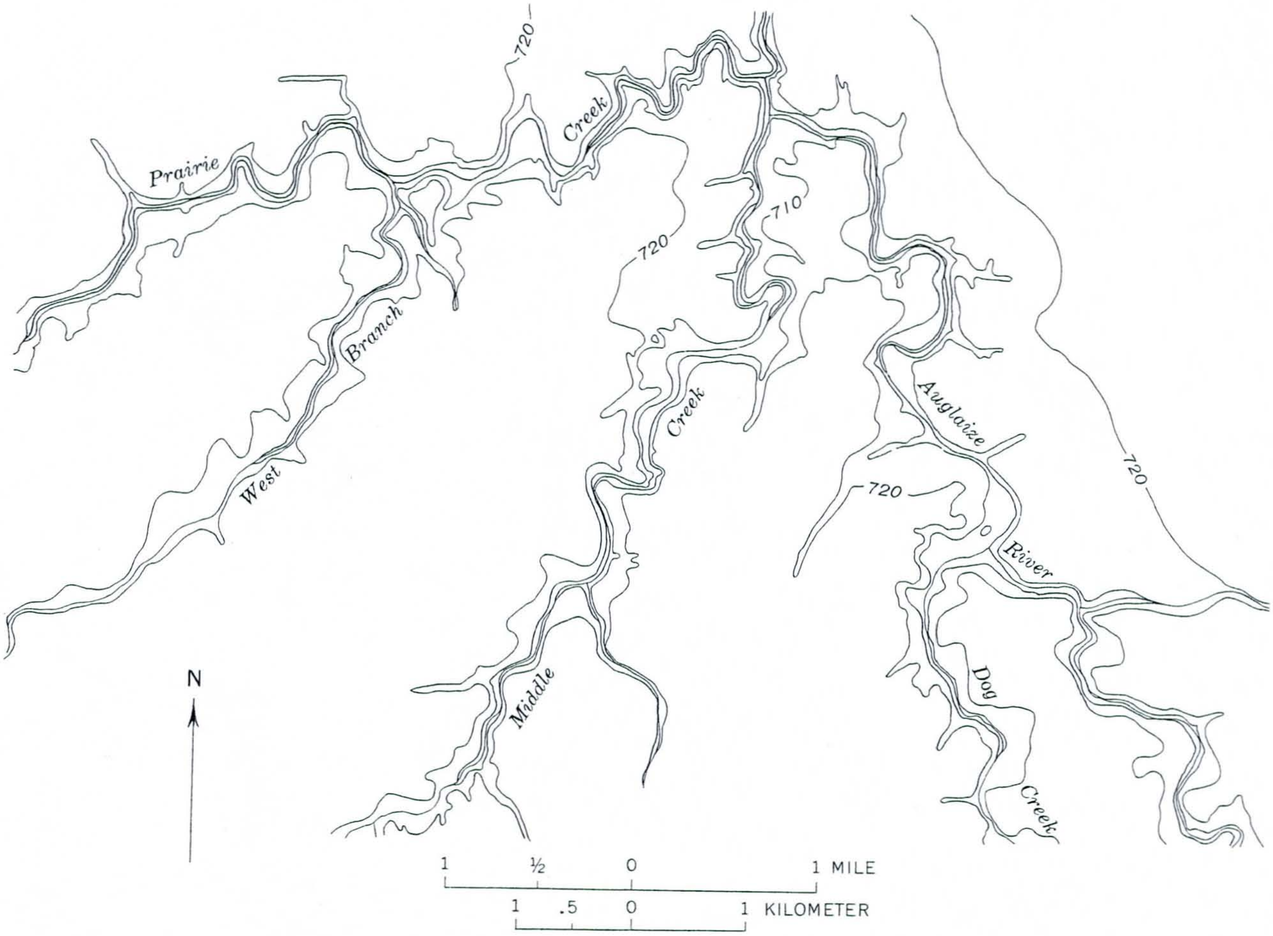


FIGURE 13.—Map of the Auglaize River system, Ohio, showing stream-channel patterns.

slightly before 12,000 years ago, with a renewed clearing of the Grand River link between the Huron and Michigan basins, the earliest date for the cutting of valley meanders into the emergent floor of Lake Whittlesey may be taken, in round figures, as 12,500 years ago, some 1,500 years after spill water had ceased to flow in the Wabash valley.

SOURIS RIVER AT MINOT, N. DAK.

The Souris River at Minot, N. Dak., provides a neat demonstration of the combined effects of the cessation of outspill of melt water and a subsequent independent reduction in channel-forming discharge. Its valley is part of a concentric system of melt-water channels that lies well within the Martin moraine and west of the former glacial Lake Souris. At Minot, where the valley swings first to the right and then to the left, former point bars now form patches of terrace on the insides of bends (fig. 16); the point bars were deposited by the broad, but meandering, stream of melt water. The present stream, which is cutting irregular meanders on the valley floor, is signally misfit; but its pre loops are arranged not in a simple meander belt bı a meander belt which itself meanders. The re structed sequence (which should be read in conjunc with that given below for the Sheyenne) is: Cut of a very large meandering channel by melt water; ting of large meanders, equivalent to valley mean in unglaciated regions, by an ordinary stream; cutting of the present meanders by the reduced str

SHEYENNE RIVER, N. DAK., AND GLACIAL LAKE AGASSIZ

The interpretation placed upon the foregoing ex ple accords precisely with that now to be given for Sheyenne River. On this river, unmistakable s exist that meanders significantly larger than thos today were developed after melt water had cease flow down a great outlet.

The Sheyenne River, which heads at an ill-mar divide that coincides roughly with the Martin mor enclosing the Souris basin, traverses some 175 mile

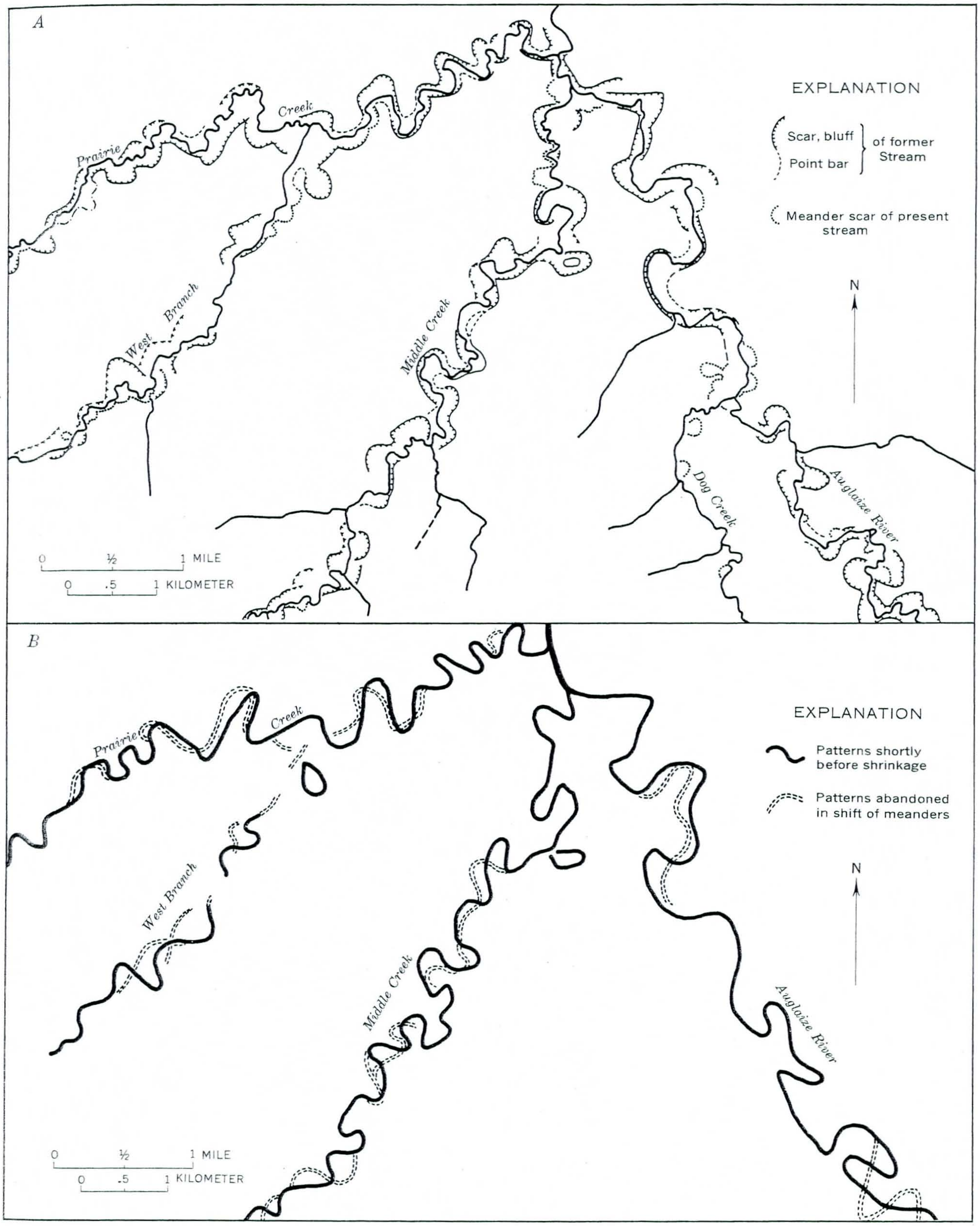


FIGURE 14.—Auglaize River system, Ohio, as mapped from aerial photographs. *A*, stream-channel patterns and lateral features; *B*, restored patterns of former streams.

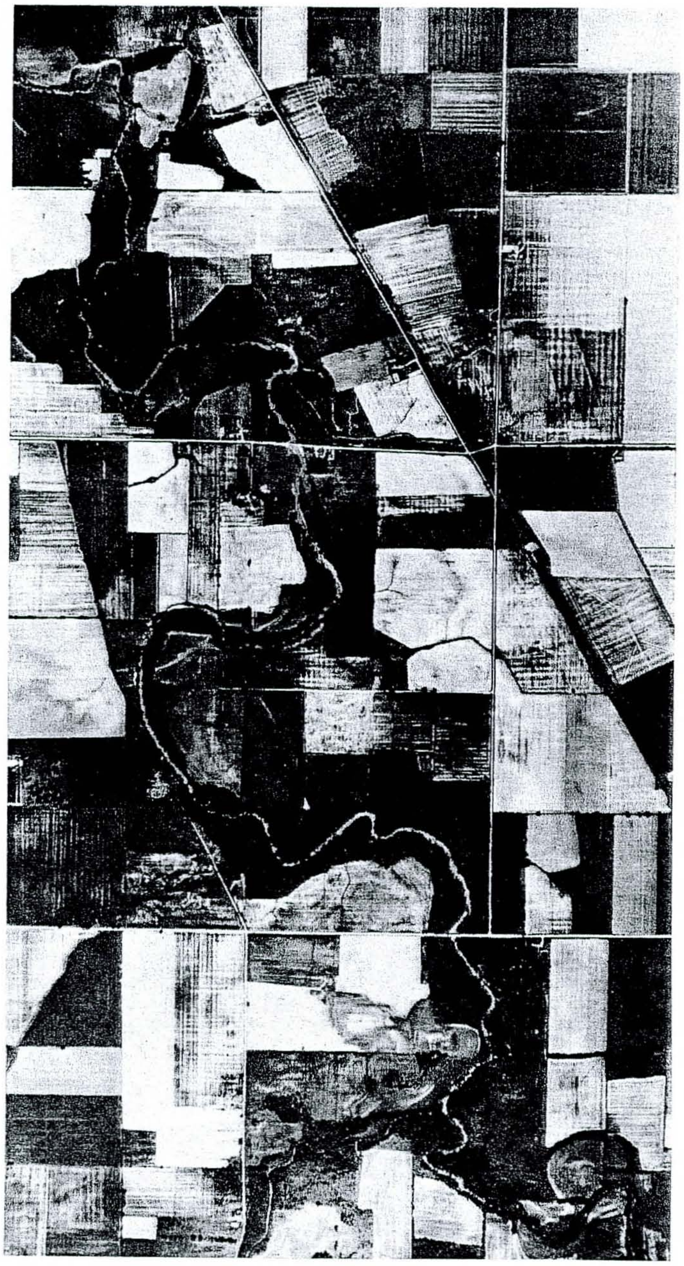


FIGURE 15.—Aerial photograph of part of the Auglaize River, Ohio, showing valley and stream meanders.

glaciated country before breaking through the last morainic barrier into the basin of glacial Lake Agassiz (pl. 2). Its upper basin includes a nexus of melt-water channels, not all of which functioned simultaneously. A tangle of complications in the sequence of events responsible for the channels arises from the competition of two ice lobes, from the variable relation between the Sheyenne and the James Rivers, and from the fluctuations of glacial Lakes Souris and Agassiz. At times, water from glacial Lake Souris spilled into the valleys of the James and the Sheyenne. At time cross connections led melt water from countr drained by the Sheyenne into the basin of the where it flowed southward, eventually reachir Mississippi at Yankton (Geol. Soc. America, But, as the Souris lobe receded and as the Agass melted back toward Devils Lake, something like ter outlet established itself. This channel, passes through the Heimdal and Hillsdale moı is distinctly trenched and meandering. Betwee raines, it tends to become braided across belts (or of outwash, but after entering the next moraine stream it resumes its meandering habit and s boldly from side to side through the whole north- reach downstream from Valley City. The great ders persist as far downstream as the point whe channel reaches the topmost beach of glacial Agassiz.

Unlike meandering valleys of unglaciated area postglacial meandering valleys of glaciated area Sheyenne channel shows little sign of increasin wavelength of its meanders except in its lowe reaches. The apparent abrupt increase in wavel at the lower end (table 1), if it is truly signif seems hardly explicable by an increment of melt supplied by the not particularly well-marked ch leading in from Eccelston Lake, 13 miles west of V City. However this may be, the uniformity of length along most of the channel suggests that the bends were being cut, melt water was being plied principally at the channel's head—for inst by overflow from Lake Souris.

TABLE 1.—*Wavelengths of meanders of the Sheyenne R spillway*

[Queried areas are interpolated, with the aid of planimetry, in the series o cited by McCabe and Crosby (1959). Total areas for localities downst Warwick include 3,940 square miles in the closed Devils Lake basin]

Locality	Drainage area, in square miles		Number of meanders	M w le in
	Total	Probably contributing		
Harvey (South Fork)	535	171	2	2
Lower North Fork	700(?)	200(?)	2	2
Below confluence of the two forks	1, 600(?)	500(?)	3	2
Above Warwick	1, 790	560	2	2
Warwick	2, 070	660	3	2.
Lake Ashtabula:				
Upstream end	7, 750(?)	1, 850(?)	4	2.
Downstream end	7, 880	1, 900	3	2
Below Valley City	8, 300(?)	2, 100(?)	3	2
Between Valley City and Lisbon	8, 400(?)	2, 200(?)	2	3
Lisbon	8, 500(?)	2, 300(?)	4	3
Kindred [1]	9, 150	2, 970	5	1

[1] The wavelengths listed for Kindred are those of ordinary valley meande of meanders of the spillway; they are inserted for comparison.

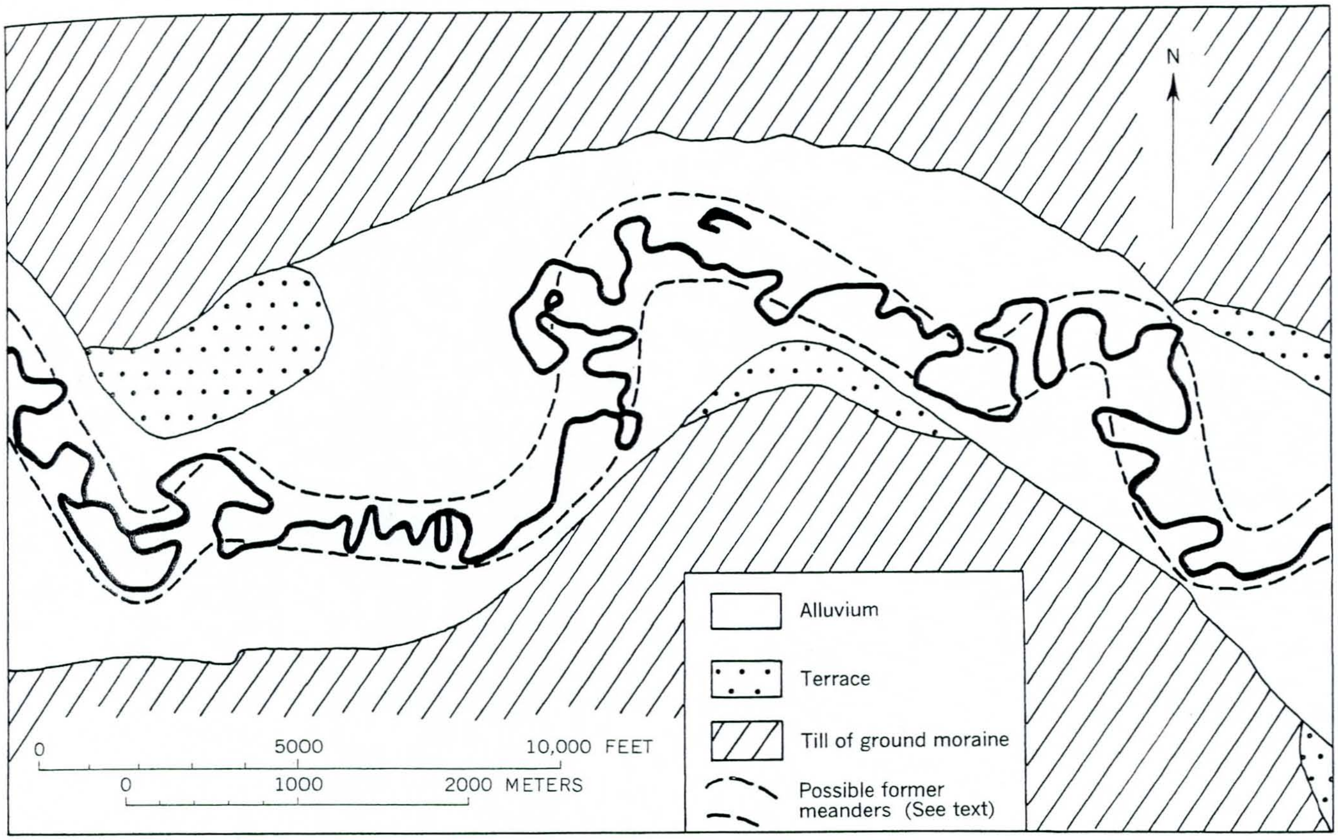


FIGURE 16.—Sketch of the Souris River valley at Minot, N. Dak.

As the meanders of the present Sheyenne increase in wavelength with drainage area, whether total drainage or probably contributing drainage be considered, the river becomes progressively underfit in the meanders of the melt-water channel as it is traced headward. Even in the reaches downstream of Valley City, it is far more underfit than is usual with rivers in nonglaciated regions: the last 5 bends of the melt-water channel contain some 100 stream meanders. But when the Sheyenne passes onto the emerged floor of glacial Lake Agassiz, it becomes an underfit stream of the usual type, with a wavelength ratio between valley meanders and stream meanders of about 5 : 1 (pl. 2).

The valley meanders of normal type occur mainly on that reach of the Sheyenne which runs somewhat north of east from the line of the Herman Beach of glacial Lake Agassiz northeast of Lisbon to and through the Wahpeton moraine and the Campbell Beach. Relative to the local chronology, the earliest possible time for the inception of this train of valley meanders can be fixed within quite narrow limits. When the lake stood at the Milnor Beach, the basin of Lake Agassiz was still almost filled with ice, and water discharged eventually to the south into the valley of the Minnesota River near the south end of Lake Traverse (Leverett, 1932; U.S. Army Map Service 1:250,000, Milbank and Fargo sheets, NL 14–6 and NL 14–9). With further recession of the ice and the formation but progressive fall in level of Lake Agassiz 1, Cottonwood Slough and the Lake Traverse outlet took over the southward outlet of water, which, after an early cessation, recommenced when readvancing ice in the north created Lake Agassiz 2. The Campbell Beach, which crosses the present Sheyenne on the inner border of the Wahpeton moraine, extends into the broad southern gateway leading to Lake Traverse; but, before this beach was cut, the huge lacustrine delta of the Sheyenne, opposite the mouth at the Milnor and Herman Beaches, was already exposed. Thus, the extension of the Sheyenne across the delta as far as the Wahpeton moraine occurred in the interval between the Herman and Campbell stands of the lake. When the water level descended still lower, valley meanders were cut by the still extending Sheyenne for an additional 5 miles, bringing them below the 950-foot contour and well within the Campbell Beach.

The lower end of the train of valley meanders on the Sheyenne is too ill-defined to show whether or not there is a sharp change from two sets of meanders to only one. If such a change were demonstrated and were found to occur also at corresponding positions on other rivers, then a useful fix could be obtained; for, in this northern region, the conversion from large to small meanders may have occurred later than in regions farther to the south. As matters stand, it is possible merely to observe that the Red River, the trunk stream flowing along the axis of the former lake basin, is not underfit. In this respect, as in its setting, the Red River resembles the Maumee. Two obvious possibilities are that the trunk streams did not begin to incise themselves until after their laterals had become underfit or that the forms of valley meanders have not been preserved in the weak materials of the bottoms of the two basins.

Some other rivers resemble the Sheyenne in having cut valley meanders across part of the lake floor. Rivers tributary from the west display the relevant features better than do those tributary from the east, and they also make possible an extension of dating. Until the ground has been carefully examined for signs of beaches below the Campbell—the McCauleyville Beach—the interpretation of possible valley meanders on the Dakota Wild Rice River, west and northwest of Wahpeton, must remain in doubt. Valley meanders on the Maple River, west-southwest of Fargo, go slightly lower (to about 920 feet above sea level) than do those on the nearby Sheyenne. If the large bends on the Minnesota Wild Rice River, about 25 miles north of Fargo, are valley meanders, then such meanders were cut into the lake bed at 850 feet; the bends on the lower Buffalo River, which enters the Red River 15 miles north of Fargo, seem likely to be authentic valley meanders, and they also are cut below the 850-foot mark. The North Branch of Elm River describes unmistakable valley bends a little above this level yet well inside not only the Campbell and McCauleyville but also the succeeding Blanchard and Hillsboro Beaches. There is room, however, in the sequence of events since ice receded across the basin of Lake Agassiz for more than one fluctuation of discharge, and it is most desirable that, in due course, the local valley meanders should be dated. As yet, all that can be said is that the local rivers have been reduced to an underfit condition, subsequent to the emergence of the lake bed; comparison between the middle and lower reaches of the Goose River, next northward from the Elm River, suggests that more than one generation of valley meanders may be present, those upstream being larger and earlier than those downstream. Because the largest of the indubitable valley meanders (meanders of spillways always excepted) are but some five times as long as the stream meanders, even these valley meanders are likely to be savagely bitten by the loops of the present streams, and any valley meanders of a lesser order promise to be identifiable only by means of very detailed investigation.

Still farther north, however, useful observatio be made on the Park, Tongue, and Pembina (U.S. Army Map Service 1:250,000, Thief Rivei Sheet, NM 14–12). The Pembina describes vall anders of a normal size, clearly distinguishable irregularities of trace associated with lake beacl least as far downstream as Neche, N. Dak.—tl within the Burnside Beach. The old lake bed stands at about 835 feet above sea level; the presen plain, some 25 feet lower, is underlain by as much feet of silt, clay, and sand—river deposits which belt as much as three-fourths of a mile in width, a cluded in the silt of Lake Agassiz (Paulson, 1951) these deposits correspond to the fills of large cha in meandering valleys, then the Pembina, when i large meanders, cut its large bed perhaps as low a feet above sea level at Neche.

Although the course of the Tongue River is festly affected by old shorelines, valley meanders a as 800 feet above sea level appear between the Os and Stonewall Beaches. But the most clearly d oped valley meanders at low level occur on the River, where they are below 800 feet above sea leve extend at least 10 miles beyond the Gladstone Beach possibly the whole way to the Red River.

In summary, the site of glacial Lake Agassiz s of the United States–Canadian border appear record the extension, with varying strength and varying subsequent preservation, of large mean across the progressively emerging floor, in place least as late as the date of the Ossawa Beach, and b the 800-foot level. The examples cited all refer to floor of Lake Agassiz 2, the history of which be with the 1,080-foot (Herman Beach) stand, date about 11,200 years B.P. (Flint, 1957, p. 347). As it is not possible to indicate what proportion of recession period of that lake—a period extending f about 11,200 to perhaps 7,500 years B.P.—included new establishment of large meanders. This matter n await the dating of relevant deposits; but the evid adduced here is at least sufficient to emphasize the tinction between the valley meanders of spillways those of rivers which did not receive spill water, e when both sets of conditions are exemplified on a sir stream.

STRATFORD AVON, ENGLAND, AND GLACIAL LAKE HARRISON

For the basin of the Stratford Avon, the convers of rivers to underfitness is clearly separable in time fr the presence of ice and the operation of spillways. So trains of valley meanders were initiated during the l interglacial (Holstein of northern Germany, Sangam of North America), whereas the shrinkage which m

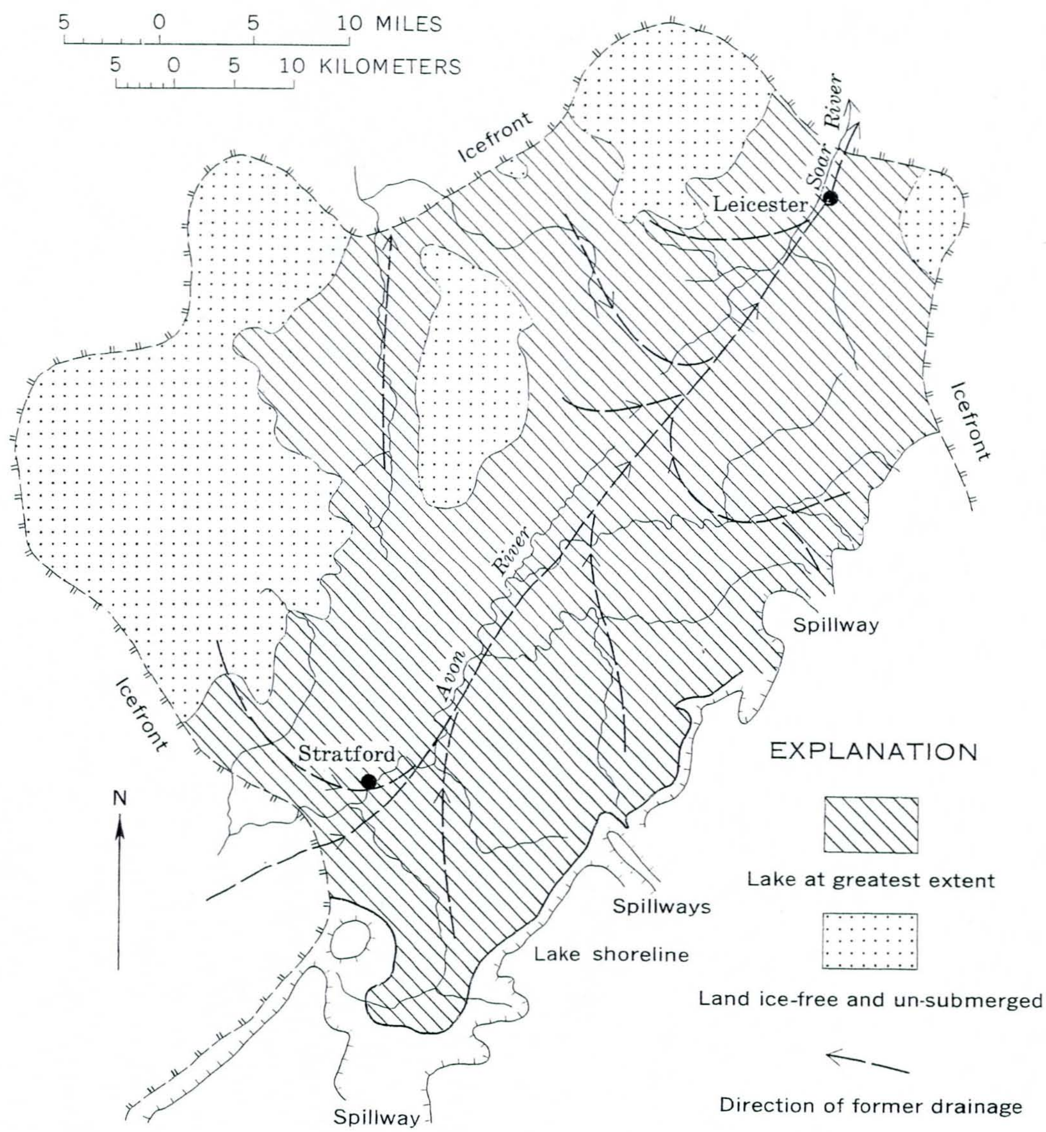
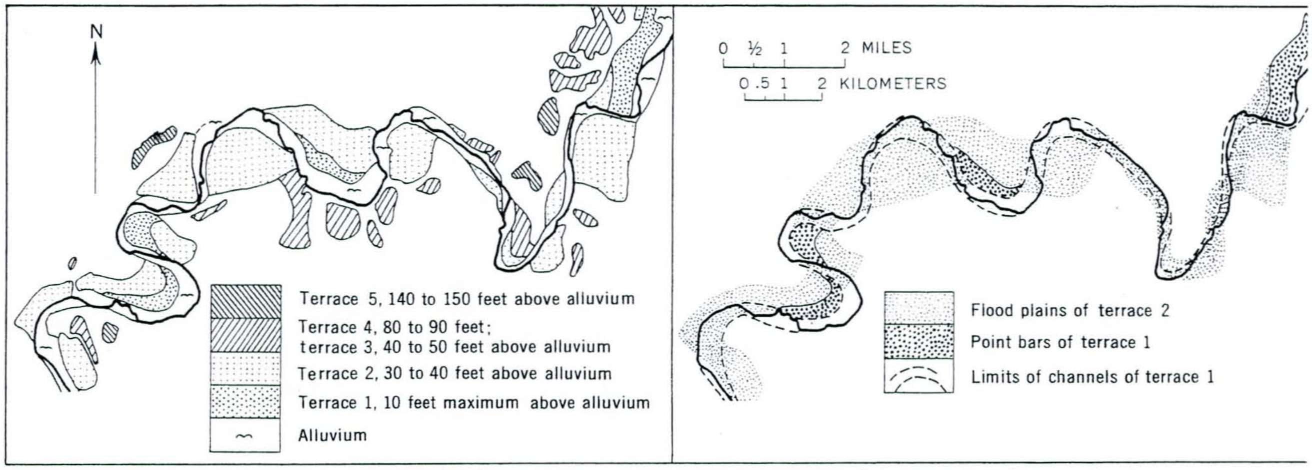


FIGURE 17.—Sketch map showing former drainage basin of the Avon River, Warwickshire, England, in relation to glacial Lake Harrison.

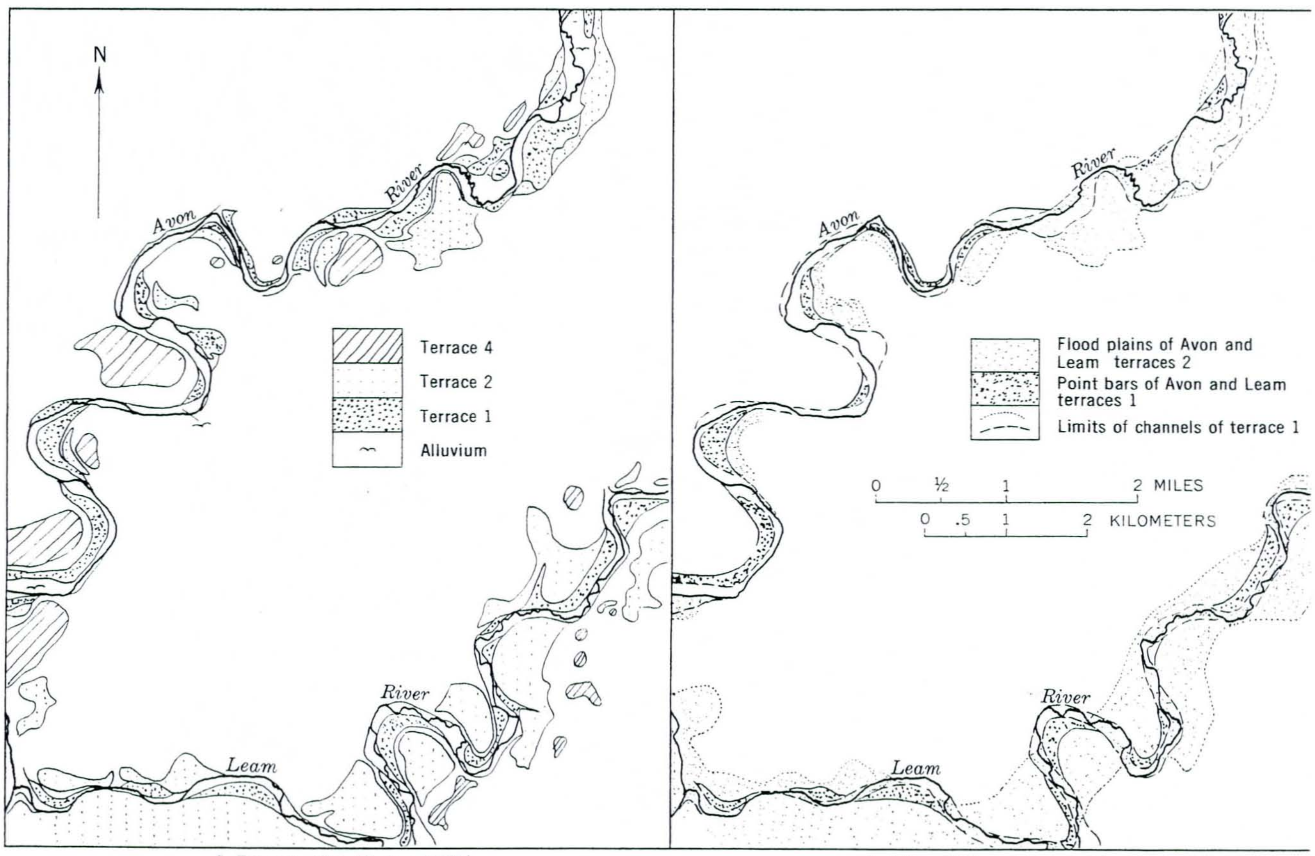
the rivers underfit was deferred until quite late in the last glacial (Weichsel, Wisconsin). Although ice-dammed lakes once existed in the region, they date from the Penultimate Glacial (Saale, Illinoian) and can have no part in explaining either the valley meanders or the reduction in discharge. Ice did not invade the Avon drainage basin during the last glacial, nor did melt water spill into it. In these circumstances, the time gap between the last local glaciation and the conversion to underfitness is readily demonstrated; the necessary outline of the evidence can be far shorter than that for the Lake Agassiz region, where, as seen above, active spillways and the induction of underfitness belong to a single part of the glacial sequence. At the same time, the extended time span applicable to events in the Avon basin demands that the expression "conversion to underfitness" be qualified. The conversion in question is that responsible for the present condition of rivers. It will be mentioned as if it were a unique event, without prejudice to the possibility that similar conversions may have occurred earlier. All that is required now is to show that valley meanders were still developing late in Last Glacial times, regardless of fluctuations of discharge that had occurred previously.

Before the Penultimate Glacial, most of the area now drained by the Avon was tributary to the River Trent. It was included in the drainage basin of the Soar, which was then considerably longer than it is today (fig. 17). Advancing ice impounded a series of lakes against the Cotswold scarp and its continuation to the northeast.



1. Terraces after Tomlinson (1925) 2. Interpretation of terraces 2 and 1

A. STATFORD AVON NEAR EVESHAM



3. Terraces after Shotton (1953) 4. Interpretation of terraces 2 and 1

B. STRATFORD AVON AND LEAM RIVERS NEAR LEAMINGTON

FIGURE 18.—Sketches of the Avon River, Warwickshire, England, showing river terraces and two interpretations of these terraces. Stratford Avon near Evesham ; *B,* Stratford Avon and Leam Rivers near Leamington.

Collectively, these lakes are called glacial Lake Harrison (Shotton, 1953) ; various extents and levels are distinguished as named stages (Bishop, 1958). The detailed history of Lake Harrison is not material here: the essentials are that while the lake existed, spill water discharged through gaps in the bounding scarp on the southeast; and when the lake was finally drained, Stratford Avon came into being on the bulky fil glacial, proglacial, and lacustrine sediments in the t porary basin.

The development of the Avon system is well recor by a series of terraces (Tomlinson, 1925, 1935; Shot

1929, 1953; Bishop, 1958). (See fig. 18.) Avon terrace 5, the highest and oldest, was assigned by Shotton and Bishop to the Penultimate Glacial; it may well consist of outwash provided by the receding ice. Terraces 4 and 3 are younger and lower than 5. Both date from the last interglacial, and both are fluvial in origin. Their somewhat dubious interrelation does not concern the present argument nor affect the circumstance that terrace 4 was deposited in a broad open valley. If significant downcutting had already occurred before the completion of terrace 4, it had been largely offset by subsequent infilling. Partly for this reason and partly because terraces 4 and 3 are but scantily preserved in most of the valley, there are few reliable signs that the meanders (valley meanders) of Avon 4 had ingrown. Terrace 2, lower and younger still, is by contrast extensively preserved. Enough remains to show that, in some reaches, Avon 2 had swept out a meander trough almost as broad as the existing belt of valley meanders, whereas elsewhere the great bends were still confined by spurs on which crescentic patches of terrace 2 represent point bars. Avon 2, that is to say, displayed the two arrays of landform which are diagrammatically illustrated in figure 4 at sites 1 and 3. Ingrowth of valley meanders continued when the river cut through terrace 2, for the reconstructed trace of Avon 1 transgresses the limits (only in part reconstructed) of terrace 2; the meander belt of Avon 1 was broader than that of Avon 2 (fig. 18*B*). On the tributary Itchen, valley meanders are cut through terrace 1 (Shotton, 1953, fig. 9; Bishop, 1958, fig. 6). A former larger stream postdates Avon 1, so that reduction of discharge and conversion to the present state of underfitness must be placed later still.

The absolute gap of time between the disappearance of ice and the appearance of stream meanders on the existing rivers cannot be assessed precisely. Something depends on the span allocated to the Pleistocene as a whole. For example, Zeuner's data (1959, chaps. 4, 6) suggest an age of at least 185,000 years for Lake Harrison and terrace 5, and age of about 125,000 years for part at least of terrace 4 (and 3?), and an age of 75,000 years for terrace 1. As terrace 1 does not represent the final incision of valley meanders, the interval of about 110,000 years between the last local deglaciation and the last conversion to underfitness is too short. Although Emiliani (1955, fig. 15) requires but half the length which Zeuner gives to the whole Pleistocene, his correlation does not greatly reduce the interval under consideration. On Emiliani's scale, the end of Lake Harrison and the deposition of terrace 5 fall at about 105,000 years B.P., whereas terrace 1 cannot be referred to anything but Emiliani's position for Würm II of the Alps and the Wisconsin of North America—say, at about 15,000 years B.P. The gap is still no less than 90,000 years, even without allowance for the persistence of large meanders after the formation of terrace 1. Glacial events in this region have no possible bearing on the regionally underfit state of the existing rivers.

The spillways on the southeast ceased to function when Lake Harrison was drained. In the Cotswolds, as on the Avon, there is good evidence that the rivers did not become underfit until much later. Although spill water discharged from Lake Harrison into the valleys of the Evenlode and Cherwell (Shotton, 1953; Bishop, 1958; Dury, 1951), these rivers are no more underfit than are other rivers which, draining parts of the Cotswold back-slope, emphatically did not carry overspill. As the maximum height of Lake Harrison was 435 feet above sea level, discharge could not have occurred except through gaps leading to the Evenlode and the Cherwell and possibly also to the east-flowing Nene. No appeal can be made to hypothetical lakes formed during glacials earlier than the Penultimate, for the valley meanders of the Evenlode did not then exist. Despite the correlation attempted by Arkell (1947, table 2), these meanders had probably been formed, and had begun their ingrowth, before Lake Harrison overflowed into the Evenlode valley (Bishop, 1958, fig. 12). The first incision of the Evenlode through a fanlike spread of gravel—the Hanborough Terrace—seems likely to belong late in the Penultimate Glacial and certainly to antedate the first outspilling of the lake. Furthermore, the main headstream of the Cherwell rises near an unbroken crest more than 600 feet above sea level and well out of reach of Lake Harrison; but the stream is manifestly underfit, just as much underfit as those reaches which occupy the spillway (Dury, 1953c). The synoptic profiles drawn by Bishop (1958, fig. 8) put the existing flood plain at 25 to 45 feet below the floor of the spillway. A descending sequence of terraces proves that erosion continued after the spillway ceased to function. Indeed, the valley meanders go below the surface of the flood plain into the so-called sunk channel, and this is an indication that they were incised by 40 to 60 feet after Lake Harrison had fallen for the last time below the col which linked it with the Cherwell valley.

Because the Cotswold rivers are equally underfit and because their condition cannot be explained by derangement of drainage, it seems likely that a date for the shrinkage of one stream would apply to all. A date is forthcoming from the Cotswold River Dorn, where the valley meanders were finally abandoned about 10,000 or 9,000 years ago (Dury, 1958). If this date applies at all widely—as, in the writer's view, it does—

then the onset of underfitness is more widely separated than ever in time from the last local discharge of spill water and from the last local glaciation. Just as with the site and borders of Lake Agassiz, glacial derangements of drainage can be seen to have no general bearing on the origin of underfit streams.

REGIONAL DISTRIBUTION AND A REGIONAL HYPOTHESIS

At the same time that the wide distribution of manifestly underfit streams in a given region conflicts with hypotheses of derangement, it supports the claim that underfitness need not always be manifest. Even where they are highly characteristic, meandering streams are rarely exclusive; but there is no purpose in contending that a stream, manifestly underfit in most of its length, ceases to be underfit in a single reach where either valley meanders or stream meanders are absent. If reaches upstream and downstream have been affected by a change in discharge, then the intermediate reach must have been similarly affected. Again, reaches are easy to locate where the present channel, meandering in the natural state, has been regularized, so that the combination of forms essential to manifest underfitness has been destroyed. Natural irregularities combine with artificial works to reduce the numbers of streams and the lengths of reaches which are manifestly underfit; distributional maps such as figure 11 tend to understate the facts.

Minor allowances for occasional reaches cause no difficulty. Regions such as eastern France demand a regional hypothesis, which cannot be other than climatic. But if a climatic hypothesis is adopted, then it becomes applicable wherever manifestly underfit streams are usual. An apparently obvious procedure is to map the distribution of manifestly underfit streams so that spatial limits can be fixed for the hypothesis of climatic change.

Practical difficulties arise here. One is that the task of distributional mapping is tedious and involves nothing more than the expenditure on routine work of time which could be more profitably spent otherwise. A second and related difficulty is that every gradation seems possible from regions where all streams are manifestly underfit to regions where none of the streams are underfit. It therefore becomes necessary to use some kind of index of manifest underfitness if regions are to be described as possessing streams that are mainly underfit. Such an index could, for example, express the total length of manifestly underfit reaches as a percentage of the total length of all streams. But, if a continuous range extends from total to zero underfitness, any such index could serve no purpose except that

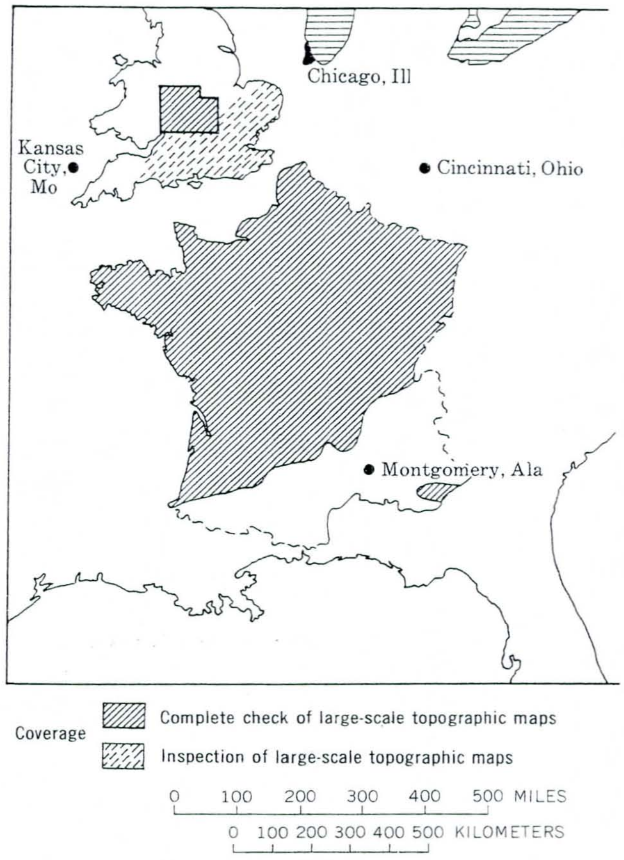


FIGURE 19.—Known areal extent of manifestly underfit stream France and southeast England superimposed on comparable area the United States.

of description. For reasons which shortly will be ol ous, no attempt has been made either to define or apply an index of underfitness.

In western Europe, manifestly underfit streams typical of large areas. In view of what has just b said about the use of an index, the word "typic should perhaps be taken to signify that an estima minimum of 50 percent of the total length of strea of the second and higher orders is manifestly under Distributional maps (fig. 11 above; Dury, 1953d, fig. Dury, 1954, fig. 2) give samples of the incidence question. Topographical maps on scales ranging fr 1:20,000 to 1:80,000 have been examined for the wh of France and for a large part of the English Pla these maps reveal that manifest underfitness charact izes many reaches of many streams in an area that e tends about 600 miles from north to south and 500 mi from west to east (fig. 19). These distances, which a roughly equal to the distances from Chicago, Ill., Montgomery, Ala., and from Kansas City, Mo., to Ci cinnati, Ohio, are thought sufficient to demonstrate th manifest underfitness in France and England can explained only by a shift in climate.

A third practical difficulty arises when attempts are made to trace the distribution of manifestly underfit streams eastward across Europe and southward toward the Mediterranean. Whereas the German 1:25,000 map is excellently suited to record the relevant combination of forms, some other surveys do less well, either because they do not purport to represent the necessary fine detail of channel pattern or because their cartographic techniques do not permit such detail to be shown. Consequently, certain rivers can be identified as manifestly underfit, but proof that other rivers are not so may be impossible. In the United States, where topographic coverage on scales no smaller than 1:62,500 is incomplete, it also is impossible to deny—or to confirm—the regional development of manifestly underfit streams in considerable areas. Moreover, even where maps exist, they can be misleading.

On occasion the forms of valley meanders are misrepresented, and not merely because the interval and incidence of contours prove unhelpful. The valley of the Kickapoo River near Soldiers Grove, Wis., is by no means well shown by the 1:62,500 map (Gays Mills quadrangle, Wisconsin, surveyed 1923–24, published 1924). The topographic sheet indicates one clear left-hand swing of the flood plain at Soldiers Grove, a broad bowing to the right on the next 4 miles downstream, and a second left-hand swing at the confluence of Bear Creek. Aerial photographs reveal that low hills rising from the valley floor on the east side of the stream are the old cores of valley meanders, that the open lower end of the valley of Bear Creek is the curve of a valley meander, that the side of Mother Lot Point is the opposing scar next upstream, and that the succeeding left-hand scar occurs on the north flank of Bear Point (figs. 20, 21). In addition, the photographs show a large scar on the right of the stream, immediately west of Soldiers Grove; although this scar is suggested by the map, the large upstanding core in its center is omitted.

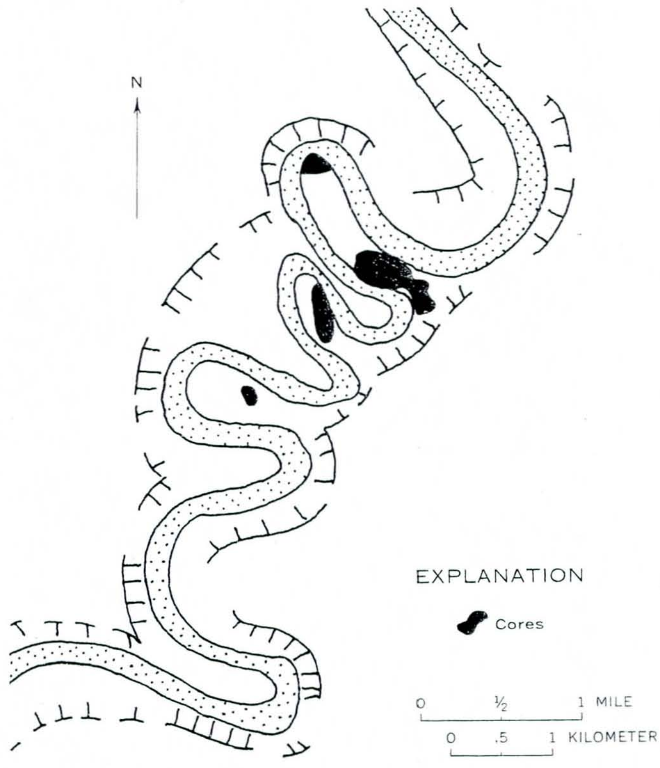


FIGURE 21.—Sketch of the Kickapoo River showing former meanders reconstructed from aerial photographs.

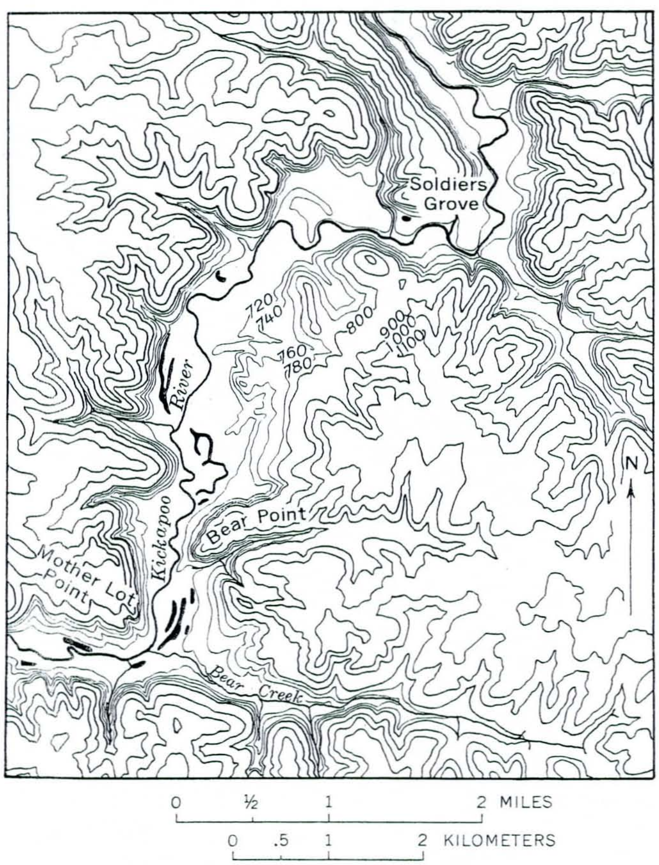


FIGURE 20.—Map of the Kickapoo River near Soldiers Grove, Wis.

Stream meanders seem most liable to omission or to misrepresentation. An example has already been given of streams on the emerged floor of Lake Whittlesey-Warren which are manifestly underfit on aerial photographs but have their present meanders obscured by the topographic sheets. Even where maps are drawn from aerial photographs, the trace of present channels is not invariably shown with great accuracy. Little streams can be wholly concealed by overhanging trees, and the fine detail of small rivers generally seems capable of becoming generalized in the process of cartography. In some localities, rivers cannot justly be identified as not manifestly underfit until their trace and the forms of the valley sides have been checked from aerial photographs and possibly also on the ground.

A further practical limitation to map evidence is that, in places, erosion has largely destroyed the forms of meandering valleys. Manifest underfitness cannot be ruled out, unless whole blocks of sheets are availabl[e] inspection. Many examples are possible. One [...] perhaps suffice to indicate how rapidly the form o[f] ground can change within a short distance.

The Delaware River of Kansas, which occupi[...]

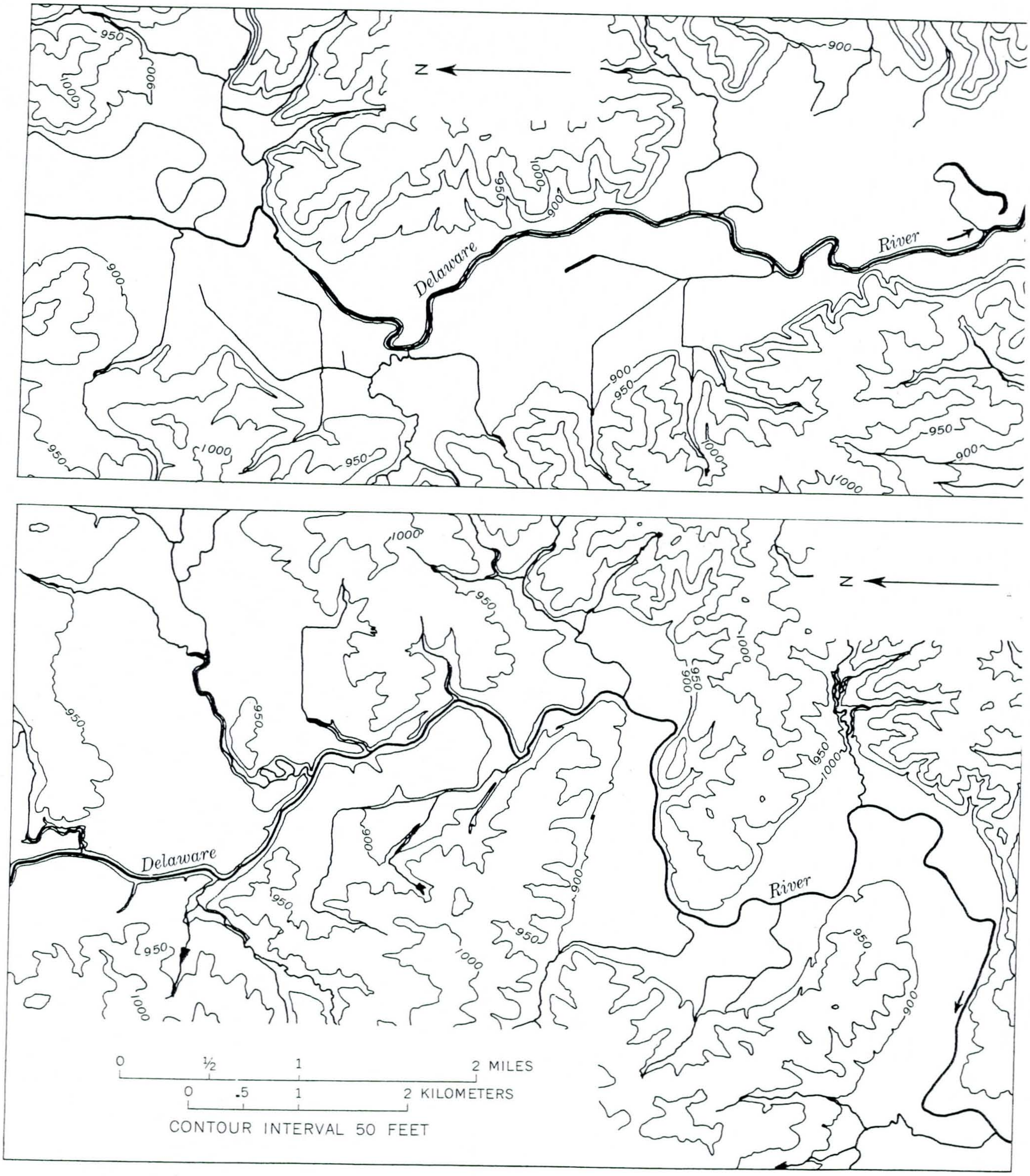


FIGURE 22.—Contrasted valley patterns on the Delaware River, Kans.

valley 65 miles long that trends from north-northwest to south-southeast, enters the Kansas River 15 miles downstream of Topeka, Kans. The valley makes an angle of some 45° with generalized outcrop boundaries; the river passes in the downstream direction onto progressively older rocks, heading in the Council Grove Group of the Permian System, traversing the outcrop of the Admire Group of the same age, and subsequently reaching rocks of the Wabaunsee and Shawnee Groups of the Virgil Series of the Pennsylvanian. (See, for outcrops, Kansas State Geol. Survey, 1937; for lithologic description, Moore and others, 1951.) Broad morphologic contrasts match the broad contrasts in rock strength.

Parts of the upper 25 miles of valley, which is incised into the Council Grove and Admire Groups and into the upper members of the Wabaunsee Group, combine valley meanders with meanders of the stream: the Delaware River is clearly underfit. In the 20 miles or so above Valley Falls, however, the valley widens, having a broad floor and gentle side slopes developed on the mainly shaly rocks of the lower part of the Wabaunsee succession. Valley bends, if present at all in this reach, are but vestigially preserved. Downstream from Valley Falls, resistant rocks reappear in the Shawnee Group and sustain steep walls which rise as much as 100 feet on the outsides of valley bends (fig. 22). The Valley Falls, Kans., quadrangle of the lower U.S. Geological Survey, 1:24,000 map, on part of which the panel of figure 22 is based, well illustrates the entry of the Delaware into a belt of outcrops where the rocks are, in the main, resistant. Near the actual entry, the 950-foot contour marks the bedrock core of a cutoff valley bend; but a little farther downstream, no very marked sweep has occurred; valley meanders are ingrown, but their intervening spurs are no more than trimmed.

Downstream again, however, resistant beds rise gradually above river level so that the valley bends are cut into increasing thicknesses of shale. The upper panel of figure 22, based on part of the Ozawkie quadrangle of the 1:24,000 map, illustrates the landforms formed in these conditions. A distance of 2 miles between the two reaches introduces a most striking alteration in the form of the valley. In this southern (downstream) reach, a number of curved recesses in the valley wall are scallops cut by the former large meanders, but the projecting spurs of the upstream reach are here replaced by the bluntest of cusps. The local rocks clearly offered little resistance to the free downstream sweep of the former large bends. Consequently, whereas the underfit character of the Delaware is manifest in one reach it would be obscure or dubious in the other if this second reach were considered in isolation.

This example returns the immediate argument to its starting point, that the absence of manifest underfitness from a particular reach is no obstacle to the claim that such underfitness can be a regional characteristic. Where manifestly underfit streams occur in widely separated regions, any climatic hypothesis invoked to explain them must be held to apply also to intervening regions. When all possible allowance is made for erosion, for the local breakdown of manifest underfitness, and for deficiencies of maps both in accuracy and coverage, it remains true that the streams of certain areas are not at all manifestly underfit or are but exceptionally so at the most. Consequently, a means must be sought for bringing rivers generally within the scope of underfitness and of climatic change.

CLIMATIC HYPOTHESIS AND UNDERFITNESS OTHER THAN MANIFEST

In conterminous United States, manifestly underfit streams have been identified in locations ranging from the Great Lakes to the gulf coast, and from the Pacific Northwest to the Atlantic coast (figs. 23, 24). The distribution shown in the figures is by no means complete. It is presented merely to emphasize that any hypothesis of climatic change invoked to explain underfitness should apply to most of the country, if not indeed to the whole. Manifestly underfit streams, however, are far less common in the United States than in France or on the English Plain. Examples described in this section will be of combination 2 of figure 4—that is, of nonmeandering streams in meandering valleys.

Hitherto, with manifestly underfit streams taken as the stereotype—indeed, with most writers, as the only type—it has been possible to regard nonmeandering streams in meandering valleys as evidence for hypertrophy of stream meanders; for the influence of structure, lithology, and crustal movement; and for the absence of underfitness. The examples described here are meant to show that some rivers in meandering valleys devoid of stream meanders are underfit. The corollary inference is that, within the limits of distribution of underfit streams in general, all streams in meandering valleys may be underfit, even though they display but one series of bends. Although underfitness cannot be demonstrated unless bed form is known, comparative measurements of bed width and of wavelength (wavelength of valley meanders) give strong general support to the corollary stated. The observations and conclusions presented here are considered to shift the onus of proof. The underfitness of streams in incised meandering valleys can no longer be denied, either ex-

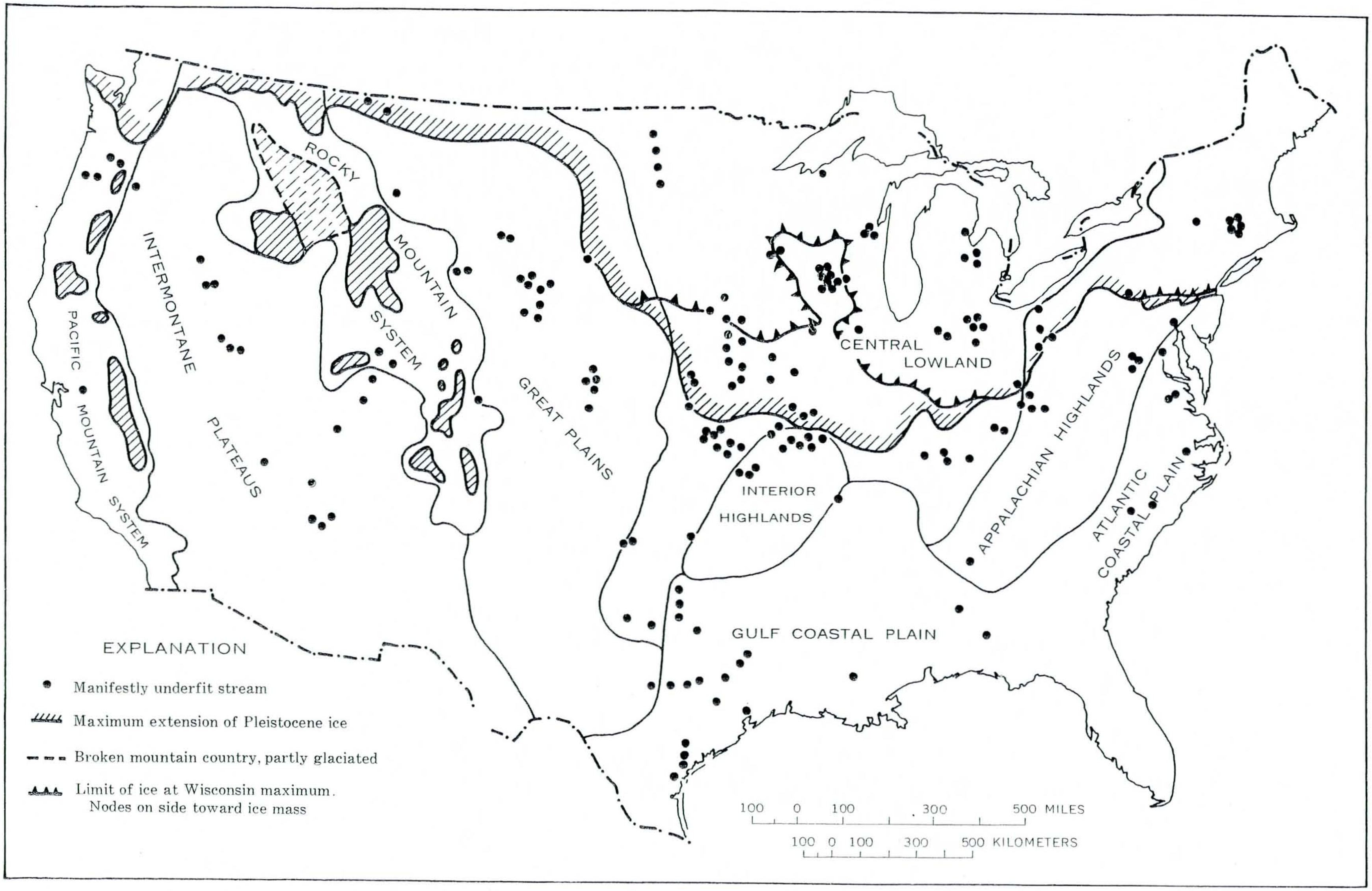
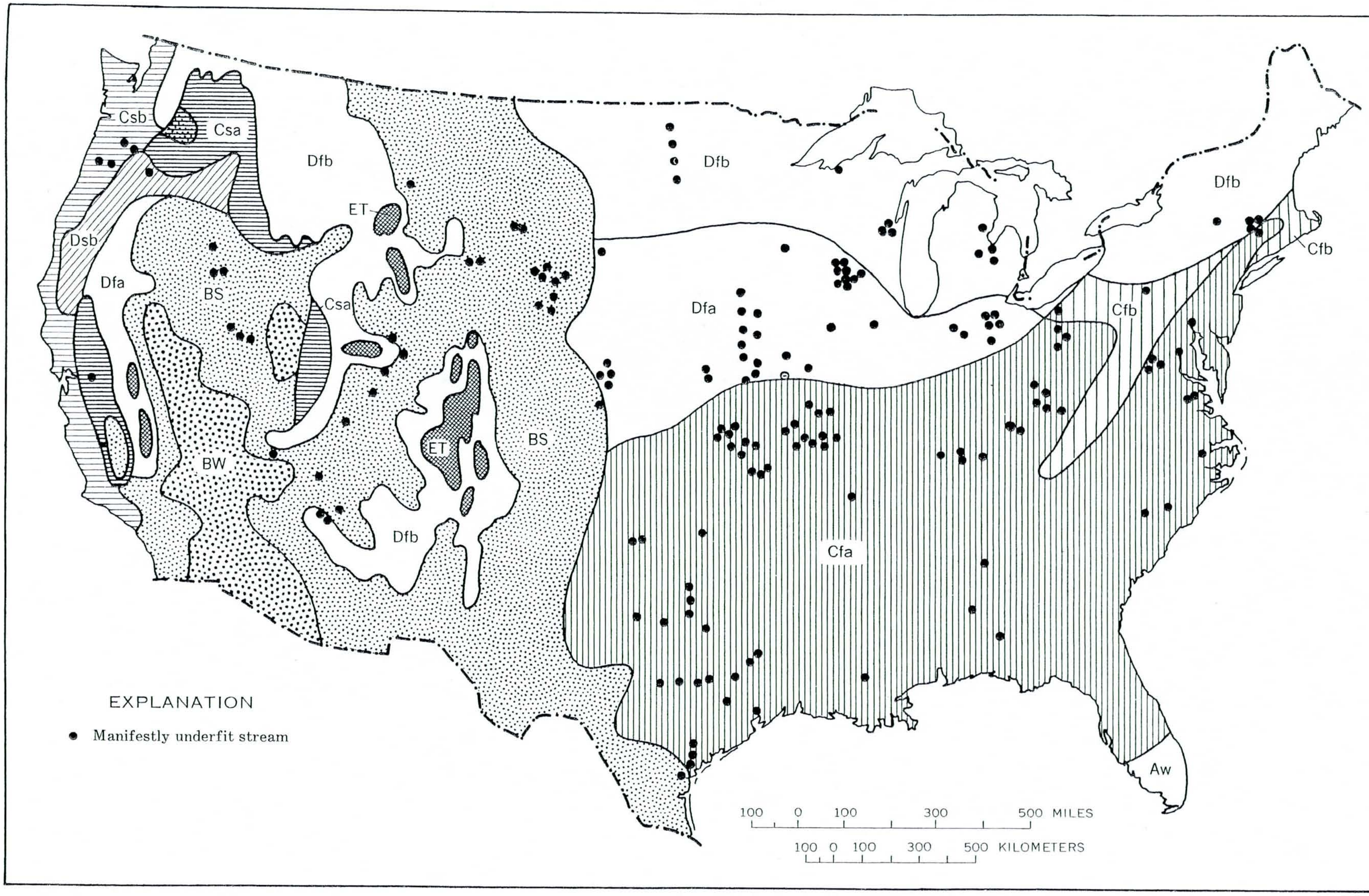


Figure 23.—Map of conterminous United States showing areal range of manifestly underfit streams in relation to physical subdivision.



AW, tropical rainy, with dry low-sun season.
BW, arid.
BS, semiarid.
Cfa, warm temperate rainy, with no dry season and hot summer.
Cfb, warm temperate rainy, with no dry season and cool summer.
Csa, warm temperate rainy, with hot dry summer.
Csb, warm temperate rainy, with dry cool summer.
Dfa, cold snowy, with no dry season and hot summer.
Dfb, cold snowy, with no dry season and cool summer.
Dsb, cold snowy, with cool dry summer.
ET, tundra.

FIGURE 24.—Map of conterminous United States showing areal range of manifestly underfit streams in relation to climatic subdivisions. After Köppen (1931).

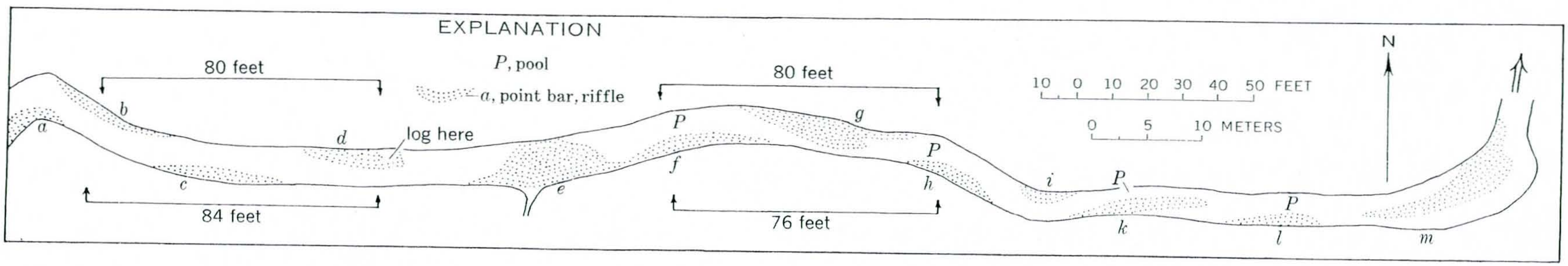


FIGURE 28.—Sketch of part of McDonald Creek, Scott County, Iowa. Letter symbols explained in text.

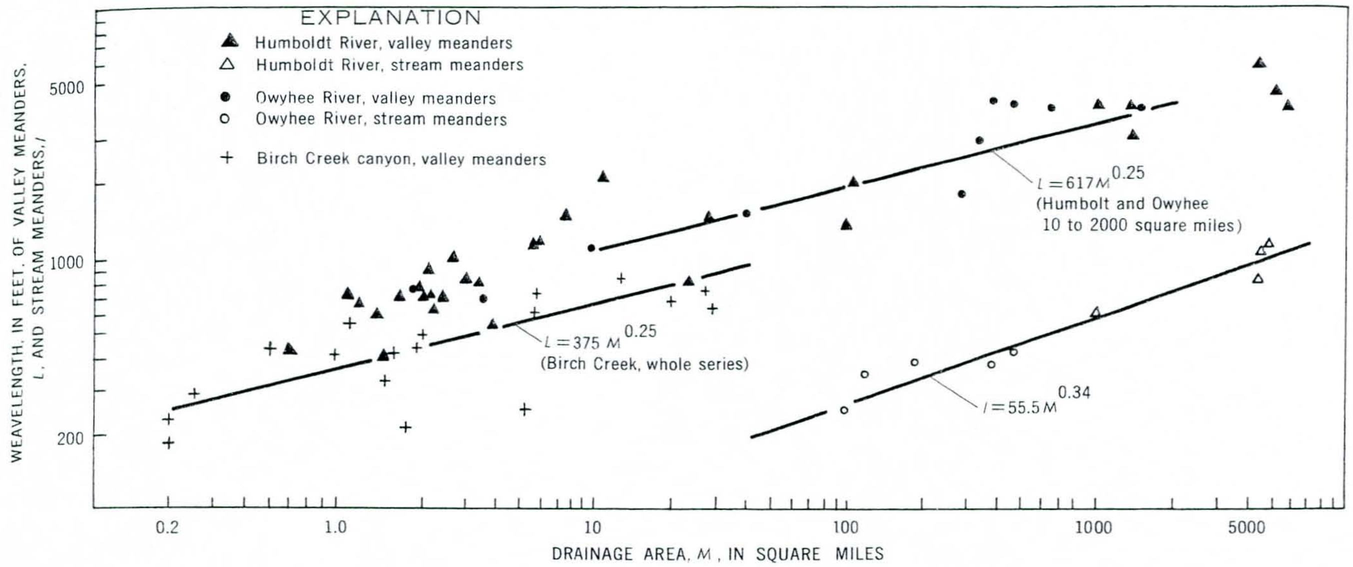


FIGURE 29.—Graph showing relation of wavelength to drainage area of the Humboldt and Owyhee Rivers and Birch Creek.

enced by factors additional to the curvature round valley bends.

McDonald Creek (Eldbridge quadrangles, Iowa, 1:24,000, T. 80 N., R. 3 E., sec. 24) was traversed along the west-east limb of a valley bend, with the results shown in figure 28. Although the channel is not sinuous in this reach, the bars are distributed according to a rough system. Bars *a* through *d* and *f* through *i* seem to display a true alternation; they permit the measurement of four wavelengths, as shown in figure 28, which average 80 feet—that is, about eight times the bed width.

HUMBOLDT RIVER, NEV.

The Humboldt River, Nev., which cuts from place to place through upstanding blocks of hills, is excellently adapted to illustrate the contrast between stream meanders and valley meanders, even though the two rarely occur on a single reach. On the open floors of basins, long reaches of the Humboldt meander considerably, with bends about 10 times as long as the channel is wide. Where the river enters a canyon, however, apparent wavelength suddenly increases (fig. 29). The streams curve round valley meanders, wherein stream meanders are unusual. The distinction of magnitude between the two series is well displayed by the regional graph (fig. 29), but separation of the two series from one another does not in itself dispose of the hypothesis that the large meanders are in some way a response to cutting into bedrock. When the present channels are inspected, however, they are found to contain pools and riffles much more closely spaced than the bends and inflection of the canyons. Such is true for the South Fork of the Humboldt where it trenches across the end of Grindstone Mountain, 8 miles southwest of Elko, Nev. (Dixie Flats quadrangle, Nevada, 1:62,500; fig. 30). Both upstream and downstream from the canyon, the river describes meanders of the size expectable from its bed width. Within the canyon, the stream is certainly braided in part, although any systematic qualities which its bed form may display cannot be detected without instrumental survey.

In Carlin Canyon, 6 miles east of Carlin, Nev., the trunk Humboldt is in places somewhat confined by highway and railroad embankments (Carlin quadrangle, Nevada, 1:62,500). Nevertheless, braiding can be observed to set in at the approaches to the canyon and to occur in places within it, whereas a meandering habit is resumed farther downstream (fig. 31). In the next succeeding canyon, Palisades Canyon, wherein the stream is much compressed by the railroad embankments, braiding again occurs near the entry of Pine Creek, where the railroads cut through a lobe of rock and the channel is unconfined. The two intervals in a succession of three braids average about one-fifth of the mean wavelength of valley bends in this reach of canyon. Toward the downstream end, a section through the local valley fill was provided in 1960 by the strip mine near Barth (Beowawe quadrangle, Nevada, 1:62,500). Gravelly alluvium was seen to extend at least 35 feet below the river bed, opposite the mouth of the lateral Safford Canyon. Before mining caused the Humboldt to be diverted, the river had a strong tendency in this reach to form meanders, as is suggested by the topographic sheet and clearly displayed on the ground. One possible inference, shortly to be confirmed by observations on the Shenandoah, is that out-

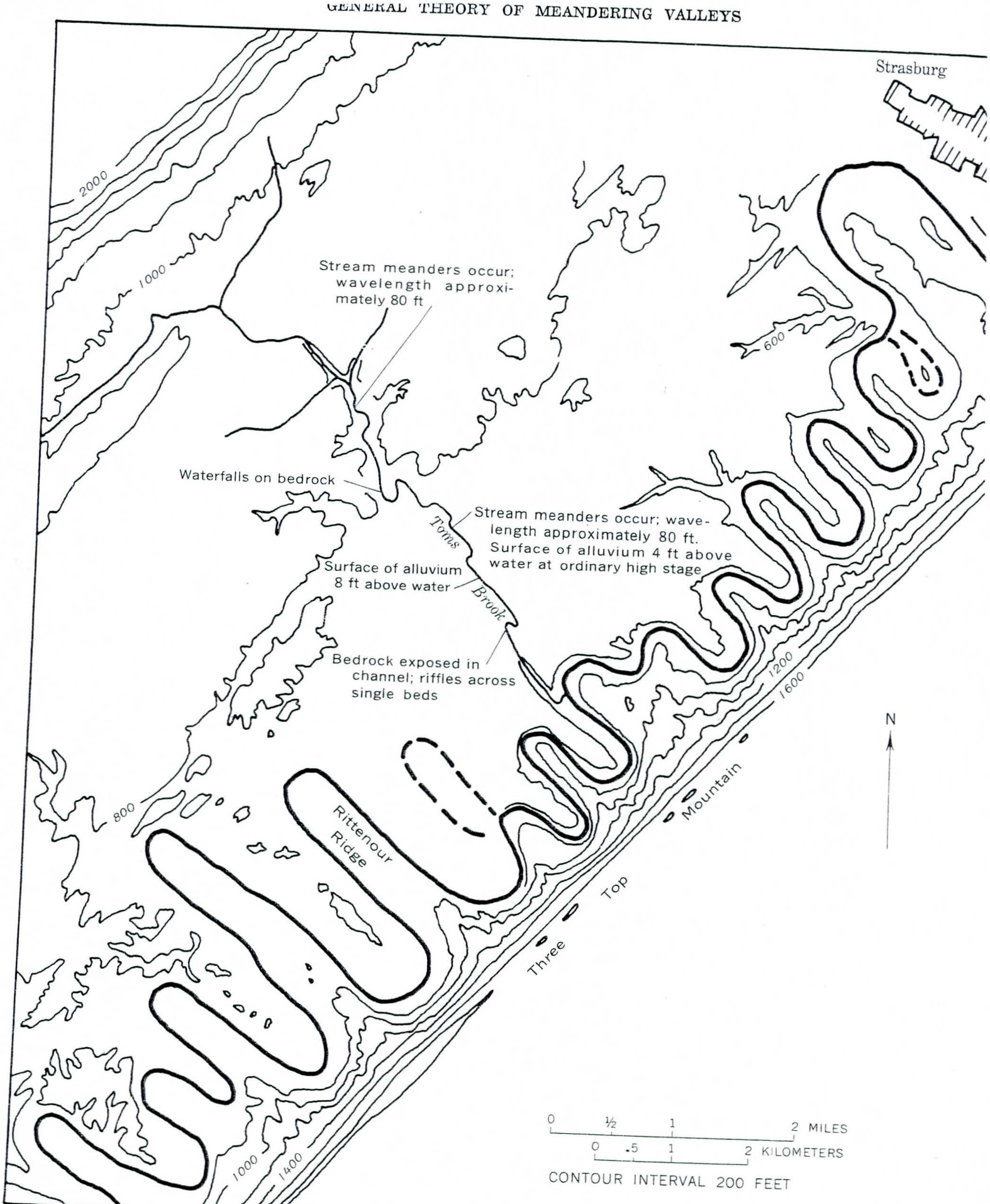


FIGURE 33.—Map of part of the North Fork of the Shenandoah River near Strasburg, Va.

which is highly relevant to the present inquiry. The main fork of Smith Creek, southwest of the Mount Jackson quadrangle, possesses actual stream meanders in addition to the valley meanders which are alone indicated by the map, as also does Holman Creek in a reach 3½ miles southwest of Mount Jackson; even on lower Smith Creek, the actual meanders of the stream are more boldly formed than the map suggests. None of the present meanders of Toms Brook (fig. 33) are recorded on the Strasburg quadrangle. Instances of this kind could be multiplied in great number; there is no doubt that stream meanders are more common on streams of modest size than the maps show. That is to say, manifestly underfit streams are far more numerous in the Shenandoah River basin than can be demonstrated without the aid of field inspection or of large-scale aerial photographs.

Although few data of wavelength of present meanders have been collected, it is clear from the observations made that the large incised bends belong to the local family of valley meanders. At the mark of 100 square miles, the valley bends seem to be about five times as long as present meanders—that is, the relevant members of the Shenandoah system have been affected in similar proportion to streams in certain other regions. In this way, the immediate problem reduces itself to one of explaining why the North and South Forks fail to possess stream meanders for much of their length.

For about a mile upstream from its confluence with the North Fork, Toms Brook does not now meander. As with other laterals of comparable size, the lowest reach has been quite strongly rejuvenated. The channel is cut in bedrock; single resistant beds crop out as bars in the channel, which is shallow in proportion to its width at medium-high stages and which, if not regular in cross section, is at least patternless. One and one-half miles above the confluence, however, the valley floor is lined with alluvium, wherein stream meanders are developed. One mile upstream again, the Toms Brook falls over a group of resistant beds; but half a mile above the fall the valley floor widens for the second time in a strip of alluvium, and stream meanders occur. As valley bends complete with valley-meander scars typify all this part of the valley, Toms Brook combines a complete train of valley meanders with discontinuous trains of present meanders. The interpretation is not that the valley bends are ordinary meanders enlarged under the control of bedrock but that Toms Brook cannot develop, or has not yet had time to develop, present meanders where its channel is formed not in alluvium but in solid rock in place.

That part of the South Fork represented on the Strasburg quadrangle is broken by numerous riffles. The stream crosses and recrosses outcrop boundaries, or single resistant beds, and is in contact with bedrock along the whole base of its channel. This channel does not meander. However, the river is by no means everywhere in contact with bedrock at the channel side; it does not press vigorously against the valley-meander scars as it presumably did when these were being eroded. The North Fork, similarly, although incised into the Martinsburg Shale of Ordovician age, reaches limestone in places. Not all the resulting shallows and bars appear on the topographic map. On the upstream side of Rittenour Ridge, the spur 8 miles southwest of Strasburg, only one set of rapids is marked; but in actuality a second bar, prominent enough to reduce the water depth to some 3 feet at normal high spring stage and to make the water surface choppy, occurs half a mile above. Like the South Fork, the North Fork appears to have retreated from a number of its spurs.

On both forks, the present stream width is disproportionately small in relation to wavelengths of the valley bends; sample measurements give the ratio $L:w$ as 47 on the South Fork and 48 on the North Fork. This second result is at variance with the findings of Hack and Young (1959) but accords with observations on numerous other streams of similar type. On the principle that ratio of length and width should normally be about 10:1, the valley bends of the Shenandoah seem to be some 4.75 times too large for the channel, a value close to the approximate 5:1 reduction of wavelength on manifestly underfit members of this river system.

OZARKS AND SALT AND CUIVRE RIVER BASINS, MISSOURI

The northeast Ozarks exemplify meandering valleys with bends distorted in many places and present streams on which meanders are unusual but not absent. The Osage River seems capable of representing the type of stream which, although not now meandering, is enclosed in valley bends and possesses a well-defined sequence of pools and riffles.

On the Meramec River, just east of Pacific (St. Louis County, Mo., Pacific quadrangle, 1:24,000), occurs a fine cutoff valley bend, with its core rising about 170 feet above the flood plain (fig. 34). Meanders of the present channel, supplemented by the recent cutoff traced by the county boundary and by the abandoned scars and channels reflected in the contours, make clear the disparity of wavelength between valley and stream. The 10-foot contours on the topographic sheet permit a likely measurement of bed width between bank tops at a generous figure, 400 feet. As the mean wavelength of valley meanders on this reach of the Meramec is about 11,500 feet, a stream of the present size could not have been responsible for the large bends unless it pos-

FIGURE 36.—Ground view of the Gasconade River in flood, May 1961.

On the Gasconade River near Rich Fountain (Washington County, Mo., Linn quadrangle, 1:62,500) is a cutoff valley bend with an exceptionally bulky core (fig. 37). Although a certain ingrowth is indicated by the contours, the steep sides of the core testify to downcutting rather than to lateral enlargement during the period recorded by existing landforms. The markedly irregular valley meanders of this whole area suggest, however, that the influence of bedrock has been strong. Meanders on the present stream are little if at all developed on this reach of the Gasconade but can be identified about 5 miles upstream.

Near their confluence in Jackson County, Mo. (Florida quadrangle, 1:62,500) the Middle and Elk Forks of the Salt River possess stream meanders well-enough developed to permit reliable averaging of waveleng exceptionally for this region, a continuous train of f meanders occurs on Elk Fork (fig. 38). Immediat upstream from the confluence, Salt Fork has w drawn from the outer curve of a large left-hand be which appears to be devoid of core. A similar lies 3½ miles downvalley. At both sites, the ri seems to have undergone rapid lateral developm after it was already well incised; this is in direct c trast to the behavior of the Gasconade near R Fountain.

The Bourbeuse and the Meramec, at short distar above their confluence 1 mile northeast of Moselle (C tral County, Mo.; St. Clair quadrangle, 1:62,5(possess highly distorted valley bends (figs. 39, 4

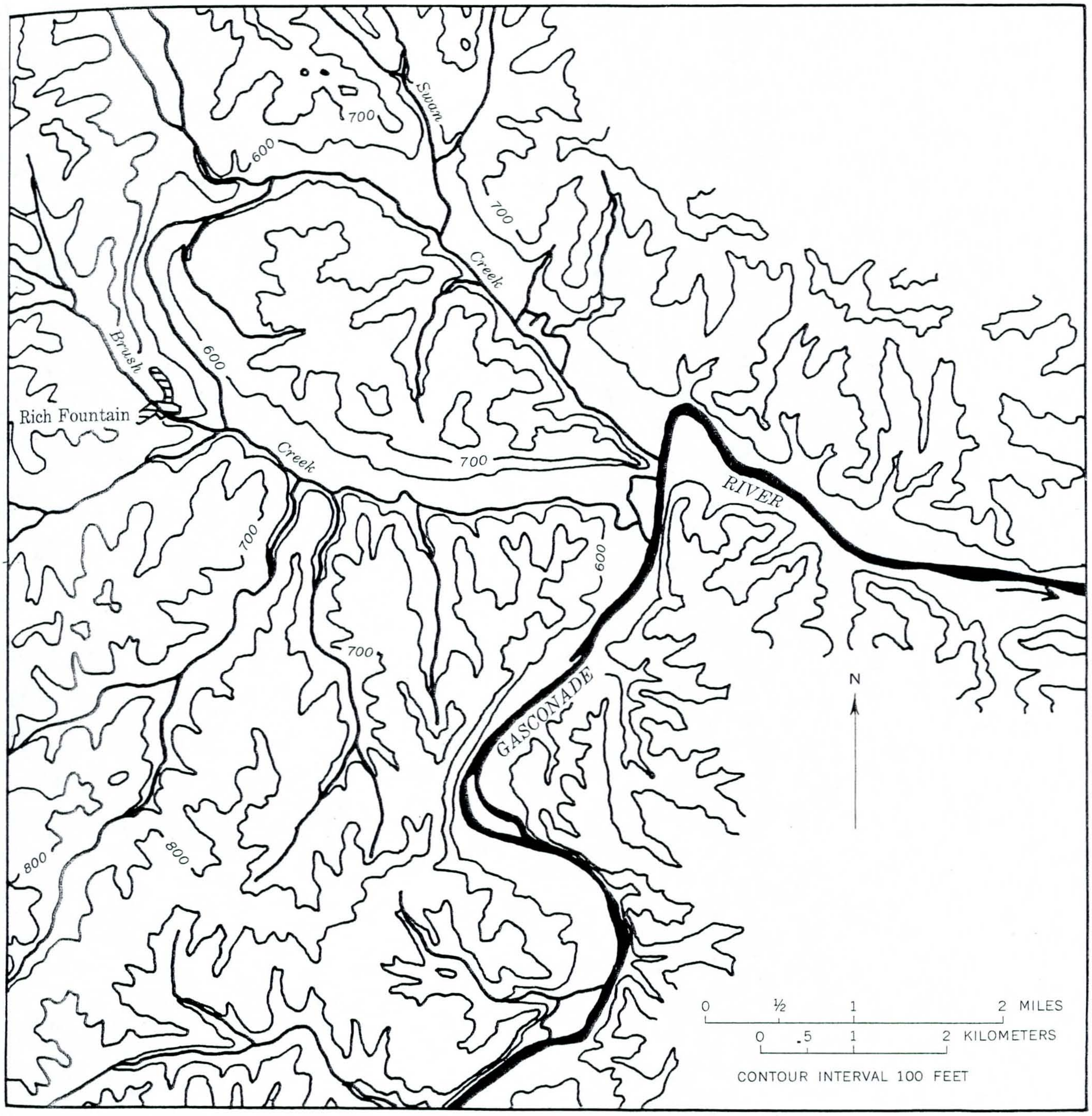


FIGURE 37.—Map of the Gasconade River south of Linn, Mo., showing cutoff valley bend.

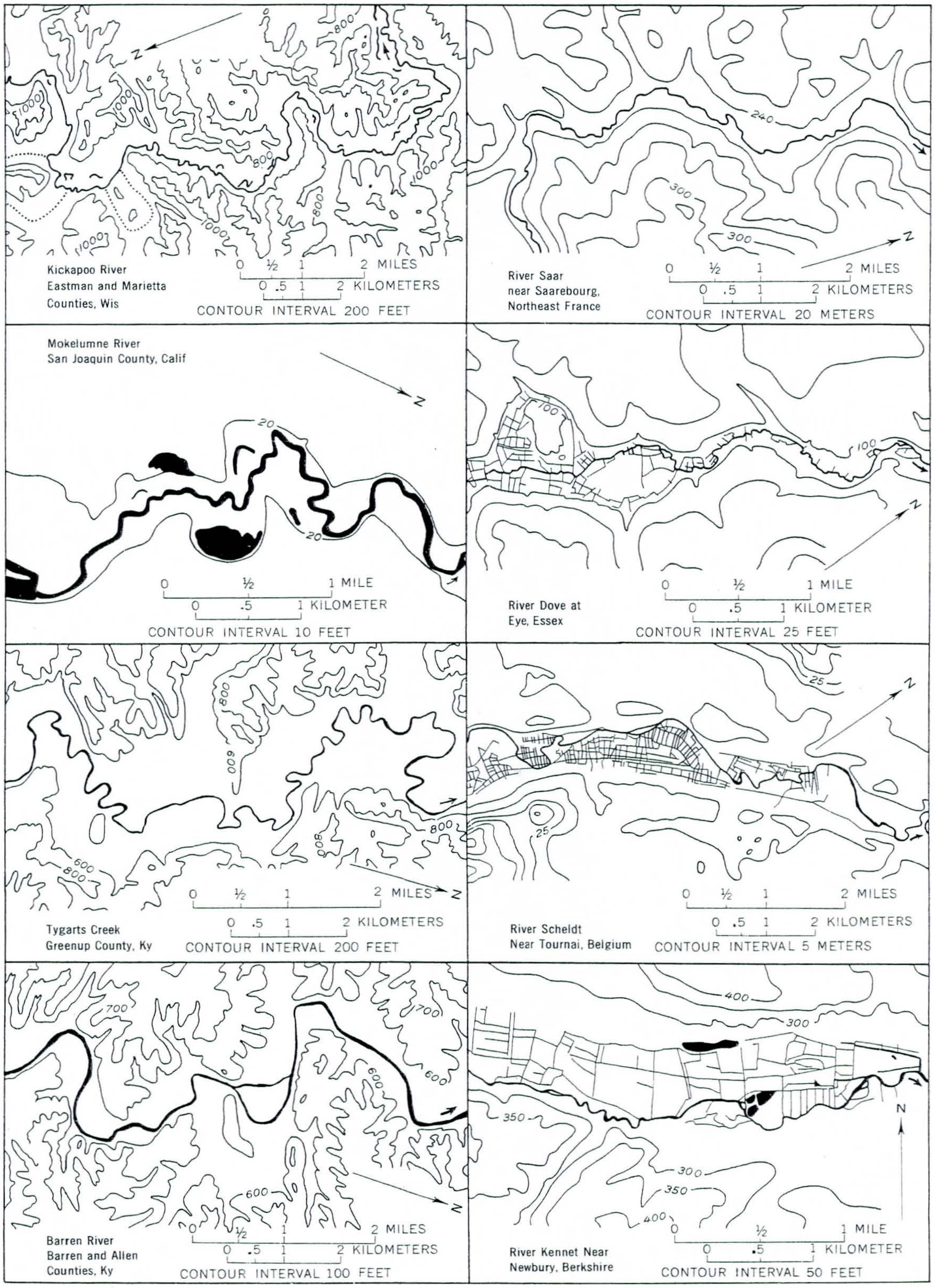


FIGURE 42.—Maps showing ranges of stream-channel and valley patterns on underfit streams. The Kickapoo and Saar Rivers, in the reaches shown, are manifestly underfit in incised meandering valleys. The Mokelumne River is manifestly underfit, although shallowly incised; and the valley meanders of Tygarts Creek have shifted considerably downstream, although through less than one wavelength. Stream meanders are very poorly developed on the Barren River. The present stream-channel pattern on the River Dove is partly obscured by numerous artificial ditches, as is that on the River Scheldt. Large bends of former meanders are identifiable on the Scheldt even though the stream occupies a former large meander trough. No trace of former meanders remains on the Kennet River, but the former large channel is proved here by excavations and boreholes.

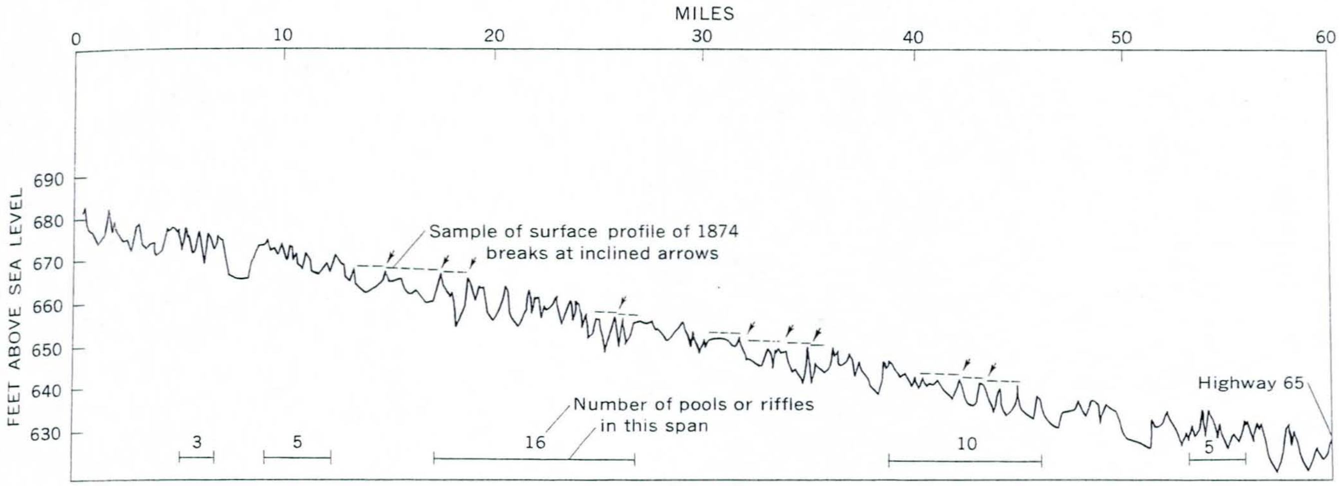


FIGURE 43.—Channel-bed profile of a reach of the Osage River, Mo., in 1874.

ıot a meandering one (Bagnell quadrangle, Missouri, [:24,000; Shawnee Bend quadrangle, Missouri, .:24,000). Furthermore, Deer Creek and Niangua River, which enter the lake from the south, are mani-:estly underfit in scarcely any reaches (Edwards quad-·angle, Missouri, 1:24,000; Macks Creek quadrangle, Missouri, 1:24,000). It seems most unlikely that the ;urveyed reach possessed trains of stream meanders be-:ore the lake filled.

The pools and riffles are far more closely spaced than ,re the bends and inflections of the valley. How much nore closely depends on how many of the peaks on the urveyed profile are classed as individual riffles and on ıow many of the troughs are classed as pools. The tally ·f hydraulic jumps in the low-water profile of 1874 is 0 for the 60-mile length, giving an average spacing ·f 1.2 miles from jump to jump. This value is, however, oo great to represent a half a wavelength, for a num-·er of low peaks were not reflected by jumps. When lose-set groups of small peaks are counted as single ›eaks, the total becomes 75, identical with that for roughs. The spacing of 0.8 mile indicates a wavelength f 1.6 miles. This is already less than half the wave-ɜngth of valley meanders, but it is probably too large ecause of the irregularities in the sequence observed, 7hich suggests imperfect development. When readings re taken on distinctive parts of the sequence (bottom f fig. 43), they give an average of 0.56 mile for 39 ıtervals, corresponding to a full wavelength of 1.12 ıiles. The wavelength of the locally distorted valley ıeanders of the Osage is measurable with difficulty but ppears to be some 3.8 miles—nearly 3½ times as great s the wavelength for the streambed. The ratio of .5:1 is within the range of regional value for meander 'avelength in the northeast Ozarks. The Osage, therefore, like the rivers of the northeast Ozarks, supports the thesis that a stream in a meandering valley need not be a meandering stream.

RIVERS IN NEW ENGLAND

Where a meandering stream in a nonmeandering valley passes into a reach of incised meandering valley where stream meanders do not occur, and where also it begins to meander again on the far side of the incised reach, the distinction between the two sets of windings cannot be gainsaid (fig. 42). But in New England, the winding valleys which dissect the upland are in many parts angular in plan, whereas stream meanders are largely confined to the irregular and drift-encumbered low ground of the coastal belt. Manifestly underfit streams can be identified in few places. Nevertheless, when wavelengths are averaged on trains of valley windings in the uplands, they plot against the drainage area in the usual fashion (fig. 45). Wavelengths of stream meanders constitute a second family, whether they are determined for lowland drainage areas or for the rare sinuosities of the channels of upland streams. Both the absolute dimensions of valley bends and their relation to the wavelengths of indubitable stream meanders show that underfit rivers occur also in New England.

Some 8 miles north of North Adams, Mass., the Deerfield River occupies a magnificently winding valley that is cut as much as 1,000 feet below the levels of nearby summits (fig. 46). The stream channel has but a slight tendency to wind, even though the valley floor is wide enough in places to accommodate stream meanders. Farther downstream, near the confluence with the trunk Connecticut, stream meanders appear in weak glacial and fluvial sediments (fig. 47) where valley

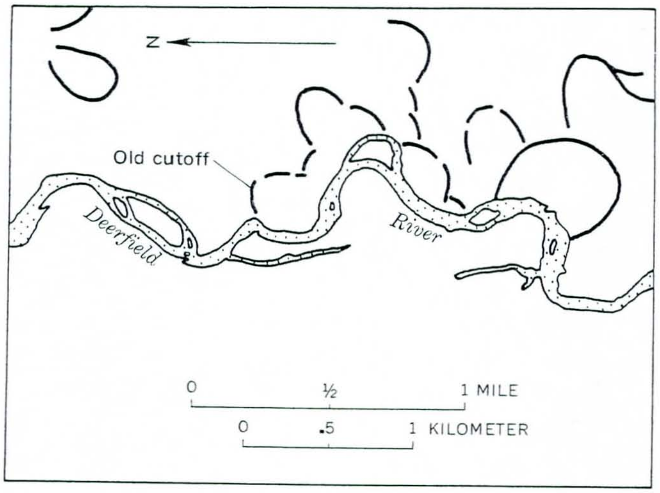


FIGURE 47.—Sketch of the Deerfield River, Mass., showing stream meanders on a lowland reach.

meanders are absent. The valley bends of the upstream reach have a mean wavelength of 6,600 feet at 300 square miles, whereas somewhat dubious stream meanders average 2,300 feet (table 3). In the downstream reach, at 560 square miles, stream meanders average 2,025 feet in wavelength. In this humid region, where bankfull discharge cannot fail to increase with increas-

TABLE 3.—*Wavelengths for rivers in southern New England*

River	Drainage area, in square miles	Valley meanders: Number of meanders	Valley meanders: Mean wavelength, in feet	Stream meanders: Number of meanders	Stream meanders: Mean wavelength, in feet
Housatonic	57	2	3, 425		
Do	132	4	4, 225		
Do	517	4	8, 050		
Eightmile Creek (to Housatonic)	21	4	1, 805		
Pomperaug	74			7	750
Hoosic	71			8	640
Pond Brook [1]	9. 6	2	3, 400		
White, Vt	101	4	4, 200		
Westfield	115	5	3, 060		
Do	160	3	4, 500		
Do	328	2	8, 350	2	2, 310
West	300	5	5, 140		
Deerfield	257	2	6, 600	3	(?)2, 305
Do	500	2	7, 900		
Do	562. 5			4	(?)2, 025
South (to Deerfield)	22. 5	4	960		
Bear (to Deerfield)	13	3	870		
Cold (to Deerfield)	16	3	1, 060		
Green (to Deerfield)	66			5	815
Connecticut	5, 400			3	8, 400
Do	10, 624			4	9, 750
Do	11, 600	2	30, 600		
Mill (to Connecticut)	53	2	3, 400	6	600
Fort (to Connecticut)	50	3	1, 625	5	400
Bachelor Brook (to Connecticut)	6. 5	2	1, 070	6	210

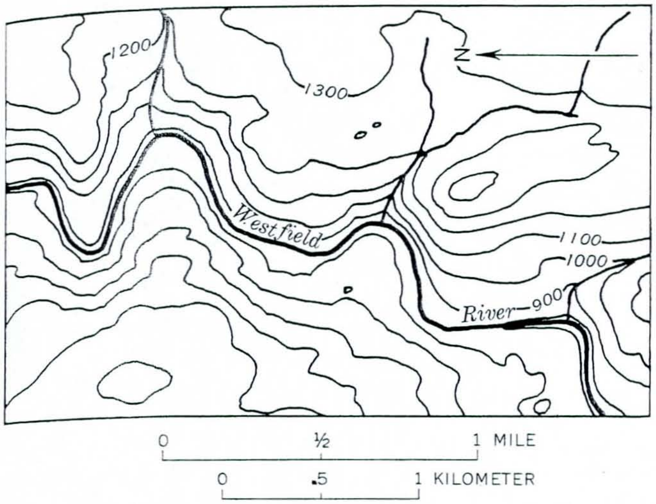
[1] See text discussion.

ing drainage area, the wavelength of valley mean appropriate to a drainage area of 560 square mil likely to be much greater than 6,600 feet; by extrap tion, therefore, the disparity between the two seri wavelengths is considerable.

Similarly, the Westfield River upstream from well-known terraces near the town of Westfield cu inside the great sweeping recesses of an incised va with a mean wavelength of 8,350 feet at 430 square n (fig. 45). Scanty readings on the stream channel a value of 2,310 feet for stream meanders here. ther upstream, about 15 miles northwest of Northa ton, Mass., the valley of the Westfield River sin leaves no room for stream meanders (fig. 48). the appearance of regional values in the regional gr leaves no doubt that valley meanders and stream m ders are separable from one another. The winding the New England valleys are strictly comparabl those of the Ozarks, despite their angularity.

Davis (1902a) chose to present the terraces of Westfield and others rivers as evidence against re tion in stream volume. Two points arise immedia First, Davis seems to have directed his main at against the changes of volume postulated by Eme (1898), who was discussing discharge of melt w rather than changes produced by climatic change sequent to the recession of ice; and second, eve Davis proved correct in maintaining that no signific change in volume had occurred since the Westfield R first began to cut into its topmost terrace, he would necessarily confute the general reduction in vol which is here claimed to have occurred since the me dering valleys were cut through bedrock. Furtherm Davis is open to challenge on his own ground, as now be shown.

Davis advocated the defense of terraces by outc ping bedrock, as opposed to reduction in volume, red tion in load, or increase in slope. His general con sion, that the several arrays of terrace fronts meander scars, is not disputed, although he is ope correction on points of detail. For example, his pers tive diagram of the terraces at Westfield (Davis, 19 fig. 82) omits several scars, as may readily be seen fr the aerial photographs now available or from ins tion of wooded parts of the terrace fronts. He m have mistaken tiny remnants of low terraces for slum masses, for he stated (p. 91) that recently abando scarps are uneven with landslides. If this were so, oldest (uppermost) scarps should be particularly even, as they have had the longest time to yield slumping; in actuality, they are nearly everywh smooth and unbroken. Again, Davis seems to h overstressed the role of bedrock in defending surviv



IGURE 48.—Map of the Westfield River, Mass., showing valley bends on an upland reach.

arts of terrace (Jahns and Willard, 1942, p. 283). Perhaps he was a little too greatly influenced by the rinciple that "observation is greatly aided by the discovery of a successful theory; for the essential facts are hen quickly acquired by well-directed research" (Davis, 902a, p. 93).

Davis' comments upon the hypothetical effects of increasing slope (1902b, p. 290–293) seem to refer mainly o Emerson's conclusion (1898) that the lake clays of he Connecticut Valley were deposited close to the sea evel of the time, so that their present altitude indicates rustal warping. Although little is yet known of the letails of late-glacial and postglacial movements of the trandline in southern New England, warp ng has certainly occurred. J. E. Upson's studies (written and ral communications, 1960–61) of bedrock valleys, Flint's assemblage (1957, fig. 14–16) of evidence for rustal movement, and tilting of the lake floor in the Connecticut Valley (Jahns and Willard, 1942 p. 272–274) show that south-flowing streams have been submerged at their mouths and uplifted in the north. Whatever the interplay of crustal movement with ustatic rise in sea level, rivers debouching along the outhern coast of New England have undoubtedly had heir downstream slopes increasingly steepened since leglaciation. Consequently, Davis' view that uplift is nconsistent with the formation and preservation of whole flights of terraces must be rejected, the more so as t relies on the elusive concept of grade—to be specific, n the indefensible notion that a meandering habit is in ome way associated with the attaining of grade, whatever the state of grade may be.

Davis' brief treatment of diminishing load (1902b,). 293–294) is little more than guesswork. In fact, according to Davis' own theory of grade, load, and therefore slope, should still be increasing on the New England rivers today as the various stream nets become progressively better organized and the laterals—which Davis regards as developing tardily by comparison with the trunk streams—come to feed increasing bulks of sediment into the main rivers.

On the subject of diminishing volume, Davis wrote (1902b, p. 288),

The best indication of the volume of the stream by which a terrace has been carved is afforded by the curvature of its frontal scarp. If the scarps of the low-level terraces have a radius and an arc of curvature similar to these elements in the existing river meanders, and significantly smaller than in the high-level scarps, while curves at intermediate levels show intermediate values, a diminution of stream volume may be fairly inferred. If the radius and arc of curvature are of about the same measure in the three cases, no change in stream volume is indicated * * * [but] a graded river on a strong slope does not develop curves of as small radius as it would when subsequently flowing with the same volume but with a finer load on a gentler slope; hence a large radius of curvature in the uppermost terraces should not alone be taken as an indication of large volume; large arc of curvature should also be found before large volume is inferred.

These comments seem to go too far in some directions and not far enough in others. Davis is clearly challenging the hypothesis of a progressive decrease in volume—at unspecified but presumably constant stage—in the context of the progressive downward narrowing of the remaining spreads of terrace. But, as is repeatedly observed, the conversion from large to small meanders which makes rivers manifestly underfit seems to occur swiftly, leaving no time for the production of scars of intermediate size. Absence of intermediate forms is thus not relevant to the present discussion. The question is simply whether or not large scars exist at high levels. However, although the reference to radius of curvature accords with the modern views of the behavior of meandering streams, Davis seems to make no allowance for the possible effects of downstream sweep in elongating particular scars or for the usual hypertrophy of single loops in weak proglacial sediments. As the curves in his diagrams do not appear to have been determined by instrumental survey, and as Davis fails to state what he means by significantly smaller, he is not equipped to decide whether or not there is a difference in the size of scars at high and at low levels. If, moreover, the coarse valley trains were deposited or partly worked over by braided streams, the relation which Davis postulates, between coarse and bulky load on the one hand and strong downstream slope on the other, ceases to be relevant to the behavior of the meandering streams which carved the terraces. In any event, it is not permissible to compare the present downstream slope of the outwash with the present downstream slope

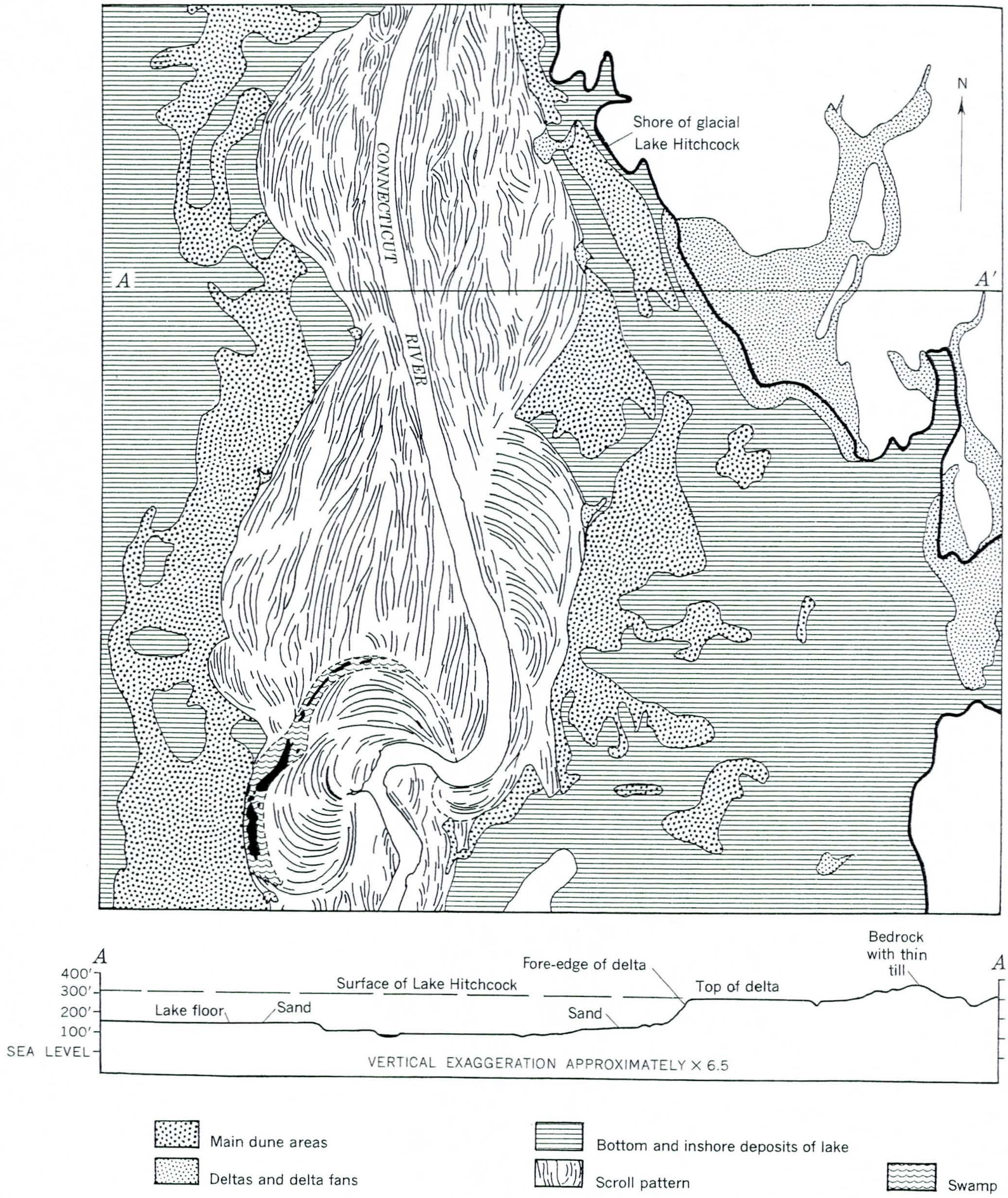


FIGURE 49.—Map of the Connecticut River valley near Amherst, Mass., showing scrolls. Simplified from Mount Toby quadrangle, scale 1 : 31,680.

the south-flowing rivers; allowance should be made r crustal warping, which cannot fail to have altered e profiles of the outwash trains. In this same connecn, Davis' claim that a steeply descending graded ver with a heavy load develops curves of unusually ge radius can also be recognized as irrelevant. The stribution of old scrolls on the Connecticut River Mount Toby quadrangle, Massachusetts, 1:31,680) [5] ggests that, at times and in some reaches, the Concticut River has been a braided stream, laying down olls more or less parallel to the axis of the valley but the same time cutting long arcs into the terrace fronts g. 49). Here also the criteria suggested by Davis are t conclusive enough for use in this area.

Although Davis did not cite the authors of the hythesis that the volume of New England rivers was ce greater than it now is, his main objection seems to directed at Emerson. Emerson (1898) made free d sometimes incautious use of the term "flood"; in dition, he was mistaken in regarding erosional teres as usually paired across the valleys (1898, pl. 25, ets A–D), as is readily seen from the results of inumental leveling by Fisher (1906). But Emerson ly comprehended the nature and origin of the valley s. Apart from lake sediments and deltas, these nsist almost wholly of outwash trains and kames, npled in many places (as Emerson observed) by tles. The valley fills were laid down at the time en local glaciers were in an advanced state of decay, d when streams of melt water carried outwash und, and over, detached blocks of ice. The melt ter streams, however effective in smoothing the tops the outwash trains, were temporary phenomena, evalent to the question of whether or not postglacial andering streams have undergone a reduction of ume.

To summarize, Davis' views on the likely effects of nges in slope and load of New England rivers are in t inconsistent with his own hypothesis of grade; his teria for separating large from small meander scars insufficiently rigorous; and, in discussing possible uction in volume, he seems to have been challenging limited concept of melt-water discharge.

t would be unreasonable to expect Davis to refer conclusions to scales of chronology comparable to se now in use, but this circumstance should not pret correction of his inferences in the light of modern lings. On the other hand, Davis can justly be critied for omitting reference to meandering valleys in rock, such as are readily visible from viewpoints overlooking the New England Peneplain. As stated, these valleys have wavelengths which relate them to the family of valley meanders. The streams in them are underfit.

The streams which cut the valley bends removed a great bulk of bedrock. Postglacial streams, operating since the last deglaciation, have not yet succeeded in removing the valley fills of unconsolidated outwash, deltaic beds, and lake sediments. The valley bends are referable chiefly to a period earlier than, and much longer than, postglacial time. As they were overridden and occupied by ice, and as they are fluvial, not glacial, features, they must be regarded as produced mainly before the last local glaciation—that is, as having a long history comparable to that of the valley meanders of the Ozarks, the Driftless Area of Wisconsin, and the English Cotswolds. This is not to say, however, that large streams capable of assuming wavelengths of the valley-bend size did not reestablish themselves in postglacial New England; this matter will be discussed presently.

The regional graphs of wavelength and area can be used to classify some of the trains of bends on the Connecticut River. The bends near Hartford and North Walpole are stream meanders, whereas those cut through bedrock near the estuary are valley meanders, even though their wavelength runs a little short by comparison to wavelengths from smaller drainage areas. The graphs also enable the anomalously large windings of Pond Brook to be interpreted. Pond Brook enters the Housatonic River 8.5 miles northeast of Danbury, Conn., occupying a valley with the usual distorted windings but also with recessed curves on the outside bends which are unmistakable on the ground. The wavelength of these bends—more than 3,000 feet near the confluence—is, however, excessive by regional standards for the present drainage area of about 10 square miles (fig. 50). As the bends cut down 300 feet into bedrock. they relate to erosion before the last local glaciation.

This may well be an instance of capture in the normal cycle (Harvey, 1920). If capture has occurred, the captor stream is Still River, which now flows northward across the head of Pond Brook valley to join the Housatonic near New Milford (fig. 50). If Pond Brook before capture drained the complex depression which now includes Lake Candelwood, its former drainage basin was about 100 square miles, a value which would locate the point graphed for wavelength and area almost exactly on the best-fit line for the region (fig. 45).

Although underfit streams can be recognized within the winding valleys of New England, the question of when the last general shrinkage occurred in this region remains open. All that has been established so far is

[5] Although the scrolls are described in the key as being shown diagrammatically, recent aerial photographs at 1 : 7,200 confirm the accuracy of the trends represented.

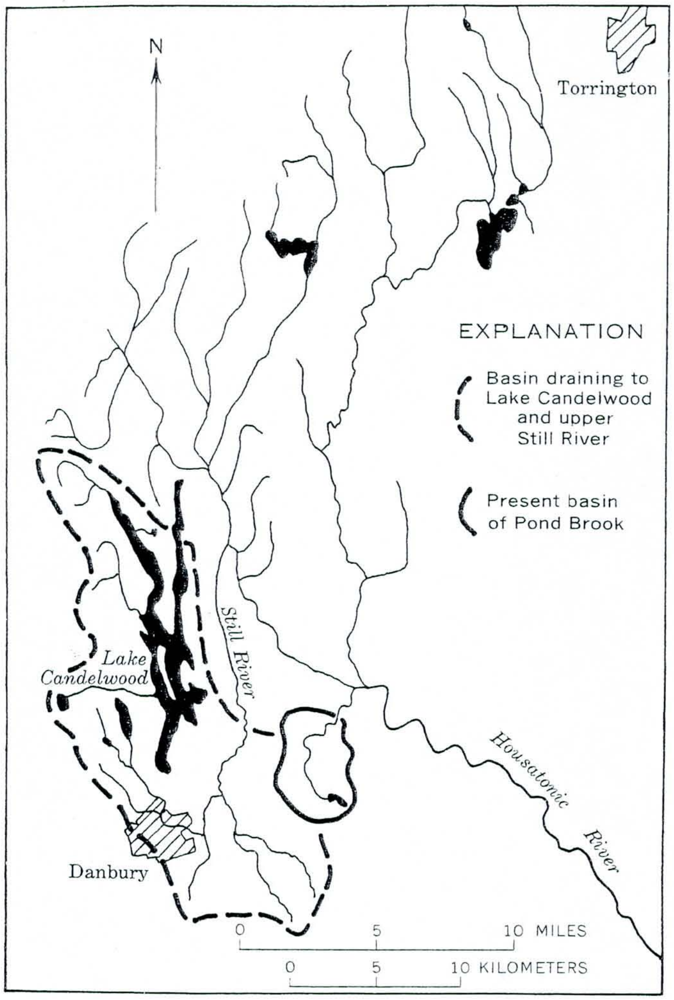


FIGURE 50.—Sketch map showing the possible capture of Pond Brook, Conn.

that the valley meanders were shaped almost wholly before the last local glaciation: postglacial erosion, apart from some clearance of surficial material, has been slight. Postglacial dates for large bends are demonstrable only if these bends are cut into glacial or proglacial deposits, and the dates bear on the argument only if they were cut by ordinary streams and not by melt water or spill water.

Relative dates of erosional features can be obtained, with some precision, from the sequence of surficial deposits. Difficulties in the low ground and in upland valleys open enough for stream meanders arise less in dating than in identification. Kettles generally, and kettles in kame terraces in particular, can produce steep slopes of curved plan like those of scallops due to meanders. Problems of identification are well illustrated by the valley of the lower Pomperaug River (Newton and Southbury quadrangles, Massachusetts, 1:31,680 and 1:24,000). Five miles above its confluence wit Housatonic, the Pomperaug is eroding kettle-pitte wash on its right bank. A recess in the outwash edge of the flood plain resembles in diameter a nı of kettles in the top of the kame and could itsel: kettle even though its enclosing slope is steep (fig In actuality, auger holes and inspection pits shov a former stream channel, now filled with humified wash, curves round the outside of the recess, wh therefore identifiable as a scallop cut by a mea Three miles downstream, immediately north o abrupt bend of the Pomperaug at South Brit: swampy recess on the right bank is neither a kett the cut of a meander but the site of a mill res that dried out when the enclosing dam failed. I second instance, the recess shown by contour lines topographic map is large enough to be a valley ı der; but independently of local reports on the hist the site, augering proves a flat surface of sand be a thin cover of humified topsoil across the whole of the recess, which lacks either the peat-filled that might be expected from a kettle or the large nel that might occur in an abandoned valley mea The two examples illustrate the need for extreme tion in the interpretation of topographic maps of England.

Aerial photographs, however, supply a check. confirm that some laterals of the Connecticut Ri the general neighborhood of the Holyoke Rang manifestly underfit (fig. 52). On the left bank, festly underfit reaches occur on the Scantic Rive dunk River, Fort River, and Bachelor Brook sys on the right bank, they occur on feeders of the han River. Where comparative wavelengths c measured, a ratio of about 5:1 is apparent, as i regional graph for small drainage areas. There is ing to suggest that the forms identified as valley ı ders are due either to the discharge of melt water the formation of kettles. Because numbers of the scallops are cut into the bottom sediments of g Lake Hitchcock, their origin must be set later tha draining of that lake. And although kettles lie to some trains, it would be a remarkable coinci indeed if kettles formed continuous winding train if the wavelengths of these trains were accidental same as the wavelength appropriate to valley mea The reaches shown in figure 52 are, then, claim display authentic valley meanders. Numbers of scallops elsewhere seem probably to be valley mea but the effects of rapid downstream shift and th tortion of some loops ensure that in many place valley constitutes a trough bordered with blunted

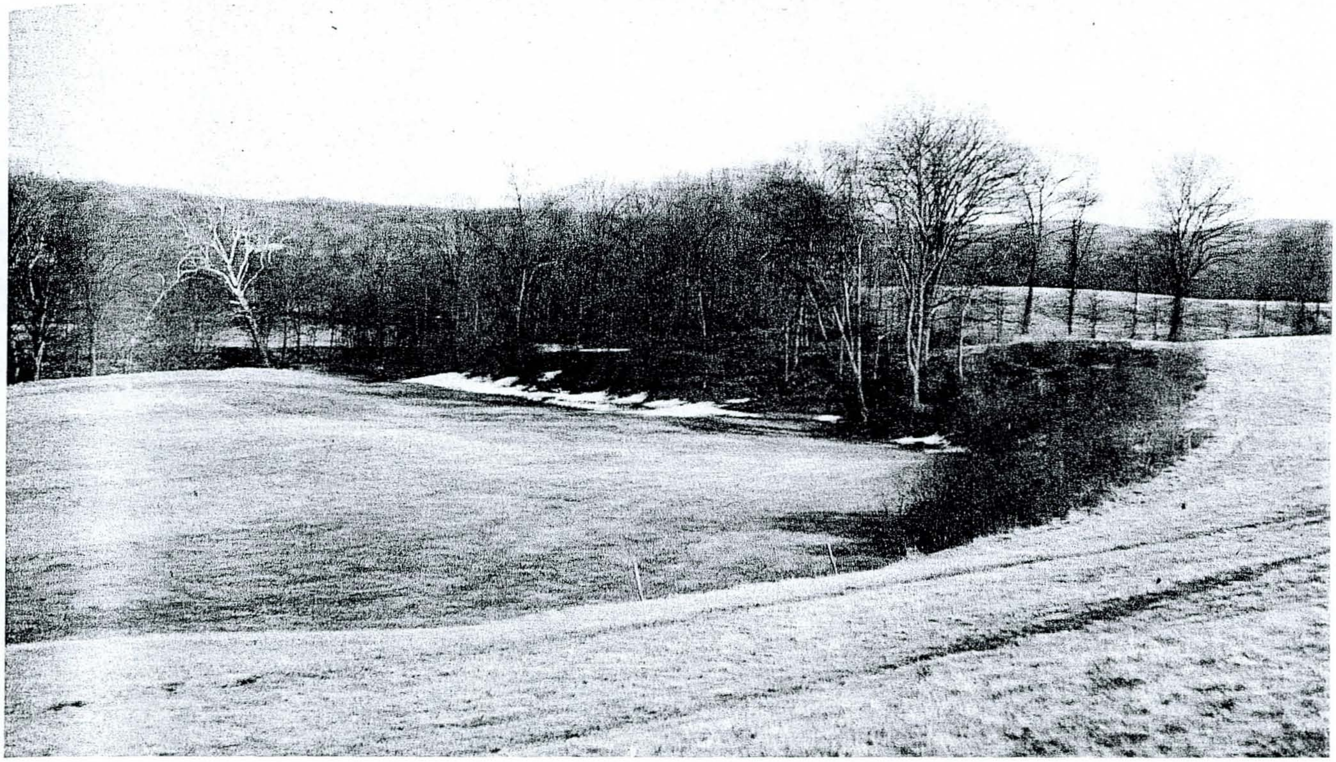


FIGURE 51.—View of scallop of stream meander in kame, Pomperaug River, Conn.

t a winding cut on which wavelength can be reliably ɪasured; in opposition to Davis, no use is here made radius of curvature.

Because some parts of some valleys are straight and ıer parts are widely opened, the evidence of manifest derfitness is fragmentary. It is nevertheless consistt with that obtained elsewhere in suggesting that the ɪt onset of underfitness came late—late enough for ke Hitchcock already to have disappeared. Davis' im that the New England rivers have not shrunken therefore rejected, independently of the reference of ıt claim to the views earlier expressed by Emerson. orly developed though they are, manifestly underfit eams exist. Because similar streams occur in the lands, shrinkage is taken to have been regional. In ite of the difficulties of its terrain, New England is ıs brought within the scope of the general theory of derfit streams.

CANYONS IN ARIZONA

Canyon Padre and Canyon Diablo, respectively ɔssed by U.S. Highway 66 at 21 miles and 33.5 miles ɪt of Flagstaff, Ariz., are tributary to the Little Colo-lo River. At the crossings they are cut into a stripped surface of the Kaibab Limestone of Permian age, which hereabouts consists largely of dolomitic limestone (Childs, 1948; Strahler, 1948). Each canyon describes large bends, with undercut slopes on the outside curves and gentler lobate slopes on the inside curves (figs. 53–56). In the bottom of each valley, large terraces or berms composed of reddish-brown sandy and silty material contrast strongly with the pale much jointed rocks of the Kaibab Limestone; the surficial deposits are presumably derived from the red beds of the Moenkopi Formation, which succeeds the Kaibab upward. The tops of the terraces or berms slope regularly downvalley; the fore-edges are slightly dissected in places by tiny gullies. The present stream channels, which cut into the deposits, are about 15 feet wide and 3 to 4 feet deep; they themselves include faintly developed berms in places. The channel in Canyon Diablo swings a little in some reaches and includes very bulky islets or braids in others.

These assemblages of features are taken to indicate reduction in bankfull discharge. The present channels, like those in the Ozarks, are far narrower than would be expected from the approximate 10:1 ratio of wavelength to bed width, when the wavelength used is that

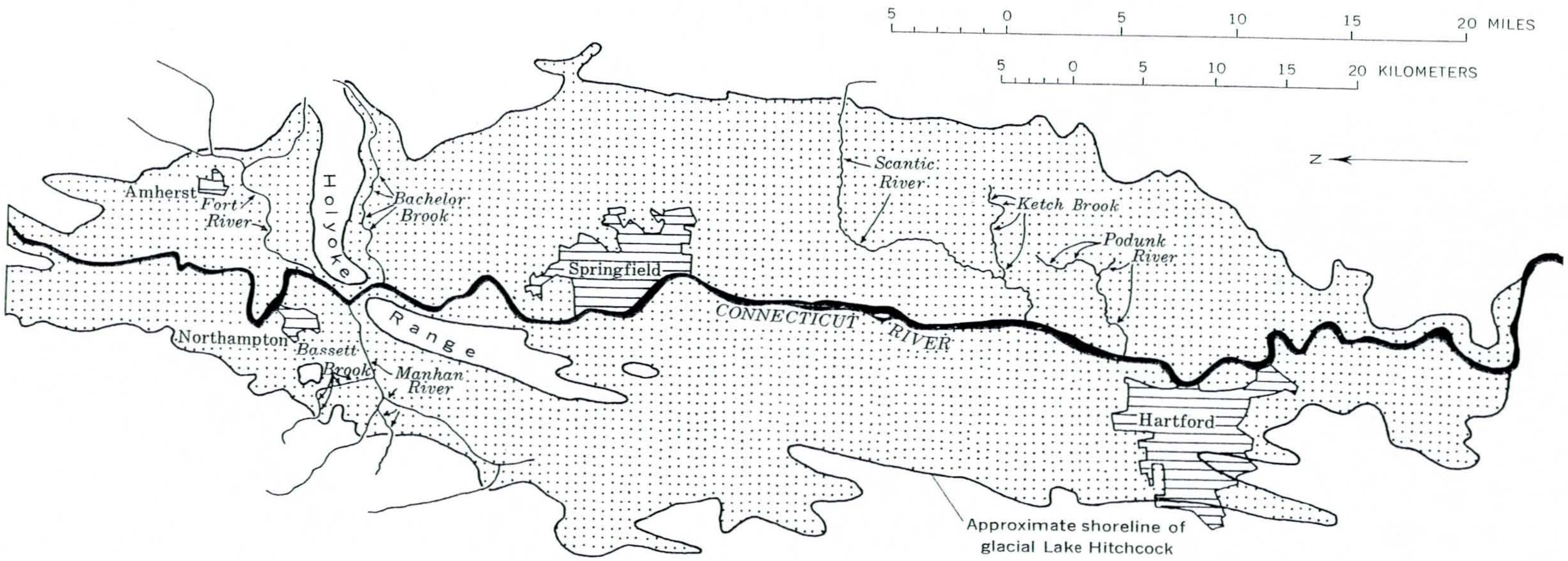


FIGURE 52.—Sketch of the Connecticut River valley showing manifestly underfit streams on the drained floor of glacial Lake Hitchcock. Shoreline drawn after Jahns and Willard (1942) and Cushman (1961) to include deltas and other littoral deposits.

FIGURE 53.—View of valley bend in Canyon Padre, Ariz.

FIGURE 54.—View of berms and stream channel in Canyon Diablo, Ariz.

FIGURE 55.—View of point bar of valley meander in Canyon Diablo, Ariz.

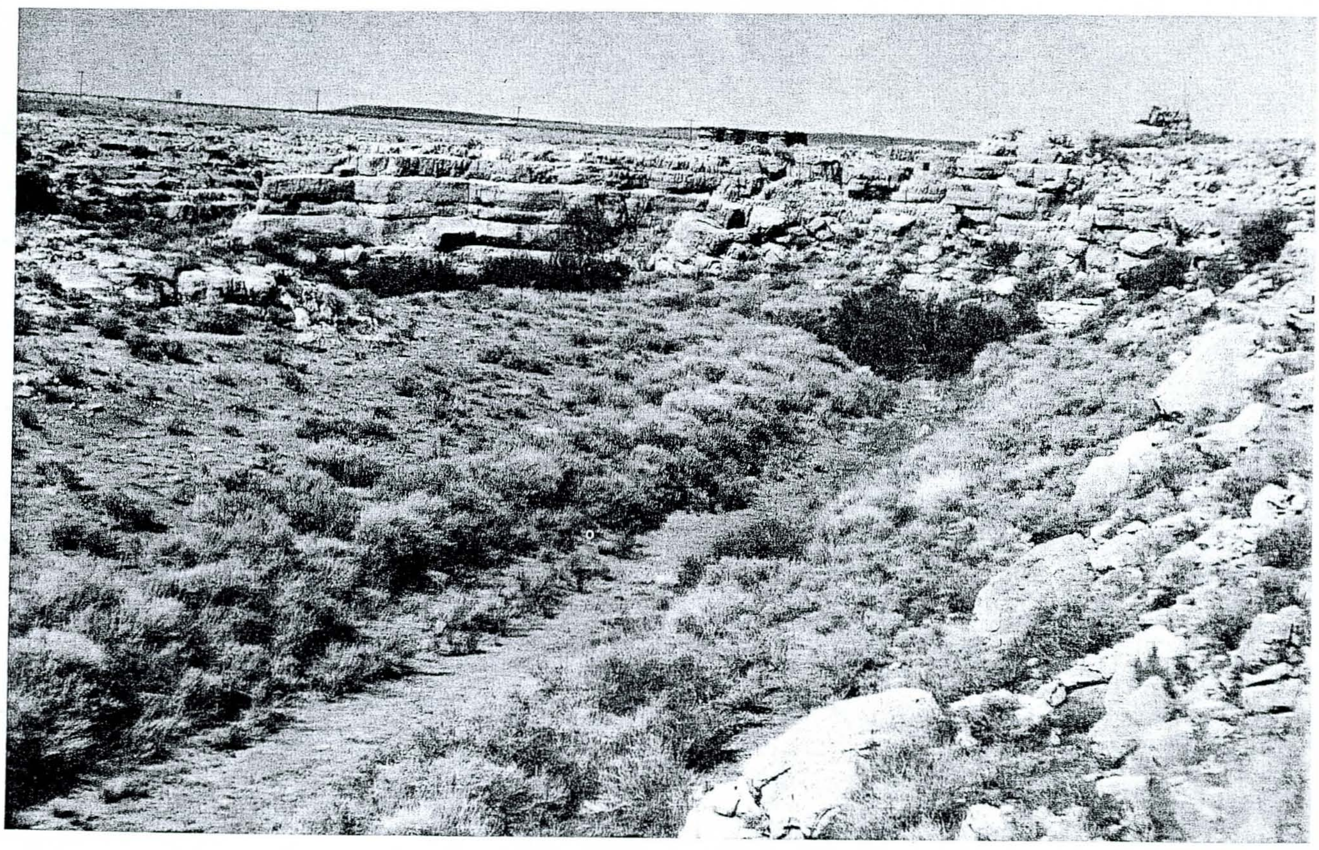
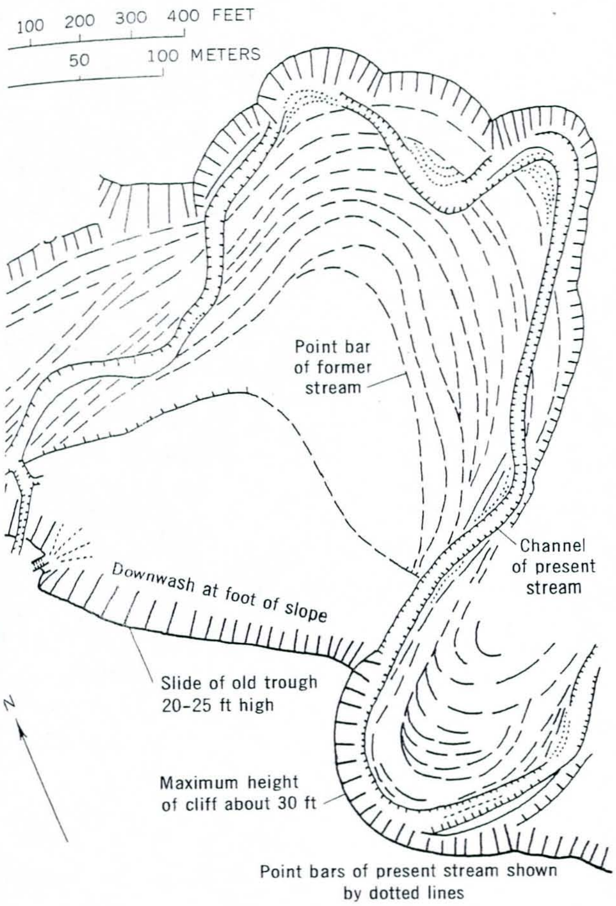


FIGURE 56.—View of point bar of valley bend, undercut outer slope, and shedding of joint blocks in Canyon Diablo, Ariz.



RE 57.—Sketch of Oraibi Wash., Ariz., showing scrolls and scars.

the bends in the canyons. The occasional faint ngs and islets of the present channels reproduce fea- :s noted for the Ozarks. Much of the reddish-brown ment is regarded as point-bar deposits of the former rs. Those narrow accumulations which line parts the outside curves cannot, however, be fully ex- ned until the whole nature and origin of berms are erstood.

he Oraibi Wash [6] in northeastern Arizona drains t of the Black Mesa and Tusayan Washes country ard the Little Colorado. At the crossing of the hway between Moenkopi and Keams Canyon, close Oraibi village, the wash is enclosed in a terraced ley where outcrops of bedrock are visible beneath es on the higher terraces but where the stream chan- appears to be underlain by some 30 feet of fill. The inage area here is about 425 square miles. Upstream, he Black Mesa, the tributaries of the Oraibi Wash in steep rock-walled canyons; downstream, the main valley opens out, and many tributaries fail to reach the trunk river. The Wash is ephemeral, usually flowing only in response to thunderstorms in July and August.

At the highway crossing, the stream is manifestly underfit. Its present meanders are superimposed on a previous trace of considerably bigger loops (figs. 57, 58), cutting in places into former meander scars and elsewhere across the former point bars. The present channel has point bars and berms of its own, but these are mostly distinguishable from the older series into which the present channel is slightly incised. As figure 57 shows, the large bends are contained in a large meander trough, not so wide that its walls are free from scalloping. Inspection of the ground accords with the record of aerial photographs to indicate that the large meanders swept downstream at higher levels but underwent ingrowth before their last abandonment. The older parts of the trough wall are now being dissected by short gullies, some of which discharge fans at their lower ends.

As Hack (1942) has shown, this whole area has experienced alternate cutting and filling of valleys such as is widely known from dry regions in the West. The site description presented here is not meant to prejudice either the chronologies obtained by Hack or by other writers or to suggest that the last conversion of the Oraibi Wash to an underfit condition was contemporary with the corresponding change affecting rivers in other regions; the dry West may offer problems which do not recur elsewhere. This is a topic to be pursued later. All that is intended at this juncture is to offer the present condition of the Oraibi Wash as essentially similar to the condition of many rivers in humid regions.

PROBLEMS OF THE INFLUENCE OF BEDROCK

As will be described later, a number of manifestly underfit streams are underlain by large channels which, winding round the valley bends, contain alluvial fills of which the present flood plains constitute the topmost parts. Streams of this kind are illustrated in generalized fashion in figure 4 at no. 1, where the large channel is marked. A similar channel can occur beneath streams which combine two sets of meanders of contrasting order of size, even though neither set is incised, as in figure 4 at no. 3. Underfit streams of either type make no contact with bedrock even at times of maximum scour, unless they impinge on the valley wall. Their meanders are alluvial meanders in the fullest possible sense.

Incised rivers of the kind exemplified by the middle Humboldt in its canyons, by much of the upper Shenan-

[6] This account of the setting is based closely upon information kindly nunicated by Richard F. Hadley, of the U.S. Geological Survey :D) Denver office, who initially suggested the site here described as y to be instructive.

FIGURE 58.—Aerial view of Oraibi Wash, Ariz.

ıah, by long reaches of the Ozark streams, by the New ıgland rivers in most upland stretches, and by Padre d Diablo Canyons—all of which possess few or no ·eam meanders—set the question of whether or not e meandering of present-day channels is inhibited by ntact of the streams with bedrock. As has been served, small feeders of the Shenandoah meander ross alluvial fills but fail to meander where they cross drock. Hack and Young (1959) take a contrasted ›w that meanders (valley meanders) on the trunk ›rth Fork are caused by structures in the Martinsburg .ale; but although structures appear to be responsible : greatly enlarged amplitude, they can scarcely ex- ıin a wavelength and area relation which is appro- iate to the region as a whole.

Outcrops of bedrock are known to occur in the bed of › Humboldt in more than one canyon, as also in cer- n rivers in the Ozarks. Logs of boreholes at high- y crossings suggest that, generally speaking, bed- ·k commonly lies close below the present-day beds of › Ozark streams,[7] whether or not it is actually exposed them. The contrast between meandering valleys and andering streams in New England corresponds to ı contrast between uplands carved in bedrock and ʼlands extensively mantled by surficial material. amples could be multiplied by reference to the Her- ıian massifs of Europe, to parts of the Appalachians litional to the Shenandoah Basin, and also to parts the Piedmont. The Meuse has been noted as failing produce stream meanders where it crosses the Arden- ;; the great loops described by the Conodoguinet :rahler, 1946) are valley meanders; and Rock Creek, ere it passes through the suburbs and city of Wash- ton, D.C., pursues a nonmeandering course along a k-lined bed, within the meanderings of an incised ley. The relevant landforms are so common that y appear to have misled some previous writers, ecially those responsible for hypotheses of the hy- trophy of meanders during incision. Although en- gement and distortion are admitted, they should be allowed to obscure the distinction between ınders of the valleys and the trace, meandering or , of the present stream. The foregoing observations w that a number of streams in meandering valleys actually underfit, although not manifestly so, and y open the possibility that many others may be un- fit, as required by an hypothesis of climatic change. ıce the problem of the absence of stream meanders m incised reaches despite their presence elsewhere. Contact with bedrock does not conflict with the de- ›pment of meanders after incision has begun. The ›le array of ingrown forms and the deep trenching of valley meanders through firm rock in place accord with the evidence of terraces in proving that neither lateral growth nor downcutting, nor the meandering habit itself, are suppressed by contact of the stream with the solid. When terraces are used to reconstruct former traces of valley meanders, they commonly show that these traces become less and less sinuous with increasing age, as would be expected from the very fact of ingrowth. At high levels and with the earliest terraces, however, the evidence is usually so fragmentary that it cannot be used either to justify or to confute a claim that, when incision began, any given river had an essentially straight channel. In the well-studied valleys of parts of Europe, the sinuosities which became incised valley meanders developed first on trains of outwash or reworked crytoturbate (Troll, 1954). The immediate forerunners of the large meandering streams were braided, and the braided deposits could have been thick enough to insulate the initial meanders from bedrock. Similar considerations probably apply to the Stratford Avon, where the highest terrace has been noted as probably consisting of outwash. It is therefore not possible to use valley meanders as proof that a river already in contact with bedrock, and not already possessing a meandering trace, can spontaneously develop meanders.

Nevertheless, the claim which is sometimes made that all trains of incised meanders are inherited from free meanders need not be conceded. Numerous observations on initially straight gullies, which assume a meandering habit as they cut into heaps of spoil, demonstrate that systematic winding before incision is not essential. (See figs. 59–60.) Any possible objection

FIGURE 59.—View of meandering gully on highway bank near Iowa City, Iowa. Pack gives scale.

[7] The records will be discussed and illustrated in a succeeding paper.

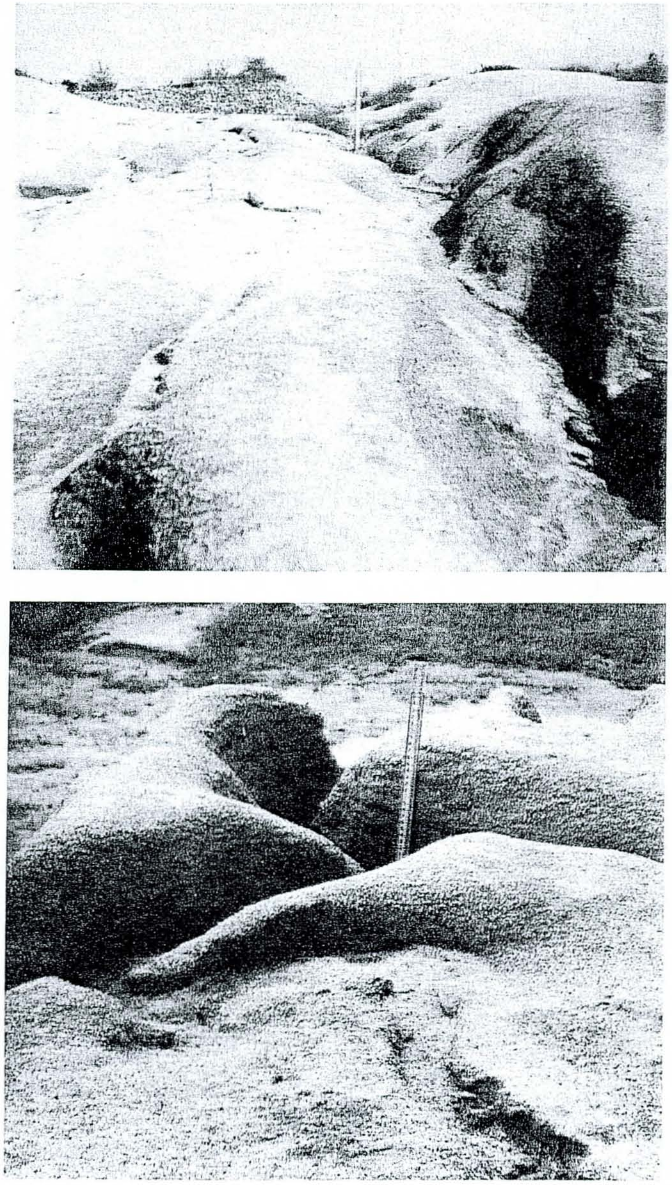


FIGURE 60.—Views of meandering gullies on mine tailings near Shullsburgh, Wis. Staff gives scale.

that the gullies are sunk into unconsolidated material—natural streams are being considered in relation to bedrock—is easily overcome by reference to natural rivulets which, originally alined on straight dikes, faults, and master joints, begin swinging only after they have cut straight trenches. Just as meanders can develop initially when the stream is in contact with bedrock, so they can persist not only when the meanders are being incised but also when a continuous flood plain has been constructed. The Garren Brook on the Welsh Border of England can perhaps be regarded as superimposing its stream meanders onto bedrock through the alluvial fill of the large underlying channel (Dury, 1954, fig. 11),

but a different interpretation is required for the wold Coln. Below a nickpoint, the Coln illustra stereotype of the alluvial trough (Dury, 1953 meanders are sweeping downstream, the botto meander pools reaching the planed-off surface o beneath the flood-plain alluvium. Mere contac bedrock is here insufficient either to prevent the tion of a meandering trace or to stop new me from forming in a reach where cutoff is frequer

Conversely, nonmeandering reaches can occ streams where bedrock lies far underground. M ald Creek, Iowa, used above to illustrate the a of the meandering pattern, is a case in point. Au proves it to be underlain not directly by till but fill of just such a large channel as elsewhere cc the flood plain of a presently meandering stream. allel instances can be adduced from rivers which, main, are manifestly underfit. Just as contac bedrock does not prevent meandering, so lack of ders does not necessarily imply contact with bedı

This is not to deny that meanders develop more in alluvium than in bedrock or that the very steep where feeders of the Shenandoah cross solid ou have nothing to do with channel habit. The diff may, in part, be one of time: meanders perhap longer to form in bedrock than in alluvium. Bu to be supplied for the general onset of underfitne a span of about 10,000 years, in which stream me could presumably have developed, if they were velop at all, on such rivers as the North Fork Shenandoah, the incised rivers of the Ozarks, a canyon reaches of the Humboldt. The very fac stream meanders occur outside the canyons, but o ceptionally inside, suggests that the incised cha of the valley and the lack of meanders on the s are in some way connected. Where a stream w elsewhere manifestly underfit passes through an ated reach of the valley, it is at least possible th trace of former large meanders has been lost alluviation. But where the valley is incised and ing, the trace of stream meanders is the one lackin contact with bedrock appears to supply no su reason, another cause of inhibition must be sough

The possibility that the walls of incised valle liver material too coarse to permit meandering attractive in that form. Meanders can be initiat velop, and persist both in coarse till and in coa luvium. The first instance is best illustrated, the writer's experience, by streams in the glaciate choked valleys of Wales and along the Scottisl der—in particular, of the Cheviot. The second exemplified by the draw, its margins boldly sca alongside the highway which descends from the

FIGURE 61.—View of joint blocks of dolomite shed from valley wall, Mineral Point Branch of the East Pecatonica River, Wis.

town of Jerome, Ariz. Moreover, the undercut slopes of some meandering valleys can scarcely have failed to shed large joint blocks when being eroded by the former rivers. Examples include the outside curves of Canyon Diablo and Canyon Padre in Arizona (figs. 53–56) and those of many winding valleys in the Driftless Area of Wisconsin, wherever the upper slopes are cut in widely joined dolomitic limestones. (See fig. 61.) At the same time, bed material in the Humboldt canyons appears to be, as reported, perceptibly coarser than that in reaches upstream and downstream where river meanders occur. As Hack (1957) has observed for another region, the caliber of loose material transported by streams declines rapidly with distance from the outcrops which supply coarse fragments. It therefore seems entirely practicable for coarse debris to be concentrated almost exclusively in the incised reaches. An hypothesis is possible that, because the lack of stream meanders in incised reaches ensures a shorter trace and a steeper slope than in meandering reaches, steepness, coarseness of bed material, and a tendency to braid are reasonably associable with one another.

Such an hypothesis must be taken with reserve, however, as soon as allowance is made for the winding thalweg which accompanies the development of pools and riffles in a straight channel (Leopold and Wolman, 1957, p. 53–55). It has yet to be proved that thalweg slopes are steeper within the canyons of the Humboldt than outside. And, if the Ozark rivers are generally in or near contact with bedrock, the change from the former large meanders to the present narrowed channels and nonmeandering traces would involve an actual reduction in downstream slope. Additional problems arise when caliber of bedload is compared from stream to stream. Numbers of lesser streams in the Ozarks fail to meander, even though they transport very little material larger than medium gravel—for example, Little Piney Creek upstream from its confluence with the Gasconade. By contrast, the Mineral Point Branch of the Pecatonica in Wisconsin, in its upper reaches where it is smaller than Little Piney Creek in that stream's lower reaches, is shifting coarse gravel along a distinctly meandering channel. The fact that the coarse gravel in Mineral Point Branch comes not directly off the valley walls but from an alluvial fill under the silty flood plain does not seem relevant to the maintenance of a meandering habit. Mineral Point Branch does move coarse gravel and does meander; the larger Little Piney Creek transports medium gravel and fails to meander.

In other respects, however, caliber may be strictly relevant to the immediate problem. Mineral Point Branch may be capable of meandering, not in spite of its coarse bedload, but specifically because its bank material consists largely of silt. If, on rivers not wholly confined by bedrock, braiding is favored by lack of cohesiveness of beds and banks and maintenance of a single channel by cohesiveness, then the lack of stream meanders in the canyons of the Humboldt may be due simply to the high erodibility of the debris present. Lack of fine-grained waste in the Ozark valleys may perhaps be widely ensured by the preponderance of limestone among outcropping rocks, which can readily supply fragments but not particles. In all the valleys so far investigated which contain manifestly underfit streams, the alluvium contains a high silt-clay fraction. This whole matter obviously requires further study, but the suggestions made are perhaps capable of pointing investigation away from the direct influence of outcropping bedrock, except insofar as the cohesive properties of bedrock may be held to have encouraged the persistence of meanders on the former streams which cut the valley bends.

CANYONS AS FLUMES

On steep hillsides in the Great Basin, and in the Cordillera generally, short canyons with steep downstream gradients are numerous. In actuality, there is a continuous range from canyons with near-vertical sides to steep-walled valleys of the form common in humid regions; the second type is well displayed in the Ruby Mountains of Nevada. The steeply descending canyons and the ordinary deep-cut valleys usually tend to wind. Their windings appear to be valley meanders, for plots

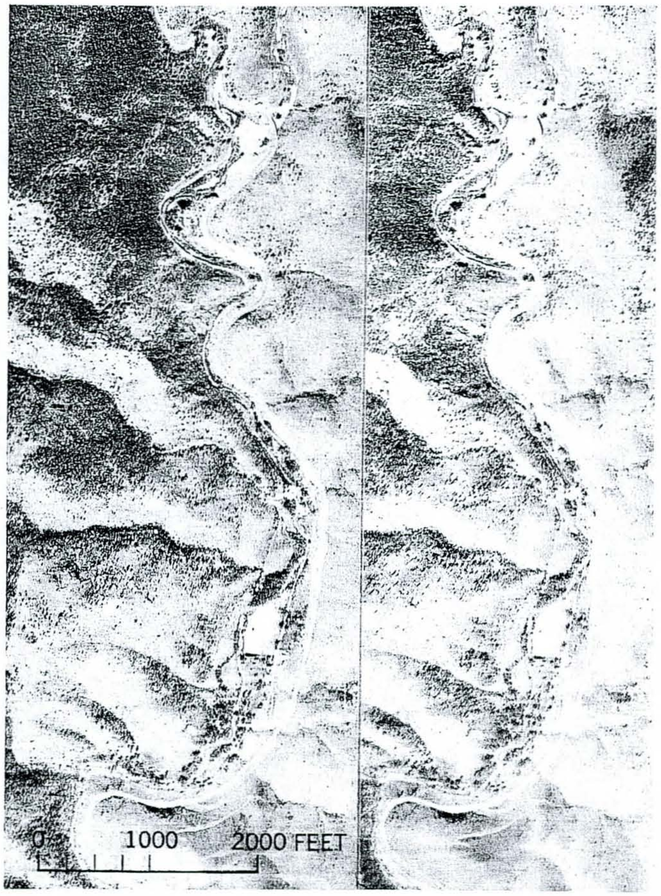


FIGURE 62.—Stereoscopic photograph of part of Bear Creek Canyon, near Denver, Colo.

of wavelength against area bring them into the family which includes, for instance, the incised systematic windings of reaches of the Humboldt (fig. 30).

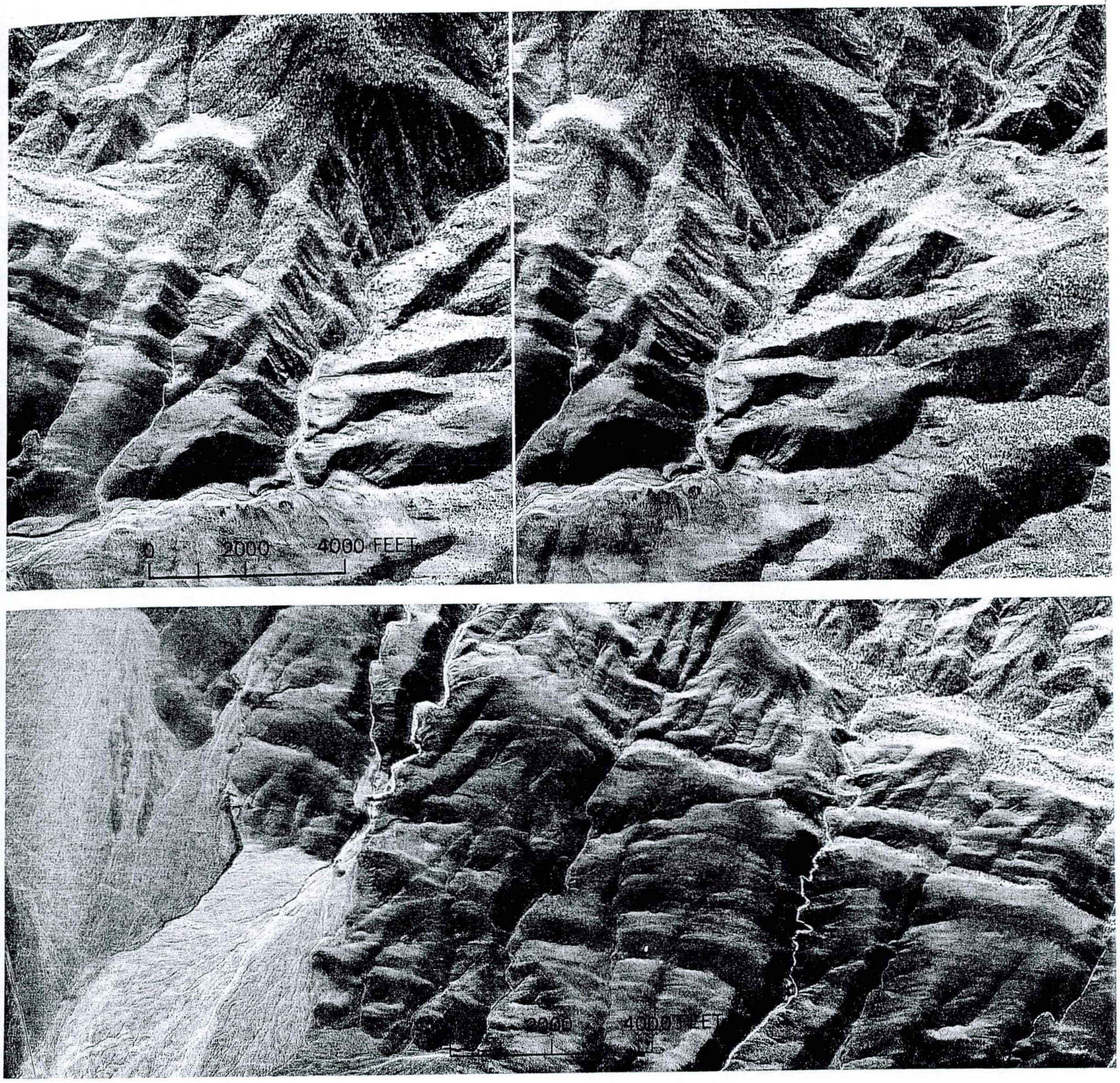
Where perennial streams flow, as in Bear Creek Canyon in the Front Range near Denver, Colo., it may be possible to compare the wavelengths of stream meanders with those of the valley; in one short reach, and exceptionally, the floor of Bear Creek Canyon is wide enough to accommodate a flood plain, on which stream meanders occur (fig. 62). Although these are too few to justify a comparative plot of wavelength, and although the limb of valley bend on which they lie appears to be elongated in response to structure, the disparity of size between the ingrown windings of the canyon and the meanders of the present stream is beyond doubt.

Where present streams are ephemeral, the floor of a canyon may be encumbered with rock waste on which braided streams flow after rain. Even then, a plot of wavelength against area may be capable of revealing a systematic relation between the two sets of data and of showing that the bends of the canyon are valley meanders. Such is the instance with Birch Creek Cany (fig. 63), which enters Deep Spring Valley, Calif., fr the west. The present stream rarely flows to the mo of the canyon; but the valley bends, although consid ably distorted by the structures of bedrock, closely cord with those of the Humboldt (fig. 30).

Numerous canyons, however, are so narrow at base that, far from providing room for meanders, tl do not contain a streambed in the usual sense. valley acts as a notch or flume, and the stream level ri on the valley wall at times of high discharge. Exa ples of this kind are provided notably by the in canyon of the Colorado, which does not wind, and the systematically winding canyons of the Wasa Front north of Salt Lake City. Cloudbursts on Wasatch Front promote torrential streams which, fl ing down the canyons, undergo very marked superele tion at bends.

The torrents resulting from infrequent heavy ra can be strongly erosive, in the sense that they ca great amounts of coarse debris from the canyons a discharge it on to the fans at the canyon mouths. Ho ever, it seems unnecessary to hold such rains responsi for the initiation of the canyons, as wavelengths bel to the regional family of valley bends. Alluvial me ders are related primarily to discharge at bankf Where streams are manifestly underfit, bankfull charges higher than those of today are held to acco for the initiation and the incision of valley meand There is no reason to seperate incised winding cany with steep downstream gradients from the general fa ily of incised meandering valleys or to postulate them an origin referable to some other type of discha than that which produced valley meanders genera At the same time, the function as flumes suggests t short canyons may, from time to time, carry dischar of an order similar to that of the discharges wh initiated them; if so, the windings of the canyons still developing in their original manner, in contrast valley meanders on lower and more gently slop ground, which are currently in a state of arres development.

An ancillary problem which still awaits investigat is that of stream density in dry regions. According the views expressed here, the existing network of vall is denser than the present climate requires. Beca canyons in well-dissected terrain provide ready-m routes of discharge, they inevitably carry streams fr time to time, but it by no means follows that the val systems were first developed in conditions like those today. Canyons which do carry streams ought to studied in association with valleys which do r Strahler (1944, 1948) infers that the systems of



GURE 63.—Views of canyons on west side of Deep Spring Valley, Calif. Upper, stereoscopic photograph of Birch Creek Canyon; lower, aerial view of nearby canyons to the south.

leys on the Kaibab Plateau result from the disap- ırance of water along lines of underground drain- ꜫ. If, however, this region has been influenced by natic change powerful enough to reduce streams to lerfitness, then its limestone hydrology is compara- to that of the Chalklands of the English Plain; in :h areas, the peculiarities of limestone are responsi- for absolute streamlessness, whereas the valley sys- ns and the rivers which cut them are related to a :mer period of high surface discharge. Rapid perco- ion superadds its effects to those of climatic change.

SUMMARY

This development of the general theory of underfit streams will be continued in subsequent professional papers, with reference to large meandering channels, to chronology, to calculations of former discharges, and to reconstructed conditions for the occurrence of those discharges. Meanwhile, the conclusions reached so far can be summarized as follows:

1. Underfit streams are those which have undergone a marked reduction in bankfull (channel-forming) discharge.

2. Derangement of drainage cannot supply the required regionally applicable hypothesis of the cause of underfitness.
3. Certain authenticated derangements serve merely to complicate the pattern of underfitness.
4. Underfit streams in more amply meandering valleys—manifestly underfit streams—are not the only underfit type, even when braided streams are left out of account.
5. Manifestly underfit streams are appropriately specified in terms of wavelength ratio between valley and stream, which gives an index of underfitness.
6. In the absence of stream meanders, valley meanders can often be recognized from wavelength and drainage area values and by comparison with the valley meanders of manifestly underfit streams elsewhere.
7. Study of nonmeandering underfit streams with unbraided single channels should include measurement of the long profiles of the beds. Subregular deformation of the bed in long profile appears capable of substituting for a meandering habit.
8. Nets of canyons in dry regions appear comparable to systems of dry valleys in limestone country, in that both series relate to high surface discharges in former times. Canyons, however, can be sufficiently narrow at the bottom to act as flumes at times of high discharge, so that the bends of some of them may still be undergoing erosion.

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UNITED STATES DEPARTMENT OF THE INTERIOR
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EXPLANATION

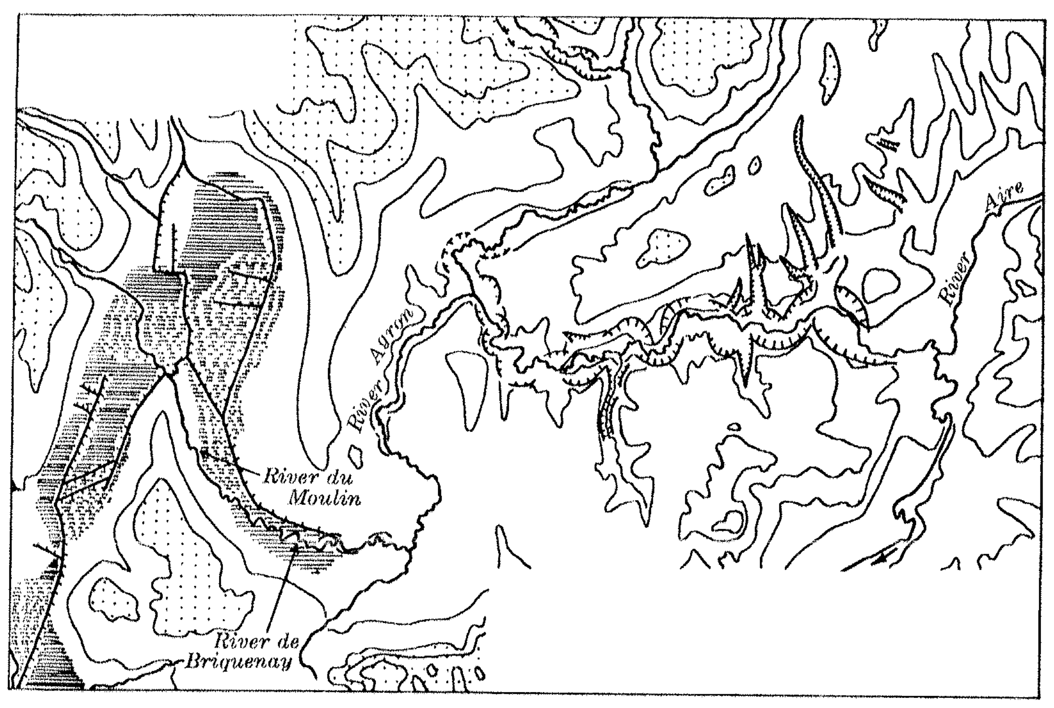
Swamp

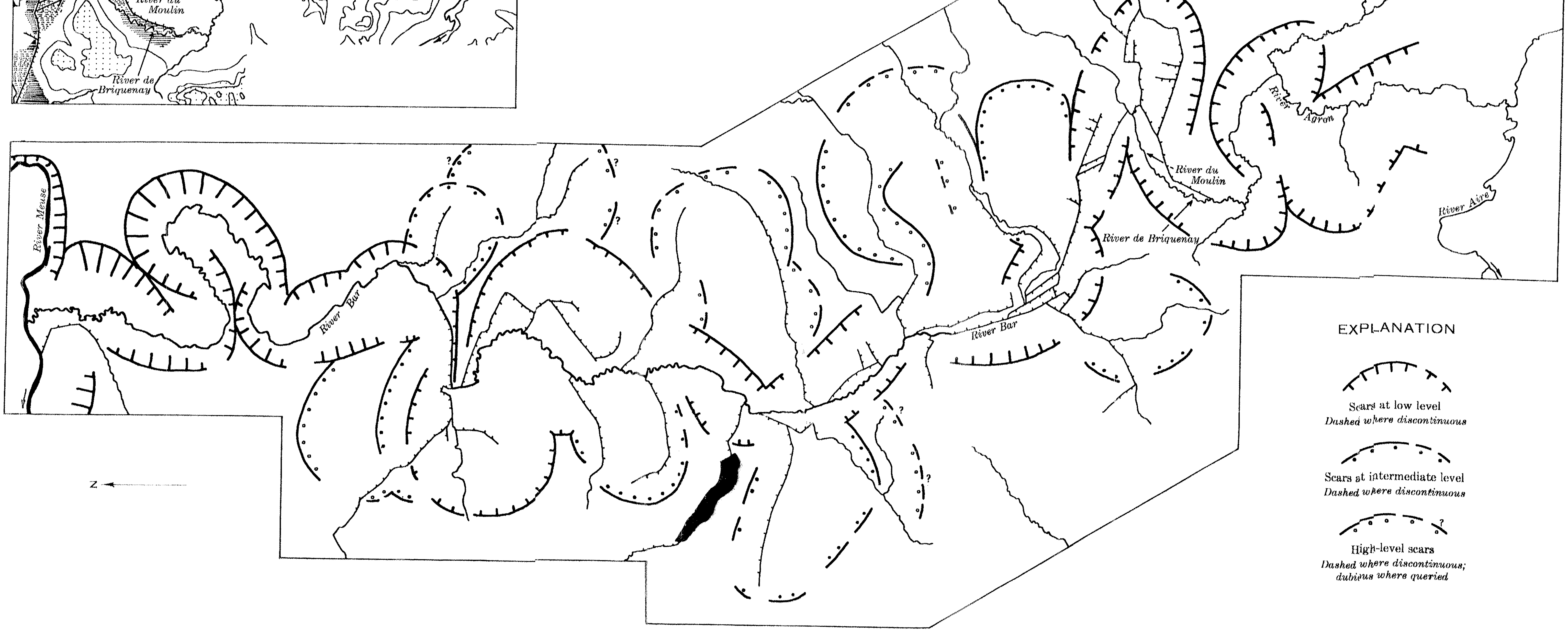
Flood plain with poor drainage (Shown near divide)

Regularized or artificial channel

Gully

Land over 200 meters marked by stipple





EXPLANATION

Scars at low level
Dashed where discontinuous

Scars at intermediate level
Dashed where discontinuous

High-level scars
Dashed where discontinuous; dubious where queried

MAP OF THE RIVER BAR AND REVERSED RIVER AGRON, FRANCE, SHOWING VALLEY BENDS

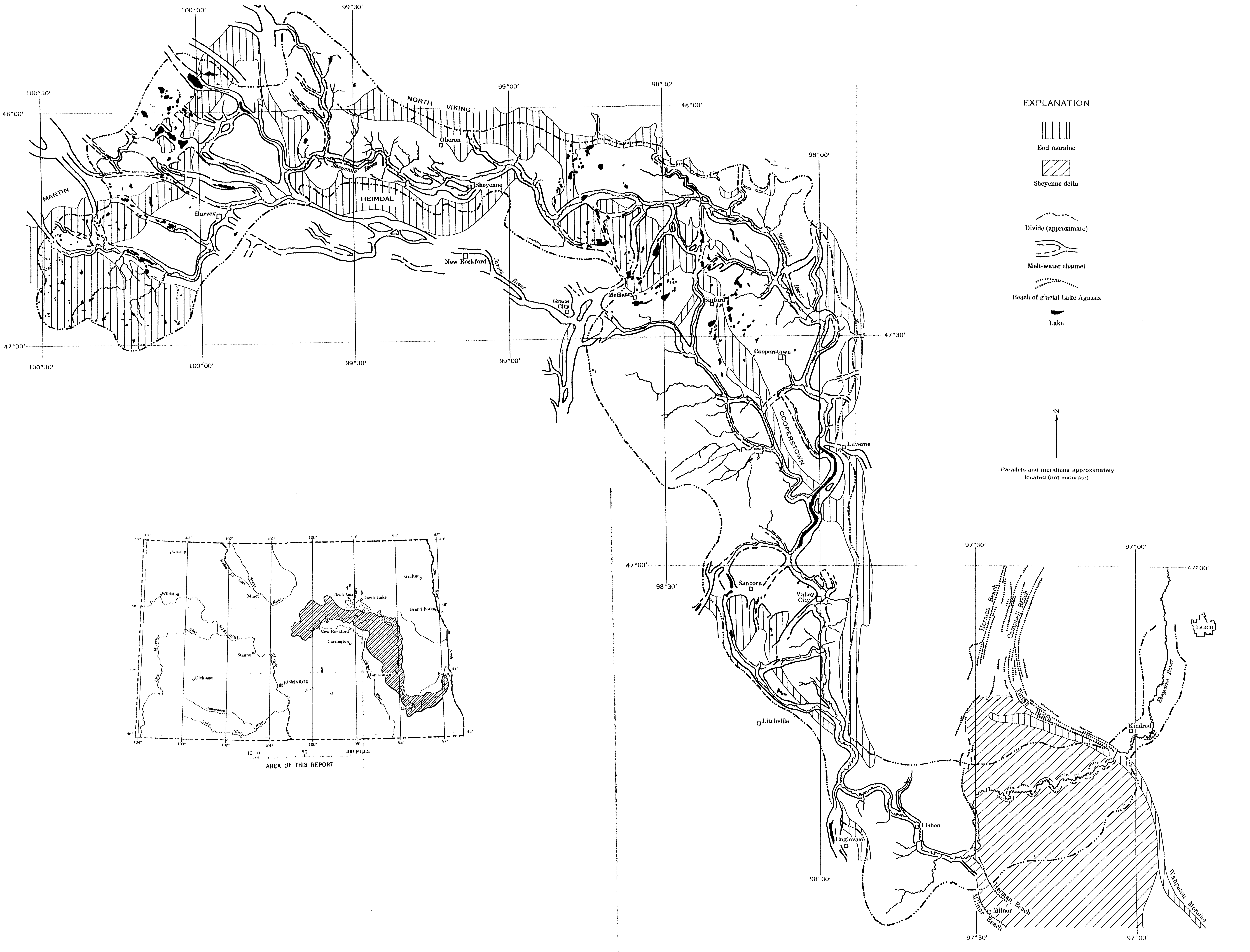
SCALE 1:96 000

2 0 2 4 6 MILES

2 0 2 4 6 KILOMETERS

CONTOUR INTERVAL 20 METERS

713-103 O - 64 (In pocket)



MAP OF THE SHEYENNE RIVER SPILLWAY AND ASSOCIATED FEATURES, NORTH DAKOTA

SCALE 1:900 000

10 0 10 20 30 MILES

10 0 10 20 30 KILOMETERS

Subsurface Exploration and Chronology of Underfit Streams

GEOLOGICAL SURVEY PROFESSIONAL PAPER 452-B

Subsurface Exploration and Chronology of Underfit Streams

By G. H. DURY

GENERAL THEORY OF MEANDERING VALLEYS

GEOLOGICAL SURVEY PROFESSIONAL PAPER 452-B

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1964

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GENERAL THEORY OF MEANDERING VALLEYS

SUBSURFACE EXPLORATION AND CHRONOLOGY OF UNDERFIT STREAMS

By G. H. Dury

ABSTRACT

'ield investigation of valley bottoms of underfit streams by ans of augering or of seismic refraction demonstrates large nnels that meander around valley bends. Information from scellaneous sources—that is, civil-engineering records from nsites—supplements and confirms the writer's findings. The ge channels are taken as the beds of former large streams; ir infilling is ascribed to stream shrinkage rather than to radation.

n areas not covered by ice at the last glacial maximum, dates the initiation of valley meanders range from early Pleisto- e to late in the last glacial. Some trains of valley meanders deeply incised into bedrock and date at least as far back as ly (Nebraskan) glacial times. Other trains, cut into till ets or littoral deposits, were initiated in the mid- istocene, whereas still others slightly postdate the outermost stand of the last (classical Wisconsin) glacial. The last ndonment of the largest channels and the valley meanders these groups appears to be associated either with the Two eks Interstade or with the opening of Zone V (Boreal), ording to locality. Areas covered by ice or by proglacial es at the last glacial maximum or during part of the last lacial display valley meanders whose date of first possible iation is perhaps as much as 2,000 years ago. Reduction to lerfitness of streams in this group is provisionally equated h infilling of minor channels by streams of the former group. 'omparison of humid regions with arid or semiarid regions questions of the cause of accelerated erosion. Generally aking, this erosion should correspond to increases in surface ter, which could, however, be effected by an increase in pre- itation or a decrease in temperature. The brief very high nds of glacial Lake Lahontan shortly before and shortly after Two Creeks interval seem to have resulted from climatic nges of both types; these high stands are correlated with last clearance of large channels, respectively, in the Drift- Area of Wisconsin and its margins and in southern Eng- d. The hypsithermal maximum (Zone VII) was a time of reases both of temperature and of precipitation; precipitation reased sufficiently to renew erosion both in humid regions l in large parts, at least, of dry regions. Infilling of large nnels in humid regions during Zone V (Boreal) resulted m decreased precipitation, which more than offset a simul- eous reduction in temperature.

)own to the limits of magnitude involved in the partial re- avation of channels in Zone VII, the respective successions of ting and filling in dry and humid regions appear synchro- is and parallel. Lesser sequences of cut and fill, on the other id, may be out of phase.

INTRODUCTION

This paper continues the development of the general theory of underfit streams begun in Professional Paper 452–A (Dury, 1964). That account was meant principally to review terminology, to establish the widespread occurrence of underfit streams, to demonstrate that not all underfit streams need possess meandering channels at the present time. and to show that derangements of drainage cannot supply the general hypothesis of origin which facts of distribution and chronology require. Neither the introduction to the series nor the acknowledgments of extensive help will be repeated here, except for a general statement that many individuals—in particular, both full-time and part-time members of the U.S. Geological Survey—have been most generous with assistance in the field, with discussion, and with constructive criticism.

The following text extends Professional Paper 452–A by reference to subsurface exploration and to dating. The two matters are closely related, for alluvial fills proved and sampled during exploration of the subsurface provide certain critical dates. The large channels that contain these fills support the interpretation of valley meanders as the products of erosion by former large streams. In addition, observations on valley meanders in relation to records of terracing, sedimentation, and the draining of proglacial lakes combine with results obtained in various connections by other workers to elaborate the sequence of incision and channeling. Although the presented material is organized under subheadings, there is in actuality considerable overlap from section to section, but the correlation eventually reached is meant to give a synoptic view of the whole.

FIELD EXPLORATION OF FILLED CHANNELS

ENGLAND

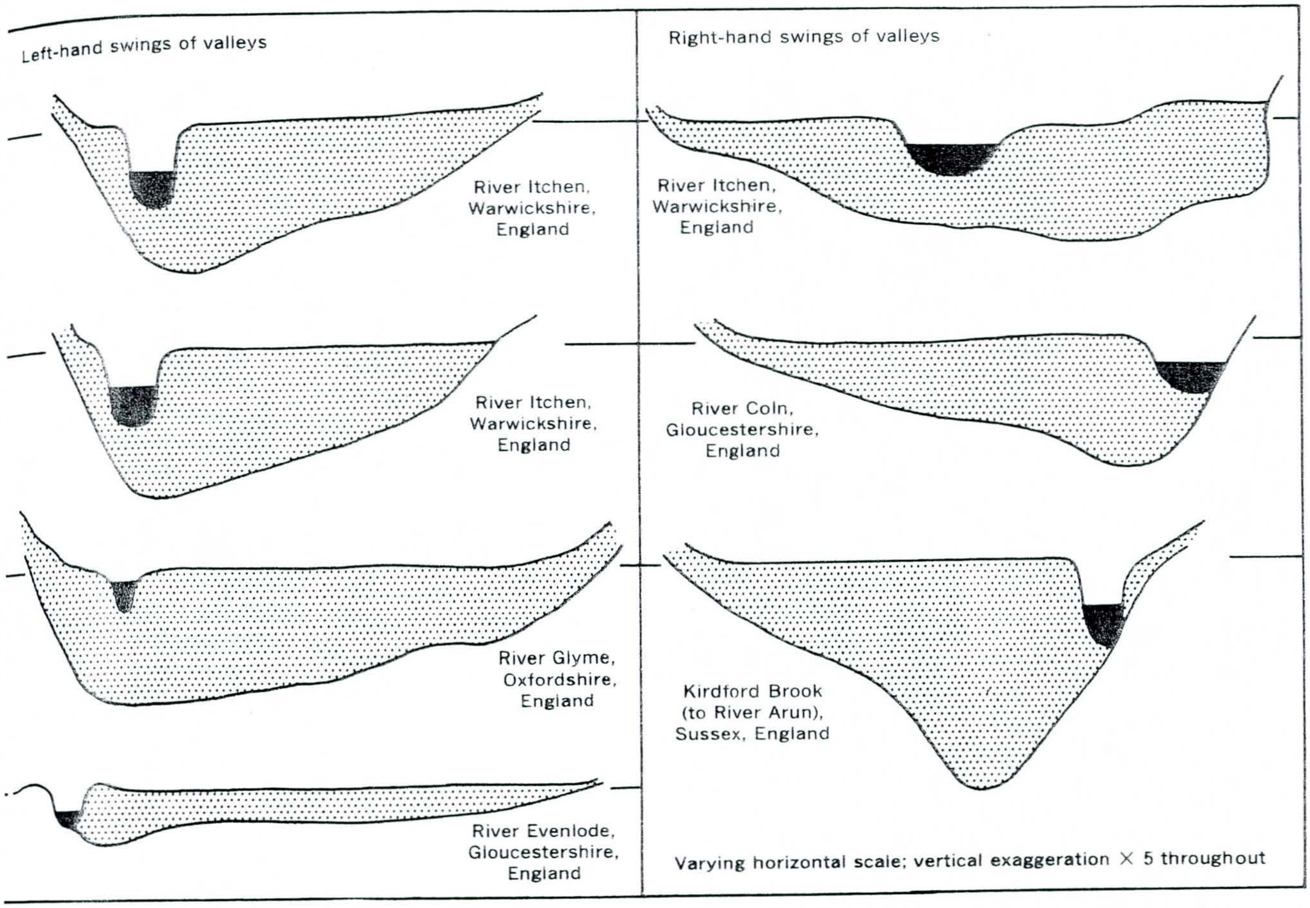
Other than the mapping of landforms, hand augering has been the writer's principal technique of investigation at sites on the English Plain and its borders. This technique is well suited to the local conditions of fine-grained alluvium. The valley fills contain little material coarser in grade than sand; they consist mainly of silt and clay, varied in places by peat, sapropel, tufa, and malm (earthy, amorphous calcium carbonate). A person working alone can auger to depths of 30 feet with a screw auger 1½ or 2 inches in diameter and to depths of 20 feet with a bucket auger 3½ inches in diameter. Reaching depths greater than 16 feet with the screw auger or greater than 12 feet with the bucket auger requires however, usually some dismantling of the shafting each time the auger is drawn up. No form of sheerlegs or derrick was found necessary. The normal bit of the screw auger can be made to penetrate a mixture of gravel and mud and can be screwed through self-cemented limestone gravel, but this auger is checked by uncemented gravel of resistant material, even when a special bit is used. Both the screw and the bucket types of auger seem capable of extracting clean samples from coherent deposits, although great care is needed in cutting away the crumbs and sludge that cling to the outsides of all samples. Experiments with a specially constructed corer, fitted with a piston to prevent collection of sediment during lowering, were discontinued when the piston was found to jam repeatedly. There seems to be no advantage in using corers of the type made for sampling peat, for they cannot be kept free of water and mud during descent. The screw-bit and, probably to a lesser extent, the bucket types clean themselves as they are driven in.

At the outset, samples were logged in detail at depth intervals of a few inches. As work progressed and the alluvial sequence was established for a given valley, it became possible to use the screw auger as a probe which could be forced through the weak fill and turned for a short distance at the base for a sample of bedrock. The junction between the base of the fill and the underlying rock almost everywhere was found to be sharp. First trials in the valley of the Itchen River of Warwickshire fortunately involved an abrupt transition from moist, gleyed alluvium to dry well-indurated shale or marl. Cemented bedrock can in places be identified from small fragments obtained with the bucket auger or from grains adhering to the pilot screw of the screw bit. For example, the alluvial fill of the valley of the River Perry in Shropshire consists of dull brown clayey silt, thin layers of intercalated gravel, and peat, all contrasting strongly with the bright-red grains of dry sand obtained from the underlying Triassic sandstone, whereas dry well-indurated oolitic fragments in the Cotswold valleys differ markedly from the moist yellow grains of decemented ooliths.

In a first series of field explorations, 53 cross profiles were determined by means of 290 boreholes in 6 valleys that contained manifestly underfit rivers (I 1952; 1953a, b, c; 1954). The work was designed tially to test the hypothesis that if valley meander homologues of stream meanders, then homologu stream channels could be associated in nature with ley meanders. Such homologues were identifie every site explored.

Subsequent work provided a further 27 profiles termined from 216 boreholes in an additional 6 va (Dury, 1958). This second field exploration sh that the degree of underfitness is constant throug the Cotswold region and permitted the onset of un fitness to be dated relatively to the beheading of River Evenlode, to the spreading of congeliturbat the floor of the Evenlode valley, and to the infil of the large channel beneath the River Dorn. Rec in addition to those published bring the writer's t of profiles to 93, determined by some 600 borehol 18 English valleys; work now in hand by grad students raises the total to more than 100 profiles i valleys, without appeal to incidental records obta by researchers in other studies or by civil engin Not all the additional material need be presented l as part of it merely duplicates what is otherwise kno the record of channeling and filling will, howeve amplified below when chronology is discussed. general outcome of subsurface exploration is to s that manifestly underfit streams on the English P are characteristically underlain by large channels w wind round the bends of valleys.

The large channels reach their greatest depth or near the extremities of valley bends. At these they are asymmetrical in profile, sloping steeply d from the outside of the curve and less steeply ac the inside curve—that is, they are identical in h with the beds of meandering streams (fig. 1). channels are much wider than the present streamb even when all possible allowance is made for diffic in identifying their banktops, and they widen con erably at some valley bends; for use in compari widths have accordingly been taken at inflections tween bends, where they are least. The average r of width between the large channels and the pre channels, for a first series of nine determination 11.5:1 (Dury, 1954, table 1). Such a ratio, or the venient but approximate value of 10:1, immedia suggests discharges for the large channels of an o entirely different from that applicable to present cl nels. A firm suggestion that bed widths of the l channels have not been overestimated comes from 13:1 ratio between the wavelengths of valley mean and the determined bed widths (Dury, 1955). A w length and bed-width ratio of about 10:1 seems the n



IGURE 1.—Profile sections of filled channels beneath manifestly underfit streams showing asymmetry at valley bends. Views are downstream.

:ely to apply to rivers in general (Leopold and Wol-an, 1960; Bagnold, 1960). As wavelengths can be easured with some accuracy, the difference between :1 and 10:1 could be explicable by values of bed idth that are too small. On the other hand, the meas-ed channels are cut into coherent material, so that the gh wavelength to width ratio may reflect nothing ore than proportionally narrow channels with a some-hat high depth to width ratio.

Records for the Itchen River valley are now avail-le from four reaches and have been used to make graphic comparison of bed width and drainage area ig. 2). It seems at least possible that the bed widths the large channels on the Itchen system vary with ainage area in the form $W \propto M^b$,

here

W=width of channel bed, in feet;
M=drainage area, in square feet;
and b is a numerical constant.

ariation of the same type seems to apply also to the d width and to the wetted perimeter of the present annels. More evidence of this kind, if it were avail-le, could usefully be employed, but these observations do at least indicate that the large channels can be traced far up the valleys toward the heads. They also suggest that the disparity between bed widths increases headwards, as the disparity between wavelengths will later be seen to do. If so, the unusually high value of 11.5:1 for the ratio between respective bed widths of large channels and present channels may result, in part, merely from the smallness of the drainage areas involved. Even within the low range of observations on the Itchen, the bed-width ratio appears to fall from about 15:1 at 3 square miles to 9:1 at 40 square miles.

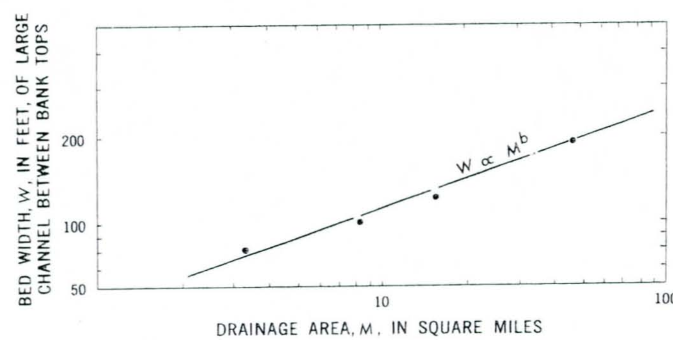


FIGURE 2.—Graph showing relation of bed width to drainage area of the River Itchen, Warwickshire, England.

However, as wavelengths and bed width are closely related and as the wavelength ratio between valleys and streams is normally found to be systematic, there is no great point in determining the bed width of large channels merely for purposes of dimensional record. Determination of cross-sectional area, on the other hand, has uses of its own, whereas analysis of sedimentary fills can obviously be instructive.

CHARACTER OF THE FILLS

On the Itchen, the sole consistent change in the vertical succession of alluvium is that from dull-brown cloddy columnar or prismatic clayey silt at the top to moist, blue-gray, gleyed clayey silt in the lower part. When the first records were made, the significance of gleying was not taken fully into account, and the two types of deposit were recorded as distinct. Subsequently, it has become clear that the difference is one between permanent waterlogging and seasonal waterlogging. In the absence of lithologic change it is impossible to define the base of the present flood plain, although the base of the large channel lies well below the maximum depth of scour in present conditions. Aquatic plants rooted in the streambed are not swept away by discharges at high stage, nor do the banks undergo general erosion at times of flood: the bed is not scoured severely enough to permit the stream to reach the base of the underlying filled channel, nor does the present channel widen sufficiently to provide the width to depth ratio which, on any reasonable view of channel form, would be needed if scour were to go down to the underlying bedrock.

In some other valleys the distinction between the alluvium of the flood plain and the underlying remainder of the fill is abundantly clear, especially where the fill beneath the flood plain is stratified. Such is the case with the Cotswold Rivers Glyme and Dorn, where the alluvial succession includes calcareous deposits that could not remain uncontaminated unless they were now out of reach of the streams which are reworking the silty clay of their flood plains and are carrying bedloads of sand. Sandy layers that extend horizontally through the fill at about the depth of the bottoms of pools mark the bases of the flood plains and indicate the present limit of scour. The fact that the fill of the Dorn valley remains undisturbed below the flood-plain base is confirmed by the fill's pollen content, which includes much birch, pine, and hazel, with elm becoming evident toward the top but with alder absent. Infilling began late enough for birch-pine forest to be established and, as shown by elm pollen, extended into the Boreal phase; but as alder is lacking, the sedimentary record does not reach the top (end) of the Boreal. Whether sedimentation did continue into the lat[e] Boreal, to be partly obliterated by later erosion, is [i]material at this point. The preservation of a poll[en] assemblage notably different from that of today sho[ws] that the remaining sediments have not been rework[ed] since they first accumulated.

Confirmatory observations come from the Engl[ish] River Rib, in Hertfordshire. The Rib flows off [the] back slope of the Chalk which rims the London Ba[sin] on the north. In its upper reaches, the valley is [cut] through or into thick outwash gravel that dates fr[om] the Penultimate Glacial—at least 30 feet, and possi[bly] 40 feet, of gravel is exposed in a disused and overgr[own] working in the side of the valley. Four profiles (fig[. 3]) indicate a rather shallow filled channel occupied [by] clayey silt and floored by gravel. Natural change[s of] course and artificial straightening complicate the [s...] but at the same time they make it possible to disc[over] that the fragmented bed load in an abandoned nat[ural] channel is separated from the underlying gravel [by a] layer of fine-grained fill (fig. 3, line 2). Here ag[ain] the signs are that scour in present conditions does [not] reach the floor of the large channel. Three and [a] half miles downstream, where both the valley and [the] stream are larger, the Rib has entered a train of va[lley] bends cut through the gravel into the underlying s[...] formation of Chalk, which is exposed in small qua[rries] on both sides of the valley. A line of boreholes ac[ross] the curve of a valley bend prove muck and pea[t to] depths of nearly 30 feet below the surface of the f[lood] plain (fig. 4). In some holes the auger was used ma[inly] as a probe, but detailed logging at seven points re[veals] the wide extension of what seems to be a horiz[ontal] layer of peat. This, like the assorted fills of the C[ots]wold valleys, could scarcely have remained intact [if it] had been scoured by the present-day stream.

MODE OF INFILLING

To apply the term "aggradation" to the infillir[ng of] large channels is to beg the question. Reduction [of a] stream to a manifestly underfit condition necess[arily] lengthens the trace and thus tends to reduce the s[lope]. Reduction of slope could be offset in part by deg[rada]tion of downstream reaches and by aggradation o[f up]stream reaches, but changes of this kind could not [com]pensate in full, in numerous valleys, for the length[ening] of trace. Lengthening must occur when st[ream] meanders are superadded to the trace of valley m[ean]ders. The magnitude of lengthening is given b[y the] ratio between distances measured along the curv[es of] the valley and those measured along the present m[ean]ders of the stream; alternatively, straight-line dist[ances] from bend to bend or from bend to inflection to

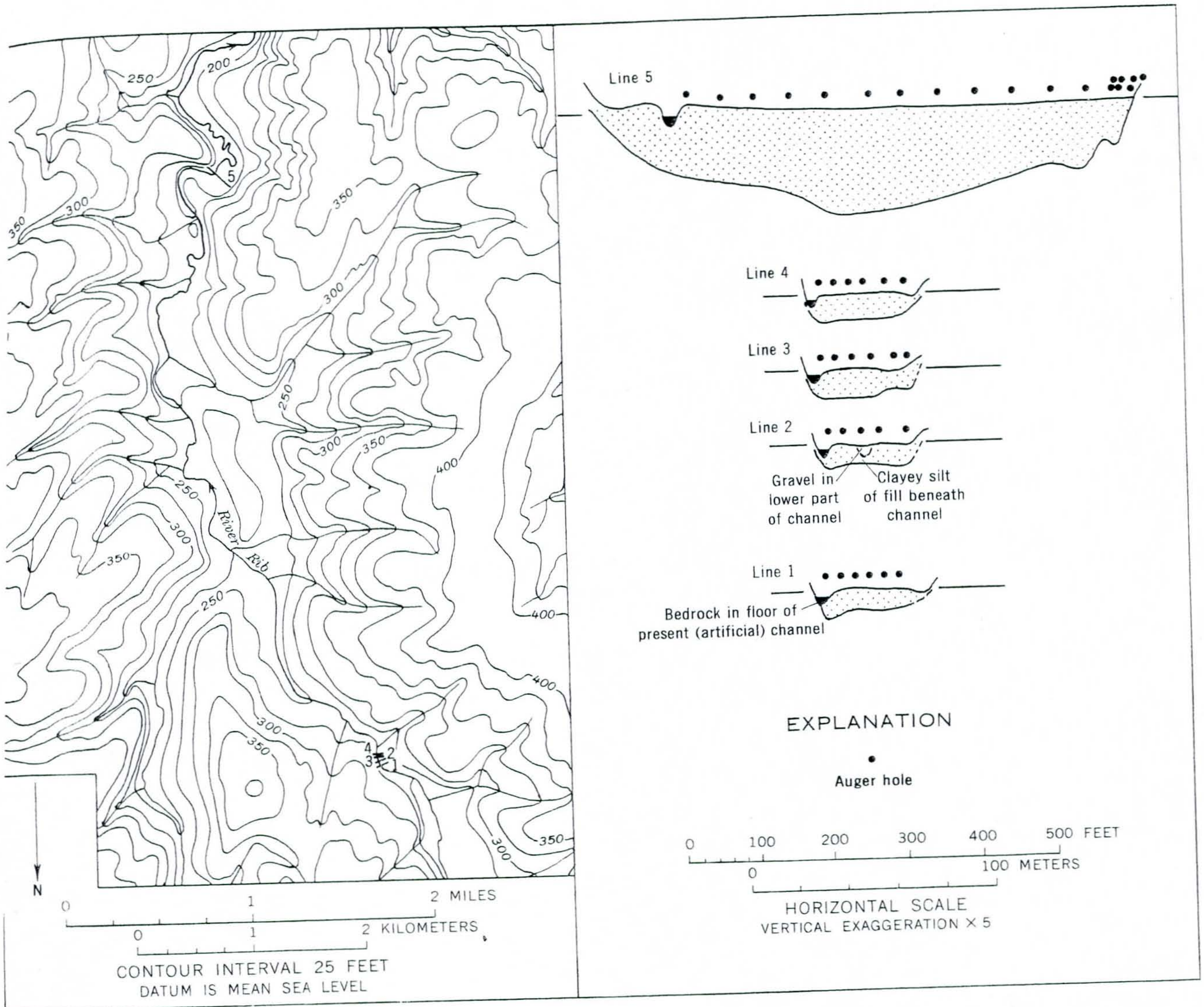


FIGURE 3.—Index map of the River Rib, Hertfordshire, England, and profile sections of channels, arranged in downstream order.

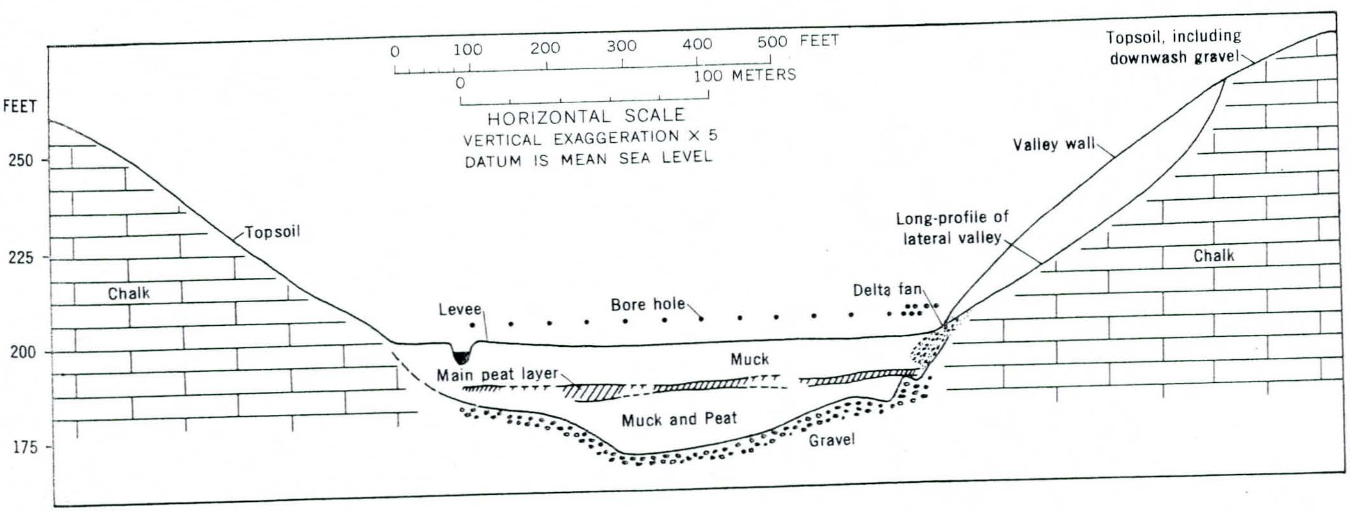


FIGURE 4.—Profile section of valley and channel of the River Rib. See figure 3, line 5.

may be used, according to understanding of the term "thalweg." A third possibility is to use axial distances, measured on straight lines where the axis of the valley is reasonably straight and on curves of the valley where the axis is that of an existing train of meanders. On any of the three reckonings, a modest estimate of the lengthening of Cotswold streams is that specified by a factor of 1.2; to compensate fully for this lengthening, the vertical difference between sources and mouths would need to increase in the same proportion (fig. 5).

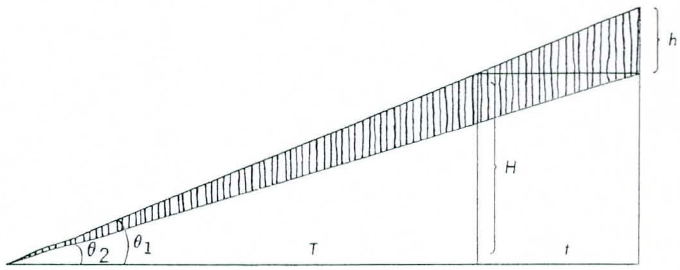


FIGURE 5.—Diagram illustrating hypothetical infilling as compensatory for lengthening of trace. Let T and H be, respectively, the length of trace and the total descent of a stream before the onset of underfitness; then $H/T=\tan\,\theta_1$, where θ_1 is the angle of downstream slope between source and mouth. If the onset of underfitness lengthens the trace to $(T+t)$, the angle of total downstream slope is reduced to θ_2, which i less than θ_1. The slope will be fully restored if H is increased by h where $H/(H+h)=T/(T+t)$. The required depth of infilling increases in the upstream direction, as shown by the shaded segment.

Now, there is no scope for degradation at the downstream ends, where levels are controlled by confluence with the trunk Thames, and all compensation would therefore need to be performed by infilling on the upper reaches. As shown in table 1, the depths of fill required by full compensation for lengthening of trace are great.

TABLE 1.—*Depths of infill required to compensate, on Cotswold streams, for the lengthening of trace induced by reduction to underfitness*

Stream	Approximate height, in feet above sea level		Vertical difference, in feet	Required depth of infill at source,[1] in feet, determined as ×0.2 vertical difference
	Confluence	Source [1]		
Cherwell	190	575	385	77
Churn	260	700	440	88
Coln	240	650	410	82
Evenlode	200	450	250	50
Dorn (to Evenlode)	200	650	450	90
Glyme (to Evenlode)	200	650	450	90
Leach	235	600	365	73
Windrush	210	750	540	108
Dikler (to Windrush)	210	750	540	108

[1] Present source: use of heights at the actual valley heads, upstream from the presently dry-head reaches, would increase the vertical differences and the required depths of fill.

No such deep fills have been located in the field. Indeed, the fills in some Cotswold head valleys are distinctly shallow (Dury, 1958, figs. 3–6). The fills on the Itchen increase in depth in a downstream dir tion and appear, moreover, to do so regularly. In tuality, there is a considerable range in depth of from valley to valley and also in the level of the floor the present channel relative to that of bedrock (fig.

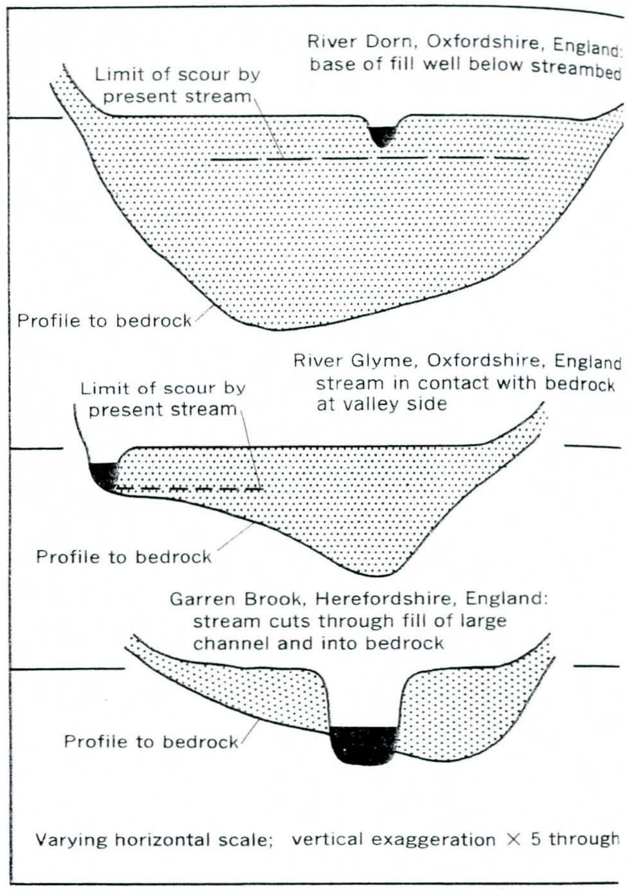


FIGURE 6.—Profile sections of English rivers showing relation of channels to bedrock. Views are downstream.

The Cotswold Dorn is separated from bedrock in valley by 14 feet of sediment, whereas more than 20 of sediment intervene between the base of the pr channel of the Rib and the gravel that lines the un lying large channel (fig. 4). The Cotswold Coln a downstream end of the investigated reach is now sc ing the bedrock at the base of its flood plain, anc Garren Brook of Herefordshire has cut through its ley fill and occupies a bed floored, and partly walle bedrock in place (fig. 6).

In the Rib valley the peaty layer, which is some 10 feet below the surface of the flood plain, s to indicate a pause of filling during which swamp e lished itself. In this respect, the investigated can be called aggraded. In general, however, a dis tion between vertical and lateral accretion does seem justified by the evidence. Vertical filling in s seems to be no more than the counterpart of The reduction in discharge at bankfull—y, in this context, at the frequency associated with ull flow—that made rivers underfit is thought to involved constriction of the large channels in both s while still providing scope for additional filling continued or subsequently resumed downcutting, ding to site.

rtical and lateral accretion are both possible where ll is remarkably weak. Epping Forest, on the ern outskirts of London, provides illustrations of circumstance every fall. Dead leaves shed copi- by hornbeam and beech collect in the channels all streams in the forest; where fallen timber acts ms, the leaves accumulate both at the bottoms and e sides of the ponds, lining channels very similar mensions to those immediately above and below onded reaches. The beds and banks of channels ed in this manner consist, indeed, not so much of es of leaves as of a loose mixture of leaves and r. Although infilling here results directly from ming and may thus be classed with aggradation, it rtheless provides useful illustrations of possibili-

Parallel illustrations come from a drainage ditch, miles to the northeast on the till plains of Essex, re the stream nourished by base flow has been ob- ed to meander between banks of algae. Given a ked reduction in bankfull discharge, vertical accre- on the bed seems as practicable as lateral accretion he banks. Accordingly, the infilling of the large nels ought not to be identified as aggradation unless tops of the fills surpass in height the banktops of the e channels.

UNITED STATES

he fieldwork of 1960 began in the general neighbor- d of Washington, D. C., with augering trials in ll valleys in the Piedmont, in the eastern borders the Appalachians, and on the Coastal Plain. The ial object was to test equipment and technique inst a new set of conditions and to compare those ditions with the conditions of the English Plain. sults of the first trials were in the main limited and elpful, partly because the selected streams of the dmont and the Appalachians are working close to lrock and partly because their alluvium includes vel, cobbles, and boulders. Because the fills are llow, any filled channels that may underlie small ers cannot easily be distinguished from simple flood ins. Bed material of streams 10–20 feet wide com- nly includes fragments measurable in inches along e *b* axis, whereas temporary sections revealed cobbles inches in diameter in the alluvium of the Piedmont and boulders 1½–2 feet across (originating as congelifracts?) in that of the Appalachian foothills. Only near the heads of small estuaries of the Coastal Plain in the Delmarva Peninsula was the hand auger passed significantly lower than the base of the stream channels. The greatest depth below this base reached in any one borehole was 12 feet. Further penetration was prevented by gravel, and the base of the fill was not reached. Hand augering, so well adapted to the English Plain, was accordingly discontinued in the Washington region.

Trials were next made in the Driftless Area of Wisconsin, where manifestly underfit streams are numerous and where deep valley fills are reported. The Driftless Area is a plateau cut in Cambrian, Ordovician, and Silurian rocks, among which dolomite and sandstone are predominant. Dips are gentle; many of the steep-sided valleys are well capable of preserving the forms of valley meanders, and stream meanders occur in trains and are actively migrating along many flood plains. Debates on the number, identity, and origin of erosional platforms (see Trowbridge, 1921; Martin, 1932; Bates, 1939) do not immediately concern the present investigation.

Two examples of manifestly underfit streams from this region had already been cited by the writer (Dury, 1954, figure 1b, c), and the Kickapoo basin had been discussed by Bates (1939), who though recording a fill 50–125 feet deep and describing terraces, supposed the underfit character of the Kickapoo River to result from the sudden introduction of a flood plain by aggradation. Reconnaissance in 1960 and reference to well logs suggested that fills are indeed bulky, but many streams were found to resemble those of the Washington district in that they transport bed loads of gravel. Although the flood plains of the Driftless Area are typically silty and penetrable without difficulty by either type of hand auger, the cherty gravel that occupies the bottoms of many valleys cannot be drilled without power tools. A bucket auger will collect and raise fragments of rock, but boreholes cave rapidly below depths of 1 foot or so; no experiments were made with casing. Powered rigs seemed likely to involve difficulties of negotiation with farmers and also practical difficulties of access, for the determination of cross profiles usually requires drilling on both sides of a stream and, in some places, drilling into swamp. In these circumstances, a series of tests was performed with seismographic equipment light enough to be carried by hand across the streams. First results justified extension of the work, which, as will now be described, enabled four profiles to be determined. The seismic equipment also proved effective in valleys where the fill consists partly of quicksand; this material, although penetrable by auger to depths of about 20 feet, makes working very difficult by closing-in the boreholes and by clinging to the shaft.

SEISMIC REFRACTION

The principal seismic work was executed in the valley of the East Pecatonica near Argyle, Wis., and the valley of Mineral Point Branch near Mineral Point, Wis. (South Wayne quadrangle, Wisconsin-Illinois, 1:62,500, and Mineral Point quadrangle, Wisconsin, 1:24,000). As the tabulated results have been given elsewhere (Dury, 1962), the present account will be limited to a description of the equipment used, an outline of the two sites, and a summary of the results obtained.

The equipment consists essentially of a portable oscilloscope powered by a 12-volt wet battery. Operation requires a team of two—one to observe and record and one to operate the tamper by which shock waves are generated. The tamper is connected by cable to the oscilloscope; a jump switch in the handle closes the circuit when the tamper is beaten on the ground, causing a wave train to flash across the screen of the instrument. Delay time to a selected point on the wave train is read from a dial that controls the position of a marker on the screen. Shock waves are generated at selected distances along a graduated tape pagged on the line of sounding, which normally runs upvalley or downvalley—that is, at right angles to the desired line of cross profile.

In the vicinity of Argyle, the flood plain of the present river is some 200 feet below the level of the nearby plateau top. Immediately northwest of the town is a valley bend devoid of upstanding core, succeeded upstream by a second bend in the center of which the core survives (fig. 7). This second bend, in the angle made by the East Pecatonica and the tributary Mud Branch, was the one chosen for seismic exploration. As there is little or no channeled surface drainage, this abandoned loop seemed likely to retain in a sensibly intact condition any fill that might occupy a large channel. B cause bedrock is accessible both on the meander co and on the steep outer slope of the curve, the site pro ised to allow some augering as a check on seismic o servations. Three lines of sounding were run as a fi trial, and 2 complete lines of cross section were subs quently determined by means of 22 lines of soundi and checked by 8 boreholes (fig. 8).

The sections indicate that a large channel is prese descending as low as 35 feet beneath the present su face of the ground. Its fill consists partly of fine- medium-grained sand and partly of silt; where auge ing was possible, the sand appeared to underlie the si Infilling of the old loop appears not quite comple for the ground is slightly hollow along the very botto of the cutoff; however, it is difficult to decide what a lowance should be made for downwash either from t hillside directly or from the remnants of silty terra that line some parts of the outer curve. Like certa patches of terrace on Mounds Creeks (to be describ below), these remnants appear to have originated loess, although the precise mode of their emplaceme is in doubt. On any reasonable view of the effects downwash, the large channel seems to be not less th 800 feet in width at the minimum—that is, it seems be more than 7 times as wide as the present strea between banktops.

No augering was possible on Mineral Point Branc because the silt of the flood plain is underlain by coar cherty gravel. Immediately upstream from the sound reach, the valley is widely opened and parallel side but at the upstream line of sounding, at an inflecti of the valley, it narrows to a width of 330 feet acr the surface of the flood plain. The downstream line near the extremity of a valley bend, where the flo widens to about 600 feet. After 3 trials, 12 lines sounding were run; and most were checked by reve spreads, connecting spreads, or both.

On the upstream section, where bedrock is expos on both flanks of the valley, a large channel some 3 feet wide descends as low as 39 feet beneath the surfa

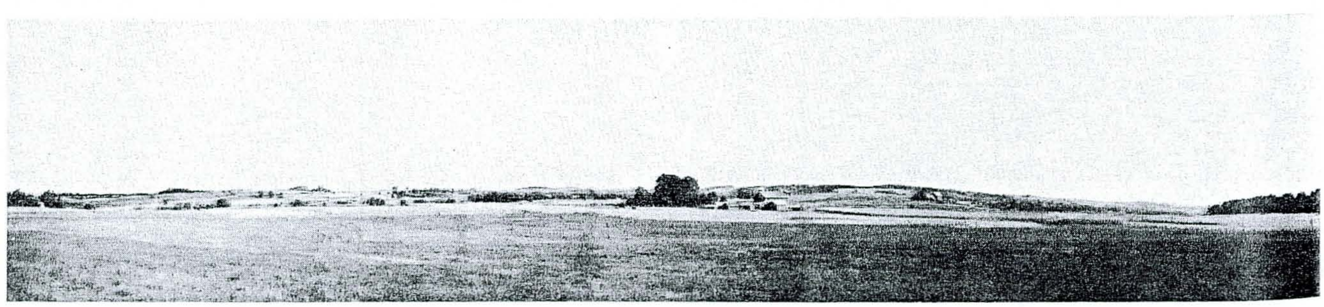
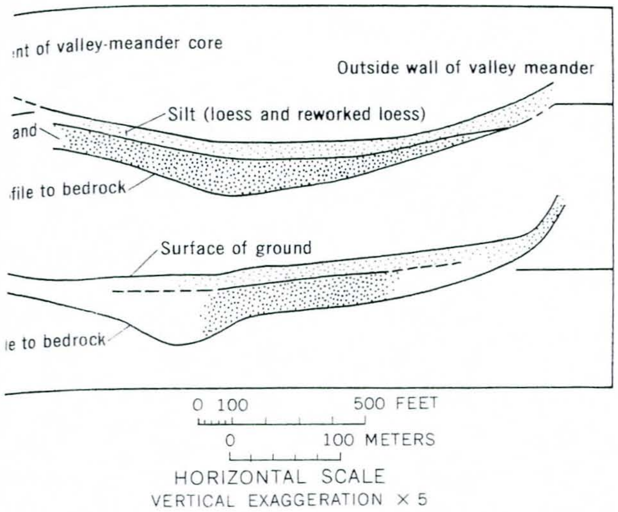


FIGURE 7.—Panoramic view of cutoff valley bend near Argyle, Wis. Hill in center, middle distance, is meander core of bedrock; foreground is floor of cutoff.



ɪE 8.—Profile sections of large channel of the East Pecatonica River near Argyle, Wis. Views are downstream.

he flood plain (fig. 9). The present limit of scour ears to be some 10 feet below the flood plain surface, ging by the depth of a pool excavated in 1960 by h discharges at a particularly abrupt bend. If, how-:, the dark bed in the silty alluvium that overlies the vel at a depth of about 4 feet below the flood-plain ɪl corresponds to the buried first-bottom soils iden-ɪd by Happ (1944) on the Kickapoo River, then maximum depth from the flood plain of the base the large channel was but 35 feet in the middle of ; century. This is still not to say that the flood-plain face of 1850 corresponded to the bankfull level of large channel when it was occupied by the former ɜam, but a width of a little less than 300 feet appears ɔbable, no matter where the former bankfull level ɜstablished.

On the downstream section (fig. 9), the large channel broadens, deepens, and becomes directly asymmetrical in cross section. Its maximum depth below the surface of the flood plain is 45 feet, reached close to the outside of the valley bend. The cross section defined in figure 9 by the profile to bedrock undoubtedly includes a massive point bar similar to those on other rivers where the filled channels widen greatly at valley bends.

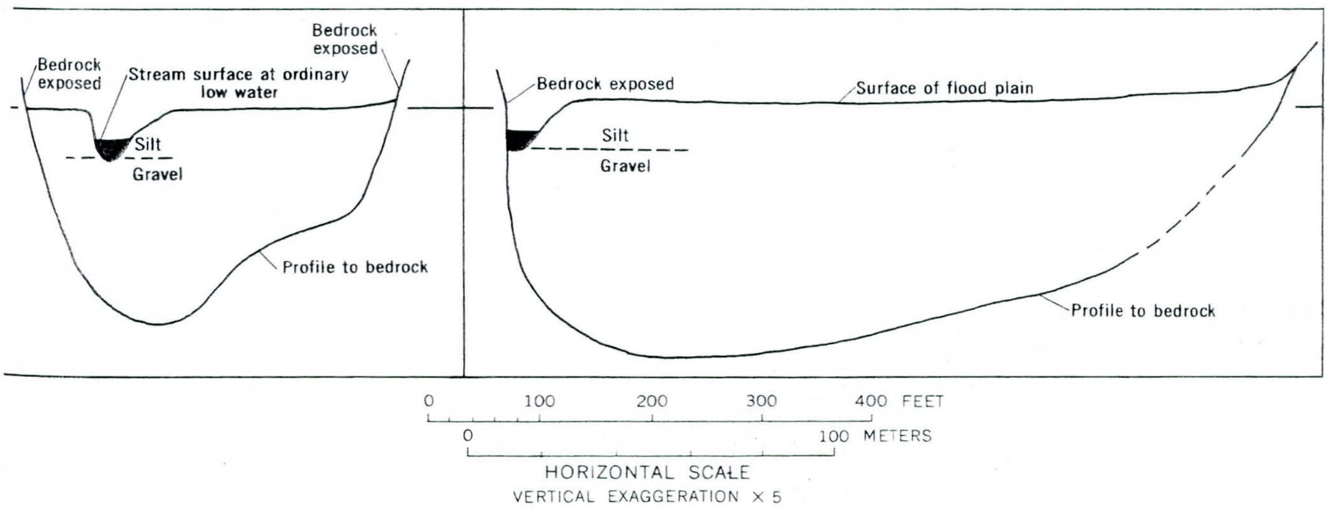
Both Mineral Point Branch and the East Pecatonica near Argyle are manifestly underfit. Large channels occur on both that are similar in their relation to valley bends and their disparity with present channels to the large channels proved on the English Plain. No stream now flows in the cutoff loop near Argyle; but on Mineral Point Branch, where a stream exists above the valley fill, it seems impossible that the fill is subject to reworking except in its topmost part, which supplies the present bed load. Here, as in most of the investigated valleys of England, conversion to underfitness has been accompanied by infilling and by insulation of the stream from the underlying bedrock.

ALLUVIAL SECTIONS

BLACK EARTH CREEK, WISCONSIN

Hand augering proved effective in the valley of Black Earth Creek and the neighboring valley of Mounds Creek, Wis. Ten alluvial profiles were determined from 93 boreholes on Black Earth Creek in the 6 miles between Cross Plains and Black Earth (fig. 10).

Whereas many valleys in the Driftless Area describe bold well-formed incised meanders, the valleys of Black Earth Creek does not. If the valley of Black Earth Creek contains large meandering forms in depth, these are concealed by a deep fill of surficial material.



ᶠIGURE 9.—Profile sections of large channel of Mineral Point Branch, East Pecatonica River, near Mineral Point, Wis. Views are downstream.

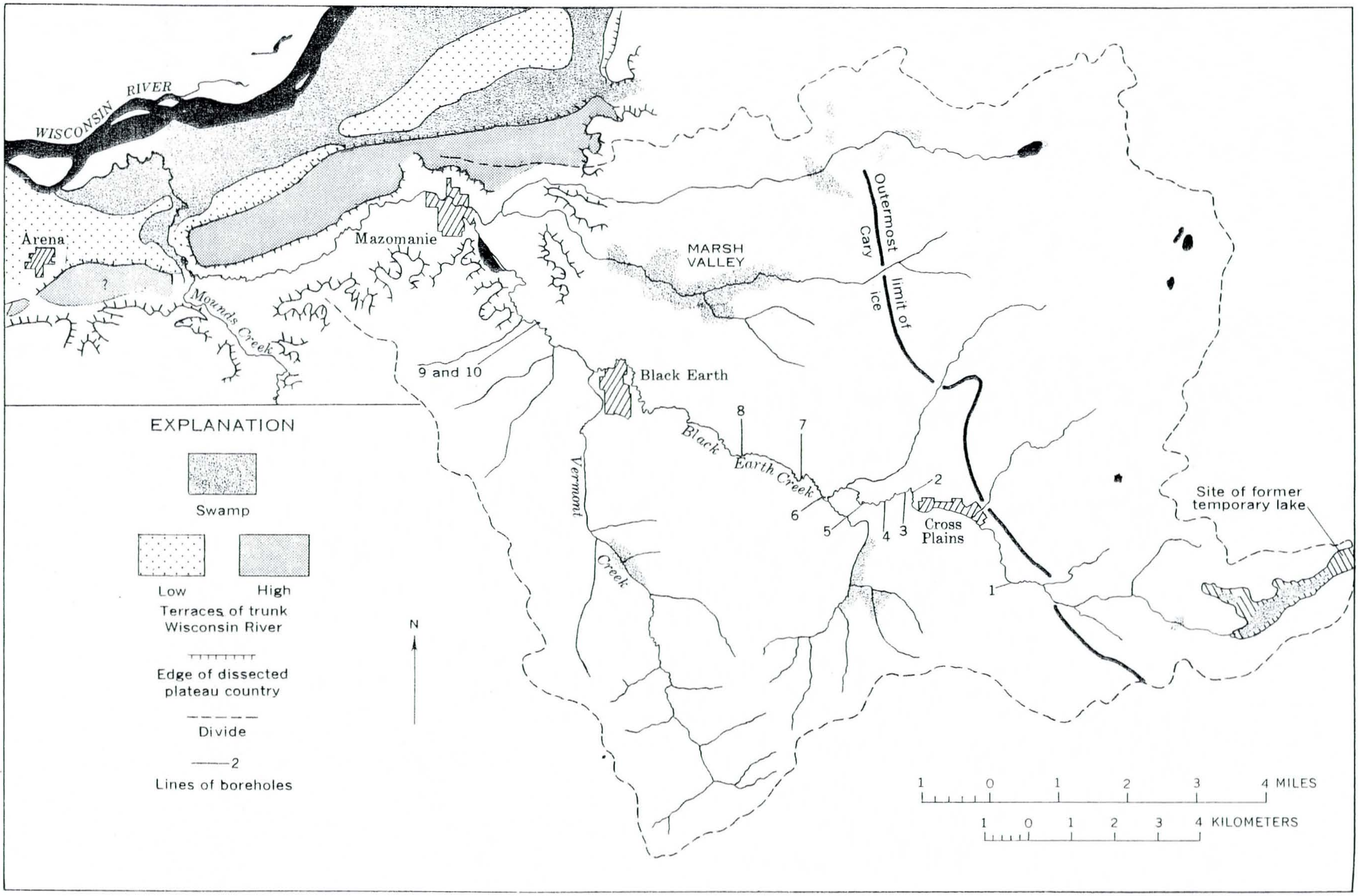


FIGURE 10.—Index map of the area of Black Earth Creek, Wis.

aight elements in the local valley system probably ılt from structures that strike roughly southwest, st, or west-northwest (Judson and Andrews, 1955; yl and others, 1959, p. 31 ff.). As the Precambrian ; at most a few hundred feet below sea level, linear ments in the structures of exposed Cambrian and dovician rocks are not surprising; subdued hills ıking Black Earth valley rise but to 1,050 to 1,200 t. The bounding divide is capped in places by Ordoian dolomite, but the bulk of the hills consists of dovician or late Cambrian dolomite and sandstone ich are carved into steep valley walls that rise ruptly from the valley floors. Although its valley is nost straight, Black Earth Creek displays two sets of anders. The existing river winds on a flood plain ıt is itself sinuous (fig. 11). Black Earth Creek is ınifestly underfit.

Now certain fixes on the local scale of chronology, ıich are easily obtained by reference to the sequence d distribution of surficial deposits, make Black Earth eek particularly suitable for detailed study. During e maximum of the Wisconsin Glaciation, the trunk isconsin River was a major outlet for melt water and twash, which it led across the Driftless Area to the ississippi River near Prairie du Chien. The outerost limit of Wisconsin ice is marked locally by the hnstown Moraine (Alden, 1918), which is referable the Cary Stade as understood by Thwaites (1943) ıd probably to the Valparaiso Moraines in the Woodrdian substage of Frye and Willman (1960). Names of glaciations and stades are less significant here than the placing of the Johnstown ice stand not long before the Two Creeks Interstade. Ice standing on the line of the Johnstown Moraine discharged melt water and outwash not only along the Wisconsin River valley but also along certain feeder valleys, including the Black Earth valley (fig. 12). Backwater flooded the downstream ends of some laterals, depositing sediment in them. The topmost surface of the fill constitutes the High Terrace of the Wisconsin River (MacClintock, 1922).

The Johnstown Moraine crosses the Black Earth valley about 2 miles east of Cross Plains, and the valley downstream of the moraine is thickly infilled with outwash. Alden (1918) conjectures that the fill may be 250 or 300 feet thick in the Wisconsin valley near Mazomanie; the grestest known depths of fill in Black Earth valley are 130 feet in a well at Black Earth village, 45 feet (unbottomed) at Cross Plains, and 50 feet reported for the gravel workings between Cross Plains and Black Earth. Because all three sites are close to the valley wall, the stated thickness are minimal, as is the 30 feet somewhat dubiously indicated by seismic readings (fig. 13). The seismic-sounding sites were near the left-hand [1] edge of the flood plain, about halfway between Black Earth and Mazomanie, and as shown on line 10 in figure 10.

Not all the fill of Black Earth valley need be of Cary (Valparaiso) age, but the topmost part certainly

[1] Right hand and left hand, here and throughout, are referenced to the downstream direction.

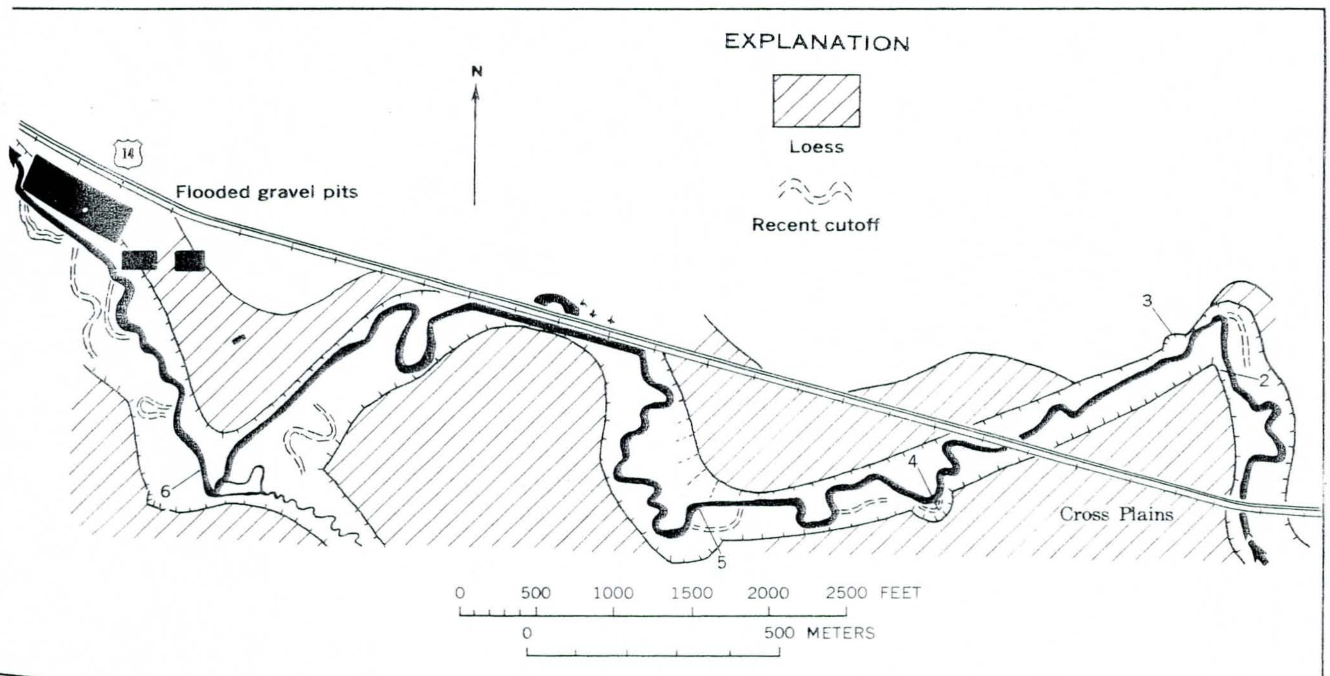


FIGURE 11.—Sketch map of part of Black Earth Creek showing valley meanders. Mapped from aerial photographs and planetable survey.

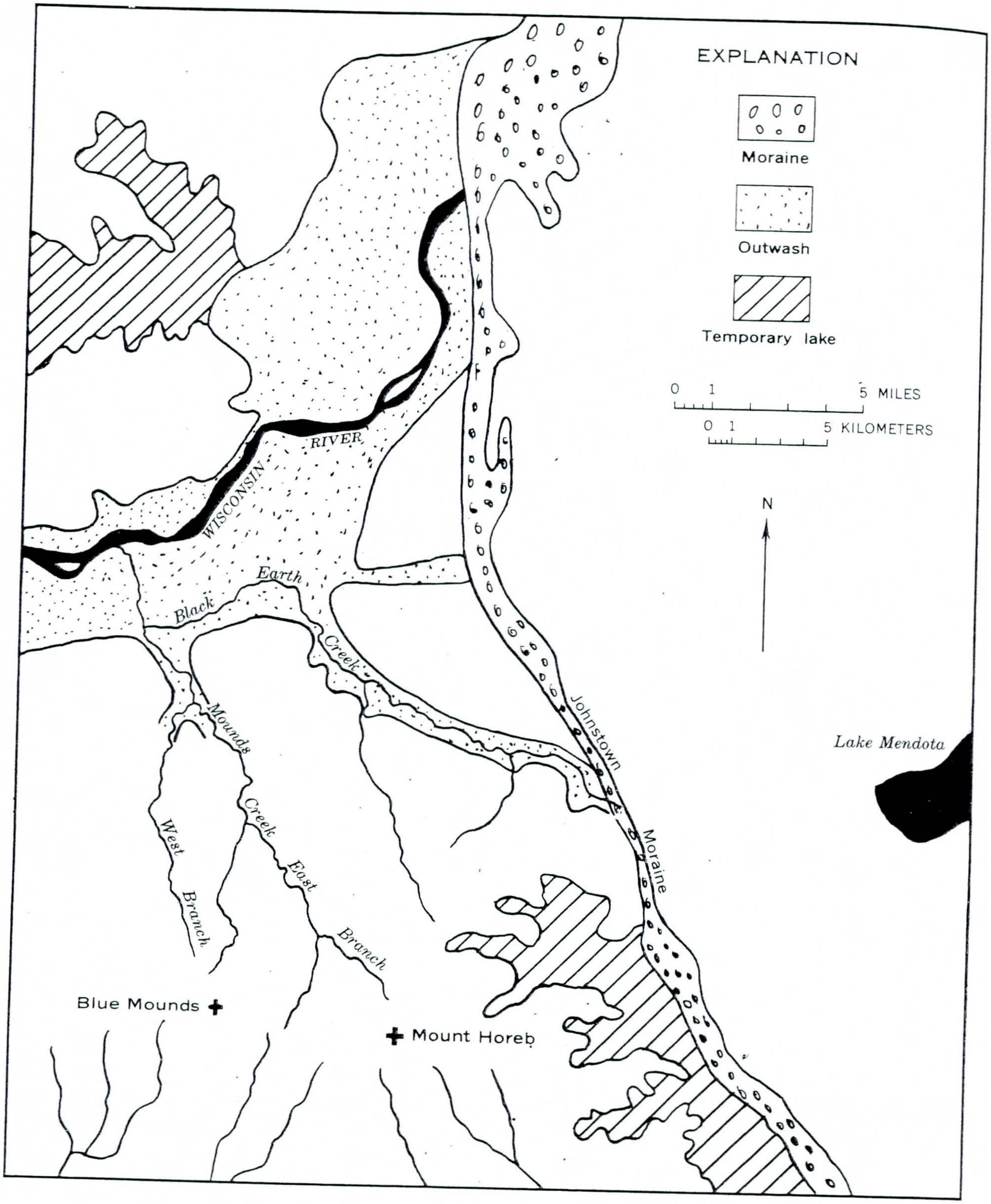
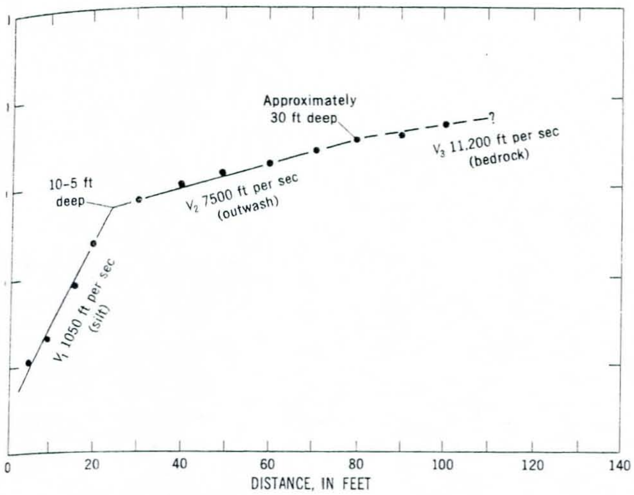


FIGURE 12.—Sketch map showing relation of ice front to Black Earth Creek and Mounds Creek, Wis.



JRE 13.—Graph showing relation of time to distance in seismic fraction sounding at Black Earth Creek, 55 feet from left-hand lge of flood plain midway between Black Earth and Mazomanie.

The continuity of Johnstown Moraine with the out- sh train was well shown in 1960 in the gravel work- on the north side of U.S. Highway 14. So much wash passed down Black Earth valley that the uths of laterals were choked. Silt accumulated ckly behind the obstructing sand and gravel and ntually provided the foundations of swamp. In rsh valley, which is tributary at Mazomanie, 8 feet peat has been proved to overlie 33 feet of silt, which urn rests on surficial sand (Mississippi Univ., 1957). e middle reaches of the valley of Vermont Creek notably peaty, and a sizable peat bog occurs in the ge feeder valley next downstream from Cross Plains the left-hand side of Black Earth Creek. The mouth the Black Earth valley may have been similarly oked as the trunk Wisconsin valley was infilled, but wash in Black Earth valley was deposited thickly ough to provide a continuous downstream slope from Johnstown Moraine at about 900 feet above sea level the top of the High Terrace near Mazomanie at about) feet above sea level. The one possible sign of ob- uction is the right angle that Black Earth Creek kes where it leaves the plateau, but this angle may ult from nothing more than the growth of a levee on Wisconsin River—that is, it could be associated th a deferred tributary junction.

So far as is known, the topmost surface of outwash Black Earth valley is remarkably flat in cross section cept beneath the flood plain, where it has been slightly oded. The flatness is probably referable to the re- rking and deposition of outwash by a braided stream nilar to many of the streams that now discharge from fronts. Cessation of outwash and abandonment of e braided habit cannot be dated precisely to the withdrawal of ice from the Johnstown line but cannot have been long delayed thereafter. The first withdrawal produced a temporary lake at the head of the valley about 4 miles east of Cross Plains (fig. 10); the site is now marked by a drained bog that is crossed by the present divide near its east end. Although melt water could still have passed into the valley by way of the lake, and presumably did so, there is nothing to show that the flat outwash surface relates to the lake level of about 920 feet above sea level rather than to the stand of ice at the Johnstown line. The outwash is covered across the whole width of the valley by loess which, lying banked against the bases of the valley walls, rises onto and passes across the hummocks of the Johnstown Moraine. At least in part, therefore, this loess postdates the withdrawal of Wisconsin ice from its local extreme limit. By the time loess deposition was completed, the ancestral creek was flowing not in braids but in large meanders. These are developed in, and cut slightly through, the loses. They are thought not to have been cut by a melt-water stream and thus to postdate in origin the last discharge of melt water and outwash. If a braided stream flowed from the temporary lake, they postdate the lake also. The view that the creek which developed and incised the large meanders was not a stream of melt water is supported not merely by comparison with existing melt-water streams but also, as will be seen shortly, by the behavior of the neighboring Mounds Creek and the mode of development of certain laterals of Black Earth Creek itself.

Because the outwash train in the Black Earth valley is correlative with the High Terrace of the Wisconsin River, it is neither necessary nor possible to follow Alden (1918) in ascribing the High Terrace to a substade earlier than that responsible for the Johnstown Moraine. Reference of the large meanders of Black Earth Creek to a time rather later than the Johnstown ice stand is supported by the relation between Black Earth Creek and the High Terrace near Mazomanie. Immediately northwest of this settlement, a large abandoned bend cuts into the High Terrace. Traces of other comparable loops may exist farther downstream, but the lower part of the creek is so strongly rejuvenated that little of such loops remains. The floor of the single intact bend, however, lies more than 30 feet below the level of the High Terrace and falls below the projected line of the profile above Mazomanie (fig. 14), evidently having been cut after the lower part of Black Earth Creek was rejuvenated by the fall of the trunk Wisconsin from the High Terrace level. Without very detailed leveling it is impossible to say what traces of the Low Wisconsin Terrace may exist along the lower, strongly rejuvenated part of Black Earth Creek, but the

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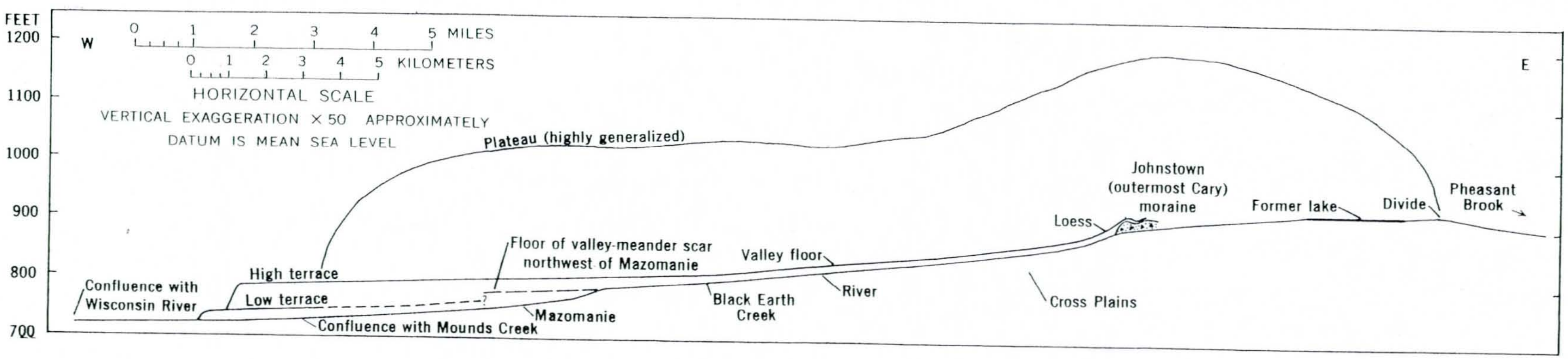


FIGURE 14.—Longitudinal profile of Black Earth Creek, Wis.

idence available is at least sufficient to show that the rge meanders persisted here after the High Terrace d been abandoned.

The large meanders of Black Earth Creek are more gular than many of the windings of the present chan-l (fig. 11), and their slip-off slopes can be identified places, mainly in the reach midway between Black irth and Cross Plains. The trough that contains the esent flood plain is regarded as modified from the rge channel associated with the large meanders. That annel can have been about 10 times as wide as the esent channel, just as its meanders are about 10 times long as the meanders of the existing stream, area for ea of drainage. The 10:1 ration ensures that the esent meanders fit rather neatly into the trough pro-led by the large channel.

Rejuvenation affects the present stream almost as far stream as Cross Plains, so that it is only near Cross ains itself that the base of the former channel can demonstrated to survive. Lines and groups of bore-les drilled at intervals from a point 2 miles upstream om Mazomanie to 1 mile downstream from Cross ains all show the present stream in direct contact th the outwash and tending at most places to shift e outwash (fig. 10; fig. 15, lines 5–10). At lines 9 d 10 the present meanders are sweeping downvalley ile at the same time tending slightly to incise them-ves. Consequently, the surface of the flood plain scalloped by the tiny curved bluffs of recent cutoffs, d the top of the outwash is distinctly irregular in tail, as shown by the profile on line 10. Although ctions through recent cutoffs suggest that fragments insported through the present meanders are as much 2 inches across, much of the outwash is far coarser an this and includes blocks 1 foot or more long. Cob-es removed from some existing pools accumulate as rs on the downstream limbs of meanders. Islet build-g or incipient braiding occurs 2 miles above Mazo-inie and again at 2 miles and 3 miles above Black irth village. Within the large right-hand loop at is last site—that is, between lines 6 and 7 (fig. 15) d opposite the flooded gravel pits (fig. 11)—come the st obviously ingrown meanders of the present series, th the outer bank perceptibly higher to the unaided e than the inner bank. This is not yet the upstream nit of rejuvenation, however. On line 5, at the apex the large left-hand bend next below Cross Plains, the ol of a stream meander descends into the gravel; the rface of the gravel lies 4 to 4½ feet beneath the top the flood plain, whereas the pool goes down to 5½ et. Still farther upstream, on line 4, the present ream has cut into and again retreated from the side the trough. Instrumental leveling discloses a faint slip-off slope and suggests that slight rejuvenation is felt as far upstream as this line.

On lines 6 and 5 (fig. 15), the base of the flood plain is so nearly flat that the present alluvium seems to be contained in a well-defined meander trough. On line 4, however, details of a new kind appear. Borings reveal a broad spread of medium- to coarse-grain sand at the base of the peaty alluvium of the flood plain. Beneath this sand, in a shallow depression, is a layer of clayey silt that reappears in greater thickness on line 2. The vertical succession in this reach is striking—dark peaty silt above is followed downward by jet-black somewhat silty spongy peat; next comes a thin layer of sand that is yellowbrown to gray, wet, and incoherent; and then comes the clayey silt, which is dark on line 4 but light gray, compact, and no more than moist on line 2.

These various deposits are interpreted as follows: The clayey silt is the remnant of a valley fill, contained in the pool of a large former meander; the sand represents the base of the present flood plain; and the high proportion of peat in the lower flood-plain alluvium results largely from growth in place, whereas the silty content of the upper alluvium indicates deposition by the spilling river. This last contention is supported by the presence of faint levees indicated by instrumental survey.

On Black Earth Creek, therefore, as on numerous other streams of the manifestly underfit type, the alluvium of the flood plain rests unconformably on the fill of a large channel. What here remains of the former pool and its fill seems to have been preserved by a combination of favorable circumstances—the very slight extent of the downcutting now in progress, and the resistance of solid rock at the apex of the large bend. At the right-hand end of line 3, the present stream has cut a meander scar into the solid rock of the valley wall, the channel perhaps shifting across a bedding plane. The former large stream was compelled to make an abrupt turn through 90°. It is possible, therefore, that this particular large pool was unusually deep and also that it has been to some extent defended by firm rock in place.

No additional pools have been found higher upstream, but numerous boreholes in the very swampy flood plain immediately above Cross Plains have penetrated gravel at a fairly uniform depth.[2] A line across the left-hand side of the flood plain on the last possible large bend 1 mile southeast of Cross Plains (line 1) produced inconclusive results. Although the present stream bed stands a little higher than the base of the alluvium, powerful

[2] Leveling was impracticable at this site, and no transverse line of boreholes could be run.

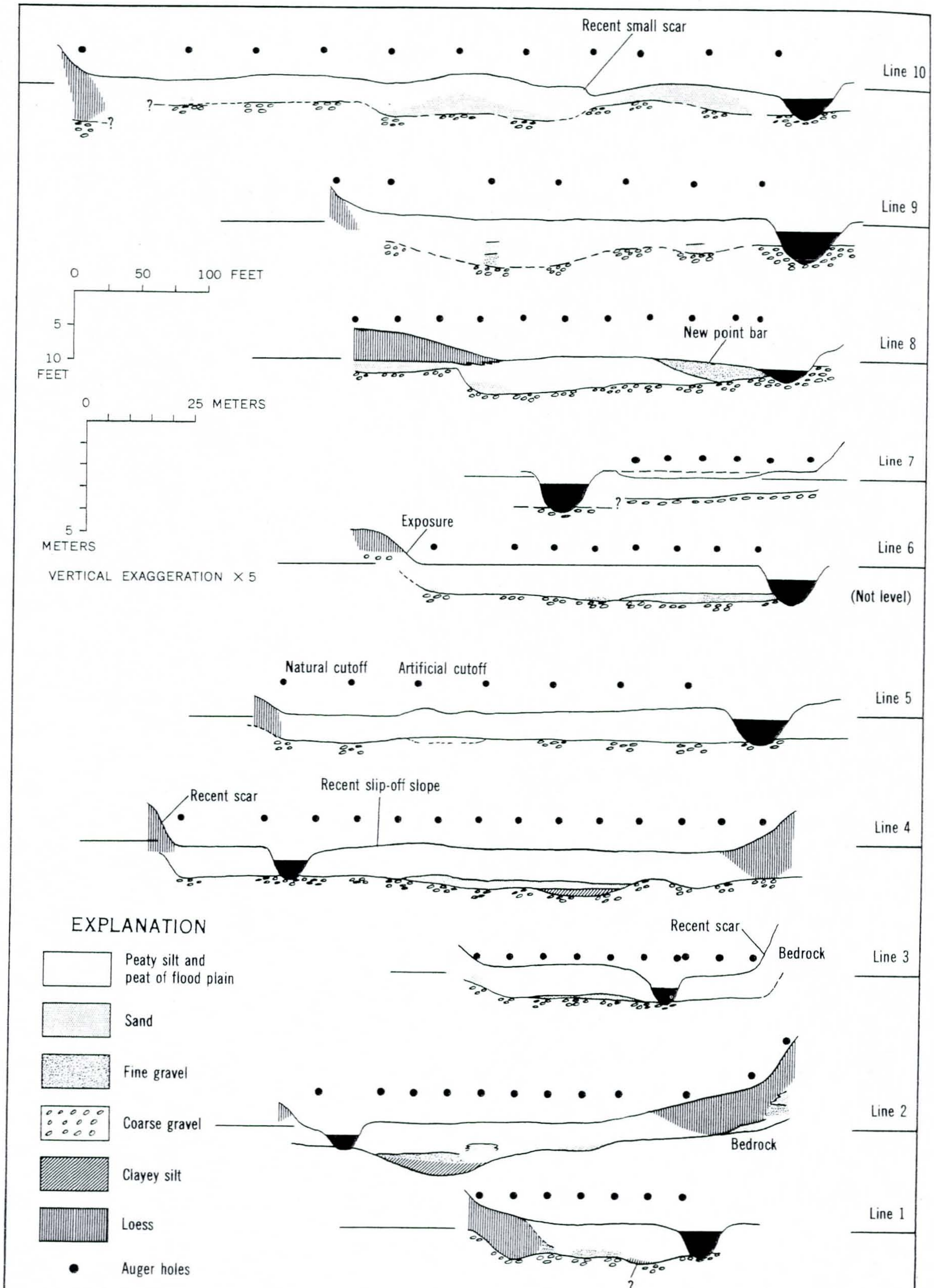
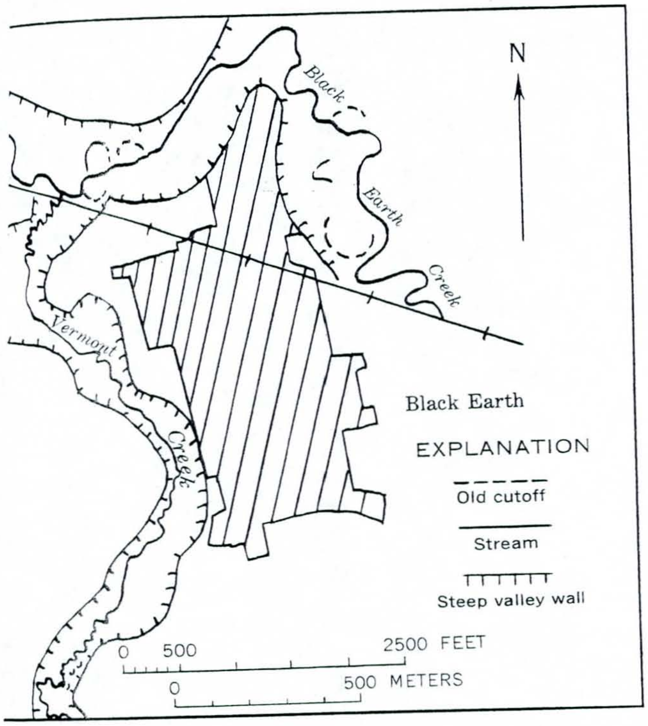


FIGURE 15.—Profile sections showing flood plain and associated features, Black Earth Creek. Views are downstream.

ige from an adjoining hill makes the site very npy; and it is impossible to say whether or not the peaty clayey silt penetrated near the stream repre- s the fill of the old channel or whether it results contamination of present alluvium by loessal ma- l from the nearby hillside. If part of the old chan- loes survive here, then the present stream is fluming ole gravel through a silt-lined bed.

arge meanders occur on some of the laterals of ck Earth Creek, in valleys which were emphatically invaded by ice during the Wisconsin Glaciation[3] . 16). Any suggestion that these valleys may have



GURE 16.—Sketch map showing entry of Vermont Creek, a lateral stream, into Black Earth Creek. Valley bends of Vermont Creek are cut through loess.

een invaded by pre-Wisconsin ice and received large ncrements of melt water at some early time is not elevant to the present investigation; for their large neanders are cut through lateral extensions of the latest oess down to the level of the existing flood plain of Black Earth Creek. Like the large meanders of the runk stream, those of the feeder streams postdate the Johnstown ice stand and the subsequent episode of loess leposition. Moreover, the wavelength of valley mean- lers in the Driftless Area as a whole, at 50 square miles of drainage, is about 8 times that of present meanders; to explain the large meanders of Black Earth Creek, there is no reason to appeal to direct outwash. Those meanders, like the large meanders of lateral streams, constitute a sample of the phenomena typical of the whole region, which are dissociable from the discharge of melt water.

The minimum outline sequence required by the observed landforms is therefore the following:

1. Discharge of melt water and outwash from the Johnstown ice front, with construction of the High Terrace on the Wisconsin River and the correlative outwash train in Black Earth valley; choking of the mouths of valleys lateral to Black Earth Creek.
2. Recession of ice from the Johnstown line, with formation of temporary lake; deposition of loess, continuing after the draining of the temporary lake; conversion of the ancestral Black Earth Creek from a braided to a meandering habit.
3. Considerable lateral growth and slight incision of large meanders.
4. Rejuvenation of the lower reaches, referable to a fall in level of the trunk Wisconsin. (See below.)
5. Reduction in volume (at bankfull), with conversion to small meanders and formation of existing flood plain; continued rejuvenation of lower reaches and slight rejuvenation along much of the river.

MOUNDS CREEK, WISCONSIN

Mounds Creek heads in the high ground near Blue Mounds and Mount Horeb. Its drainage basin was not invaded by ice during the Wisconsin Glaciation (fig. 12) nor is it thought to have been invaded during earlier glacials. In consequence, Mounds Creek serves to demonstrate the landforms produced by periglacial conditions, those which occur very close to the ice front but in the absence of outwash and melt water. Like the valley system of Black Earth Creek, that of Mounds Creek hints very strongly at guidance by structures in the solid; many of its elements are alined from south-southeast to north-northwest or from south-southwest to north-northeast. In their middle and lower parts, the valleys both of West Branch and of East Branch are steep sided and wide bottomed. There is room—but perhaps not ample room—for large meanders, to the trace of which the present meanders are superadded. Like Black Earth Creek, both branches of Mounds Creek are underfit, although they cannot be described as flowing in meandering valleys unless the semblance of valley bends on the upper reaches of East Branch is authentic.

As Mounds Creek received no melt water or outwash from its valley heads, the High Terrace simply invades the valley from the lower end. The Low Terrace

[3] This statement does not prejudice, nor is it prejudiced by, the revisions which Prof. Robert F. Black, University of Wisconsin, is making in the location of former ice fronts on the margins of the Driftless Area.

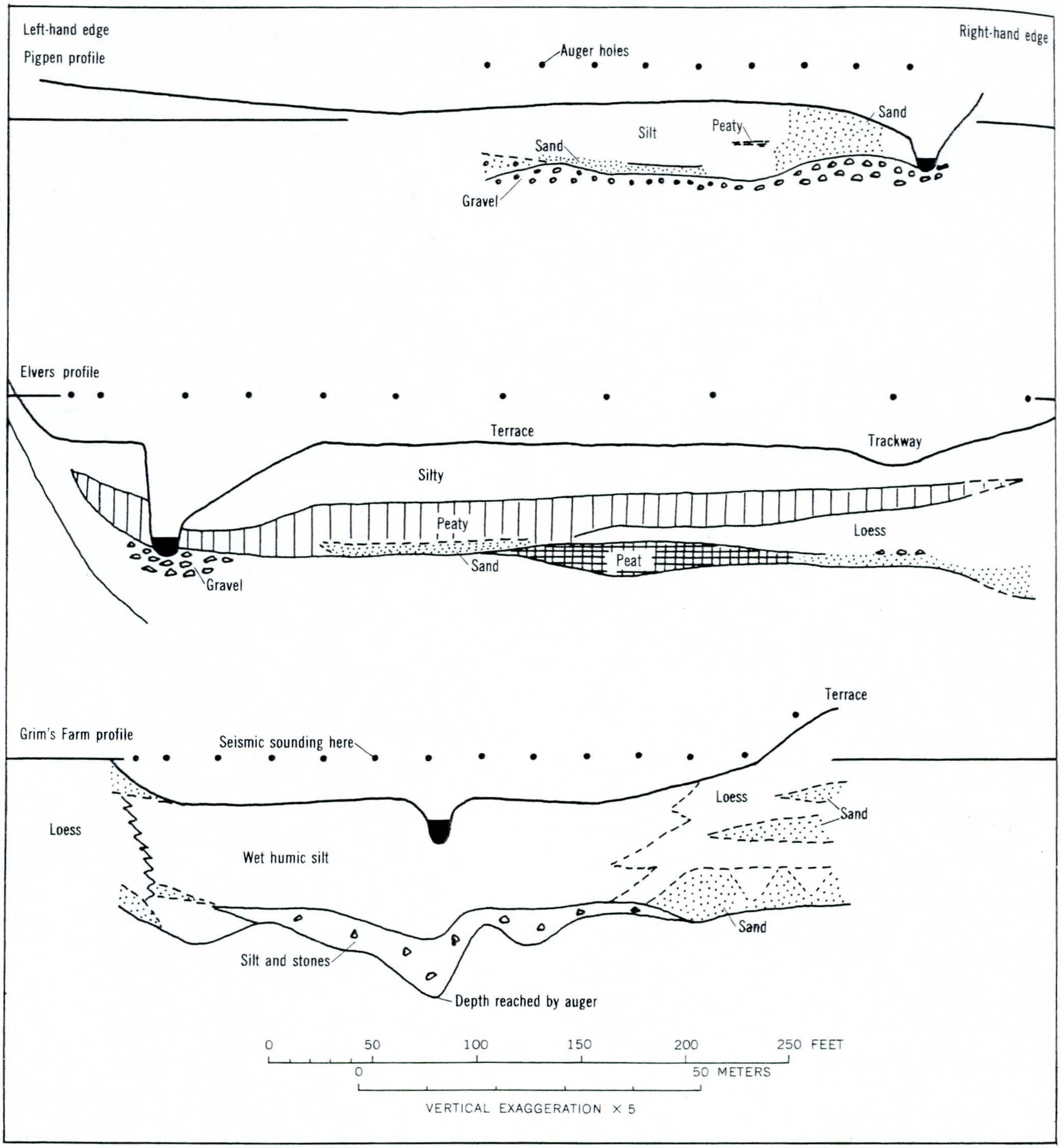
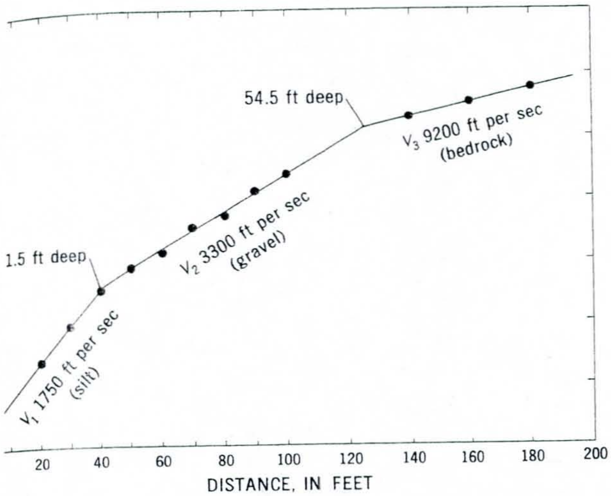
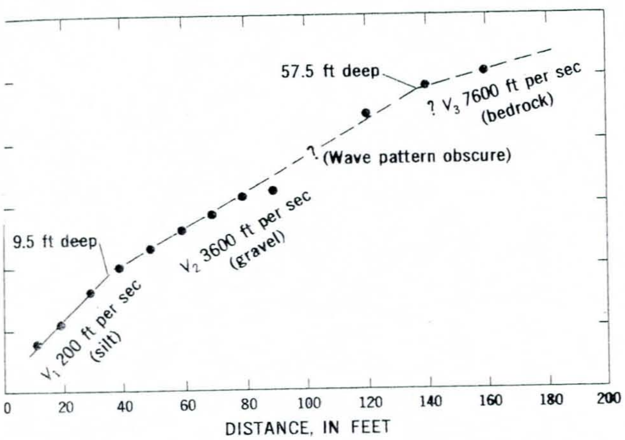


FIGURE 18.—Profile sections of East Branch Mounds Creek, Wis. Views are downstream.



ɛ 19.—Graph showing relation of time to distance in seismic action sounding at Mounds Creek, from valley bottom near n's Farm. Second borehole from left-hand edge of channel, re 18.



RE 20.—Graph showing relation of time to distance in seismic fraction sounding at Mounds Creek (second example). First bore- ıle from right-hand edge of channel, figure 18.

:rrace. Immediately upstream of Orcutt Farm, more- r, a tiny right-hand lateral has cut large meander rs into its own strip of Low Terrace almost down to ı level of the present flood plain (fig. 21). On ›unds Creek, that is to say, the former large meanders ı be seen to postdate the Low Terrace.

Exploratory boreholes at the site of the highway .dge near Orcutt Farm reached a depth of 40 feet fore penetrating bedrock. As the bridge is close to ɜ valley side, the bedrock surface could lie much lower the center of the valley. Both in the main valley of ɔunds Creek and in the valley of the tiny lateral, icksand greatly handicapped augering; it closed in so pidly that no borehole went deeper than 21 feet. owever, it was possible with the aid of 22 boreholes to ace the sharp descent of the right-hand wall of rock depths well below the level of the present streambed ıd to discover that quicksand occurs also beneath the Low Terrace and underlies the lowermost part of the lateral stream to a depth of more than 8 feet. In all probability this quicksand was swilled into the valley of Mounds Creek from the Wisconsin River when the latter was building up its High Terrace.

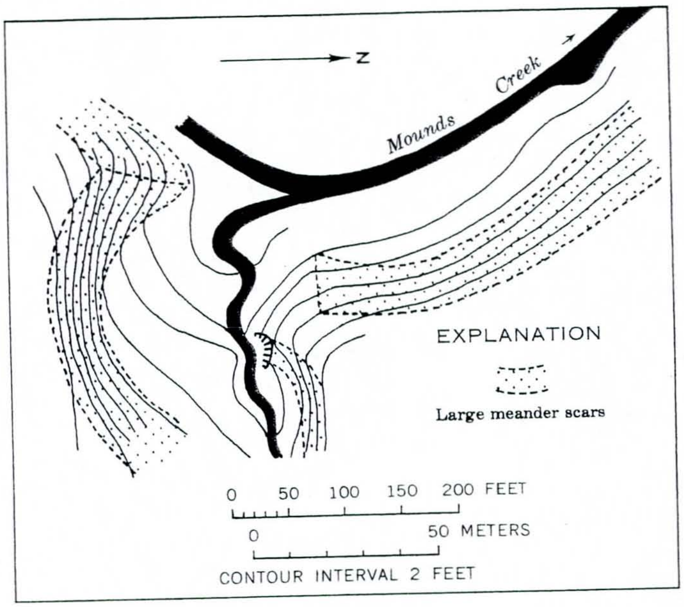


FIGURE 21.—Sketch map showing entry of lateral stream into Mounds Creek.

The outline sequence of development of Mounds Creek thus becomes:

1. Deep cutting of valley in bedrock.
2. Onset of periglacial conditions with the approach of Cary ice; discharge of frost-shattered gravel from the valley sides.
3. Infilling of the Wisconsin River valley and the lower part of Mounds Creek valley by sandy outwash—that is, formation of the High Terrace.
4. Deposition of loess (compare sequence established for Black Earth Creek).
5. Deposition of peat near Elvers.
6. Completion of the Low Terrace by erosion in the lower reaches, but probably also by considerable reworking of loess.
7. Entrenchment of large meanders through the Low Terrace.
8. Reduction of channel-forming discharge, with reduction in size of meanders; scalloping of the fore edge of the Low Terrace by small meanders, accompanied by slight rejuvenation at least as far upstream as Orcutt Farm.

The two sequences, for Black Earth Creek and Mounds Creek respectively, will be referred in a later section to the scale of absolute time.

MISCELLANEOUS ADDITIONAL RECORDS

Reconnaissance trials on the till plains of Iowa provided further evidence of large filled channels, although some locations were not easy to explore. Crooked Creek, near Lime City (Lime City quadrangle, Iowa, 1:24,000, T. 79 N., R. 2 W., sec. 5), describes stream meanders within a meandering valley. Eight inches of light silt covers the flood plain and overlies dark material which in its stratigraphic relations resembles the alluviated first-bottom land described by Happ (1944) in Wisconsin. Below the streambed is dark muck containing medium- to coarse-grained sand and humified fragments of wood; this muck is penetrable with difficulty because of a gravel band at about the level of pools in the present channel. The dark muck, however, continues beneath the gravel and, in one borehole, changes after another 3½ feet of depth to light uniform light-brown loess. Gravel at this location prevents definition of a complete profile; but the indications, as far as they go, match those of the two sections next described.

The upper profile in figure 22 relates to an unnamed right-hand lateral of North River near Winterset (Winterset quadrangle, Iowa, 1:62,500, T. 76 N., R. 30 W., sec. 6, bordering R. 29 W., sec. 1). On the line of profile the stream is cutting slightly into the right-hand bank, where downwash probably obscures the bedrock that a natural section exposes farther downstream. A quarter of a mile from the augered line, limestone and shale are visible to 8 feet above stream level, mantled by weathered till with abundant Kansan erratics; loessic slope wash drapes the whole valley side. Augering proves a large channel that is filled with dark muck, silt, and sand; contains the present stream channel and old cutoff channels; and bottoms either in loess or against bedrock. Taken at the inflection of a valley bend, this profile reveals a disproportion between large channels and present channels similar to that on the upper Rib (see above).

Certainly on this stream, and apparently also on Crooked Creek, the large channel is trenched into loess. The loess of Iowa, however, permits a wider range of possible dates for trenching than does the loess at Black Earth Creek. All that can be said is that the general relations of large and small channels to the spread of loess are similar in all three places.

McDonald Creek (Eldridge quadrangle, Iowa, 1:24,000, T. 80 N., R. 3 E., sec. 24) was used in Professional Paper 452–A to exemplify a pool-and-riffle sequence on the limb of a valley bend. The incomplete profile on the lower part of figure 22 herewith lies near the downstream end of that bend, which cuts through sandy outwash. The steep slope on the right-hand side of the profile in figure 22 is the right-hand turn.

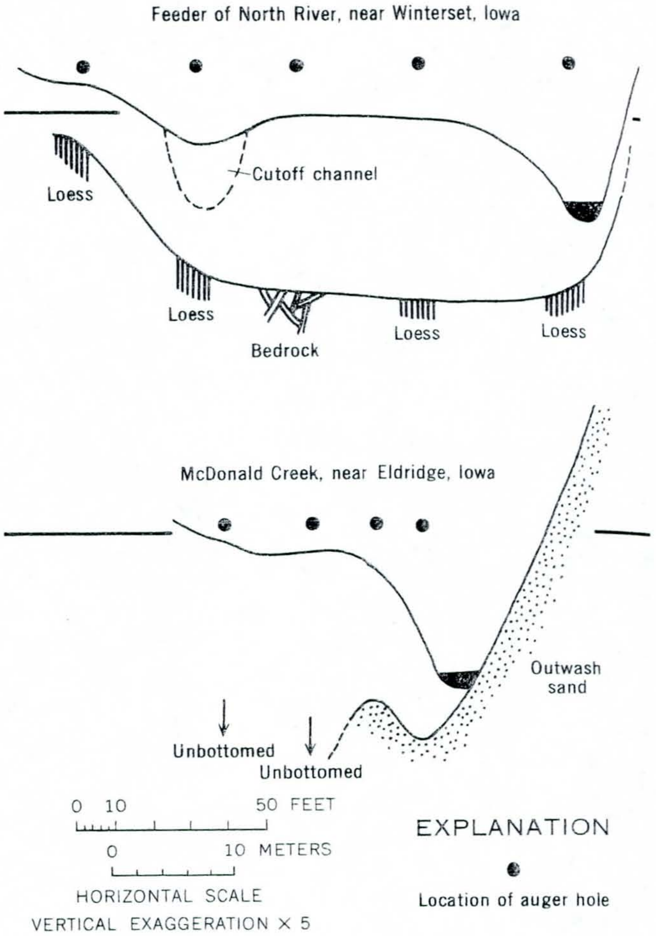


FIGURE 22.—Profile sections of large channels of two streams on till plains in Iowa. From field exploration.

Whereas the outwash sand is bright orange, the fill is dark and in part silty. On the line of profile, the outwash appears to include a berm or point bar of the former large stream. Although two boreholes failed to reach the bottom of the fill, they descended well below the level of the present streambed. This site appears more complicated than a number of others, but it once again reveals part at least of a filled channel that was much larger than the channel of the existing stream.

Exploration by civil engineers, particularly at damsites, is obviously capable of revealing the depth and form of alluvial fills in the valley bottoms of underfit streams. But because many of the records remain unpublished and relate in part to irrelevant instances or to inappropriate lines of cross profile, engineering work supplies less information than might be hoped.

Borings undertaken in conjunction with highway construction in the Ozarks are mainly confined to the centers of valleys. Nevertheless, their results suggest that a number of head valleys contain alluvial fills, even though streams are in contact with bedrock elsewhere, even though recorded depths from surface of flood n to bed rock in the filled valleys are nowhere great. ure 23 illustrates a common type of situation, where th to bedrock beneath the streams is certainly iter than the depth on the step valley sides. Each he two outline profiles resembles profiles determined other rivers by close-spaced augering; at least on nch of Woods Fork Creek, the fill appears too deep the existing stream to scour to bedrock at the bank-stage.

'he Yellowstone damsite in Lafayette County, Wis. s. 24, 25), lies in a meandering valley with steep s but shallow fill. Here again, however, the 20 feet o of fill seems too great for the present stream to ır. As the explored site lies at the inflection of a ey bend, either the former large stream had an un-ally high width to depth ratio or it widened its bed he course of downstream shift.

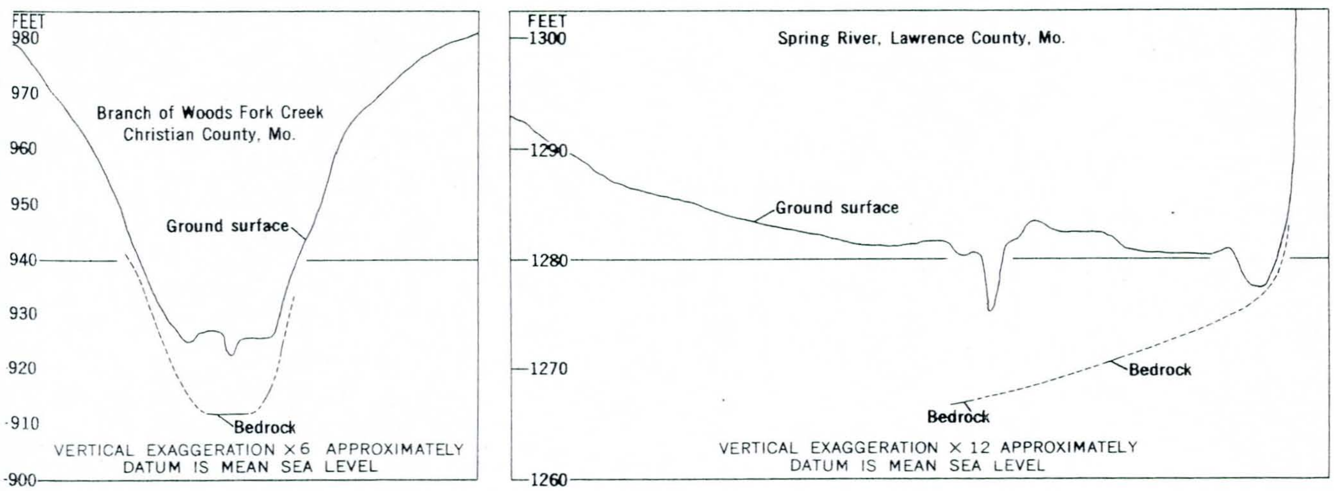
)rilling logs from the damsite at Governor Dodge te Park, Wis. (fig. 26), are too few and too general-l to supply a detailed profile. Nevertheless, the ıl conditions strongly resemble those described else-ere and become more clearly understandable by com-ison with the Mill Creek site illustrated below. In State Park, a large channel trenches the St. Peter ıdstone; the alluvial fill of the channel contains al gravel and interfingers with the loess or loessic vnwash of the valley walls. Unless the present ɜam assumes the improbable cross section that would mit it to scour 20 feet below the flood plain, it can-now reach bedrock.

Records from the Mill Creek damsite, Wisconsin (fig. , are especially valuable in identifying loess and luvium, showing how the colluvium wedges into the in the valley bottom, and recording that the loess ms low terraces. This site recalls profiles taken on Mounds Creek (Elvers; Grim's Farm) and near Argyle. (See figs. 8, 18.) The large channel beneath Mill Creek is partly filled with loess. In addition, however, the likely profile of the base of the loess suggests a channel intermediate in size between the channel cut in bedrock and the channel of the present stream. A possible sequence of development is the following:

1. Incision of valley, cutting of large channel into bedrock.
2. Shrinkage of stream, partial infilling with colluvium, cutting of intermediate channel (by somewhat re-enlarged stream?).
3. Loess fall, filling of intermediate channel (implying renewed shrinkage of stream); loessic alluvium spread as flood plain.
4. Slight rejuvenation, fill of intermediate channel terraced.

Variations on this sequence, or elaborations of it, are easy to make. Suggested correlations with the outline sequences for Black Earth Creek and Mounds Creek occur in a later section. For the present, it suffices to observe that Mill Creek illustrates the common circumstance that by the time of the last heavy loess fall in Wisconsin the deep incision of valleys into bedrock had already taken place. The slightly rejuvenated condition of the present stream belongs, perhaps, with the widely observed epicycle of slight erosion now in progress.

Logs for the Brighton damsite, Maryland, on the Patuxent River, supply a complete profile across the valley (fig. 28). In the valley bottom is a trench, about five times as wide as the present channel of the Patuxent, cut into residual rock waste and containing sand and gravel below an upper layer of finer alluvium. Although the Patuxent, in common with many other rivers that cross the Piedmont, displays poorly developed or



'IGURE 23.—Profile sections of valley bottoms of two streams in the Ozarks. From civil-engineering records. Views are downstream.

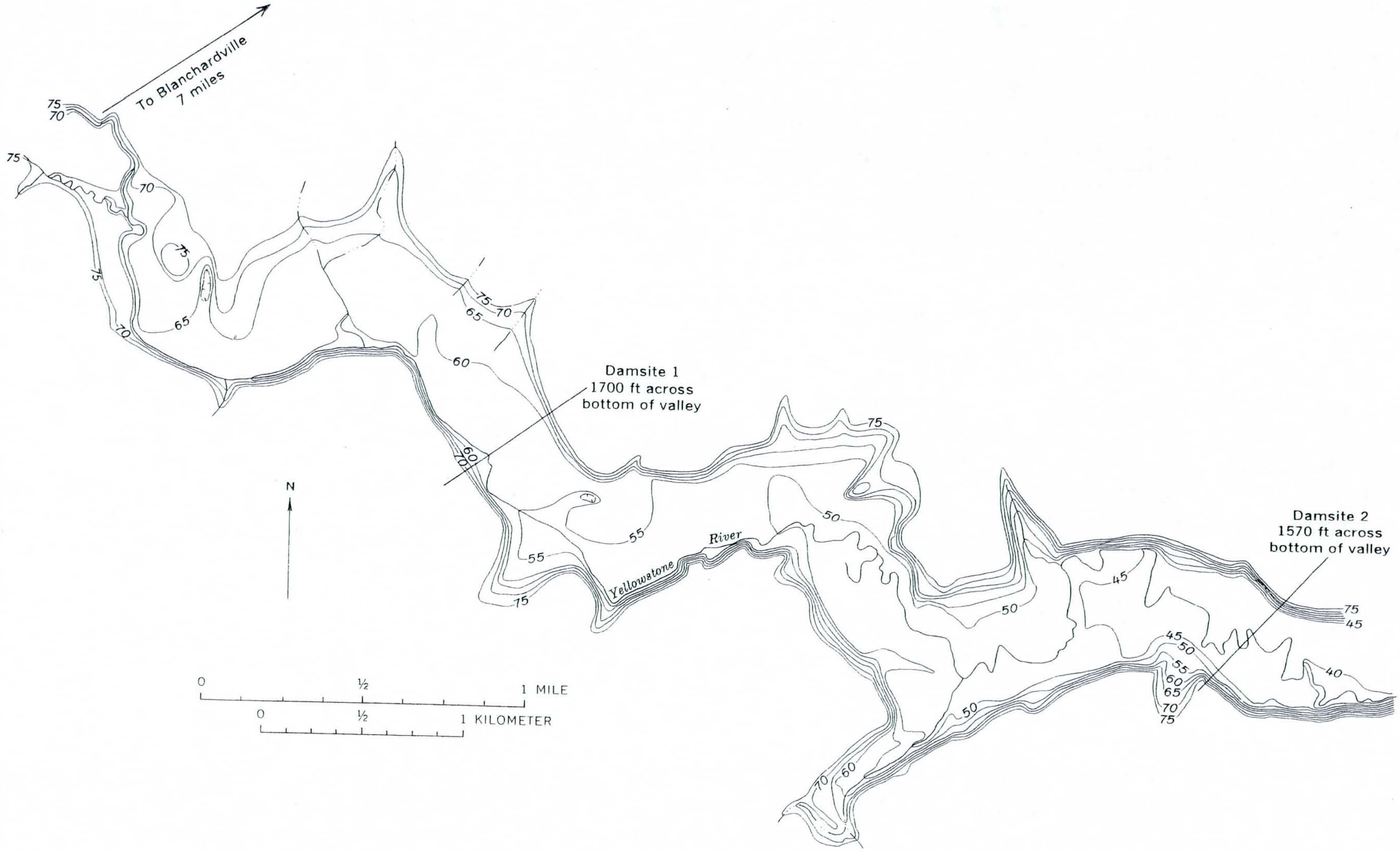
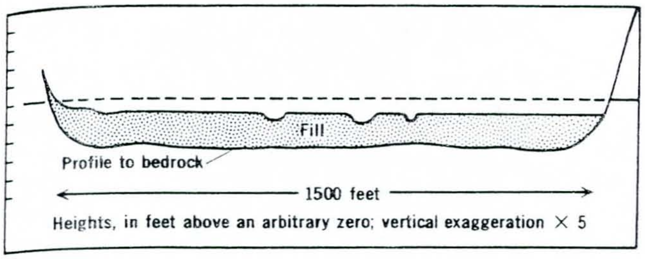


FIGURE 24.—Contour map of the Yellowstone damsite, Wisconsin. Data from Wisconsin Conservation Dept.



:GURE 25.—Profile section of damsite 2, Yellowstone area, Wisconsin. See figure 24 for location of section. View is downstream.

o stream meanders, its valley bends in some reaches re cut boldly and incised deeply. The 5:1 ratio beveen the width of filled trench and that of present iannel is identical with a widely distributed wavength ratio between the valley meanders and stream ieanders of manifestly underfit streams. The filled ench is accordingly taken as a former stream channel, it when bankfull discharge was greater than it now ;. Without study of the present river, the coarse basal art of the fill cannot positively be distinguished from ie existing bed load—which, in fact, it could supply, ist as outwash gravel supplies bed load to the present Black Earth Creek. But the difference in caliber etween this material on the one hand and the bulk of ie uppermost alluvium on the other at least invites omparison with the widely reported coarseness of ormer alluvium and the fineness of material underoing transport today. The residual material beneath he fill suggests that, in this reach, the large Patuxent f former times failed to scour to bedrock in the latest tage of its history.

None of the instances specified here relates to infilled pillways. All six rivers drain basins that remained e free at glacial maximum. The profiles exemplify hat, in the writer's view, should be a common circumstance: if underfit streams, whether manifestly underfit or not, are regionally developed, then alluvial fills in former channels should be widespread. The profile described by Lattman (1960) is a case in point. That author shows that Beaverdam Run, Cambria County, Pa., occupies a valley where the fill is 20–25 feet deep whereas the stream is but 4–4½ feet deep at bankfull. A deposit of rounded boulders as much as 1 foot in diameter, resting directly on the bedrock floor, can represent the bed load of the former large stream. The writer consequently inclines to reject Lattman's view that infilling resulted from aggradation, especially aggradation resulting from forest clearance. Beaverdam Run closely resembles Mineral Point Branch of the East Pecatonica where, upstream from the reach explored by seismic means, a reach widened and straightened by the downstream sweep of valley meanders has been infilled; Lattman's cross profile closely resembles the profiles described here for the Brighton, Governor Dodge, and Mill Creek damsites. Conditions of this sort appear to be very widely represented in head valleys throughout the Driftless Area and can be matched repeatedly in small valleys on the English Plain.

CHRONOLOGY

Three specifications of time apply to underfit streams: that of the initiation of large meanders or large channels, that of the onset of underfitness and the abandonment of large meanders or channels, and that of duration between initiation and abandonment. In practice, initiation can often be dated solely by reference to an erosional platform or a depositional spread, with no certainty that large meanders existed upon the platform or on the sheet of sediment. Streams that now possess incised valley meanders need not have been meandering

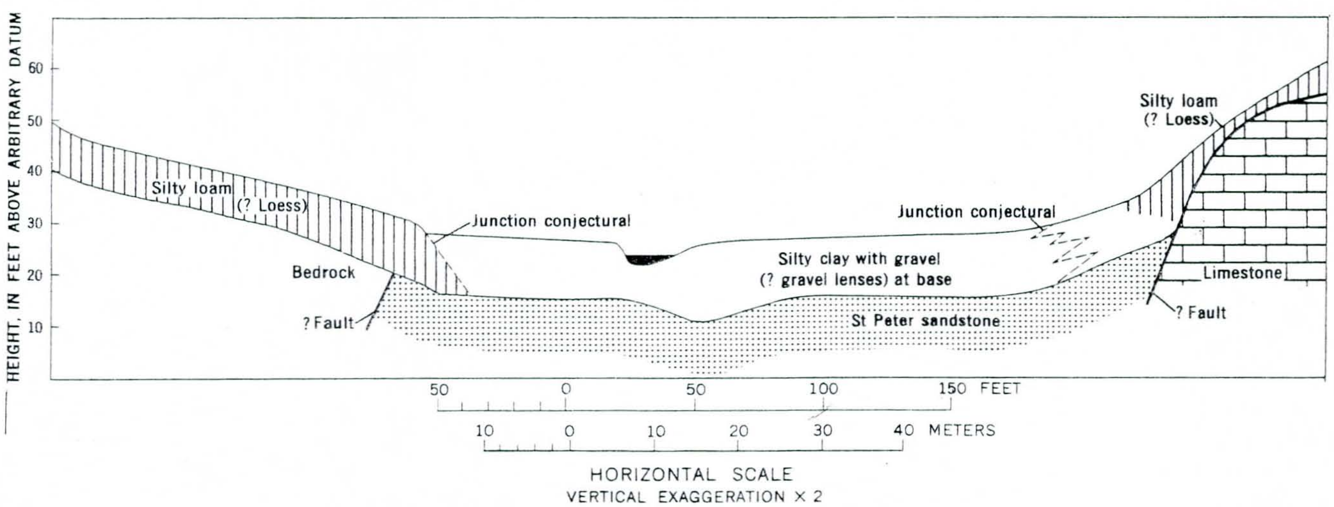


FIGURE 26.—Profile section of damsite in Governor Dodge State Park, Wis. View is downstream.

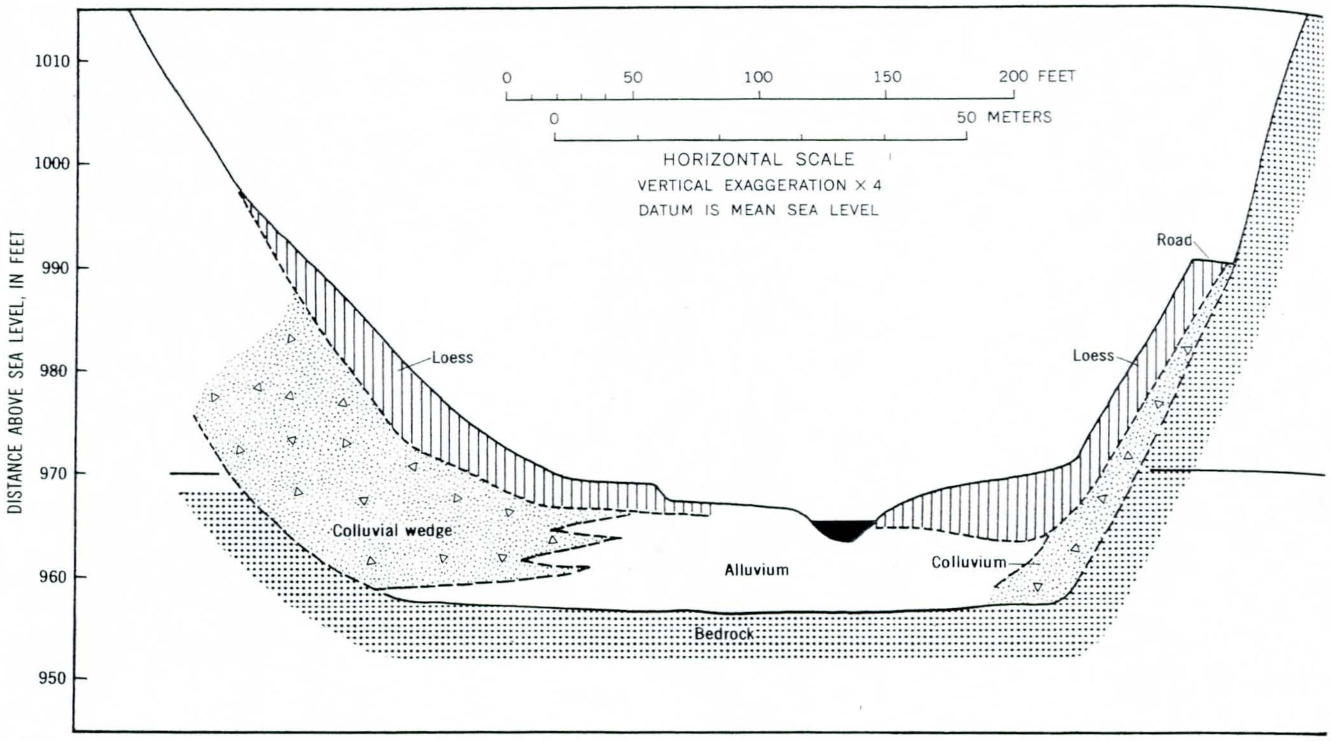


FIGURE 27.—Profile section of Mill Creek, Richland County, Wis. Site 10, drainage area 2.10 square miles. View is downstream.

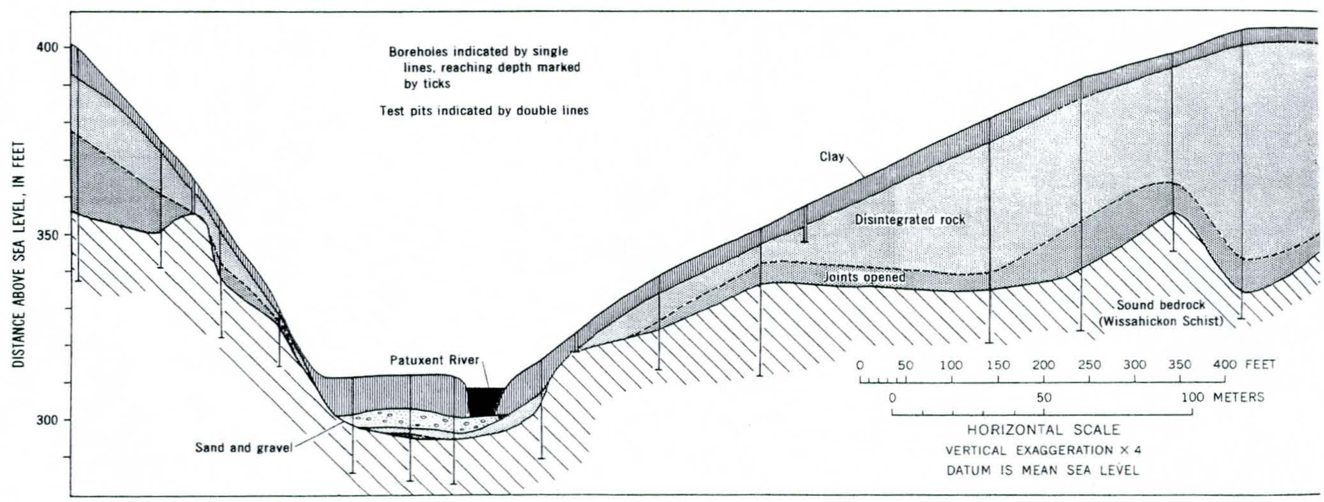


FIGURE 28.—Profile section of valley of the Patuxent River, Md., at Brighton Dam. Data from records of core borings and test pit View is downstream.

streams when they flowed at (or near) the level of plateau tops. Platforms and sediments, therefore, frequently do no more than set limits before which the relevant trains of valley meanders could not have existed. On the other hand, terraces contained in meandering valleys, especially those formed as point bars on the ingrowing loops, can fix points on the time scale when large meanders were certainly in being. In this way the dubious gap between incision of stream and appearance of large meanders can often be mu narrowed.

Duration, expressed as the span between initiati and abandonment, does not imply that large meand or large channels continued in use throughout any sp defined. On the contrary, the climatic hypothesis gi strong general reasons to suppose that the last onset underfitness was but one episode of several. Especial is this so, as the event which has hitherto been treat single will now be shown to have been multiple; the ord of cutting and filling in a number of valleys re-res the term "last onset of underfitness" to connote, actuality, a complex series of fluctuations.

Not all the instances described below supply dates—ether relative or absolute—for initiation, duration, l abandonment of large meanders, but overlapping uences are useful in this context as in stratigraphy general. To relate one sequence to another, however, ne common scale of reference is needed; and because span of time involved in abandonment is that from well-marked glacial maximum to the present day, stions of nomenclature, succession, and absolute ing at once arise. The scales of stratigraphy and ie used in this report derive respectively from the al scheme of Blytt (1876) and Sernander (1910) l from radiocarbon dating. As shown in table 4, scales allow for parallelism, if not for strict simul-eity, between events in Europe and those in North nerica. In the form stated, they rely on the findings Firbras (1949), Deevey (1949, 1951), Gross (1954), iiliani (1955), Suess (1956), Barendsen, Deevey and alenski (1957), Wright (1957), Deevey and Flint)57), and Fairbridge (1961). The use of a general ema is not meant to prejudice any evdience of op-sed or compensatory change according to location, for instance in northwest Europe as against the Med-rranean; but, as will appear, considerable similarity ms to be indicated for the main episodes of channel-; and filling, respectively, in northwest Europe and the United States.

In addition to scales of reference, discussion also re-ires some rather general terms indicative of the lo-ion of events in some bracket of the whole sequence. r example, Godwin (1956, fig. 29) places the Lower = Older) Dryas, the Allerød, and the succeeding pper (= Younger) Dryas in the late glacial, extend-; the postglacial from the beginning of the pre-real to the present. Deevey and Flint (1957) reject terms "postglacial climatic optimum," "thermal ximum," "altithermal," and "megathermal" and pro-se "hypsithermal" for the time interval embracing llen Zones V through VIII of the Danish sequence or real through sub-Boreal in the sequence of climatic ases. Morrison and others (1957) defend the use of e term "Recent," whereas Frye and William (1960) ply Recent to their post-Wisconsin time—that is, to e last 5,000 years. Cooper (1958) calls the whole span om Zone III (Younger Dryas) onward the Neother-l (see also Antevs, 1953), subdividing it into Anather-al (Zones III and IV), Hypsithermal (V through [II), and Hypothermal (IX). In the text which fol-ws here, the well-documented rise of temperature during hypsithermal times (the hypsithermal maximum) will be regarded as belonging to peak interglacial conditions. Accordingly, the whole span between the last preceding maximum of glaciation on the one hand and the hypsithermal peak on the other will be styled "deglacial." No difficulty will arise from questions of the number and relative severity of stadial maximums within the last (Wisconsin, Weichsel) glacial, as all that is needed is a term indicating the net trend of amelioration through some 15,000 years or through a shorter period during which ice sheets or proglacial lakes persisted in northern latitudes. It is principally in the context of deglacial time that the last recorded abandonments of valley meanders and large channels will be discussed.

EXAMPLES OF EARLY INITIATION OF MEANDERING VALLEYS

The examples immediately following are those of meandering valleys where the first incision has been dated to early Pleistocene times, where a sequence of terraces extends well back into the Pleistocene, or where a dated record of slight excavation contrasts with the presumably much longer span of time needed to carve the whole trains of deeply ingrown bends. In all instances, the total possible duration of cutting is long, on the Pleistocene time scale.

In discussing the Belgian River Ourthe, Alexandre (in Macar and others, 1957) holds that free meanders appeared above the level of the highest terrace—that is, at the beginning of Pleistocene time and upon the last incompletely developed erosion platform of the local Tertiary sequence. His opinion that this platform formed in a subarid climate need not be examined; as Alexandre observes, the earliest series of meanders is incompatible with such a climate, whatever the conditions immediately before their appearance. At the present time, the Ourthe lies 300 feet or more below the flat adjacent summits. Its history includes the downstream sweep of some large meanders, although sweep through a whole wavelength characterized single loops rather than complete trains. The valley, therefore, retains much of its original meandering trace, modified by ingrowth. Alexandre neglects to state that the present stream in some reaches is manifestly underfit (his fig. 1).

Seret (in Macar and others, 1957) distinguishes 11 successive flood plains on the River Lesse, a tributary of the Meuse; he correlates his succession with that on the trunk stream and accepts shifts in climate as the dominant cause of spasmodic downcutting. Lateral migration of the whole river, distortion by structural guidance during incision, and certain derangements of drainage obscure the earliest traces of the large meanders; but they had certainly appeared by the time Seret's terrace 4 was formed (his fig. 6)—that is, at a height of 200 feet above river level in the lower reaches.

Troll (1954) shows that certain incised meandering valleys in the Hercynian massifs of Europe and in the Alpine Foreland came into being early in the Pleistocene. Kremer (1954) refers the highest terrace of the Moselle to infilling during the Gunz (=Early, Nebraskan) glacial; she places the initiation of the large cutoff loop near Kommlingen in this glacial, so that the history of incised valley meanders on the Moselle spans much of glacial time. Ingrowth rather than downstream sweep ensures that the meandering trace of the incised Moselle is in general well preserved, although modified in places; terraces provide the means of dating single episodes of cutoff (fig. 29). Both on the Moselle and on the other rivers that Troll examined, the record is one of alternate braiding in times of cold and meandering at other times; braiding was independent of the presence or absence of ice in the headwater basins. Although Troll calls the great incised bends valley meanders, he appears not to use this term in contradistinction to stream meanders and to overlook the essential disparity between the two series, perhaps because most of the rivers which he considers are not manifestly underfit. He seems, by implication, to ignore the possibility that braiding in cold periods may have been a response to infilling by congeliturbate which, charging the valley bottoms with incohesive and coarse material, inhibited the retention of single meandering channels.

According to Peltier (1949), the Susquehanna River scours during high stages without cutting down to bedrock. Matters are different in the lowermost 30 miles above tidehead, for which reach Mathews (1917) describes islets of bedrock in midstream and long spoon-shaped depressions in the channel floor. In neither part is a meandering trace well, or at all, developed on the present channel. Conditions in the reach discussed by Peltier seem to resemble those described above for the Patuxent at Brighton Dam.

Terraces, preserved chiefly on the insides of valley bends, record little but slight ingrowth since an episode of filling correlated by Peltier with the Illinoian glacial maximum. Greater coarseness of terrace material than of the present alluvium indicates, in Peltier's view, a reduction in the magnitude of floods since the terraces were formed. Such an inference is independent of invasion of the upper basin by ice, which occurred at least once during the Illinoian and three times during the Wisconsin Glaciation; for the 15-foot terrace of North Branch, which was not fed directly from ice, is also coarser than the present alluvium. Pelt cludes that filling occurred during cold peri which he posits ET (Tundra) climate (Köppe fication) over much of the Appalachians. A the widespread evidence of severe frost action with the charging of coarse debris into the bottoms, it does not suggest a means of trans the materials so delivered; transportation req change of regimen, such as that which Peltier Conclusions about the Susquehanna thus acco those about European rivers, especially when al is made for the already incised condition of the hanna at the time when Peltier's earliest (=Ill terrace was formed.

There is no reason to think that the geomorp tories of the North and South Forks of the Shen have differed in any major way. If not, then t incised bends on the North Fork extend the tions of King (1949) on the South Fork near Va. King finds that the valley floor of the Sout on the east side of Massanutten Mountain, mainly of a series of gravel-topped benches, w refers to alternate erosion and deposition dur Pleistocene. Three series of gravels occur: a gravel, 300 to 700 feet above river level, that t laterally into coalescent fans at the foot of the tain; a thin intermediate sheet resting on benches; and a younger sheet, also thin, that terraces 50 to 75 feet above the river. Becaus residual clays are present beneath the gravel separates the origin of the existing loops from ting of the so-called Valley Floor Peneplain; cause many of the benches occupied by the y gravels lie within the loops of entrenched (meanders, he considers that the meanders did n until near the end of the period when the inter gravels were laid down. However, the Nort above Strasburg is cut as much as 250 feet be floor of the Great Valley of Virginia—a floor although now much dissected, appears formerly been sensibly intact and gently sloping. Be airline miles upstream from Strasburg, remnant floor extend along spurs in the incised bends, s ing that the loops already existed here—althou developed laterally than they now are—when t flowed very little below the general level of the floor. (See also Hack and Young, 1959.) Inc a depth of 250 feet or more along the whole le a sizable reach indicates a long history of Nevertheless, a local history of ingrowth by the North Fork does not oppose the sweeping a large meander trough on part of the South Fo does it deny the possibility that both rivers alt

	NAME OF TERRACE	SERIAL No	CORRELATIONS WITH GLACIAL SEQUENCES		
			ALPINE	CONTINENTAL U.S.	GENERAL
	HOHEN	7	GÜNZ	NEBRASKAN	EARLY
	OBERE HAUPT	6			
	MITTLERE HAUPT	5	MINDEL	KANSAN	ANTEPENULTIMATE
	UNTERE HAUPT	4			
	OBERE MITTEL	3	RISS I	}ILLINOIAN	PENULTIMATE
	UNTERE MITTEL	2	RISS II		
	} NIEDER (in part subdivided)	}1 b a	}WÜRM	}WISCONSIN	LAST

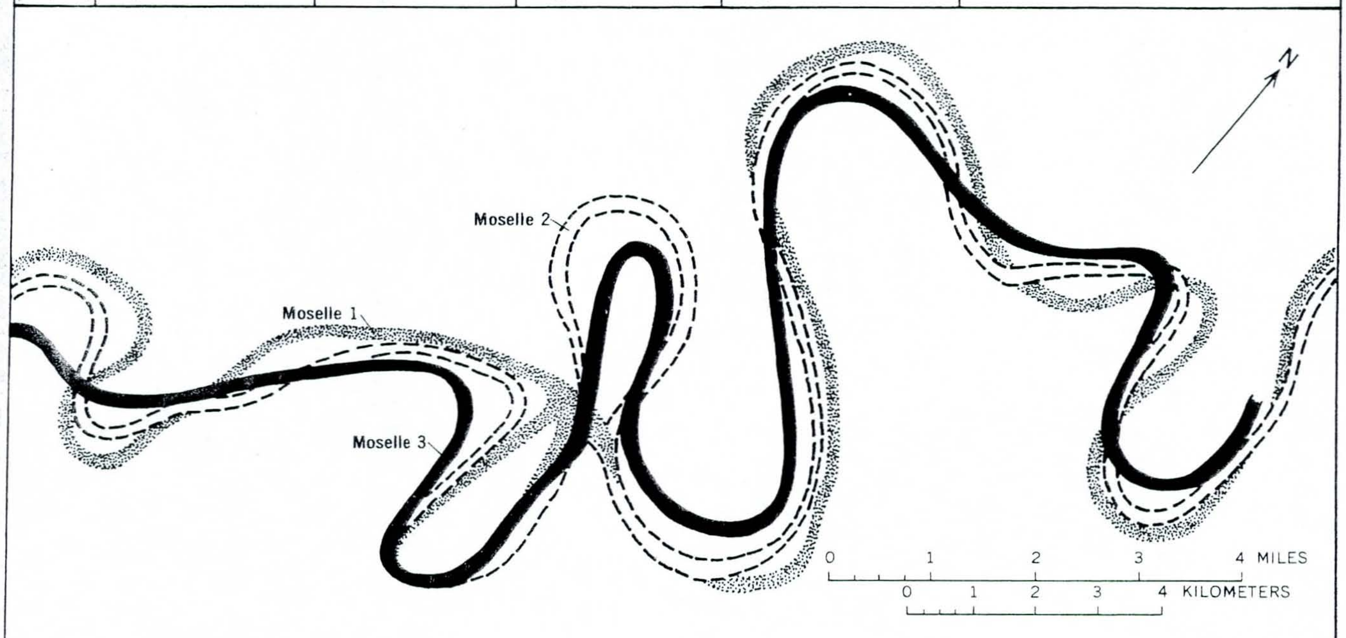


FIGURE 29.—Sketch map of the middle Moselle River. Above, arrangement of terraces; below an interpretation of ingrowth from end of deposition of terrace 3 to end of deposition of terrace 1. Data after Kremer (1954).

in time between braiding and meandering. On the Shenandoah, as elsewhere, the most deeply incised valley bends seem to have been initiated well back in Pleistocene time. Again, like many similar bends elsewhere, they have failed to modify their original wavelength during their long history of ingrowth.

In Wisconsin the Driftless Area and its margins support the customary doubt surrounding the age of summit platforms, though clearly exemplifying the contrast between the extent of downcutting in the last deglacial and that recorded in incised meandering valleys. No review of the disputed platforms in Wisconsin seems necessary, for definitive conclusions of identity, number, age, and origin have yet to emerge. Similar uncertainties apply to the Ozarks, where, however, a thick regolith with ghosts of original bedding seems to record prolonged weathering and conditions unlike those responsible for the incised valleys and advanced dissection which typify parts of the region. In the Driftless Area, as in the Ozarks and in the Hercynian massifs of Europe, deeply cut valleys contrast powerfully with subdued tops. Whatever hypotheses of planation be adopted, and whatever sequence of platforms be distinguished, the existing valleys have undergone marked rejuvenation.

When it rose to form the High Terrace, the Wisconsin River caused choking of lateral valleys, as noted previously for Black Earth Creek and Mounds Creek. Mounds Creek has cut through some 60 feet of weak material in its lower reaches, (since the formation of the High Terrace), whereas its valley descends 350–400 feet through solid rock. No means exist to extrapolate from shallow incision through a weak fill to deep cutting through rock in place, especially because changes in discharge are involved; but the disparity of effect is at least great enough to require a long span for the development of the whole valley. The High Terrace dates from the last (classical Wisconsin) glacial—say, from about 14,000 years B.P., if the Johnstown moraine is correlative with the Valparaiso system.

Identical signs come from the Kickapoo River, where the High Terrace occupies valley meanders (fig. 30). All the six valleys named in the diagram are cutoff valley meanders; Haney Valley, Steuben Valley, and Pine Creek Valley still have upstanding cores of bedrock, but Barnum Valley and Posey School Valley do not. Citron Valley may be not one cutoff, but two. The inflection on its west side may well be a left-hand valley bend, formed on the downstream limb of a greatly hypertrophied northward swing. This reach of the Kickapoo illustrates with unusual freedom the modification by cutoff of an incised meandering trace. The contrast between cored and coreless loops corresponds to the difference between prolonged ingrowth on the one hand and rapid lateral enlargement at low levels on the other. All six loops contain patches of High Terrace; all six, therefore, predate High Terrace times, whether or not they had already suffered cutoff by then. Piney Creek, tributary along the southern limb of Piney Creek Valley, and Citron Creek, tributary near the downstream end of Citron Valley, are manifestly underfit. As their valley bends cut through the High Terrace, some time must have elapsed between the formation of that Terrace and their reduction to underfitness. This part of the record, as far as it goes, corresponds with that established for Black Earth Creek and Mounds Creek On the Kickapoo, as on Mounds Creek, the incomplete removal of the High Terrace contrasts with the bulky excavation of bedrock from the river valley.

INITIATION OF MEANDERING VALLEYS DURING MID PLEISTOCENE TIMES

Certain meandering valleys were first cut before th last (Wisconsin) glacial but later than the early (Ne braskan) glacial. Their occurrence shows that hydro logic conditions generally similar to those responsibl for the earliest dated meandering valleys obtained i mid-Pleistocene times, in the broad sense of this term Moreover, since the wavelength to area relations o meandering valleys initiated during the mid-Pleisto cene very closely resemble those of early Pleisto cene valleys, region for region, there is nothing to choos between the conditions responsible for the earlier serie and those responsible for the later.

Trains of valley bends on the River Avon, Warwicl shire, England, were noted in Professional Paper 452– (Dury, 1964) as originating in the last interglacial- not long, perhaps, after the formation of the outwas train which, predating them, belongs to the recessic phase of Penultimate (= Illinoian) ice. Incised benc on the competing Evenlode, also described in the earli report, date from at least as early as the valley meande of the Avon, and probably from the Penultimate Glaci itself. The Avon, its tributary Itchen, the Evenlod and neighboring streams on the Cotswold back slo are all highly underfit, with a wavelength ratio betwe valley and stream of about 9:1. (See Dury, 1964 Their shrinkage was no less marked than that of strear in the Driftless area of Wisconsin. Although the wav length of valley meanders is difficult to compare fro region to region because of variable wavelength to ar relations, comparative degrees of underfitness sugge that, climate for climate, valley meanders initiated about the end of the Penultimate Glacial were as lar as those dating from earlier times. The Avon is esp cially instructive in this connection. Working in we

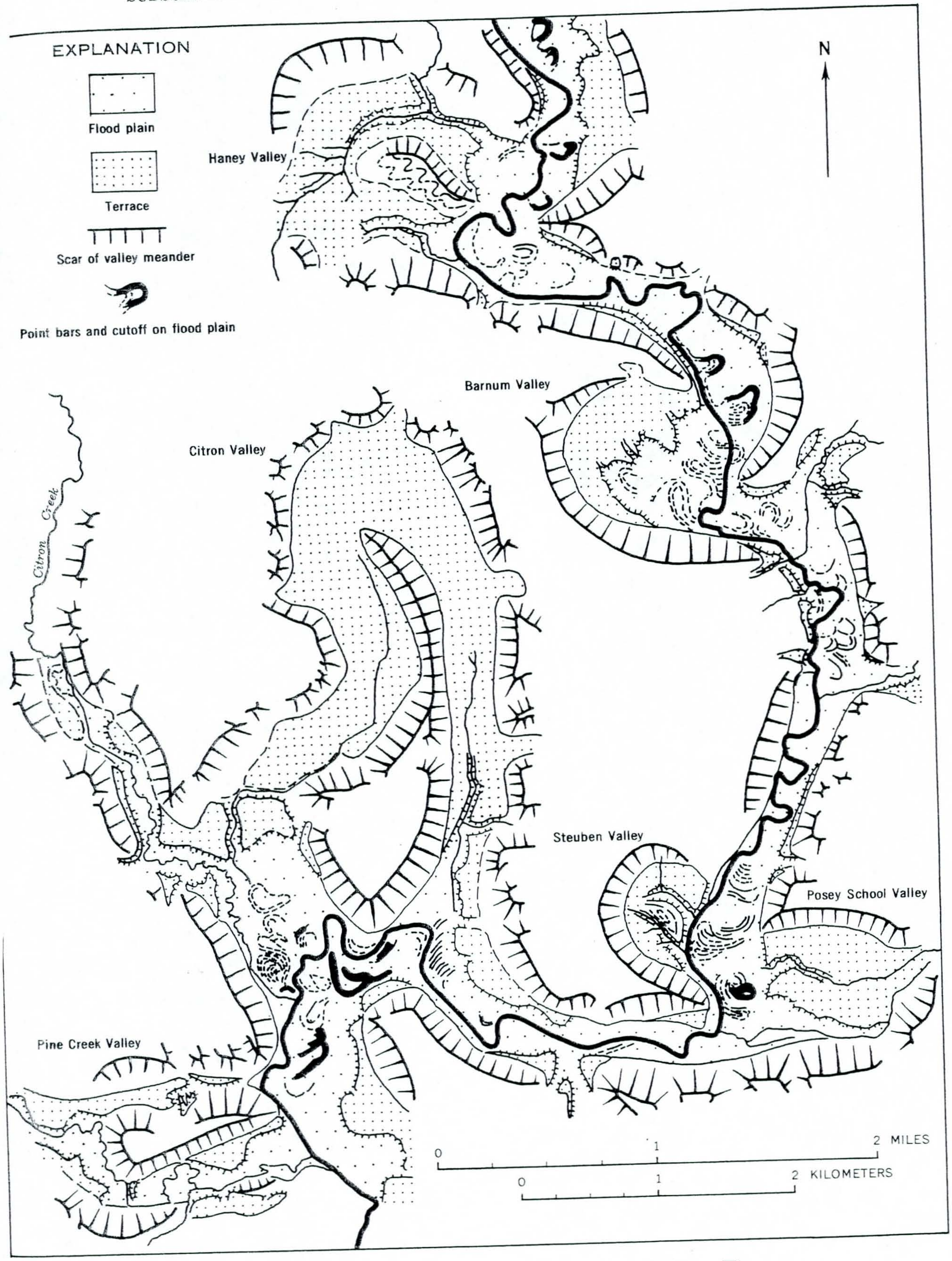


FIGURE 30.—Morphologic sketch map of part of the lower Kickapoo River, Wis.

rock where confinement by resistant valley walls does not come in question, the Avon should have been able to accommodate its wavelength to changes in regimen; in actuality, the wavelengths recorded by the lowest terrace are identical with those at the highest level known for the great bends. Each episode of high discharge at the bankfull stage in valleys with long histories of ingrowth seems to have produced rivers comparable in size with those produced in earlier episodes. A similar set of circumstances exists for channels; Sandford (1924; written communication, 1955) regards the channels beneath terraces in the Oxford district as similar in dimension throughout the range of terrace stages, including that stage when the filled channel beneath the present alluvium was cut.

Streams on parts of the Atlantic Coastal Plain underwent extension during the Pleistocene, developing valley meanders on the emerging sea floor. Relevant fixes come from the work of Doering (1960) in the range Georgia to Virginia. Doering identifies six offlapping surface formations—the Citronelle, Sunderland, Wicomico, Penholoway, Talbot, and Pamlico, listed in order of ascending age and descending height. He assigns an early Pleistocene (preglacial) age to the Citronelle by reference to plant fossils from Alabama and cold-water Foraminifera from Louisiana and tentatively places the Pamlico Formation and its associated shoreline in a mid-Wisconsin interstade.

Because the Foraminifera belong specifically to the Calabrian, the Citronelle formation correlates with the 600-foot shoreline of the European Mediterranean and with the base of the Pleistocene succession. Fairbridge (1961), in a broadly ranging review of published material, gives reasons for regarding the Pamlico as Sangamon, thus reducing the interval spanned by the whole sequence to slightly below that inferred by Doering. Even so, much of Pleistocene and glacial time is involved.

Valley meanders trench the whole series of six formations, descending well below the Pamlico level near the coast. Doering's figure 12 shows the Roanoke River entering Albemarle Sound at Plymouth as a manifestly underfit stream, 50 feet lower than the Talbot Shoreline and 25 feet lower than the Pamlico shoreline; indeed, valley bends continue below present sea level. Whatever the earliest history of excavation, valley cutting through the local Quaternary succession is mainly the work of former large streams. These, lengthening across the Coastal Plain in response to emergence, continued in being until the time of low sea level during the Wisconsin Glaciation. As in other regions, the known record does not require that the large streams persisted continuously from initiation to abandonment of their meanders. But because there is nothi choose between extended and original reaches in of wavelengths to area value or degree of underfi there is no evidence for large meanders of pre-Pl cene date on the original reaches; and because th of origin of extended reaches decreases down without disturbing wavelenth to area relations valley bends, some trains of valley meanders are to have come into being earlier than others. The native view that all originated simultaneously insuperable difficulties and, moreover, conflicts di with the long-continued use, or repeated re-use, of meanders elsewhere. The record of initiation of meanders on the Coastal Plain is therefore tak include part, and possibly all, of the glacial Pleis and to refer in part to trains first developed in Pleistocene times.

Till sheets in the Midwest set earliest possible for the origin of streams which now drain them. is no suggestion that the reconnoitered streams, tioned in a previous section and in Professional 452–A (Dury, 1964), occupy former spillways; contrary, they represent common habits of n drainage. Reasoning similar to that applied Coastal Plain holds good here also. Unless the tion of valley meanders be referred to about the the last glacial, numerous meandering valleys Midwest have histories beginning before the Wis advance.

INITIATION OF MEANDERING VALLEYS DURIN LAST DEGLACIAL

Just as valley meanders initiated in the mid-P cene testify to the continuance or reestablishm hydrologic conditions similar to those of the early tocene, so do valley meanders initiated later th last glacial maximum indicate the similar conti or reestablishment of similar conditions late in the tocene sequence as a whole.

Professional Paper 452–A (Dury, 1964) uses carbon dates for glacial Lake Whittlesey to da initiation of valley meanders on the Auglaize System as not earlier than 12,500 years ago. Sim the draining of Lake Agassiz II sets an earliest lir the development of valley meanders on the She Elm, Park, Tongue, and Pembina Rivers wher cut into the lake floor; the limit coincides with cession of Valders ice (Elson, 1957). But where ley meanders can have formed on upstream r within the once-flooded area soon after draining those at low levels must be later. The date o draining is yet uncertain, but it lies between th ars B.P. mentioned in the foregoing paper and the ;est limit of 3,600 years B.P. set by Elson.

Valley meanders on the site of glacial Lake Souris are ;o relevant here. They occur, for example, on Deep ver and Cut Bank Creek, within 2 miles of the Souris ver; both streams are manifestly underfit (Paulson d Powell, 1957, pl. 1). Although dates are not forthming, the location of these particular trains denies eir appearance at least until very late in the history the final draining of Lake Souris.

Manifestly underfit streams on the Shropshire-ıeshire Plain of England cannot fail to postdate the eet of till into which they cut. Barendsen, Deevey, d Gralenski (1957) report a radiocarbon date of ,670±130 years for peat in a kettle near the outer argin of the sheet; so that, in round figures, the relent streams can be taken as not older than 11,000 years P. Valley meanders on the floor of glacial Lake itchcock, in the valley of the Connecticut River (see ury, 1964), can be as early as those of the Auglaize; ubin and Alexander (1960) give 12,200±350 years for e base of a bog formed, when Lake Hitchcock drained, ı a delta near Suffield, Conn. By contrast, some of e trains near Green Bay, Wis., date from late in the scending sequence of lake levels in that region.

In certain respects, the country around the head of reen Bay is well documented. Recent maps on the ales of 1:62,500 and 1:24,000 clearly portray the eanders of existing streams and, in addition, the windg valleys in which these streams are contained (Applen, Chilton, Denmark, De Pere, Green Bay, Neenah, ıd New Franken, Wis., quadrangles, 1:62,500; see also e Pere, Wis., quadrangle, 1:24,000). Prepared as ey were with the aid of aerial photographs and conured at intervals of 10 or 20 feet, the maps are comendably detailed. Although the trunk Fox River, ading from Lake Winnebago north-northeastward to Green Bay, is broad and shallow, irregular rather ıan sinuous in plan in its lower reaches, and, moreover, fected by dredging and by damming, numerous lateral reams are manifestly underfit. They drain areas here the glacial geology has been mapped in detail, ıe East River crossing the shorelines of three former kes above the level of the present shore.

Wavelengths of valley and stream meanders have en measured on stereoscopic photographs and plotted gainst drainage area for three tributaries of the Fox iver—Plum Creek, Apple Creek, and East River—hich enter the Fox respectively 10, 8, and nearly 2 iles above its mouth at Green Bay (figs. 31, 32). The itio of wavelengths between valley and stream meaners is almost exactly 5:1; but the two clusters of lotted points are less regular than usual, probably mainly because the meanders themselves are irregular, having been developed rapidly in weak sediments. Cutoff valley meanders, distinguishable by upstanding cores, are not uncommon; although the reduction of parts of the sides of the large trenches to blunt cusps suggests, in combination with the irregular plan of the side slopes in general, that cutoff was quite frequent on the former large streams. The 5:1 ratio of wavelength associated with hypertrophy of some existing loops ensures that valley-meander scars are considerably eroded, but the clear separation of the two meander series and the almost perfect parallelism of the two best-fit graphs are reassuring.

The valley meanders near the head of Green Bay cannot have originated before the recession of Valders ice, as is demonstrated by the work of Thwaites (1943) and Thwaites and Bertrand (1957). Flint (1957), Hough (1958), Frye and Willman (1960), and Zumberge and Potzger (1956) supply relevant dates, which the following discussion incorporates (see also Crane and Griffin, 1960).

In the advance before the Two Creeks Interstade, ice in the Michigan basin reached points 40 miles beyond the present southern tip of Lake Michigan and 130 miles beyond the head of Green Bay (fig. 33). During its recession, this lobe blocked an extensive lake, early Lake Oshkosh, which drained to the Wisconsin River through an outlet near Portage at about 800 feet above sea level. Such a level indicates that most of the country now drained by Apple and Plum Creeks and the East River was drowned. Further recession of the ice front back to the straits of Mackinac allowed the water in the Michigan basin to fall below its present level, to the Bowmanville low-water phase which corresponds to the Two Creeks Interstade. The closing date for this interval is about 11,000 years B.P.

When ice readvanced during the Valders Stade, the Green Bay lobe eventually reached a limit at least 60 miles southwest of the head of Green Bay. It seems likely on general grounds that water was impounded in front of the spreading Valders ice (compare Bishop's (1958) account of glacial Lake Harrison in the English Midlands); Thwaites and Bertrand do in fact identify pro-Valders lake beds in places. The best-known body of water referable to this part of the sequence is, however, later Lake Oshkosh, which established itself against the eventually receding Valders ice. Later Lake Oshkosh possessed a first outlet near Portage, once again at the approximate 800-foot level, but was in time drained by successive spillways across the Door Peninsula until the ice front cleared Sturgeon Bay, at which time the southern half of the Green Bay basin was united with the main water body in the

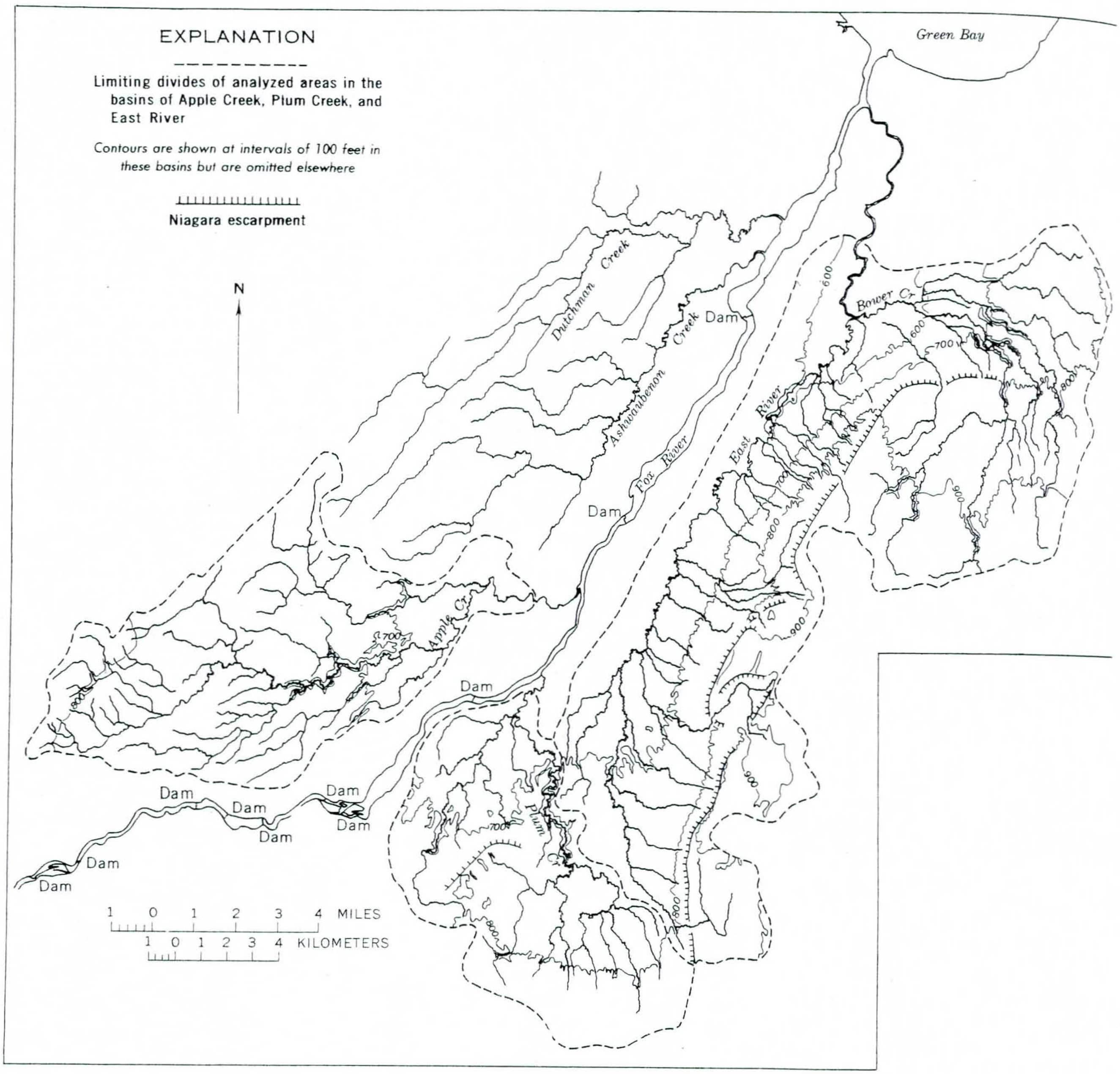
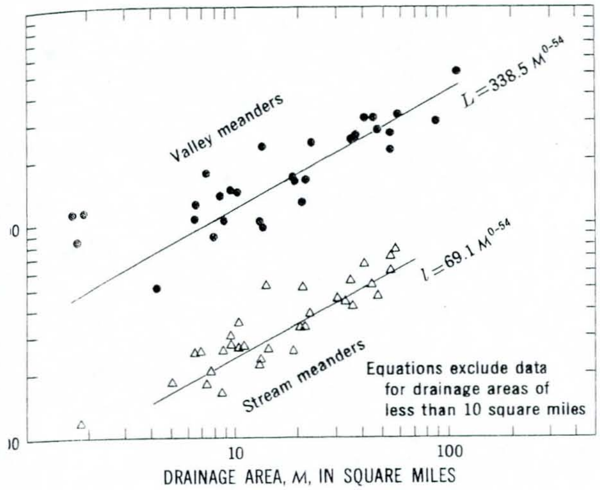


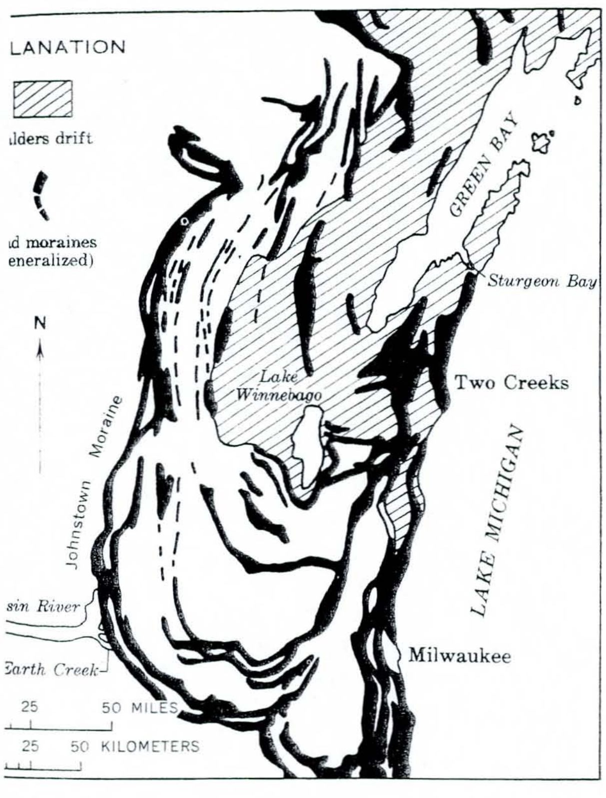
FIGURE 31.—Index map of streams near the head of Green Bay, Wis.

Michigan basin. Meanwhile, this main lake had fallen from the Calumet level of 620 feet to the Algonquin (=Tolleston) level of 605 feet. The first late-glacial exposure to subaerial processes of much of the basins of Plum and Apple Creeks and the East River—that is, the land between altitudes of 800 and 605 feet that was drained by these streams—occurred not earlier than about 10,000 years B.P. However, as will now be shown, parts of the meandering valleys began to form much later than this, so much later that their cutting did not begin until long after the large bends on Black Earth Creek on the margin of the Driftless Area w already disused.

Although the flat interfluves of Plum and Ap Creeks near their confluences with the Fox River 1 respectively above altitudes of 660 and 640 feet, scars of valley meanders on these streams are cut do past the 620-foot level. On Ashwaubenon and Dut man Creeks, left-bank laterals entering the Fox betw De Pere and Green Bay, corresponding scars are below the 600-foot contour; indeed, on Ashwauber Creek they are cut below the 590-foot contour. Si



32.—Graph showing relation of wavelength to drainage area ·eams (Apple Creek, Plain Creek, and East River) near Green Wis. See figure 31 for location of streams.



33.—Sketch map showing glacial setting of country near the head of Green Bay, Wis.

', large meanders on the East River reach along loor of the wide, lowermost valley almost to the ıence with the Fox, if indeed they do not run the e way down. These large bends must, it would , have been cut not only after the recession of ers ice but also after the fall of the lake from the oot level. Any allowance made for warping of the Algonquin shoreline would increase the extent of submergence indicated for Algonquin time by present-day contours; without such allowance, a stand of the lake at 605 feet would inundate more than 3 miles of Dutchman Creek, nearly 4 miles of Ashwaubenon Creek, and nearly 10 miles of the East River. However, as the present streams tend to cut into the floors of the enclosing valley meanders, measurements related to interfluves are preferable to those referred to the valley bottoms; a lake standing at 605 feet would inundate more than 1 mile of the whole valley of Dutchman Creek, 1½ miles of the valley of Ashwaubenon Creek, and 7 miles of the valley of the East River.

The Algonquin phase ended about 8,000 years B.P. with a spasmodic fall in water level that eventually reduced the water body in the Michigan basin to Lake Chippewa, which stood at about 230 feet above sea level and is dated to about 5,000 years B.P. The 605-foot level was later reestablished when the Nipissing Great Lakes came into existence as warping raised the North Bay outlet of the whole Superior-Michigan-Huron system. As a group, the Nipissing beaches are strongly developed, indicating a lengthy stillstand; they can be distinguished from the Algonquin beaches, and thus compared in respect of strength of development, in those northern areas where the two series are upwarped in different degrees. Because of the inferred long stillstand, the radiocarbon date of 3,656±640 years for peat from a Nipissing beach is merely a sample from a long span of time. In view of its lateness by comparison with Algonquin dates, and especially in view of the presence of the Nipissing shoreline in the East River basin, this date, or span of dates, is significant in the present context. The succeeding beach in the Michigan sequence, the Algoma Beach at 596 feet, dates from about 2,500 years B.P. and may have continued to develop even as late as 2,000 years ago. Very little of Dutchman Creek or Ashwaubenon Creek would be flooded at the Algoma level, but the valley of the East River would be occupied by water for at least 7 miles above the river mouth. It seems unlikely that large meanders on the East River have been neatly re-excavated after the long stand of water at the Nipissing level. Furthermore, the patches of swamp (indicated on the 1:24,000 map) alongside the lowermost reaches of the stream are seen on aerial photographs to be associated with large point bars and with large cutoffs downstream from the 590-foot contour (fig. 34). The inference is that these particular large bends were cut in the emergent floors of Lakes Nipissing and Algoma. This means that the necessary climatic conditions for large meanders were prevalent here as late as about 2,000 or 2,500 years B.P. Subsequent slight upwarp-

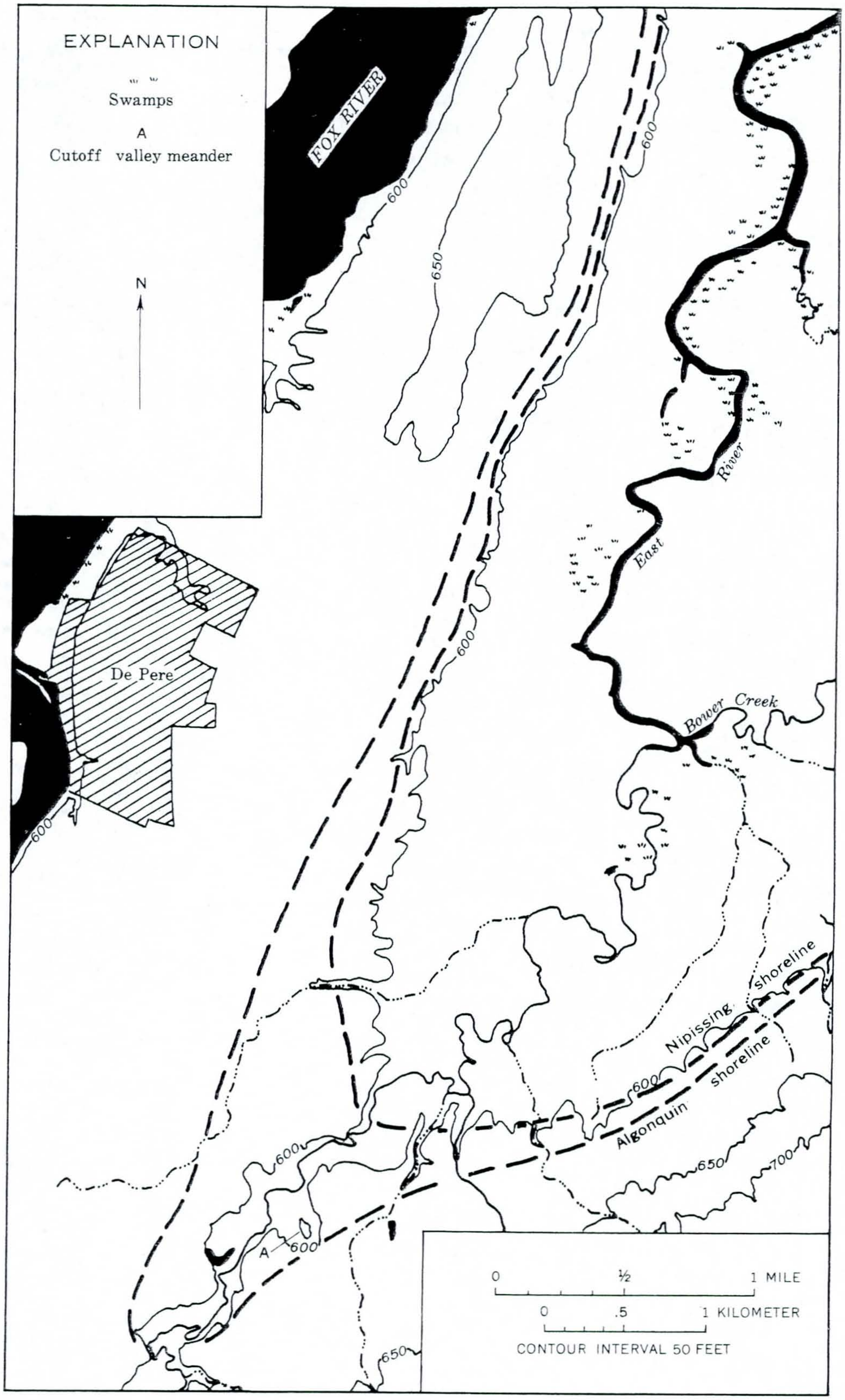


FIGURE 34.—Sketch map showing relation of the lower East River to glacial-lake shorelines.

ı be held to account for the apparent drowning East River, perhaps as far upstream as the con- of Bower Creek 5 miles from the mouth of East and for the drowning of the extremities of nan Creek and Ashwaubenon Creek. Drowning responsible for the failure of the streams to con- ı their lowermost reaches from large to small ers.

ATES FOR THE ABANDONMENT OF VALLEY MEANDERS AND LARGE CHANNELS

this context, abandonment means infilling suf- to separate a stream channel from underlying :k, always provided that the infilling is associated stream shrinkage. Alternatively, of course, aban- ent of valley meanders merely constitutes reduc- f wavelength sufficient to produce stream meanders the ample meandering trace of the valley. In ice, the dates available are those supplied by s of channel fills. In practice, also, these dates ıd largely upon the analysis of fossil pollen and rrelation of pollen data with the data of radio- n analysis.

is section will review the evidence from Wisconsin from a number of well-documented sites on the ish Plain. The Mission River of Texas will be to exemplify valley meanders drowned by the ıcial rise of sea level. The general outcome of part of the discussion will be that, although great nations between cutting and filling are recorded ıumid regions, the last abandonment of large chan- –that is, the end of the last episode of complete ring—occurred not later than some 9,000 years ago, pt in regions such as parts of the Lake Borders re still-receding continental glaciers were able to ıence the regional hydrology until much later dates.

nless a streamless interval be postulated for Black th valley between the outwashing of a braided am of melt water and the initiation of the stream ; cut large meanders into the loess, it must follow ; the change from braids to large meanders was ning more than a change of habit by a stream which tinued in being. Changes of this type are well wn from other regions. Fisk and McFarland (1955, 2) show the Mississippi Trench at the head of the taic plain as filled by late Quaternary (=late Wis- sin) deposits; the lower part of the trench is re- red to sedimentation by a braided river. Their text 292) states that the channel characteristics of the ssissippi at the time of late Wisconsin low sea level re probably similar to those of the braided present- y Willamette River. Change from braiding to flow in valley meanders was noticed above for certain rivers on the European mainland and, in Professional Paper 452–A (Dury, 1964), for the English streams Avon and Evenlode. The postulated braided ancestor of Black Earth Creek was probably capable of dealing with silt, because it could transport large blocks. Conversion from braids to meanders should therefore be placed at about the time that loess fall began; the top of the outwash supplies an earliest possible fix for the origin of large meanders.

The cause of change in habit may be somewhat more complex than a simple retreat of the ice front and a cessation of the discharge of melt water. Some of the rivers described by Troll were not supplied from ice fronts. It is possible, therefore, that Black Earth Creek continued to braid after the ice had withdrawn so far as to discharge no more sediment or melt water into the valley head. On the other hand, because the train of outwash is stratigraphically continuous with the Johnstown Moraine, the change of habit appears to have followed swiftly upon the first withdrawal of ice. If the Johnstown ice stand equals the Cary of the traditional subdivision of the Wisconsin Glaciation, then it belongs in the interval 14,000–12,000 years B.P. (see Horberg, 1955; Ruhe, Rubin, and Scholtes, 1957); if it belongs with the Valparaiso Moraines, the earlier rather than the later date applies. The base of the present flood plain seems to date from not later than 10,000 years B.P., giving a maximum possible span of 4,000 years for the duration of large meanders. The duration was probably less than this, for the large stream had time merely to cut through the shallow loess and to make slight excavations in the underlying gravel; although its meanders greatly extended themselves in the lateral sense, they failed to migrate down-valley to any appreciable extent. The spread of loess on the valley floor does not seem stratified; if it were proved to be stratified, it would seem referable to deposition by a braided stream, as it extends wholly across the valley floor in a belt much wider than the large meander belt of Black Earth Creek. In that event, the span available for the initiation and development of large meanders would be reduced below the possible 4,000 years.

The date of 10,000 years B.P. for the base of the flood plain depends on gross determination of pollen in sediment samples taken on line 2 (fig. 15), where part of the fill of the old pool remains. Determinations were made through the courtesy of Dr. Estella B. Leopold of the U.S. Geological Survey Paleontology Laboratory, Denver, Colo. Although the sediment in the old pool is almost barren of pollen and thus fails to provide adequate material for dating, the sandy layer at the base of the present flood plain (USGS paleobotany loc. D1559–A, D1559–B) seems to relate to Zone III of Jelgersma's pollen sequence for east-central Wisconsin (Jelgersma, 1962). The top of Jelgersma's Zone I (nonarboreal pollen) is fixed by radiocarbon dating at 12,000±350 years B.P. Next comes Zone II (*Picea*), then Zone III (Mixed *Picea* and *Pinus*), followed in turn by Zone IV (*Quercetum mixtum* and birch) with its top at 9,300±350 years B.P. Zone III would seem to bracket an interval of about 11,000–10,000 years B.P. These dates are thought to give lower and upper limits for the deposition of the base of the present flood plain, which unconformably truncates the fill in the pool of a large meander, and thus to locate the conversion to the small meanders of the present day.

Gross determinations of pollen have been made, also through the courtesy of Dr. Estella B. Leopold, for two samples from the valley of Mounds Creek (USGS poleobotany loc. D1558–A, D1558–B). The samples were taken from the lowermost part of the fill which overlies the gravel on the Elvers profile (fig. 18). They contain mostly spruce pollen with some pine near the base of the section and mostly spruce pollen at the base, immediately above the gravel. Both are tentatively assigned to Jelgersma's Zone II (*Picea*). As noted above, the beginning of this zone is fixed at 12,000±350 years B.P. The date confirms the interpretation of the gravel in the Mounds Creek valley as a sludge train of congelifractate provided by weathering in a periglacial climate near the ice front at the glacial maximum. If conversion from braiding to meandering on Black Earth Creek coincide cessation of sludging in the valley of Mou and if the base of the flood plain of Black E is dated to 11,000 years B.P., then the larg acquire a maximum duration of only 1,000

The approximate date of 12,000 years B.P resembles the opening date of the Two Cr stade. It is entirely possible that the larg of Black Earth Creek were abandoned duri terval and that a gap of time separates thei ment from the deposition of dated material the flood plain (table 2). On this view, t flood plain accumulated after Two Creeks t

On the assumption that the movement of into Mill Creek valley was contemporaneou delivery of gravel into the valley of Mounds that the first channeling of the gravel of Mo at Elvers coincided with the appearance of l ders on Black Earth Creek, a further minor becomes possible—namely, that loess was a cumulate in Black Earth valley because the st ing at the time of loess fall was a small one. of this kind depends on some such correlati presented in table 2, which must remain tent further dates are forthcoming—that is, for t at Mill Creek.

The dating assigned to the oldest channe these Wisconsin streams (see table 4), howev parable to that established by Daniels, Rubi monson (1963) in a paper that was published present text had been drafted. These three w

TABLE 2.—*Possible correlation of events on Black Earth Creek, Mounds Creek, and Mill Creek, Wis.*

Event	Black Earth Creek	Mounds Creek	Mill Creek	Time correlati
7	Present slight downcutting; epicyclic(?), independent of control by Wisconsin River?		Cut terrace formed?	----------------
6	Base of present flood plain	----------------------	-------------------------	11,000–10,000 ye
5	Large meanders abandoned		Intermediate channel filled	Two Creeks Int
4c 4b 4a	Large meanders cut through low terrace Low terrace formed (episode of stability?) Large meanders formed and incised through high terrace		Intermediate channel cut	Identical with 2
3	Stream shrinkage and loess fall			----------------
2b	Wisconsin high terrace and correlatives		Colluvium fed into valley	Johnstown ice sta years B.P. appr
2a	Outwash train accumulating	Sludge-train formed		
1	Valleys excavated in bedrock		Largest channel cut in bedrock	----------------

for the Thompson Creek watershed, Iowa, that alluviation began some time before 14,300 years B.P., and that the oldest member of the formation studied may be related to climatic changes associated with the retreat of Cary ice. Although they state that alluviation was apparently continuous from before 14,300 B.P. to some time before 2,000 B.P., the several members of the De Forest Formation are described as separated by disconformities, and these disconformities are ascribed to erosion. The evidence adduced by Daniels, Rubin, and Simonson thus lends general support to the view here advanced, that the last major episode of infilling—that is, the episode when many streams lost contact with bedrock—was associated with end-glacial climatic changes. Their observations are especially important to the present thesis, as relating to valleys that appear to represent type 5a in figure 4 of Dury (1964)—that is, valleys that are too greatly eroded to preserve valley meanders, and where the trace of the present stream channel is merely irregular.

The Mission River of south Texas represents that group of manifestly underfit streams which possess drowned valley meanders in their coastal reaches. Streams of this type are common on the south Texas coast, just as manifestly underfit streams are well displayed farther inland; despite the claims of Stricklin (1961) that the Brazos River is not underfit, his own diagrams (figs. 3, 4) show the familiar combination of stream meanders with valley meanders. The Brazos system, containing an alluvial sequence dated with reference to the fall of Pearlette Ash Member of the Sappa Formation in Kansan or early Yarmouth time, incidentally extends the list of streams known to have long histories of ingrowth.

The adjacent Woodsboro and Rockport (Texas) quadrangles (1:62,500) clearly show the relation of meandering valleys to tidewater. Drowned valley meanders occur at the mouth of the Aransas River at the west end of Copano Bay, on the border of Refugio and Aransas Counties; others lie at the east end of the bay, where the county line runs inland along Copano Creek. A third and larger train has been cut by the Mission River (fig. 35).

Each side of Mission Bay is a valley-meander scar that rises 10–25 feet above sea level; this level is but slightly affected diurnally by a mean tidal range of less than 0.5 foot. Scars similar to those on the flanks of the Bay run inland up the valley, uniting in series of blunt cusps and enclosing a valley floor where the present levees, possibly with the aid of artesian water, create pools and swamps. The history of development of the Mission River, is, in general terms, easy to read. The scars of valley meanders, now partly drowned and truncated by low receding cliffs of 10–20 feet at the seaward end, relate to a time when the river was distinctly larger than it is now and when sea level stood below its present mark. Drowning seems to have promoted infilling about as far upstream as Refugio, 11 miles from where the Mission debouches from its embanked channel onto the flats at the head of Mission Bay.

Alternate cutting and filling on the lower Mississippi are known to correspond to the alternation of glacial and interglacial conditions (Fisk and McFarlan, 1955). Whatever effects may have been produced in and near the delta by crustal warping, the south Texas coast seems to have remained crustally stable during the last deglaciation; so that the drowning of its valley mouths is referable to glacioeustatic recovery of sea level. Radiocarbon dates from nearby sites indicate rates and extent of drowning without, however, preventing some uncertainty about the date of recovery from low glacial levels to a level near the present; dating, discussion, or both, appear in Shepard and Suess (1956), Deevey, Gralenski, and Hoffren (1959), and Bray and Burke (1960). There is fair agreement that sea level stood about 200 feet below the present mark at 12,000 years B.P., and 120 feet below at 10,000 B.P. But whereas Shepard and Suess would defer completion of recovery until about 2,000 B.P. or even later, Godwin, Suggate, and Willis (1958) regard it as sensibly completed by about 5,000 years B.P. This earlier date agrees with the schema presented by Fairbridge (1961) which is broadly accepted here. (See also McFarlan, 1961; Broecker, 1961.)

Until more is known of the sedimentary fill of the Mission River valley or comparable neighboring valleys, the record of drowning and abandonment of large meanders will remain imprecise. Information already available shows that the cutting of the lowermost drowned bends dates from not earlier than the last glacial maximum, whereas their inundation and filling are deglacial. Thus, the evolution of the Mission River since the last glacial maximum in no way contradicts the histories of better known streams. Occurring far beyond the extreme limit of land ice and indeed beyond the limit of permafrost at last glacial maximum, the Mission River will subsequently buttress the general hypothesis that changes in temperature and precipitation are primarily responsible for underfitness.

Southern England lay south of the ice front at the last (=Wisconsin) glacial maximum. For this reason, its rivers, which became underfit at end glacial or in deglacial times, require explanation by some climatic hypothesis. Detailed studies of fossil pollen in this region not only serve to confirm that reduction to underfitness postdates the last glacial maximum but also to

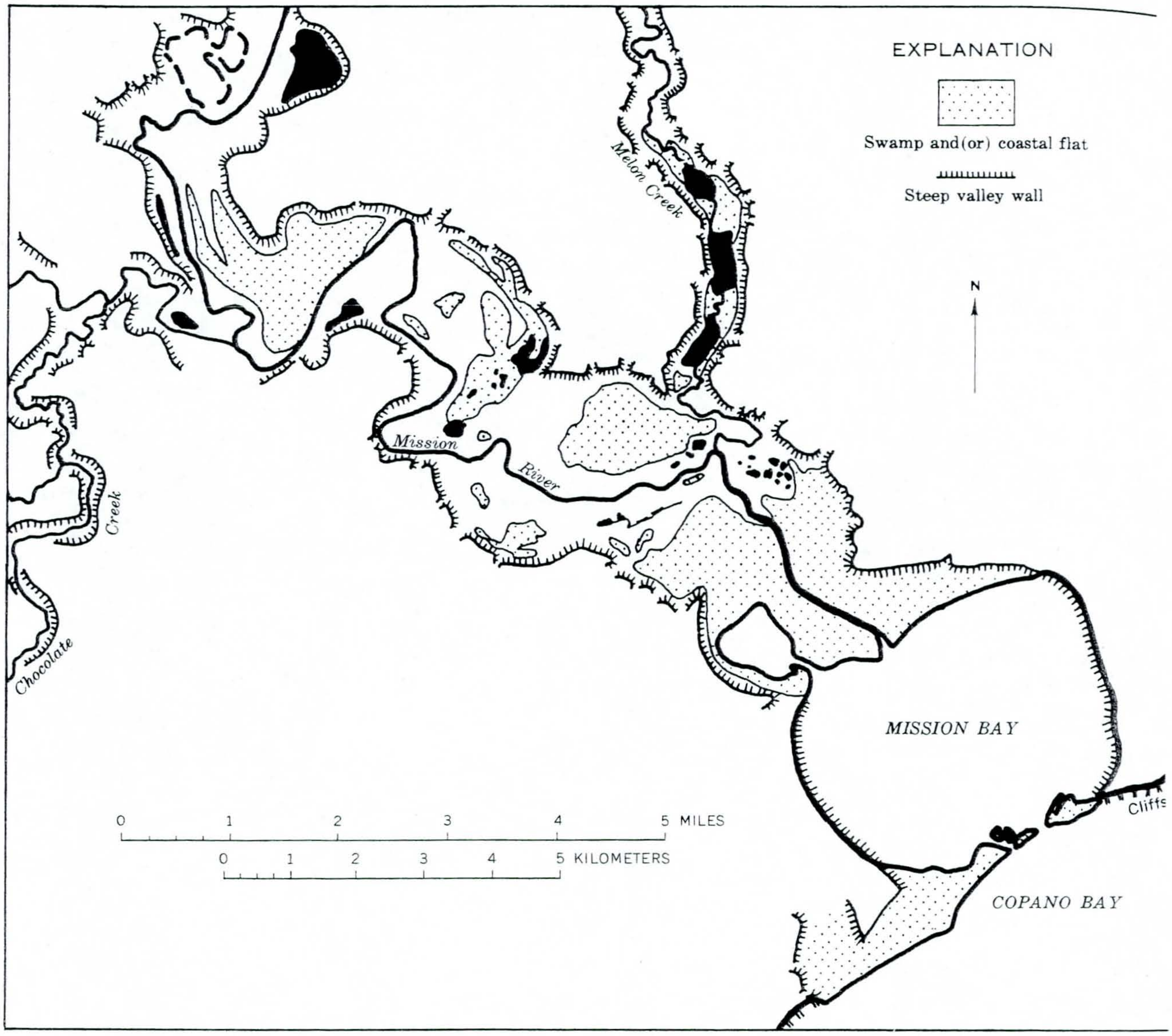
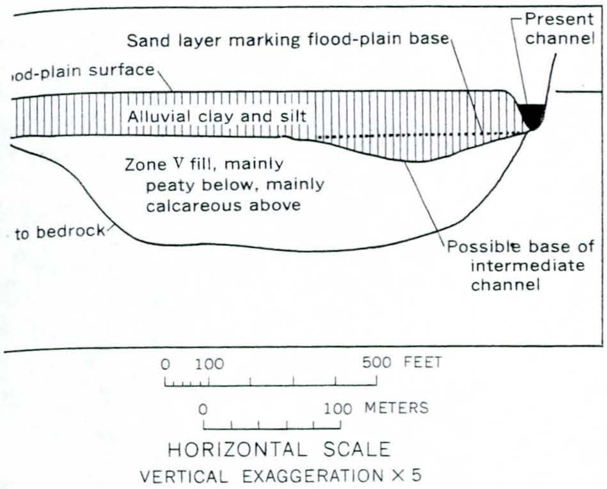


FIGURE 35.—Morphologic sketch map of part of Mission River, Refugio County, Tex.

show that channeling and filling went on long after the date of 10,000–11,000 years B.P. obtained for the base of the present flood plain on Black Earth Creek.

The peaty flood plain of the English River Kennet constitutes the top of a composite valley fill. Both the large channel that contains the fill and its first infilling predate the Atlantic phase (Zone VII), when partial reclearance took place (Dury, 1958). Slight reexcavation may have affected the Zone V fill of the Dorn Valley, for part of the alluvial silt and clay of the topmost layer descends below the sandy base of the present flood plain. Not two channels, but three, may occur here (fig. 36).

The Nazeing site lies in the valley of the Lea at Nazeing, Essex, close to the northern outskirts of London, England, and 7 miles downstream from the conflu of the River Rib. Standing about 80 feet above level, the present surface of the flood plain at Naz is well below the mark of a proglacial lake which du the Penultimate Glacial (= Illinoian) remained some time at about 230 feet above sea level (Clayton Brown, 1958). A gravel fill in the valley bottom re not to the Penultimate but to the last (= Wiscon glacial, when the nearest ice front lay far to the n of the Lea basin. At this time the lower reaches of Lea were controlled by a sea level lower than the pre and gravel moved into the valley in response thawing-and-freezing. Within the gravel occurs Arctic Bed, with its content of dwarf willows, a plants, and remains of *Elephas primigenius*



ꜱ 36.—Profile section showing possible interpretation of the sedi- tary sequence on the River Dorn, Oxfordshire, England. View ownstream.

ιoceros tichorhinus (Warren and others, 1912; ren, 1916; Reid, 1949).

complex sequence of channeling, filling, and veg- onal change at Nazeing has been disentangled by son, Godwin, and Warren (1952), who identify the ιging character and the floral and faunal content eposits both within the channels and upon the ad- nt valley floor. The earliest of three channels post- s the gravel of the valley fill, wherein the Arctic supplies a radiocarbon date of 28,000±1,500 years . (Godwin and Willis, 1960). The second chan- contains near the base a sediment that is referred ollen Zone III (Younger Dryas), whereas the cut- of the third channel took place in Zone VI (Transitional) and filling was completed in Zone VII (Atlantic).[5]

The relation of the channels to one another, to the valley fill of gravel, and to valley-floor deposits is diagrammatically shown in figure 37; figure 37, drawn by the present writer, relies on the text of Allison, Godwin, and Warren and incorporates features recorded in their cross sections. Channels are numbered for convenience of reference in the present text. Allison, Godwin, and Warren refrain from dating the cutting of channel 1 except to place it later than the deposition of the Arctic Bed and the immediately succeeding valley fill and earlier than channel 2. The fill of channel 1 consists of gravel at the bottom but mainly black organic peaty muds with some sand throughout the remainder. Pollen and faunal remains in the muds resemble the content of the Arctic Bed, indicating tundra of either the grass-sedge or the park type.

Above the fill of channel 1 is found the calcareous nekron mud of a lake (lake marl in the diagram), still with fauna of the cold type and with an old land surface at the top. The lacustrine stage was succeeded by the cutting of channel 2, during which much of the nekron mud was cleared away. Infill of channel 2, above bottom gravel, begins with organic mud and continues with detritus mud, with contents indicating Zones III and IV (Younger Dryas and pre-Boreal). While infilling was taking place, closed birch woodland replaced the herbaceous vegetation of the surrounding countryside. Shallowing of channel 2 promoted the

[5] Zonal numbering used in this and following sections accords with the general schema of table 4.

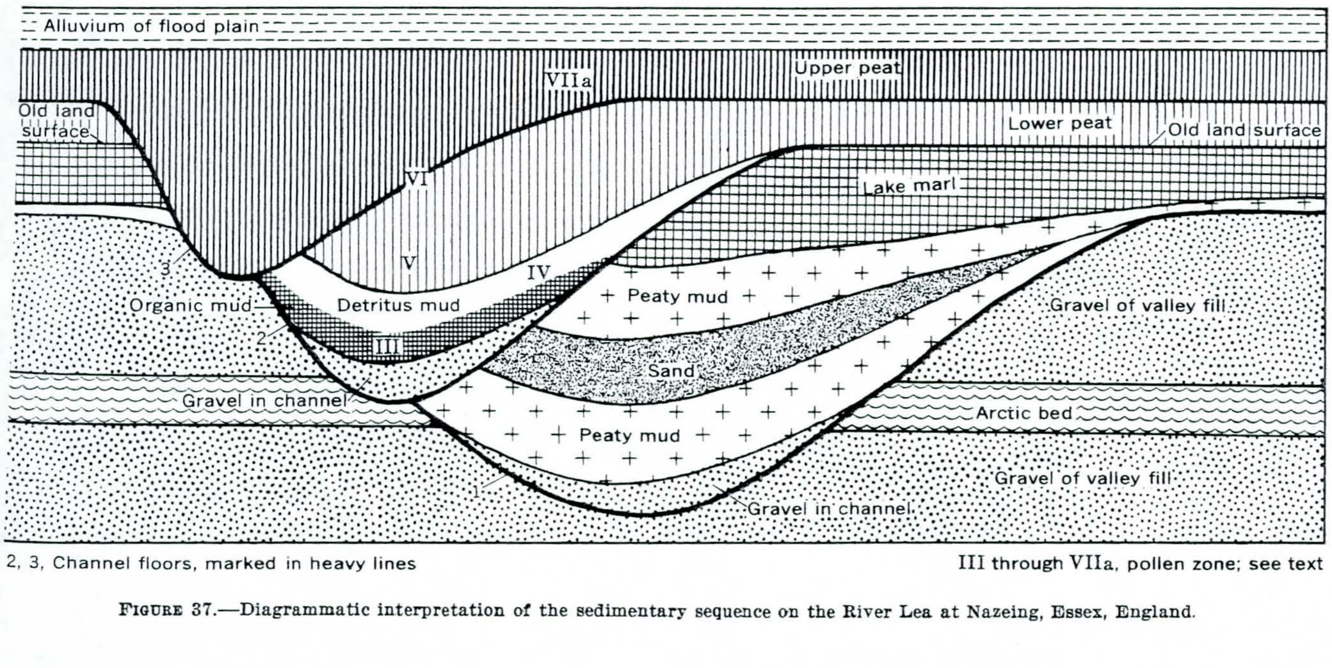


FIGURE 37.—Diagrammatic interpretation of the sedimentary sequence on the River Lea at Nazeing, Essex, England.

growth of sedge peat or reed peat in an ameliorating climate that saw the spread of pine and the arrival and extension of hazel. Peat continued to form through Zone V (Boreal), but eventually the channel and the valley floor almost dried out.

Environmental changes set in during Zone VI (Transitional), causing waterlogging of the valley floor and the renewed erosion which produced channel 3, and brought a mixed-oak forest with hazel to the region. Fen peat grew on the valley bottom and in channel 3 until undated erosion of forest soils in the headwaters supplied clay which forms the existing flood plain and seals in the organic muds and the peats.

Allison, Godwin, and Warren (1952) point out that the record suggests some element of cyclic repetition in physiographic history and thus indicates also a corresponding—and causal—alteration of climate. There could accordingly have been additional cutting and filling to that proved by the available findings. In particular, nothing in the Nazeing record proves the occurrence of the park tundra which may be supposed to have established itself during the Bölling phase, (Ib), or the woodland, which elsewhere is widely indicated for the Allerød (II).

Channel 3, cut at the end of a period of relative dryness and warmth and associable with the change from birch-pine-hazel forest to mixed-oak forest, can scarcely be referred to anything but the onset of the Atlantic phase. Channel 2 predates much if not all of Zone III sediment and postdates the lake marl. A distinct possibility seems to be that the lake and its ultimate drying-out belong in the Allerød, in which event the cutting of channel 2 could be linked with the increasing rigor of climate at the end of the mild Allerød fluctuation. Similarly, the cutting of channel 1 might belong near the base of Zone Ic in the Older Dryas phase and between the preceding Bölling and the following Allerød. The correlation of the whole sequence on these assumptions is schematized by the present writer in table 4, which implies that the cutting of channels 1 and 2 was induced by increasing cold and wetness whereas that of channel 3 was due to increased wetness in conditions of rising temperature.

A general difficulty arises from the indubitably low temperatures—low, that is, compared with those of today—which existed when channels 1 and 2 were cut. Quite apart from possible gaps in the floral record, low temperatures tend to make that record imprecise by comparison with the needs of the present discussion. However, the Nazeing record of channeling spans an interval not shorter than that bracketed by Zones I through VI. Dates listed in table 4 indicate a minimum span of 4,500 years for this interval, the length of which in terms of climatic phases and is not affected by any revisions to absolute d may prove necessary in future.

The Broads are shallow lakes in the va East Anglian Rivers Bure, Yare, and Wav three rivers unite near Yarmouth in Brey an estuary enclosed on the seaward side b spit. Spasmodic growth and erosion of the changes in sea level, slight crustal warping building by the rivers combine to make th sedimentation a complex one. In a first stud (1952) regards the Broads as natural lake amid the backswamps of naturally embank but also notes certain discrepant evidence th by Jennings and Lambert (1953) and late and reinvestigated by Lambert and others accounts which reveal that the Broads or medieval turf pits. The revised origin doe ever, affect the chronology established by J the channels beneath the Broadland rivers 1952, 1955).

These channels closely resemble those id the present writer beneath manifestly under Indeed, the Bure, Yare, and Waveney are festly underfit, so that the presence of large channels beneath their flood plains is in n prising. As the chief concern of Jennings a was to study the Broads as water bodies and the peaty basins which contain them, their li profile do not everywhere run completely valley bottoms nor lie at the inflections of v where the widths of filled channels are mo determined. However, the block diagram Jennings for the Bure valley (Jennings, 19 clearly displays the asymmetry of a filled two valley bends and the near-symmetry at tion of the valley.

The block diagram is here redrawn as in which the excavations of turf pits flanking have been arbitrarily restored. Zonal n serted in the diagram agree with Jenning's tion into climatic phases but not with his forest zones; they have been revised to acco sequence employed here.

The *Phragmites* peat at the base of the sec fore-edge of the block is Boreal, with a very of nonarboreal to tree pollen in its lower suggesting that it may have begun to form beginning of Zone V. That part of the tre it occupies is therefore earlier than Bore when the trench was cut, the Bure was cont sea level perhaps as much as 70 feet below t (Jennings, 1952, p. 49). The Boreal-Atla sition occurred before deposition of the peat was completed, and a slight marine incursion during the Atlantic phase brought estuarine mud some way up the valley. This mud appears immediately above the basal peat on the fore-edge of figure 38; farther upstream, beyond reach of estuarine sedimentation, nekron muds continued to accumulate as they had done previously. (See central and rear profiles, fig. 38.) Part of the fill of brushwood peat belongs to Zone VII (Atlantic phase), but its deposition continued into the succeeding Zone VIII (sub-Boreal). Infilling was interrupted during the Sub-Atlantic by channeling, after which renewed deposition of clay occurred. The final part of the record, which includes a fresh establishment of *Phragmites* peat across the top of the fill, is greatly obscured by the effect of peat cutting.

Although the profiles on the Bure do not reveal channeling during the Transitional or Atlantic phases, they do not conflict with the possibility of channeling at this time in other valleys; for events in the Broadland were certainly then affected by rising sea level that may either have obscured the record of any channeling which took place or have prevented cutting altogether in that downstream reach of the stream for which the record is the most detailed. In outline, then, the sequence of cutting and filling is:

1. Channeling dated earlier than Boreal;
2. Filling by *Phragmites* peat, beginning early in Zone V and continuing into Zone VI, interrupted in lower valley by

2a. Incursion of estuarine clay (Zone VII);

3. Channeling? (partly or wholly offset by effects of transgression?);
4. Bulky accumulation of brushwood peat (Zone VIII);
5. Renewed channeling in Zone IX, followed by a second and farther reaching incursion of estuarine clay;
6. Spread of *Phragmites* peat across the valley bottoms, with levee formation alongside the channels.

As in the Lea valley at Nazeing, the largest, deepest and oldest channels are older than Boreal. If the deepest channel under the Bure correlates with channel 2 at the Nazeing site, and if, as suggested above, the equivalent of Nazeing channel 2 is obscured in, or missing from, the Bure sequence, then the Zone IX channel of the Bure brings the combined record to four successive filled-and-cut channels. The Bure profiles are especially useful in demonstrating minor channeling as late as Zone IX, when it can be referred to a change in climate toward increased cold and wetness.

Sparks and Lambert (1961) record a fill on Willow Brook, Northamptonshire, England, that they describe as not occupying a deep buried channel. However, Willow Brook in the reach described is manifestly underfit (Sparks and Lambert, 1961, fig. 1; Ordnance Survey of Great Britain, 1:25,000, Sheet TL/09), and the explored profile is entirely similar to many of those described by the present writer for sites elsewhere. (See fig. 39.) The fill beneath Willow Brook begins either at the very end of Zone III or early in Zone IV, continuing with Zone IV and Zone V deposits, displaying the results of disturbance and erosion in Zones VI and VII, and recommencing with sedimentation in Zones VIII and IX. The general import of this sequence resembles that of the previously described sites with satisfactory closeness, although it suggests that details of the basal parts of some other fills may have been undetected. On Willow Brook, infilling of a large channel began about 10,000 years B.P. and was interrupted by partial reexcavation at about the time of the hypsithermal maximum.

The 1,300 square miles of Fenland in eastern England constitute a region of drained swamp separated from the coastal inlet of the Wash by a low broad compound baymouth bar of littoral and estuarine silt (Skertchly, 1877; fig. 40). The history of artificial drainage is very long, going back at least to the 3d century, although the principal works date from the 17th century or later (Dugdale, 1662; Wheeler, 1868; Miller and Skertchly, 1878; Darby, 1940a,b). The trace of extinct natural waterways remains visible on the ground and even more clearly so on aerial photographs (Fowler, 1932, 1933, 1934a,b). Their principal type is the roddon, a low bank of clay-silt which stands a little higher than the surrounding peat. Godwin (1938) rejects Fowler's original view that the peat alongside the roddons has been lowered by shrinkage and also denies the efficacy of compaction and wastage, maintaining instead that the roddons are natural levees.

In their lower reaches, as Godwin shows, the roddons result in part from estuarine sedimentation, much of their total bulk having accumulated in Romano-British times—roughly speaking, in the first eight centuries. The natural channels that the roddons border are, however, underlain in places by two earlier and large channels, dating, respectively, from the Neolithic division of prehistory and from some pre-Boreal part of the deglacial climatic sequence (Godwin, 1940; Godwin, 1956, fig. 8; Godwin and Clifford, 1939, fig. 34; fig. 41 herewith).

Although channeling may not equate at all simply with the record of floral change, it seems to correspond sufficiently well to the established sequence of deglacial events and to specific happenings elsewhere. Godwin and Willis (1960) show by radiocarbon dates that track-

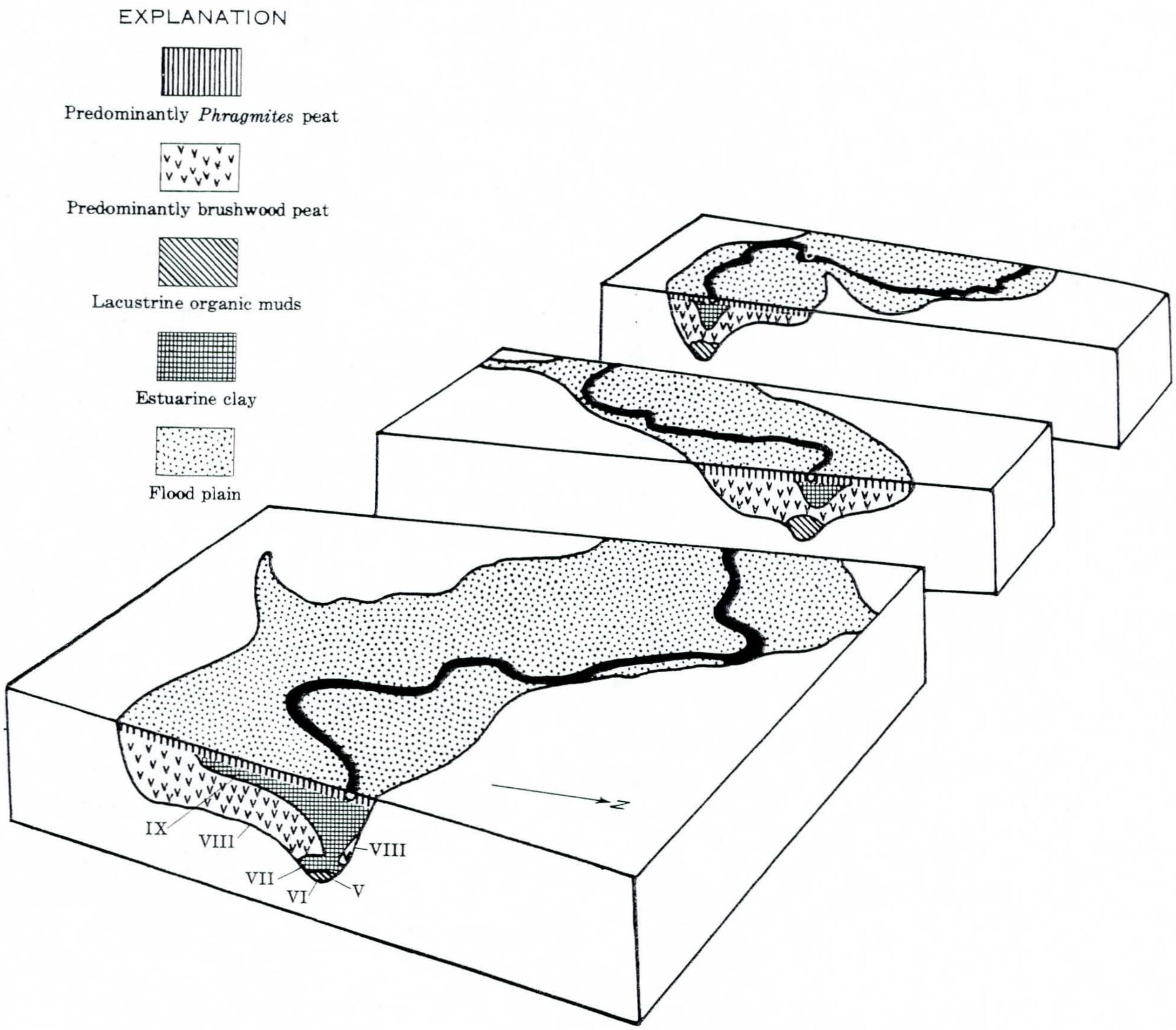


FIGURE 38.—Block diagrams of the filled channel beneath the River Bure, Norfolk, England. Redrawn from Jennings (1952) with excavations restored. Zonal numbers from Jennings (1955) to accord with Firbas (1949).

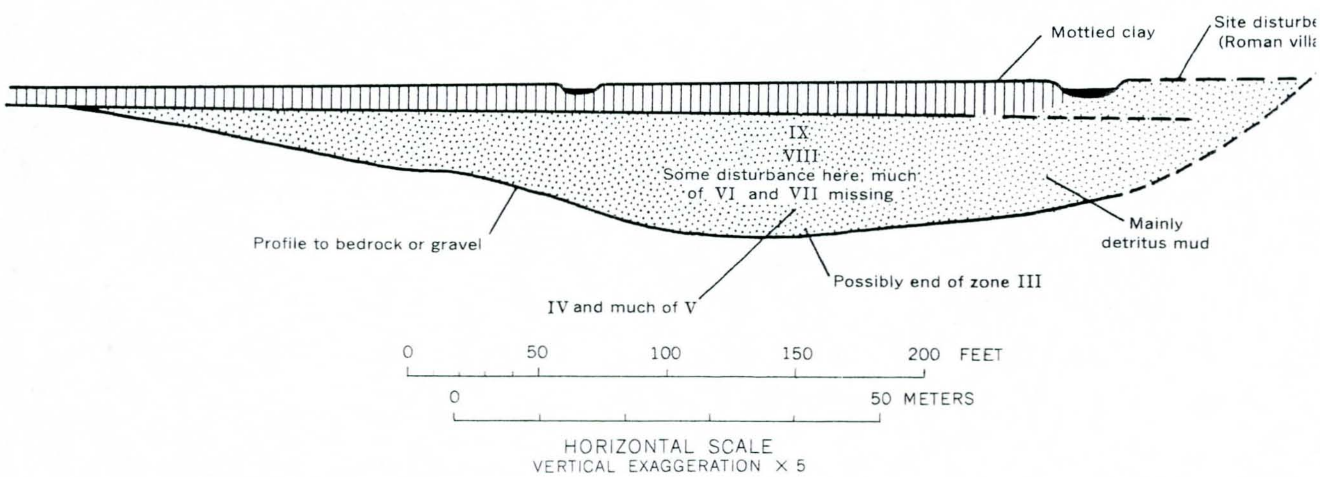
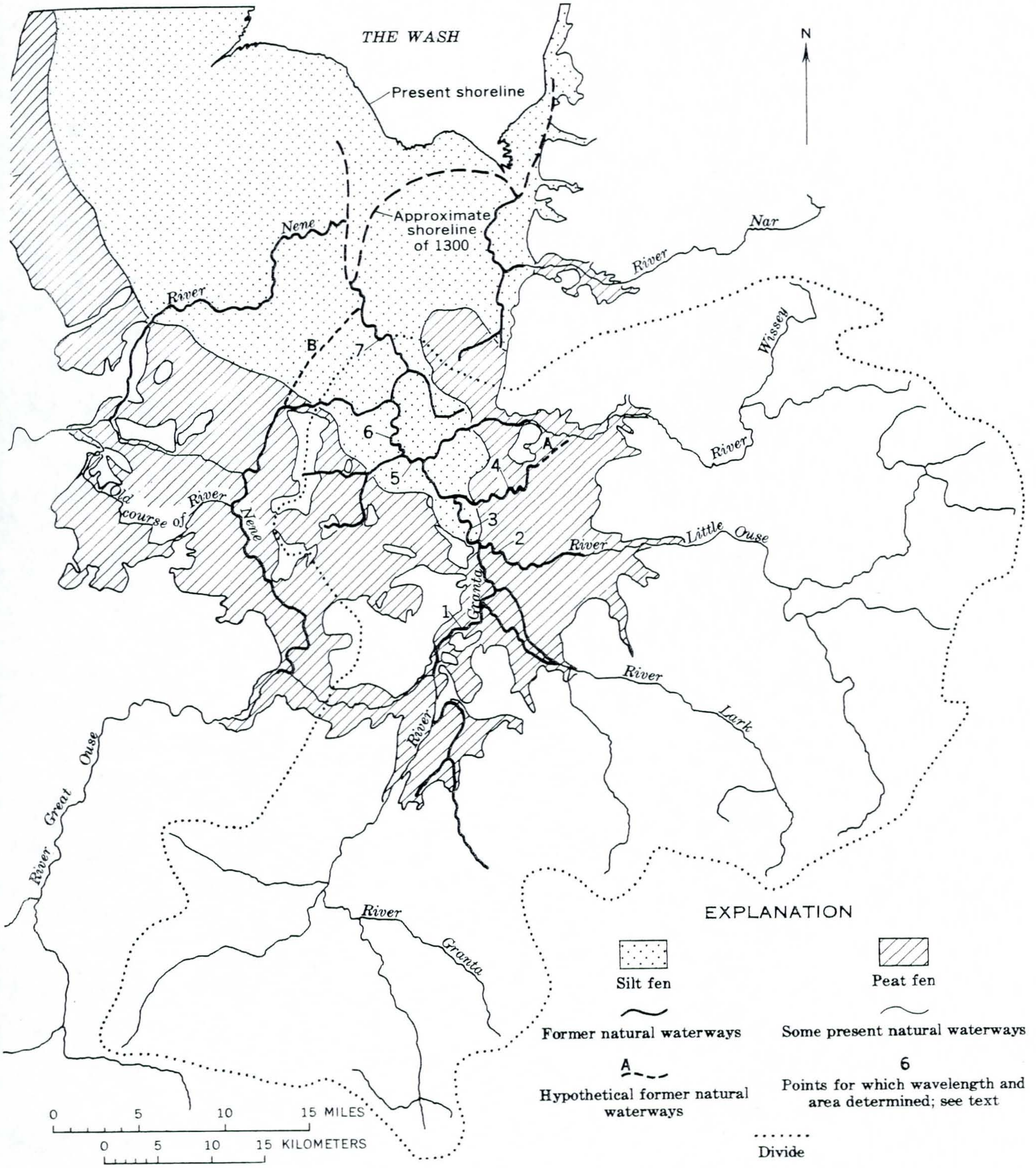


FIGURE 39.—Profile section across the valley floor of Willow Brook, Apethorpe, Northamptonshire, England. Redrawn and zones renumbe from Sparks and Lambert (1961). View is downstream.



'IGURE 40.—Map of the Fenland of eastern England with reconstructed early drainage pattern. After Fowler (1932, 1933, 1934a, b).

ays were laid across the increasingly moist and grow-g bogs of Somerset, on the other side of England, at mes which indicate increased wetness underfoot for te Zone VII (Atlantic) and for about the beginning of Zone IX (sub-Atlantic). For the Fenlands, Godwin and Clifford observe that peat growth was controlled to some extent by profile drainage, which may have worsened during episodes of drying climate. However,

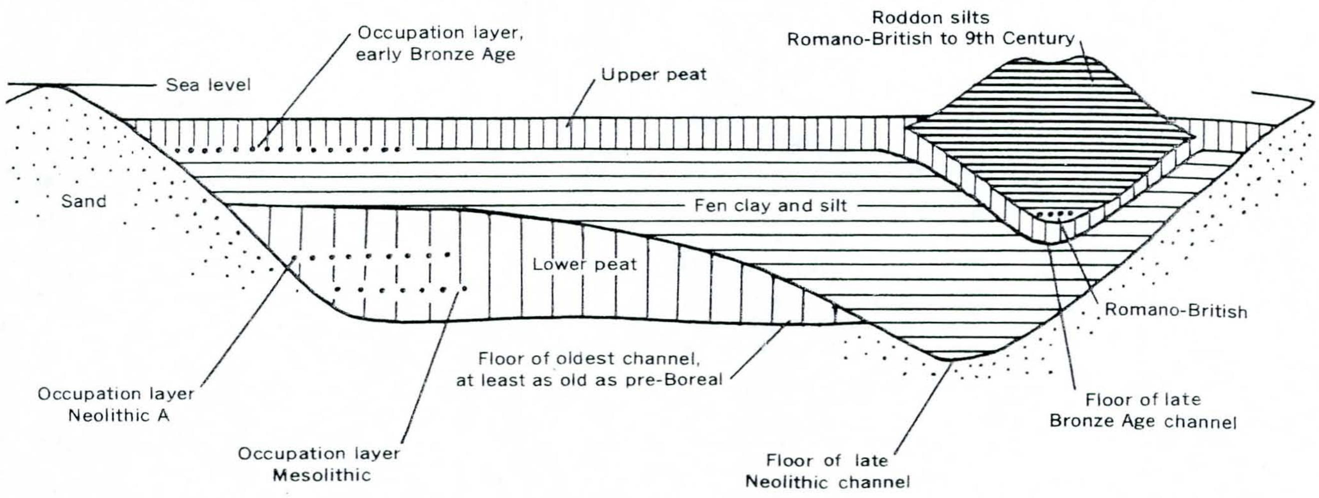


FIGURE 41.—Profile section of channels, sediments, and archeological correlation in the Fenland of eastern England. Redrawn from Godwi and Clifford (1939). View is dowstream.

a reasonable outline of the alternation of channeling and filling in the Fenland seems to be:

1. Largest channel, cut at least as early as Zone IV (pre-Boreal);
2. Infilling;
3. Second channel, intermediate both in age and size, cut in Zone VII (Atlantic);
4. Infilling;
5. Third channel, cut in Zone IX (sub-Atlantic), with levees which now form roddons completed in historic times;
6. Shrinkage of streams in third channel.

Roddon building need not have ceased entirely when shrinkage began.

The channels with which the roddons are associated describe large meanders that are distinctly larger than the meanders of existing streams, area for area of drainage, but distinctly smaller than the valley meanders on reaches upstream from the Fenland. In figure 42, wavelengths for roddons are plotted against drainage area for the River Granta, the largely extinct stream now called indifferently the Cam or Granta in its upper reaches and largely superseded in the Fenland by artificial waterways (fig. 40). As shown in the diagram, the meander wavelengths for roddons on the reconstructed Granta plot consistently against drainage area at about one-third of the extrapolated values for valley meanders on the Nene and the Great Ouse.

Certain difficulties arise from the former cross connections between the Great Ouse and the Nene and between the Great Ouse and the Granta that, being uncertainly known, make uncertain the former drainage areas of downstream reaches. But the eastward channel between Great Ouse and Granta, just within the limits of the fens, and the west-east reach linking t Nene to the Granta may not be earlier than mediev (See Darby, 1940b.) Points plotted in figure 42 the reconstructed Granta suggest that the waveleng in question relate to a stream extending itself across Fenland swamp independently of the Ouse.

Because the old course is marked by roddons, extension appears to belong to Zone IX. (See fig. 4 The 3:1 ratio of wavelength between the exten and the present streams means that bankfull discha on the restored Granta was less, area for area of dra age, than the earlier discharge responsible for the

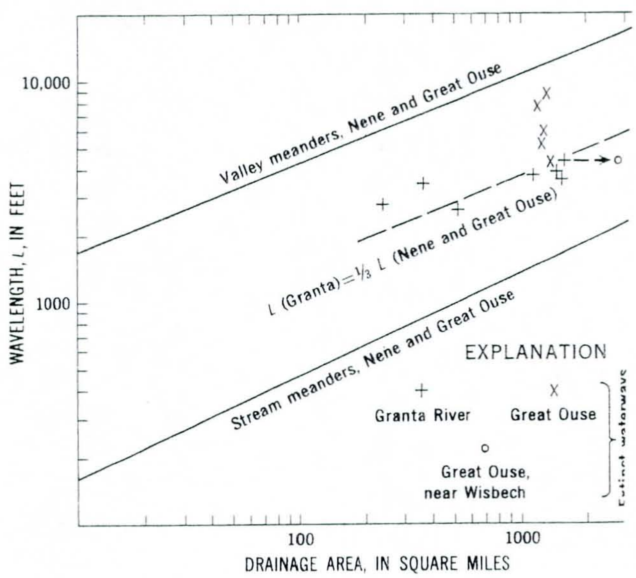


FIGURE 42.—Graph showing relation of wavelength to drainage Fenland and adjacent areas, England. Values for the Nen the Great Ouse from Dury (1958); determinations for the G according to reconstructed course in figure 41; for discussion former course of the Great Ouse in its lower reaches, see text.

the landward reaches of the Great Ouse and the It would perhaps be stretching the evidence to hat the apparent progressive downstream re- of wavelength on the large meanders of the l Great Ouse corresponds to a progressive re- in bankfull discharge per unit area as the stream ned its established channel across the swamp; observations are at least enough to show that the plotted for the Granta belong to that river rather the Ouse, which, greatly changed by artificial is now the main waterway of the region.

roblem arises with the point plotted for location ure 40 if the Wissey is taken at the relevant time taken the course marked by the broken line A. ypothetical early course, B, for the lower Nene, other hand, makes uncertain the placing in figure the lowermost downstream point for the Ouse; dition of the Nene drainage would shift the point position indicated in figure 42 by an open circle. tainties aside, the wavelength ratio between the systems and present streams accords with evi- from dimensions of channels: channels cut dur- ne IX are distinctly smaller, both in width and in ength, than those referred to Zone IV or earlier. hough fixes on the scales of time, stratigraphy, loral succession remained few, the reduction of ns to underfitness could be regarded as an abrupt happening. More specifically, the last reduction derfitness could be placed at the end of the last l without prejudice to partial clearance—for in- e, in Zone VII. The foregoing summaries of suc- e channeling and sedimentation require a more lex view. Nevertheless, simply because successive nels are preserved at more than one site, the younger maller being contained in the older and larger, the ipal shrinkage below the size of stream appro- e to valley meanders is still to be put as early de- al. If, after their first deglacial reduction, streams rally had regained their former volumes, former nels could scarcely have survived renewed erosion. is sense, that reduction which caused certain rivers se contact with bedrock is the most significant of But because ice persisted north of the Great Lakes on long after other districts were far into deglacial , the first main reduction was not everywhere simul- ous. Flint and Rubin (1955) give 6,000 years B.P. the time when the receding ice front reached Coch- e. Southern England, the Great Plains, and the dilleran country in the Southwest were by that e closely approaching their hypsithermal maximum. ereas the large channel beneath Black Earth Creek abandoned before 10,000 years B.P. and that be- th the Dorn was infilled by 8,000 years B.P., parts of the Great Lakes shores were still drowned at these times. Therefore, the initiation of their valley meanders postdates the main reduction of streams in more genial climates. If long-distance correlation is possible, then the main channeling and main reductions of streams in the Lake Border country correspond to minor channeling and minor filling—for example, in southern England.

The alternative view that climatic belts moved across country as the ice front receded, giving similar but not synchronous changes in different regions, runs counter to much recent work. Broad synchrony throughout midlatitudes seems to characterize certain events, at least; one such event is the hypsithermal maximum, even though various authors place it at various positions in the range Zone VI through Zone VIII. The fact that the country near Cochrane still experienced an ice-marginal climate as late as 6,000 years B.P. involves a truncation rather than a deferment or compression of the local climatic sequence.

The upper part of table 4 summarizes and correlates the foregoing observations on the abandonment of large channels in southern England and Wisconsin. As noted in table 4, the principal known episodes of channeling in the Driftless Area predate the Two Creeks Interstade and the Valders ice advance, whereas most of the dated large channels in southern England seem to have remained clear as late as the Pre-Boreal. To some extent the evidence here, provided by the absence of deposits earlier than those of Zone IV, is negative; but the Zone IV fill of the bedrock channel beneath the Dorn and consistent reports of channeling before Zone IV throughout the Fenland, when viewed together, suggest that erosion was indeed marked in Zone III. Results obtained by West (1961) suggest a possible reason for the difference between the two regions; West uses floral evidence to show that a continental climate maintained itself in Wisconsin close to the Valders ice, whereas New England experienced a maritime climate. If the south of England, like New England, still had a maritime climate at the relevant period, then vigorous channeling may have been possible there in Zone III, when it was inhibited by drier conditions in the Driftless Area.

So far as they go, the records for the Kennet and the Bure agree with the more complex record from the Fenland. Slight marine transgression during Zone VII affected the Bure, which cannot therefore be expected to show channeling at this time; otherwise, with the aid of the early pollen reported from the base of this large channel, the revised sequence for the Bure (Jennings, 1952, 1955) accords with that for the other rivers. Correlations with the Lea at Nazeing are less easy, especially since the inference that channeling and filling were cyclic rules out the influence of evulsion or other shifts of channel. In some reaches, the present River Lea is given in the natural state to anastomosis, which is perhaps a response to the deep infilling of its lower valley caused by the deglacial rise of sea level. But the inference of cyclic change combines with the fact that the Lea is manifestly underfit at flood-plain level in some reaches to point to widely operating causes rather than local accidents as responsible for the proven record. Future studies must be relied on to resolve the apparent conflict between the Nazeing sequence and the indications from the Bure, Kennet, Dorn, Willow Brook Rivers and the Fenland rivers that their last time of complete channel clearance was Zone III.

EVIDENCE FROM ARID AND SEMIARID REGIONS

Channels in humid regions, many diminishing progressively in size with increasing lateness of origin, demand comparison with successive cuttings and fillings in drier country and with the histories of pluvial lakes. Literature about cut and fill in the Great Plains and in the Cordilleran country embodies opposing views on the mechanisms involved: some authors claim that accelerated erosion results from increasing wetness; others consider it a response to increasing dryness. Caliche formation, in particular, can be taken in either of the two ways. Debate on causes is all the keener because the sequence of cut and fill appears sensibly identical throughout a very large area (for data, references, and discussion, see Leopold and Snyder, 1951; Miller and Leopold, 1952; Leopold and Miller, 1954, 1956; Schumm and Hadley, 1957; Albritton, 1958; Miller, 1958; Hadley, 1960; Schumm, 1961; Martin, Schoenwetter, and Arms, 1961).

In one respect it is immaterial to the present argument whether aridity or humidity causes erosion to accelerate. Once conceded, shifts or climate link themselves to changes in the erosional or depositional tendency of streams either directly, as by altering the discharge occurring at any given point on the frequency scale, or indirectly, through the medium of vegetation cover. In a sense, also, redistribution of rainfall in time (see Leopold, 1951a, b) does not affect the present discussion, even though it can involve shifts in weather which do not appear in the average values defining climate. Whatever the mechanisms, certain regional trends of erosion of deposition result from external causes, either climatic or meteorologic. They are thus similar in kind, although not necessarily in direction, to those of humid regions. The progressively diminishing cuts of the West and Southwest broadly resemble those described above for rivers in humid climates, lending general support to the hypothesis that underfit streams owe their condition primarily to climatic change.

Grave difficulties are foreseeable, however, if the question of parallel or opposite sequences of cut and fill, respectively, in moist and in dry regions is altogether overlooked. With increased cutting in moist regions ascribed to increased discharge, and that in turn to increased precipitation, 4 items produce 2 contrasted pairs of combinations (table 3). If channeling is everywhere a response to increased wetness, and if simultaneous increases affect climates of both kinds, then the records of cut and fill should be parallel. They should also be parallel if moist regions become increasingly humid while dry regions become increasingly arid and if increasing aridity in dry regions causes accelerated erosion. In this event, however, intermediate areas should be unaffected—a logical inference at variance with the facts of distribution of underfit streams. If increasing humidity affected moist and dry regions at the same time but caused infilling in dry climates, then channeling in humid regions should be on the time scale where filling appears for arid regions. If humidity increased in moist regions while aridity increased in dry regions, and if increased wetness promoted channeling in whatever climate, then the two kinds of sequence would again be out of phase. In these last two instances, evidence of a highly ambivalent kind should be available from areas intermediate in location between distinctly moist and distinctly dry regions.

TABLE 3.—*Possible interrelation of channeling in dry and humid regions*

<table>
<tr><td colspan="2"></td><td colspan="2">Channeling in dry regions caused by—</td></tr>
<tr><td rowspan="5">Direction of change, to or from pluviality or aridity</td><td rowspan="3">Same for dry and humid regions</td><td>Increased wetness</td><td>Increased dryness</td></tr>
<tr><td colspan="2">Humid regions and dry regions alike become more humid</td></tr>
<tr><td>Channeling synchronous in dry and in humid regions</td><td>Channeling not synchronous</td></tr>
<tr><td rowspan="2">Opposite, respectively, for dry and humid regions</td><td colspan="2">Humid regions become more humid; dry regions become drier</td></tr>
<tr><td>Channeling not synchronous</td><td>Channeling synchronous in dry and in humid regions</td></tr>
</table>

In the broad view, many streams in dry parts of the United States resemble streams in humid parts in being manifestly underfit (Dury, 1964). One well investigated instance of manifest underfitness in the arid Levant is described by Voûte (1955); that author ɜply incised valley meanders on the Orontes .at cut through bedrock to well below the level ·esent streambed and include a large channel characteristic manner is asymmertical in cross valley bends, being deepest at the outsides of The valley fill of the Orontes rises above lood-plain level, to which it descends in ter-The present river, winding in alluvium, is mani-derfit, with a wavelength ratio between valley ım channel—measured on Voûte's figure 3—

The former alluvium which constitutes the ıe terraces is coarser than the present alluvium, ıg that regimen has changed. Although not ıltogether to exclude tectonic movement, Voûte ·limatic change as the preponderant cause of ı regimen.

r the condition of the Orontes nor the wide-·currence of underfit streams in the dry parts ıited States accord with the concept of opposed ɜhavior in arid and in humid regions, respec-As streams in humid midlatitude regions are y underfit at the present time, the concept quire existing streams in arid regions to be g unusually large volumes of water; this they). In the longish term, dry and moist regions ɪerienced similar effects—namely, the effects d desiccation. Furthermore, the principles of ·ionship between bankfull discharge and down-ope require slope to decrease (or, at the least, ·o decrease) as bankfull discharge increases. ·ed erosion should correspond to increased dis-ıd increased wetness.

; point a caution becomes necessary. Erosion l areas can obviously correspond to deposition eas, so that deposition alternating in time with .eed not always mean quite the same as filling ıeling. Langford-Smith's observations (1960) ·urrumbidgee River, Australia, usefully illus-relevant topographic context. Dealing with ıtion on the riverine plains of the Murrum-.e associates it with greater-than-present dis-hat occurred in former times. As for the ;t United States, climatic fluctations are ad-; the causes of change in stream habit, but ·orkers take the direction and effect of change nt ways. Butler (1950) and van Dijk (1959) that sedimentation along the Murrumbidgee ıded to arid phases, whereas channeling oc-ıring humid phases. Langford-Smith, by con-.intains that sedimentation by prior streams, ı channels larger than those of existing rivers, ·rom increased discharge at a time when head-·eams deepened their valleys in the uplands and supplied increased quantities of sediment to the plains. To associate accelerated erosion in the uplands with accelerated deposition on the plains and to associate both with increased discharge seems to the present writer entirely reasonable. The instance of the Murrumbidgee points to the need for precise use of the term "deposition": if it connotes valley filling, then it is the antonym of channeling; but if it means sedimentation at the junction of mountain and plain, then it is the essential complement of erosion (=channeling) in the uplands.

The references cited above show that the deposition which alternated with erosion in the dry parts of the United States took place mainly in valley bottoms and not at mountain feet. Scattered radiocarbon dates place in the range 7,000–11,000 years B.P. an interval of filling that lies at the beginning of the best-studied sequence but is later than the major period of cutting responsible for the largest channels of all (Martin, Schoenwetter, and Arms, 1961). That is to say, the well-authenticated succession of deposits and the channels which they occupy, dating stratigraphically upward from the Ucross gravel and its correlatives, correspond in time to the lesser fills and channels of English rivers (table 4). The main reduction to underfitness of streams in the dry Southwest is to be sought not in this division of time, but earlier.

Although it is not yet possible to equate in detail the cut-and-fill sequence of the Southwest with that of southern England, two sets of events promise to supply fixes in both sequences—those associated with the hypsithermal maximum, and those associated with the Two Creeks (= Allerød) fluctuation. The hypsithermal maximum of temperature is widely reported not only in midlatitudes but also far beyond. Flint and Brandtner (1961) present evidence for impressive correspondence among northwest Europe, lower Austria, the Great Lakes–St. Lawrence region, the Great Basin, and Bogotá, Colombia, all of which attained their highest deglacial temperatures in about 5,000 years B.P. Gill (1955) suggests 6,000–4,000 years B.P. for the thermal maximum in Australia, and Hough (1953) detects evidence in a Pacific Ocean core for similar effects in the same bracket. Livingstone (1957) concludes that Umiat, Alaska, experienced a temperature increase at about this time. (See table 4.) In northwest Europe, for which the hypsithermal maximum is usually taken to coincide with Zone VII, increased warmth coincided with increased wetness.

Martin, Schoenwetter, and Arms (1961) use palynologic, biogeographic, and geologic evidence to demonstrate heavy summer rainfall, rather than overall drought, for the period of extensive erosion in the

Southwest United States during middle postglacial time—that is, for a bracket which includes the hypsithermal maximum. They also cite work by Murray (1951) and Butzer (1958) to the effect that increased temperature in this interval was associated with increased precipitation in the arid Near East, and they call on Selling's work (1948) for data on the precipitation increase in New Zealand and Hawaii at the same time. Whatever the niceties of change in seasonal regimen, the hypsithermal maximum seems to have been a time not only of increased temperature but also of increased precipitation, with rising precipitation well documented for widely separated regions, both dry and moist. Just as in humid northwest Europe, so in the dry West and Southwest of the United States it was a time of accelerated erosion and channeling.

The record of rising and falling levels in pluvial lakes is somewhat coarser than that of channeling and filling in river valleys; it is also longer and earlier, in part extending back from about 10,000 years B.P., whereas the channel records date mainly from 12,000 B.P. onward. The overlap, although short, is most useful.

If rather short-term fluctations be envisaged for lake level, then difficulties in separating full pluvials and interpluvials from minor shifts are easy to imagine. Similar difficulties, of course, arise with the subdivision of glacials into stadials and interstadials. Further complications ensue from attempts to equate glacials directly with pluvials. (For discussions, see Charlesworth, 1957, and Flint, 1957. For correlations, see Gromow and others, 1960; Termier and Termier, 1960; and Fairbridge, 1961. For recasting of stadials within the Wisconsin glacial, see Frye and Willman, 1960.) Nevertheless, data from the Great Basin and from certain other regions provide the means to associate one long pluvial with prolonged glaciation and to relate rapid changes in lake level with rapid climatic and glacial fluctuations in and near the Two Creeks Interstade.

Broecker and Orr (1958), and Broecker, Ewing, and Heezen (1960) infer a broad maximum of lake level in the Great Basin between 24,000 and 14,000 years B.P., a succeeding minimum, a sharp maximum about 11,500 years B.P., and near desiccation by about 9,000 years B.P. Correlation of the last two maximums, respectively, with Zone Ic and Zone III (see Curray, 1961) requires a slight displacement of dates in table 4 but does no violence to the association with the Two Creeks recession of the minimum between the two sharp maximums. The first long high stand agrees sufficiently well in date with the findings of Ruhe and Scholtes (1956) that forest dominated Iowa between 24,000 and 11,000 years B.P., under a cold moist glacial climate with cool moist intraglacial interludes. Reduced ter peratures and increased humdity characterized the Mi west simultaneously with the Great Basin, but this lo high stand occurred during the maximum cold of Wi consin glaciation (see Flint and Brandter, 1961; And sen and others, 1960; van der Hammen and Gonzal 1960), including that point on the time scale where g cially controlled sea level was at its lowest (Curra 1961; table 4.) The very high, brief, and late stands Lake Lahontan, occurring when deglacial rise of s level was well advanced, seem referable to increased p viation rather than to increased cold. Each admitte coincided with a cold fluctuation, but nothing more w involved than temporary reversals of the temperat graph; the cold of full maximum glaciation failed reestablish itself in Zone Ic.

Flint and Gale (1958) find records of two success deep-water bodies in the basin of Searles Lake, C fornia. Each lake implies pluvial climate, with molluscan fossils indicating reduced temperatu Each was followed by deposition of evaporites and least, by near desiccation, in climates similar to present. Radiocarbon dates show the first plu as well established by 46,000 years B.P., and as beg ning to wane by 32,000 years B.P. The second plu spans the interval 23,000–10,000 years B.P., being t contemporary with the classical Wisconsin Glaciat with the long pluvial specified for Lake Lahontan, with the dominance of cold moist climate in the M west. The deglacial minimum stand of Searles L falls at the hypsithermal maximum, squarely correla to Zone VII of the general schema (table 4).

Hunt, Varnes, and Thomas (1953), discussing Lake Utah remnant of Lake Bonneville, state that m annual evaporation at the north end of the present is at least three times as great as rainfall; they inf very different precipitation and evaporation ratio Lake Bonneville times. Changing speed of soil for tion, as demonstrated in palaeosols, records chang the humidity of the environment; and facies cha in sediments indicate former prevalence of north or northwesterly winds, by contrast to the southw erly prevalence of today. Lakes existed during the tiary but disappeared or greatly shrank at the begin of the Quaternary. When they later reappeared, fluctuated in level with alternation of semiarid humid climates, giving at least four lacustrine epis prior to the desiccation of Lake Bonneville. The Bonneville episode included two maximums durin Wisconsin Glacial, for which volume of sediment indicates erosion more rapid than that now in prog

Eardley and Gvodetsky (1960) conclude that a foot core taken near Saltair in the deposits of the (

Lake records sedimentation from somewhere in ftonian Interglaciation until 11,000 years B.P., the lake withdrew from the Saltair site. Using earlette Ash Member of the Sappa Formation ıarker for early Kansan deposits, they distinguish rong pluvials, the first beginning in Kansan times, he whole series correlative with the record of cutting and filling in the valley of the lower ssippi. As their figure 2 shows, however, the sion of pluvials does not correspond exactly with ıccession of glacials. They extend the pluvial begin during the Kansan Glaciation through ıirds of the succeeding Yarmouth Interglaciation, g the uppermost one-third of the Yarmouth suc- ı relate to arid climate, with pluvial conditions blishing themselves during the Illinoian Glacia- ınd arid conditions returning during the Sanga- nterglaciation. Then come pluvial, arid, pluvial, pluvial, arid, and pluvial, giving a total since n times inclusive of six pluvials, four of them ging to the last glacial.

e and Leonard (1957), interpreting the ecology of eat Plains region during the Pliocene and Pleisto- infer replacement of semiarid by increasingly l conditions at the beginning of the Pleistocene culmination of humidity in Kansan times. Thus eir findings resemble those of Eardley and Gvodet- ut the evidence of great soil groups, offset to the y about 100 miles during the Sangamon Inter- tion, indicates humidity greater than that of the t day instead of the aridity shown for Saltair. oint will be taken up below. According to Frye eonard, the Great Plains, since the peak humidity Kansan, have been subject to a pulsating and lar trend of progressive desiccation, eventually cing a present-day climate approaching in dry- nd rigor that of the late Tertiary.

IE PROBLEMS OF GENERAL CORRELATION

heir several ways, the Great Basin, the dry South- the Great Plains, the Midwest, and the East all t signs of desiccation—greatly reduced levels in enclosed lakes, complete drying out of others, d volume of sedimentation, vegetational change, onment of large river channels, and manifest fitness of streams. But the incomplete corre- ence of pluvials with glacials, the great fluctuation ıviality within the time range of the Wisconsin ıtion, the conflicting evidence of conditions in the monian, and the contrast between increased sum- ıin in the Southwest in Zone VII with the simul- ıs fall of Searles Lake all suggest lack of correspondence between changes in temperature and changes in precipitation.

As commonly used in studies of enclosed lakes, the term "pluvial" connotes not increased precipitation but rise in lake level. In its implication that changes in precipitation are solely responsible for changes in level, the term is unfortunate. Similarly, phases intermediate between pluvials are arid only in the sense that, during them, lake level is reduced. Reduction in level can be due not only to reduced precipitation but to increased temperature. If the low stand of Searles Lake during Zone VII is ascribed wholly to increased temperature, it ceases to conflict with the widespread evidence of increased pluviality elsewhere. In this way, an actual increase in precipitation during hypsithermal maximum could be accommodated in the record for the Searles Lake drainage area, a record which is comparable with that reconstructed by Martin, Schoenwetter, and Arms (1961) for the Southwest. But as the lake itself was lowered by the postulated increase in warmth, it should follow that inflowing streams were also reduced in their lower reaches. For this reason, their record of cutting and filling should differ as between lower reaches and headwaters, supposing the headwaters to have been affected by increased precipitation and discharge.

The brief high stands of Lake Lahontan in Zones Ic and III have already been ascribed to increased pluviality rather than to reduced temperatures. Their occurrence and their cause are much less discordant with the record of erosion in the dry Southwest than is apparent from the generalized successions marked in table 4. As table 4 shows, erosive episodes in the Southwest correspond broadly with intervals of increased humidity; whereas depositional episodes coincide, again somewhat roughly, with intervals of increased aridity. The end of the chief pluvial associated with Wisconsin Glaciation and the end of the accompanying erosion are set at about 10,000 years B.P., extending beyond the Two Creeks interval, which was a time of low stand in the Lahontan Basin. But the studies of the Rampart Cave deposits by Martin, Sabels, and Shutler (1961) indicate an arid maximum which, bracketing 11,350 years B.P., lies between a cool moist interval of 35,000–12,000 years B.P. and a slight climatic reversal which probably relates to Zone III. For the Grand Canyon country, then, the climatic and vegetational changes in the known record are in phase with, and similar in direction to, those associated with the last high stands of Lahontan. Increasing dryness in Zone II times appears distinctly; a reversal toward moistness, coolness, or both during Zone III times, rather obscurely. The weak development of Zone III conditions recalls the evidence from Wisconsin that channeling and the development of valley meanders were vigorous in Zone Ic but can have been slight or absent in Zone III.

Infilling of some of the largest channels in southern England during Zone V is referable to increasing dryness. As the floristic record demonstrates, this was a time of increasing cold, which necessarily reduced evapotranspiration. It follows, therefore, that the reduction of precipitation was great enough not merely to compensate for the influence on runoff of reduced temperatures but actually to promote stream shrinkage.

Sea-level changes provide a scale of comparison for sequences of cutting and filling or of pluviality and desiccation. But just as pluviality can involve changes in temperature rather than in precipitation, and possibly changes in temperature great enough to offset changes in precipitation, so sea level is not controlled by temperature alone. Fairbridge (1961) convincingly urges that eustatic events should be assumed for every geologic period, regardless of climatic events, and presents massive evidence for overriding controls of eustasy that supersede both climatic and local tectonic influences. However, his detailed analysis for the last 15,000 years gives a very close correlation between minor oscillations of sea level on the one hand and climatic events on the other, with every recorded glacial advance during the last 5,000 years matched by eustatic lowering.

The conclusion of Fairbridge that each of the younger cool phases corresponds to a drought in the arid West of the United States but to pluvial events in humid midlatitudes emphasizes the principle that pluviality can be either in or out or phase with trends of temperature change. Once again, however, the term "pluvial" connotes net effects of climatic shift. If the additional supply of surface water in midlatitudes in these cool phases were due to temperature decrease, it may even have coincided with decrease in precipitation, provided that reduced evapotranspiration more than compensated for the latter.

The younger cool phases here in question set in during a span of time which produced minor episodes of cutting and filling. The known record of minor change in stream behavior may well be no more than partial; probable minor changes of sea level under control of temperature (Fairbridge, 1961; table 4) suggest comparison with the climatic and vegetational changes signalised by von Post for deglacial time. (For references and discussion, see Conway, 1948.) Reservations seem to be necessary about the likelihood of strict parallelism in the respective erosional-depositional sequences of humid and arid regions, within the scope of minor fluctuations, especially as the evidence of cut-and-fill is much coarser than the record of slight i regularities in sea level. Above a certain limit, ho ever, parallelism and synchrony seem to apply. T evidence for widespread pluviality, rise of temperatu and accelerated erosion at the hypsithermal maximu means that the relevant happenings lay beyond t critical limit for nonparallelism and lack of synchron The last abandonment of the largest stream chann and of valley meanders lies still higher above th limit, for it constituted events of greater magnitu than the partial reexcavation of channels in Zone VI

Minor irregularities apart, the graph of sea level f the last 20,000 years records, among other things, t replacement of full-glacial conditions by full-inte glacial conditions. The Zone Ic rise of Lake Lahont roughly matches a reversal of the rising trend in s level, whereas the renewed pluviality of Zone III co responds to another, but less marked, inflection of t sea-level graph. The major conversion from glacial deglacial sea levels and from pluvial to nonpluvial c mates was by no means steady. Although correspon ing to the main change of streams to underfitness, t conversion permits the hypothesis that the dates of la abandonment of large channels and of valley meande differed, for example, between the Driftless Area Wisconsin and southern England. The dates offer for the main conversion to underfitness of streams di tant from the still-receding ice fronts accord less we with the claim of an abrupt change of climate abo 11,000 years ago (Broecker and others, 1960) than wi the modifications proposed by Curray (1961, and refe ences therein).

SUMMARY

1. Field investigation in both England and the Unit States, supplemented by information from miscellan ous sources, demonstrates large meandering channe beneath the flood plains of underfit streams.

2. These large channels are taken as the beds former streams that cut valley meanders.

3. The infilling is ascribed to stream shrinkage rath than to aggradation in the usual sense.

4. Cutting of large channels and of valley meande may be dated in terms of initiation, duration, or aba donment. Initiation has ranged from early Pleist cene (=Early, Nebraskan, Günz Glacial) in some are to perhaps as late as 2,000 years B.P. in parts of t Lake Borders. Duration of cutting has ranged a cordingly. Measurement of duration by interval b tween initiation and abandonment does not, howeve imply continuous cutting. The evidence shows alte nate cutting and filling—certainly associated in so valleys with alternate meandering and braiding—on t

of former streams prior to the last reduction to rfitness. There is ample room for onsets of the rfit condition earlier than the last.

Channeling and downcutting—in humid regions, ·ally, and in upland parts of arid regions—are iable with increased wetness. This can have been o increased precipitation, reduced temperature, or

Pollen analysis and radiocarbon analysis supply oximate dates for the abandonment of some large nels and their associated valley meanders. The appear to concentrate themselves either near the of Zone Ic (Older Dryas and Allerød boundary; nning of the Two Creeks Interstade), or in Zones nd IV (recession of Valders ice, change to Boreal e), except in certain areas that were still covered nd ice or by lake water at the relevant times.

Simultaneous decrease in temperature and increase ecipitation are inferred for Zones Ic and III; the site effects for Zone II. Simultaneous increase in erature and in precipitation are inferred for Zone in both arid and humid regions, with increase in pitation more than sufficient to offset the increase mperature except perhaps in some parts of some y dry areas. Infilling of some of the largest chan- in southern England during Zone V was due to asing dryness which, occurring at a time of re- d temperature, is referable to a distinct decrease of pitation.

From the orders of magnitude represented by the full use of valley meanders and the last complete ring of large channels to bedrock, down to and iding the orders of magnitude represented by par- reclearing of large channels in zone VII, the ences of cut and fill appear mainly parallel and hronous, respectively, in arid and in humid regions.

Lesser fluctuations betwen erosion and filling, how- , need not have been parallel and synchronous, rding to whether or not changes of temperature and ipitation were in or out of phase and according hich type of change acted the more powerfully on large in particular climates.

. Quantitative implications of the conclusions here hed will be stated and discussed in a succeeding essional paper.

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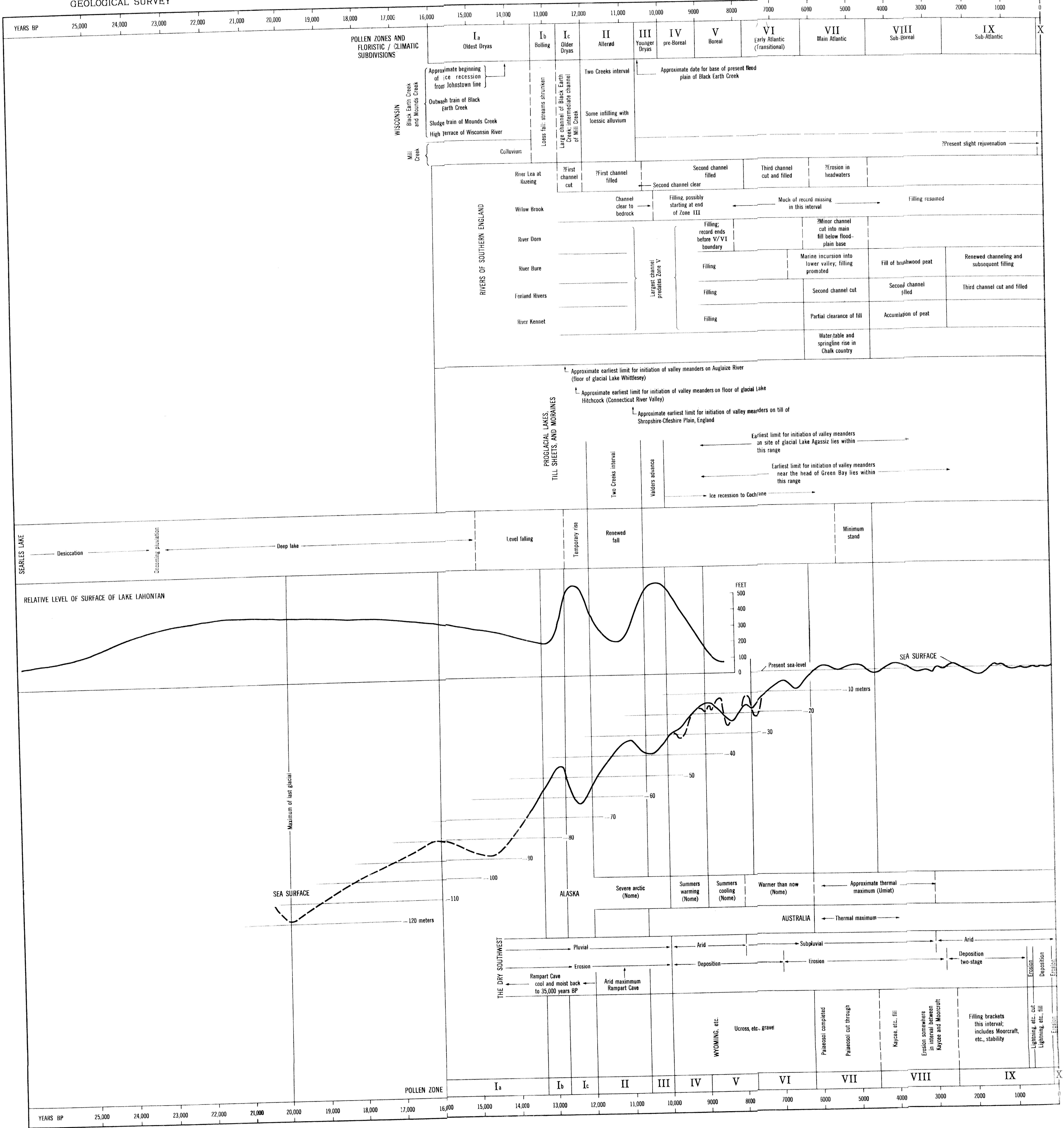

Sources and discussion given in the text

CORRELATION TABLE OF SELECTED GLACIAL, DEGLACIAL, AND POSTGLACIAL EVENTS

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Theoretical Implications of Underfit Streams

GEOLOGICAL SURVEY PROFESSIONAL PAPER 452-C

Theoretical mplications of Jnderfit Streams

G. H. DURY

ENERAL THEORY OF MEANDERING VALLEYS

EOLOGICAL SURVEY PROFESSIONAL PAPER 452-C

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CONTENTS

ILLUSTRATIONS

TABLES

SYMBOLS

a	Cross-sectional area of present channel
A	Cross-sectional area of former channel
d	Depth of present channel
D	Depth of former channel
F_q	Ratio between former and present mean annual runoff
l	Wavelength of present (stream) meanders
L	Wavelength of former (valley) meanders
M	Drainage area
n	Roughness coefficient for present stream
N	Roughness coefficient for former stream
p	Wetted perimeter of present channel
P	Wetted perimeter of former channel
q	Present discharge (usually at bankfull stage)
q_{bf}	Present discharge, specifically at bankfull stage
$q_{2.33}$	Present discharge at the 2.33-year flood
q_{rp}	Present discharge at fixed return period
Q	Former discharge (usually at bankfull stage)
Q_{bf}	Former discharge, specifically at bankfull stage
r	Hydraulic radius of present channel
R	Hydraulic radius of former channel
s	Downstream slope of present channel
S	Downstream slope of former channel
v	Present mean velocity through the cross section
V	Former mean velocity through the cross section
w	Bed width of present channel
W	Bed width of former channel
$b,c,d,$ $e,e',$ $f,f',$ $g,k,$ k'	Numerical constants

GENERAL THEORY OF MEANDERING VALLEYS

THEORETICAL IMPLICATIONS OF UNDERFIT STREAMS

By G. H. Dury

ABSTRACT

;echnique of defining discharge at the natural bankfull upon rivers subjected to artificial banking and dredging, lied to extending the data available on the relation between ɑt of discharge and wavelength of meanders. The new confirm that wavelength varies with the square root of .rge and support the contention that bankfull discharge nifestly underfit streams has been reduced by the square reduction shown by meander wavelengths.

›wances, however, are necessary for additional changes, ly those in channel form and in downstream slope. The :hannels occupied by former streams are thought to have ıer width-and-depth ratio than the present channels, thus , for instance, a 25:1 ratio of cross-sectional area where ɜd-width ratio is 9:1 or 10:1. If velocity through the r cross section at bankfull stage were identical with present-elocity at the corresponding stage, then the discharge :or highly underfit streams would be about 25:1 instead 1 or 100:1 as previously suggested by the writer. But re-n to manifest underfitness involves a reduction of down-ı slope on account of lengthened trace, for which infilling of ·ard valleys or excavation of downstream reaches do not ːnsate. The reduced slope, in turn, involves reduction in ;y, which considerably offsets the change in channel form. et outcome of revised calculations is that a discharge ratio ›ut 50:1 or 60:1 is required where streams are highly ït—that is, where the wavelength ratio is 9:1 or 10:1—and ıarge ratio of about 20:1 is required where the wavelength is about 5:1. This last ratio is widely represented in :.

ıiry into the possible hydrologic effects of climatic change into account the temperatures reconstructed for full-times and applies them in transformations of the empir-ːletermined interrelation of temperature, precipitation, and l runoff. In conjunction with increases in total precipita-y a factor of 1.5–2.0, the inferred temperature changes are .e of increasing annual runoff by factors in the approximate of 5.0–10.0 in a wide range of existing climates. Com-proportional increases in runoff rise toward increasingly ıd increasingly warm climates.

ıperature change alone is not sufficient to explain the ·ed morphological effects, especially in view of the dating .ain episodes of channeling and of the assigning of initiation .ey meanders to parts of the deglacial succession when the ı of low full-glacial conditions had already been passed and air temperatures were distinctly rising.

zen ground, whether seasonally frozen or permafrost itself, t provide a general explanation of the former discharges required by underfit streams. So much is shown by reference to the hydrologic regimens of present-day climates in Alaska and by a hypothetical translocation of seasonally cold climate from Wisconsin to the gulf coast of Texas. Manifestly underfit streams exist in this latter region, which is well beyond the extreme limit of permafrost at the last glacial maximum; similar streams in Puerto Rico are even further distant from the former lines of ice stand.

Hypothetical modifications of regimens of precipitation suggest that change in total precipitation is likely to have been more influential than change in seasonal concentration. The difference between the increase effected in total runoff both by reduced temperatures and by increased precipitation and the increase computed for momentary peak discharge is likely attributable to the short-term effects of single storms, especially those of rather high frequency and rather long duration. Their influence can readily be accommodated within the frame-work of a modest general increase in precipitation, and also within the framework of greatly reduced temperatures which do not prohibit the required increases either in total precipitation or in single rainfalls.

The general postulate of increased precipitation in early deglacial times agrees with fluctuations of pluvial lakes; the loca-tion of the postulated increased precipitation on the time scale does not conflict with extensive and persistent continental highs, postulated for full-glacial episodes. More broadly, the climatic and meteorological demands made in connection with this general theory of underfit streams accord with reconstructions of the global weather patterns of high-glacial episodes and with observations of marked changes in rates of deep-sea sedimenta-tion.

INTRODUCTION

This paper concludes the development of the general theory of underfit streams begun in Professional Paper 452–A (Dury, 1964a) and continued in Professional Paper 452–B (Dury, 1964b). The first of those papers reviewed terminology, established the widespread occurrence of underfit streams, demonstrated that not all underfit streams need possess meandering channels at the present time, and showed that derangements of drainage cannot supply the general hypothesis of origin which the facts of distribution and chronology require. The second paper reviewed and amplified studies of large filled channels in the valley bottoms of underfit streams, recorded evidence for dating the initiation and abandonment of large channels and valley meanders, and suggested the location of certain events on a scale of general chronology.

Neither the introduction to this series of papers nor the acknowledgements of extensive help will be repeated here, except for the general statement previously made in Professional Paper 452–B that many individuals—in particular, both full time and part time members of the U.S. Geological Survey—have been most generous with assistance in the field, with discussion, and with constructive criticism.

The following text extends the two foregoing papers by an inquiry into the hydrologic and climatic implications of underfit streams. Since the readiest available standard of comparison between former and present streams is wavelength of meanders, the argument relies in part upon observations of wavelength ratio. Most early work on underfit streams relied on the general circumstance that meanders on large streams are larger than those on small streams; such work, however, was not reinforced by measurement or by definition of stream size. The empiric relation $l \propto q^{0.5}$ between meander wavelength and discharge at bankfull stage is here reexamined and confirmed. It validates the inference that a pronounced reduction in meander wavelength, such as that involved in the conversion of streams to manifest underfitness, demands a reduction in discharge at the bankfull stage. In addition, it offers means of computing discharge ratios between present and former streams or of computing former discharges in numerical terms. Certain refinements of calculation lead, however, to revisions of discharge values obtained from wavelength alone. The revised values are here compared with quantities expectable in specified conditions of climate, allowance being made for the fluctuations of deglacial time and for the dates obtained in Professional Paper 452–B for onset of underfitness. The general outcome is that the required changes of discharge accommodate themselves within the reconstructed sequence of climatic change and that, to explain them, changes in precipitation are needed in addition to changes in temperature.

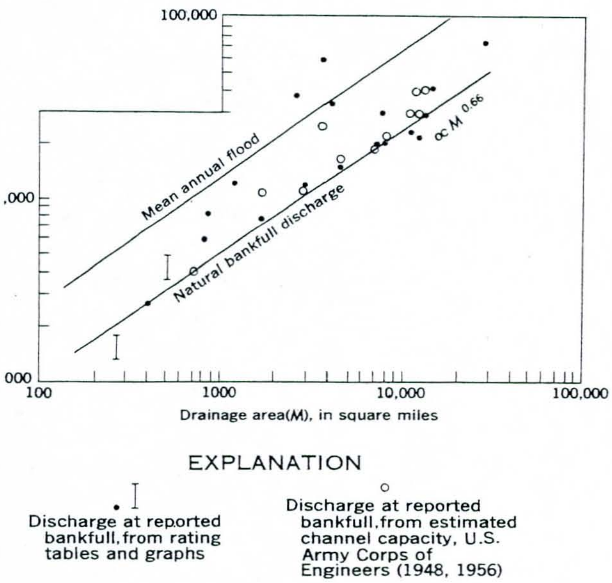
Unless the contrary is specified, calculated former discharges apply to the largest of ancestral streams—those appropriate in size to valley meanders and capable of scouring the large channels proved in numerous valleys. Similarly, discharge ratios are between present-day streams and their largest ancestral streams. Former discharges and discharge ratios alike relate to maximal underfitness, not, for instance, to the intermediate range of underfitness involved in the shrinkage of those streams which partly reexcavated valley fills in Zone VII times (table 15). Comparison of discharge from present and former streams or calculations former discharges refer throughout to discharge bankfull stage.

EMPIRICAL CONNECTION BETWEEN WAVELENG AND DISCHARGE

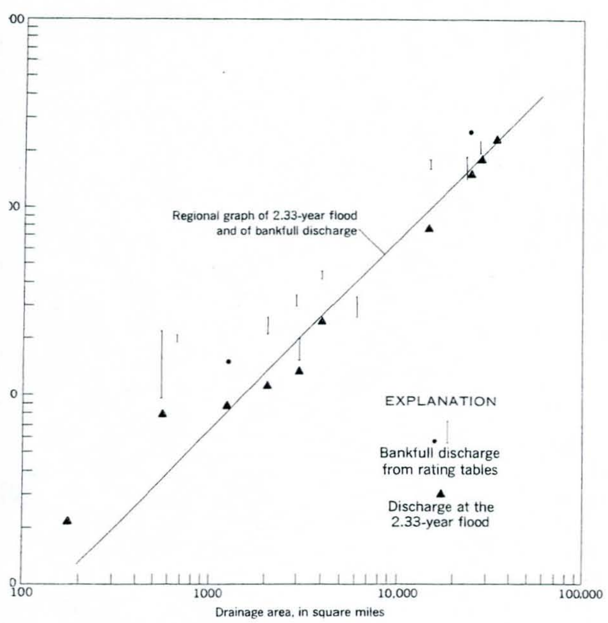
Any empirical connection between wavelength, l, a bankfull discharge, q_{bf}, may be statistical rather th causal, although this distinction is perhaps a fine o Leopold and Wolman (1957, p. 59) concluded that b width is determined directly by discharge, wher wavelength depends directly on width and thus o indirectly on discharge. Theoretical support for t direct dependence of l on w comes from Bagnold (196 who observed that, in an open curved channel, resi ance to flow descends to a sharp minimum when t curvature radius lies between 2 and 3. Since the cur ture radius is the ratio of mean radius to bed wid and since mean radius in a continuous train of meand needs differ little, if at all, from one-quarter of a wa length, wavelength should range generally from 8 to times the bed width. A higher ratio of $l:w$ than 12 used in rule-of-thumb practice by civil engineers, pr ably relates to bed widths at stages below bankf stage. (See Leopold and Wolman, 1957, p. 58–5 Analysis of observations relating specifically to bankf conditions gives a value of 9.2:1 for $l:w$ (Leopold a Wolman, 1960), well inside the range implied by t findings of Bagnold.

It is not necessary for the present purpose to inves gate any distinction which may be required betwe empirical and causal connections. A close statisti connection between wavelength on the one hand a bankfull discharge on the other will suffice: waveleng in actuality increases with increasing discharge the form $l \propto q^b$. Previous work supports a connecti between bed width and discharge in the form w and a further connection between wavelength and b width in the form $l \propto w$; the two connections comb to indicate again that $l \propto q^b$.

Formidable difficulties surround the collection concurrent data on meander wavelength and on ba full discharge. Because many streams have be artificially embanked, the natural bankfull stage hard to define. However, a possible technique definition, based on minimal discharge and draina area values, has been developed in a study of the Wh and Wabash Rivers (Dury, 1961). The technic uses discharge and area values for flow at report bankfull, flood, or flood-damage stage and also u estimates of channel capacity. When the various d charge and area values are plotted, they form irregu clouds; lines drawn for the bases of the clouds are tak as regional graphs of discharge at natural bankfull sta relevant diagram for the White and Wabash Rivers ere reproduced as figure 1. Corresponding graphs the Red River of the North and for the Sheyenne er, for a group of rivers in the northeast Ozarks for the Salt River of Missouri, and for members of Alabama River system constitute figures 2–4. kfull discharges, read off against areas for which nder wavelength is known, are listed in table 1 nst the associated wavelengths.



ɛ 1.—Determination of discharge at natural bankful: stage, Wabash and White Rivers.



. 2.—Determination of discharge at natural bankfull stage, Red River of the North and the Sheyenne River.

TABLE 1.—*Concurrent values of meander wavelength and of bankfull discharge*

[Wavelengths determined as averages for trains or groups; discharges read from regional graphs, against drainage area. Listing for each series follows order of increasing discharge]

Meander wavelength, in feet	*Bankfull discharge, in cubic feet per second*
Red River of the North	
2,600	2,600
1,250	3,000
1,800	4,000
4,100	15,000
5,000	17,750
4,175	21,000
Wabash and White Rivers	
2,050	1,650
2,700	2,900
3,600	3,900
3,300	4,000
2,800	4,150
3,400	6,600
2,950	6,600
3,000	6,750
3,900	6,800
2,700	6,900
2,900	9,000
3,550	9,400
2,800	10,500
3,250	10,600
2,800	11,250
3,000	11,250
3,550	15,000
4,400	15,100
4,000	21,000
5,050	21,500
9,000	24,000
4,750	24,250
5,300	25,250
6,500	25,750
6,000	27,000
7,900	27,000
5,900	27,250
6,200	28,500
8,000	31,500
10,000	31,500
7,400	34,000
Sheyenne River	
800	115
825	365
825	420
1,100	800
1,200	1,300
1,925	1,800
1,575	1,950
1,100	2,950
Northeast Ozarks (Meramec and Bourbeuse systems)	
1,085	1,400
1,890	3,800
1,700	4,400
1,560	5,400
1,700	5,600
2,000	6,400
1,920	8,600
1,660	10,250
1,950	11,750
2,750	18,750
3,270	22,000
3,000	27,000
Salt River, Missouri	
720	450
1,100	2,100
1,000	2,400
900	3,300
1,060	6,250
3,675	15,750
2,900	17,000
Alabama River system	
4,850	19,000
6,125	26,500
9,250	105,000
8,580	110,000
7,920	130,000
9,500	145,000

Figures 3 and 4 display a lack of parallelism between discharge and area graphs of bankfull discharge on the

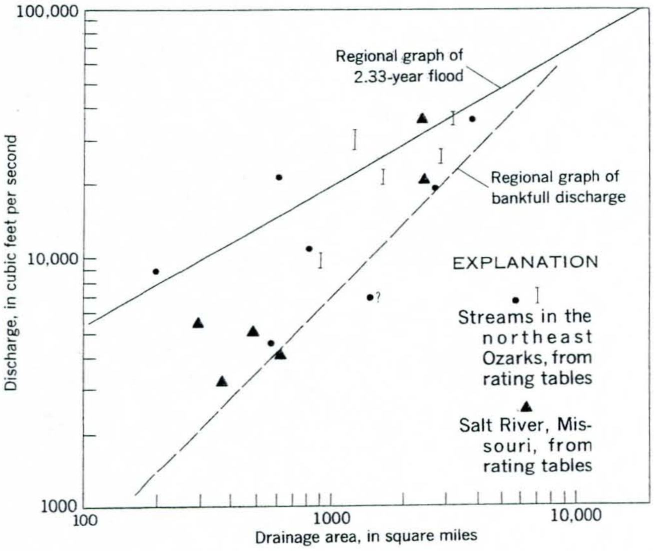


FIGURE 3.—Determination of disacharge at natural bankfull stage, rivers in the northeast Ozarks and the Salt River of Missouri.

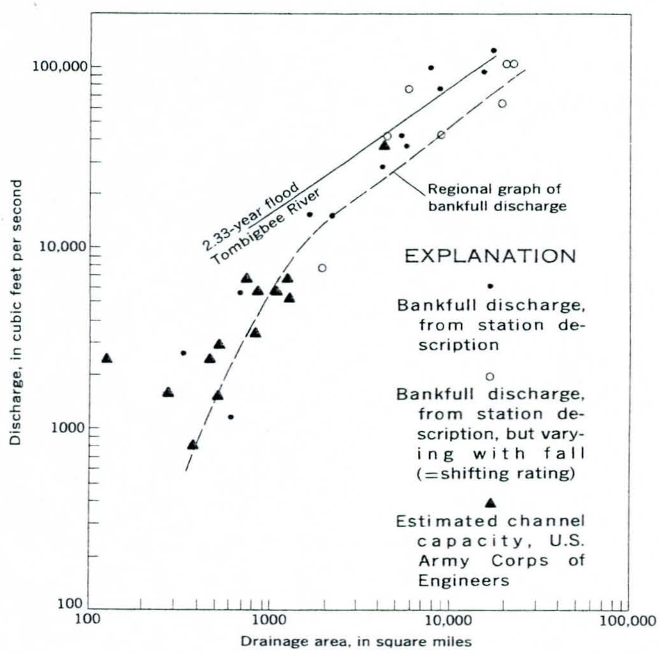


FIGURE 4.—Determinations of discharge at natural bankfull stage, members of the Alabama River system.

one hand and graphs of discharge at a fixed return period on the other. It follows that bankfull discharge cannot be expressed, in these instances, as a constant fraction of discharge at the 2.33-year flood. Regional studies of magnitude and frequency of floods normally deal with annual series of momentary peak discharges, to which the graphs here also relate, and such studies are directed in part toward defining the mean annual flood. Where the data conform to the theory of extreme values (Gumbel, 1945, 1958), the mean annı flood has a return period of 2.33 years; the mean annı and the 2.33-year floods are indeed taken as identi in many reports. Not uncommonly, floods of other turn periods are expressed as multiples or fractions the 2.33-year flood, so that their graphs on Gum paper run parallel to the graph of that flood. Dov stream convergence of the pairs of graphs in figure and 4 means a downstream increase in the return per of bankfull discharge; this downstream increase der the convenient possibility that discharge at bankf stage might be expressible as a fraction of discharge the 2.33-year flood.

The circumstance is not surprising, for many he streams are notoriously flashy, whereas channel stor tends to suppress peaks progressively in the dov stream direction. Nor does a downstream increase return period conflict with a downstream increase total duration (Dury, 1961). Again, parallel gra may appear for the middle and lower reaches o given stream despite marked lack of parallelism for headwaters. The curved graph of discharge a area, drawn in figure 4 for bankfull discharge on me bers of the Alabama River system, might well becc parallel to the graph of the 2.33-year flood, if it could extended downstream. In that event, bankfull charge would have a fixed return period for part of drainage. The rectilinear graphs drawn in figure for streams in Missouri would, if projected downstre eventually cross. But this would give bank discharge—quite anomalously for a humid region return period greater than 2.33 years. Here, also regional graph of bankfull discharge ought somewh to be inflected.

The type of relation here envisaged between gra of discharge at fixed return period and graphs of charge at bankfull stage obviously raises problem of duration—that need further study. Meanwhile, values of bankfull discharge obtained from the acc panying diagrams are both consistent among th selves and in agreement with values obtained by ot workers. Figure 5 shows the 70 wavelength and charge readings of table 1 plotted against the co sponding values assembled by Leopold and Wolr (1957, fig. 45 and Appendix E). The data of Leop and Wolman, relating to observations on rivers flumes by the two authors and by Friedkin (19 Inglis (1940, 1949), Qraishy (1944), and Brooks Eakin (unpub. data), ranged from a discharge of 0. cfs (cubic feet per second) for a meander model 1 million cfs for the Mississippi River. Though the c from the present writer's observations are more clo grouped, they still range through more than three

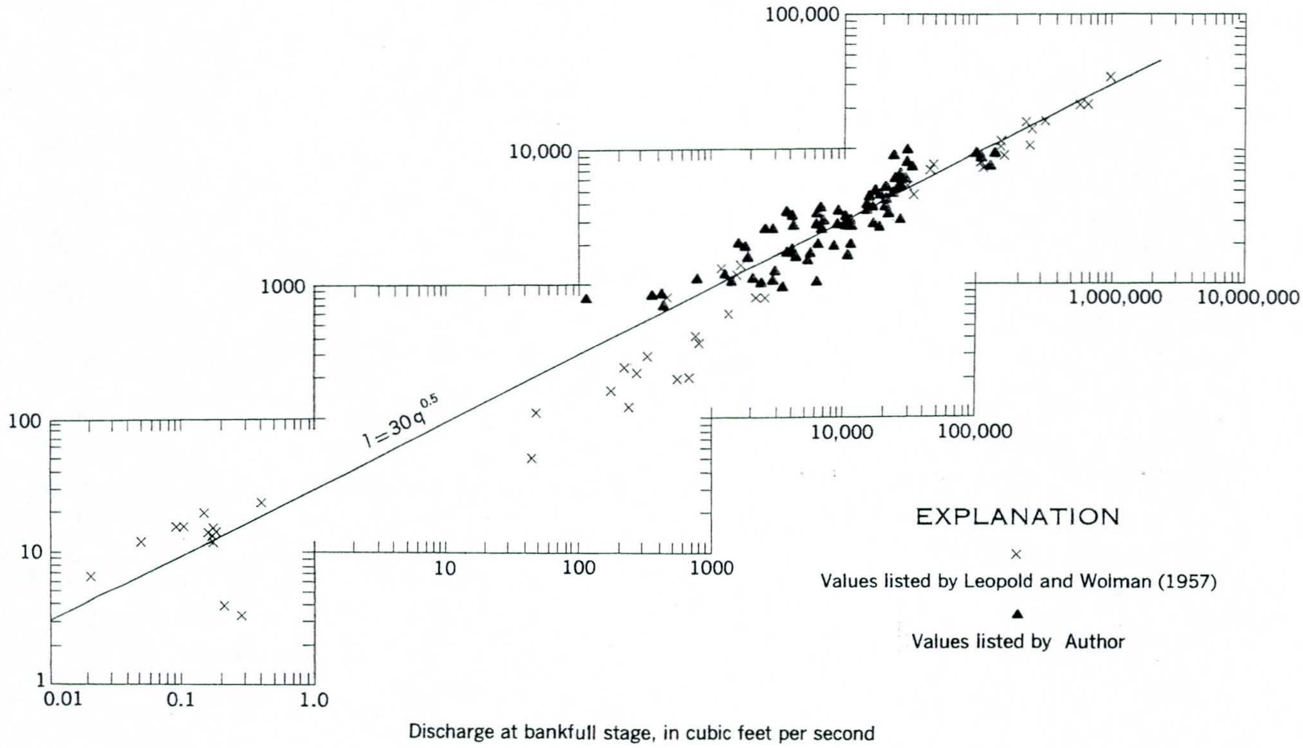


FIGURE 5.—Relation between meander wavelength and discharge at bankfull stage.

es on the scale of discharge and lie mainly in a gap by the data of Leopold and Wolman. The two ɜs of data accord well with one another—a particu- ʏ encouraging circumstance, in view of their lack eference to sediment load, roughness, channel form, e, and velocity.

ariations in load, roughness, and channel form are ıaps mainly responsible for the irregularity of ter in figure 5, although the present writer's use egional discharge and area values alongside specific ιes of wavelength may also contribute. Simons Albertson (1960) suggested that regime equations ɜrally may depend on the conditions on which they based and that they may thus be valid only within range of observed data. Specifically, these authors ned to identify variations in wetted perimeter, hy- ılic radius, and cross-sectional area according to ations in the material of bed and banks. (See also ımm, 1960.) But if a general equation for wave- th and discharge is to be obtained, variations of kind must be provisionally neglected. Slope and city may not, in one sense, be immediately relevant. ın velocity through the cross section at the bankfull e promises to vary little, if at all, in the downstream ɔtion. If that is so, then changes of slope in the ɘ direction may be omitted from the argument for time being.

The total plot in figure 5 strongly suggests a connection between wavelength and discharge in the form $l \propto q^{0.5}$—that is, an equation in the form

$$l = kq^{b}. \quad (1)$$

The best-fit equation for the readings in table 1 is

$$l = 36.1q^{0.47} \quad (2)$$

or

$$l = 26.8q^{0.5} \quad (3)$$

if the constant b is assigned the value of 0.5. This value appears in much previous work. Inglis (1941) related wavelength to bed width in the form

$$l = 6.06w^{0.99} \quad (4)$$

or

$$l = 6.06w \quad (5)$$

and bed width to discharge in the form

$$w = 4.88q^{0.5}. \quad (6)$$

Although the value of the coefficient 6.06 conflicts with the findings of Bagnold (1960) and Leopold and Wolman (1960), equations 5 and 6 combine to give

$$l = 29.6q^{0.5}, \quad (7)$$

which differs very little from equation 3. Inglis later (1949) suggested

$$l = 36q^{0.5}, \quad (8)$$

but the line for this equation runs high for the data plotted. In figure 5 is drawn the graph of the compromise equation

$$l = 30q^{0.5}; \quad (9)$$

although 30 is perhaps a slightly high value for the coefficient k, the data probably fail to justify its refinement. This last equation will therefore be adopted for subsequent use in calculation as a first approximation to the empirical connection between meander wavelength and discharge at bankfull stage.

Bed width has already appeared in the foregoing discussion. Its linear relation to wavelength is well supported by previous analysis. Inglis (1941) found that $l \propto w^{0.99}$; Leopold and Wolman (1957, p. 58) concluded that $l \propto w^{1.1}$ but later (1960) gave $l \propto w^{1.01}$ The minute departures from unity of the power functions obtained respectively by Inglis and by Leopold and Wolman in 1960 are too small to affect the present discussion, particularly since they are opposite in sign. The linear connection between wavelength and bed width, $l \propto w$, will therefore be assumed here. It follows in practice that if $w \propto q^b$, then also $l \propto q^b$.

Inglis (1941) found that $w \propto q^{0.5}$; Nixon (1959) agreed. Leopold and Maddock (1953) also concluded that $w \propto q^{0.5}$, for average variation in the width and discharge relation in the downstream direction. Their average of $w \propto q^{0.26}$ for variations at a station has no bearing on the present argument, which is limited to conditions at bankfull stage; bed width at a station, at bankfull stage, can reasonably be taken as constant if no secular change takes place in magnitude of discharge. Schoklitsch (1920, 1937) gave $w \propto q^{0.6}$, while Wolman (1955) observed a range from $w \propto q^{0.4}$ to $w \propto q^{0.57}$ within the single basin of Brandywine Creek, Pa. It may well be that variations in the power function reflect variations in roughness, load, and channel form (see Simons and Albertson, 1960); but 0.5 appears acceptable, once again as a generalized value.

CALCULATIONS OF FORMER DISCHARGES FROM MEANDER WAVELENGTH

Meander wavelength is the channel dimension that is simplest to measure. Even where maps and aerial photographs are inadequate, wavelength as mean wavelength of a meander train can be rapidly determined on the ground. There is, then, a practical advantage to the use of wavelength values, both in calculating former discharges and in determining ratios between former and present streams.

Bed widths cannot be used in the treatment of former streams unless the banktops of former channels can be identified with reasonable certainty. Where they can be so identified, the form of the whole cross section is usually known. Calculations can then be elaborated by the introduction of cross-sectional area, wetted perimeter, hydraulic radius, and a slope factor. As will be shown presently, the net result of such elaborations is to tend to reduce the discharges computed for former streams below the values computed from waveler alone. But the reductions are pronounced onl roughness is left out of account. Discharges compu from wavelength ratio appear to be of the correct o of magnitude, even though leaving considerable r for uncertainty.

The range of bankfull discharges and of mea wavelengths through which the relation $l \propto q^{0.5}$ h is large enough to justify the assumption that $L \propto$ where L is the wavelength of former (valley) mean and Q is the associated discharge at bankfull st Therefore

$$L/l = (Q/q)^{0.5}$$

and

$$Q/q = (L/l)^2;$$

in more general terms,

$$L/l = (Q/q)^b$$

and

$$Q/q \doteq (L/l)^{1/b}.$$

Examples of the relevant calculations occur in D (1958, p. 110–113; 1960, p. 230–235). Subseq observations, however, show that the observed w length ratios L/l used in these calculations are unusu high; the discharge ratios computed from them accordingly extreme. The wavelength ratios were tained in areas where streams are underfit to an unu degree. Insofar as such ratios can be defined for w regions, a value of the order of 5:1 seems quite comn With b taken as 0.5, this ratio gives Q/q as 25:1. additional points arise here: the effects of spe values assigned to b in the treatment of single regi and variations within single regions of the ratio L/l

Results of analyses of wavelength are convenie expressed in the form $l \propto M^f$, where M is drainage a numerically,

$$l = eM^f.$$

Where the discharge and area relation is known or be defined, it usually takes the form $q_{rp} \propto M^g$, w q_{rp} is momentary peak discharge at fixed return per Consequently, if bankfull discharge is assumed to h a fixed return period, then $l \propto q^{gf}$. The relation l can still hold where f is less than 0.5, for g is comm less than unity. For instance, if $q \propto M^{0.8}$ and $l \propto M$ then still $l \propto q^{0.5}$. However, analysis of wavelength area and discharge and area relations on the Eng rivers Nene and Great Ouse (Dury, 1958, 1959) sug that something more is involved. On these riv $q_{rp} \propto M$, but $l \propto M^{0.44}$. Therefore, if bankfull disch has a fixed return period, then

$$l \propto q^{0.44}{}_{bf}.$$

uation 11 then gives

$$Q/q=(L/l)^{1/0.44}; \qquad (15)$$

:e

$$L/l=9,$$

$$Q/q=150 \text{ approximately.}$$

however,

$$l,L\propto q^{0.5},Q^{0.5},$$

n

$$Q/q=80 \text{ approximately.}$$

[he higher the value of b, the lower is the calculated io Q/q and the simpler is the problem of accounting the discharges computed for the former meanders. mere wish to simplify, however, carries no logical :e.

f bankfull discharge in this region does not have xed return period on the annual series of momentary .ks, its graph must be either steeper or less steep .n the graphs of discharge at fixed return periods. steeper, so that the return period of bankfull dis- .rge increases downstream, as in figures 3 and 4, n $q_{bf}\propto M^{>1}$, which seems unlikely. If, however, $\propto M^{0.88}$, so that $l\propto q_{bf}^{0.5}$, then the return period of ıkfull discharge decreases downstream, which also ms unlikely. If a strict linear connection between velength and bed width is assumed, then $w\propto q^{0.44}$, ich is possible although not especially probable ofar as it implies a downstream increase in depth l width ratio. Unfortunately, information on veloc- is not available.

Specific difficulties of this kind cannot be resolved without further study. Meanwhile, the general equations 1 and 9, referring as they do to a wide range of observations, serve perhaps to suggest that $Q/q=80$ may be at least as likely as $Q/q=150$ for the combined drainage of the Nene and Great Ouse.

An additional reservation is now required. Although massed plots of wavelength against drainage area (Dury, 1960, fig. 2; fig. 6 of this report) show a clear division of valley meanders from stream meanders, the ratio L/l need not always be constant throughout a region. When a first regional comparison was made between the two sets of wavelengths—that for the Nene and Great Ouse (Dury, 1958, fig. 7)—the slight upward convergence of the two best-fit graphs was ascribed to accidental variability of the data. Convergence in the same direction has subsequently appeared in several regional graphs, whether wavelengths are plotted against drainage area (fig. 9) or whether the respective wavelengths of valley meanders and of stream meanders are plotted against one another. In some areas, that is to say, the ratio L/l decreases in the downstream direction.

The second (equation 16) of the best-fit equations for the connections

$$l=eM^f \qquad (14)$$

and

$$L=e'M^f, \qquad (16)$$

where $e'/e=L/l$ must be revised to

$$L=e'M^{f'}, \qquad (17)$$

where $f'<f$.

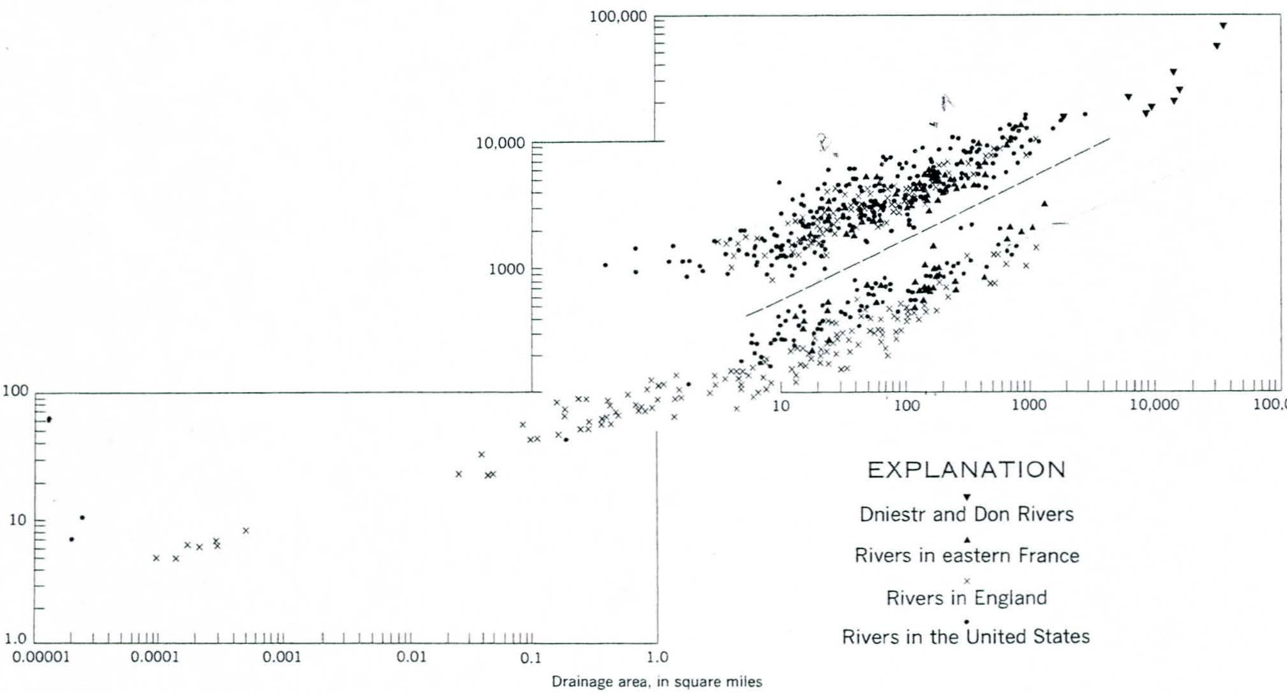


FIGURE 6.—Massed plots of wavelength against drainage area. Upper band, wavelength of valley meanders; lower band, wavelength of stream meanders.

If the downstream decrease in the ratio L/l is not merely apparent from variable data but is actual, then it may be due to former channel storage. Storage is in part responsible for the frequently observed reduction, in the downstream direction, of discharge per unit area for a flood of given frequency. Channel storage may have been more pronounced on the former rivers than it is on the present rivers simply on account of former large channel dimensions. But if this is universally so, it then becomes remarkable that some pairs of wavelength and area graphs, respectively for former and for present meanders, run parallel. If the parallelism is not due to variability of the data, it may be due either to the selection of areas too small to reveal the effects of former progressive storage in the downstream direction or to the offsetting of the effects of storage by slopes steeper than those of the present day. Too little information is available to permit choosing one of these possibilities or a combination of them.

Again, contrasts in climate and hydrology between headwater basins on the one hand and middle and lower basins on the other might be suggested as a cause of downstream decrease in the ratio L/l. Here too, not enough is known for thorough discussion, although the likely general effects of both relief and noncontribution can be sketched. In areas of strong relief, particularly if the present climate is somewhat arid, it seems entirely possible that former pluvial conditions affected the headwaters more strongly than they affected the lower reaches. The Dirty Devil and Virgin Rivers, Utah, appear to have ceased to augment the wavelength of their former (valley) meanders, in the downstream direction, within the range of height 4,000–5,000 feet above sea level (fig. 7). The implication is that runoff

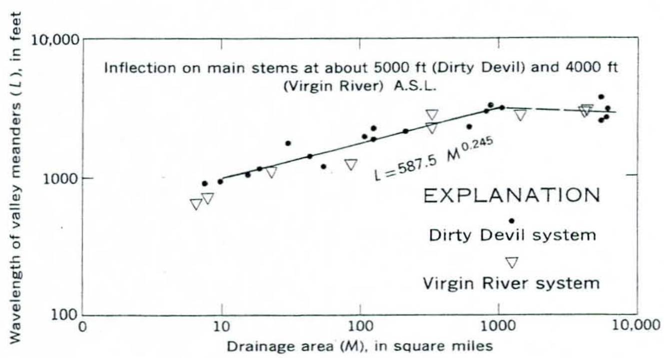


FIGURE 7.—Relation of wavelength of valley meanders to drainage area, Dirty Devil and Virgin Rivers, Utah.

in this height range was formerly counterbalanced by losses to evapotranspiration: at higher levels there was a water surplus and at lower levels there was a water deficiency when the rivers ran at bankfull stage in their former meanders down to 5,000 or 4,000 feet abov level. Increasing aridity, with change toward exis climates, probably shifted the zone of counterbal vertically upward. Available maps for the two systems unfortunately do not permit a compa between former and present wavelengths. On Humboldt River, Nev., and Owyhee River, ((fig. 8), the scanty evidence collected is not suffi

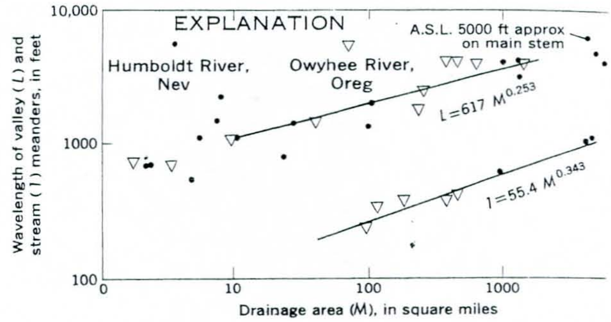


FIGURE 8.—Relation of wavelengths of valley and stream meanders to d area, Humboldt River, Nev., and Owyhee River, Oreg.

to show whether or not the graph of stream mea becomes inflected at a level different from that o inflection in the graph of former meanders. How it seems highly likely that the change toward a would reduce the total area of contributing draina that the effective basins are now smaller than once were.

The areas used in the regional graphs of wavel against drainage area have been planimetered maps marked with the physical divides; no allov has been made for noncontribution. Whatever allowance is that ought to be made for former c tions, the present allowance is likely to be greater

Where the whole of a given basin was formerly tributing, but parts of it now fail to contribute plot of former wavelengths against drainage a correct. But the plot of present wavelengths s be adjusted by reductions in the values of area ag which wavelengths are set, and the slope of the would consequently steepen. Even if the two length graphs, as first drawn, ran parallel, they then converge: the adjusted plot of present wavele would ensure a downstream decrease in the rati

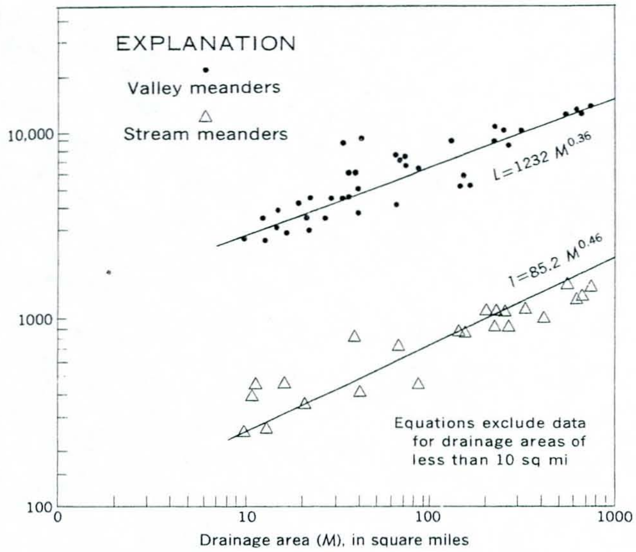
The regional graphs of wavelength for the Dri Area of Wisconsin, a humid region, no allowance made for noncontribution, take the forms

$$l=85.2M^{0.46}$$

and

$$L=1{,}232M^{0.36}$$

that is, there is upward convergence (fig. 9). Bu the likely view that the former meanders were c



ɪURE 9.—Relation of wavelength of valley and stream meanders to drainage ɪrea, rivers in the Driftless Area of Wisconsin (Kickapoo, Platte, Galena, and ᵖecatonica Rivers).

mes of great cold, there seems little scope for increasing ɪe steepness of the graph of former meanders. Any lowance made with respect to present conditions, for ɪncontribution in narrow bands along the divides or r percolation, would steepen the graph of present eanders and would thus increase the downstream re- ɪction in the ratio L/l. It is difficult to imagine that ɣdrologic contrasts between headwater and other ɪrts of drainage areas were greater in former times ɪan they are today, and hypothetical climatic con- asts—for example, heavy snow on high ground—are ɾoblematic, especially if they are required to account ɪr downstream changes in stream behavior greater than ɪe changes which now occur.

It therefore seems necessary to conclude that, in ɪrtain basins, a downstream reduction in the ratio L/l ɪes occur, for whatever reason. The ratio Q/q, cal- ɪlated from L/l, then also decreases downstream. Bank- ɪll discharge can no longer be specified as having been ɪduced to a set fraction of its former value throughout ɪ area, and interregional comparisons can no longer ɪ made unless numerical values of area are given. omputed values of Q/q, derived from a variable ratio $/l$, are increased above the regional value in the upper ɪsin and reduced below it in the lower basin.

Discharge ratios between former and present streams ɪay therefore be computed in one of two ways, each ɪ which is open to some objection. The regional ɪlues of wavelength, as specified by best-fit equations, ɪay be accepted, and discharges may be computed for ɪem by means of the general wavelength and discharge ɪuation 9. In this event, discharges must be referred ɪ specific areas—that is, near the upper and lower ɪnds of the observed range of drainage. Alternatively, it may be assumed that $q \propto M$ and that $l \propto q^{0.5}$, so that $l \propto M^{0.5}$ and also $L \propto M^{0.5}$. The best-fit graphs, adjusted accordingly, can again be made to supply values of discharge and of discharge ratio.

Discharges for the bankfull stage in table 2 derive from the application of

$$q=(l/30)^2 \tag{20}$$

and

$$Q=(L/30)^2 \tag{21}$$

to unadjusted regional graphs of best fit, near their upper and lower ends. Variation in the ratio L/l ensures that computed values of Q/q shall also vary—within the range of calculation—from as high as 132:1 to as low as 20:1. Where meanders on the present channel are poorly developed or are poorly recorded by available maps, the computation is possible only for Q; a number of relevant entries appear in table 2, and the best-fit graphs that have not already been presented appear in figures 10–12.

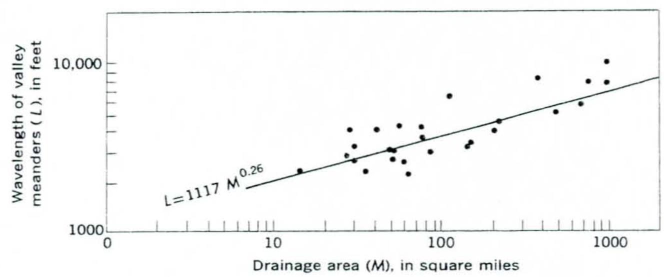


FIGURE 10.—Relation of wavelength of valley meanders to drainage area, in part of Minnesota.

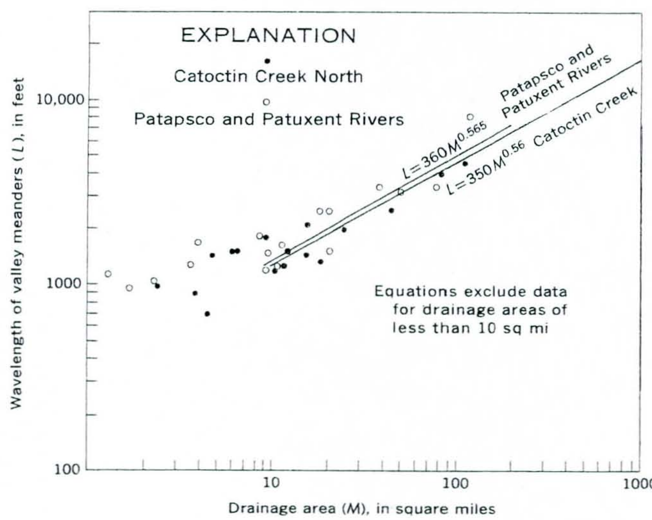


FIGURE 11.—Relation of wavelength of valley meanders to drainage area, Patapsco and Patuxent Rivers and Catoctin Creek, Md.

A possible reason for converting the best-fit graphs of wavelength of present meanders to the general form $l \propto M^{0.5}$ is that for two of them f in $l \propto M^f$ is greater than 0.5: for the group of streams at the head of Green Bay,

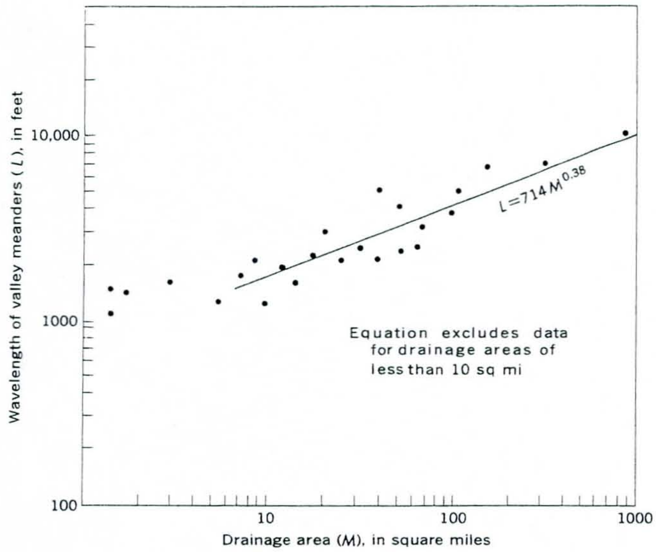


FIGURE 12.—Relation of wavelength of valley meanders to drainage area, Monocacy River, Md.

the best-fit value is 0.54, and the corresponding value for southern New England is 0.56. Also, in these two areas $L \propto M^{0.54}$ and $L \propto M^{0.56}$, respectively, so that the ratio L/l is constant in each area. Unless bankfull discharge in these two areas is related to drainage area in the form $q \propto M^g$ where $g > 1.0$—on the face of it, an improbability—the results of best-fit analysis may be due merely to accidental variability of the data. But if so, then the fact that f in $l \propto M^f$ is elsewhere less than 0.5 could equally be due to accidental variation, notwithstanding what has been said previously about relation of bankfull discharge to discharge at fixed turn period. At the same time, the difference betw f' and 0.5, in the expression $L \propto M^{f'}$, is so well mar that a wholesale conversion of best-fit graphs former meanders to the general form $L \propto M^{0.5}$ is diffi to justify. The values obtained from such conver are here required merely for discussion.

Table 3 lists bankfull discharges obtained from application of equations 20 and 21 to adjusted regi values of meander wavelengths, on the assumption t $Q,q \propto M$ and that $L,l \propto Q^{0.5},q^{0.5}$ respectively. The hig ratio of Q/q is now reduced to 68.5:1 in place of 132:1 previously obtained, while the lowest ratio mains at 20:1. As in table 2, the highest disch ratios apply to the Nene and Great Ouse or to Driftless Area of Wisconsin. This latter region trasts with the adjoining drift-covered tracts of Mi sota, west of the Mississippi River, where fo meanders are considerably smaller, area for a Thus, computed former discharges for Minnesota 4,600 cfs at 10 square miles and 50,400 cfs at 1 square miles, against 9,270 and 243,000 cfs for Driftless Area. Both the Driftless Area and sout England, in which the Nene and Great Ouse basin included, are regions where streams are highly und Discharge ratios ranging from 20:1 to 25:1, as culated from wavelength ratios for the Green Ri for the Green Bay country, and for New England,

TABLE 2.—*Bankfull discharges, calculated from unadjusted wavelength and area values*

[Computed from $q=(l/30)^2$, $Q=(L/30)^2$, where wavelengths are given by the best-fit equations $l=eM^f$, $L=e'M^{f'}$]

Basin or region	M, in square miles	Steam meanders			Valley meanders			
		e	f	q, in cubic feet per second per square mile	e'	f'	Q, in cubic feet per second per square mile	
Nene and Great Ouse Rivers, England	1,000 10	59	0.44	1.7 2.9	708	0.38	106 320	
Driftless Area, Wis	1,000 10	85	.46	4.6 7.0	1,232	.36	243 927	
Green Bay area, Wisconsin	100 10	68	.54	7.4 6.1	338.5	.54	184 153	
Southern New England	1,000 10	58	.56	8.5 4.9	260	.56	172 99	
Green River, Ky	500 10	135	.41	6.6 13.4	663	.41	160 325	
Eastern France	1,000 10	49	.53	4.0 3.8	400	.50	178 178	
Humboldt River, Nev., and Owyhee River, Oreg	1,000 10	55	.34	.4 1.6	617	.25	13.4 13.4	
Cotswolds, England	100 10	---	---	---	680	.37	155 282	---
West side of glacial Lake Agassiz area	1,000 10	---	---	---	1,155	.30	94 618	---
Minnesota, west of Driftless Area	1,000 10	---	---	---	1,117	.26	50 460	---
Mission River, Texas	1,000 10	---	---	---	176.5	.62	181 60	---
Catoctin Creek and Patapsco and Patuxent Rivers, Md	100 10	---	---	---	355	.56	243 185	---
Monocacy River, Md	1,000 10	---	---	---	714	.38	108 326	---
Dirty Devil and Virgin Rivers, Utah	1,000 10	---	---	---	587.5	.245	18 119	---
Deep Spring Valley, Calif	100 10	---	---	---	373	.25	16 49	---

ıought to be much more nearly representative of idespread conditions. As will be now demonstrated, oreover, the whole series of discharge ratios obtained om wavelength ratios is susceptible of reduction.

ABLE 3.—*Bankfull discharges, calculated from adjusted wavelength and area values*

[Computed from $q=(l/30)^2$, $Q=(L/30)^2$, where $L, l \propto M^{0.5}$, $l=eM^{0.5}$, $L=e'M^{0.5}$]

Basin or region	Stream meanders		Valley meanders		Q/q
	e	q, in cubic feet per second per square mile	e'	Q, in cubic feet per second per square mile	
ne and Great Ouse Rivers, England	41	1.8	355	125	68.5
iftless Area, Wis	66	4.9	515	295	60
eenBay area, Wisconsin	79	7.0	397	175	25
ıthern New England	84	7.8	377	158.5	20
een River, Ky	82	7.5	403	180	24
stern France	71	5.6	400	178	32

CALCULATIONS OF FORMER DISCHARGES FROM VALUES OTHER THAN MEANDER WAVELENGTH

When form and dimensions are known for the former ıannels, calculation can involve cross-sectional area, etted perimeter, and downstream slope. In Manning's uation, let v, n, r, s, and a be, respectively, the mean locity (at bankfull stage), the roughness, the hydrau- radius, the downstream slope, and the cross-sectional ea of the present stream, and let V, N, R, S, and A the corresponding values for the former stream. hen, since,

$$v=\frac{1.5r^{2/3}s^{1/2}}{n} \quad \text{(22, Manning's Equation)}$$

$$V/v=n/N(R/r)^{2/3}(S/s)^{1/2}. \quad (23)$$

it is assumed for the moment that $N=n$, equation 23 comes

$$V/v=(R/r)^{2/3}(S/s)^{1/2}. \quad (24)$$

ıt since $Q=VA$, and $q=va$, then

$$Q/q=A/a(R/r)^{2/3}(S/s)^{1/2}. \quad (25)$$

, a, R, and r can be determined if the cross sections former and of present channels are known. S/s can adily be derived from comparative sinuosities: $=H/T$ and $s=h/t$ where H and h are the vertical falls the respective horizontal distances T and t.

Professional Paper 452–B notes that, on becoming anifestly underfit, a stream lengthens its trace; nple calculation shows that, on many streams, filling on the headwaters could not possibly compensate for the reduction of slope involved in a lengthıed trace, and field observation reveals that the tendıcy of compensatory infill is either negligible or absent altogether. In point of fact, any such tendency has yet to be demonstrated. If anything, the trend in southern England is precisely the reverse of compensatory infill—numbers of streams are working close to bedrock in their uppermost reaches, but have experienced drowning near their mouths. There can be no objection to assuming, for general purposes, that H and h are identical. The value of S/s for a given manifestly underfit stream is given by t/T—that is, by the comparative lengths of the two traces. Examples of comparison appear in table 4.

Figure 13 is a nomogram for determining the ratio Q/q, where cross-sectional areas A and a, wetted perimeters P and p, and the slope ratio S/s are known. Two examples are worked, one for an actual case although somewhat generalized, and one for a hypothetical, but probably representative, case. The first example (line 1 in the diagram) accords with the approximate 25:1 ratio of cross-sectional areas, 8:1 ratio of wetted perimeters, and 1.3:1 ratio of downstream slope observed on the East Pecatonica and Mineral Point branches of the Pecatonica River (Dury, 1962, 1964b) and on the Warwickshire Itchen in England (Dury, 1954, figs. 4–6; these observations are confirmed by subsequent additional work). A line joining the appropriate points on the scales A/a and P/p is projected onto the combined scales of R/r and $(R/r)^{2/3}$. From the point so determined, a second leg of the line passes through the appropriate point on the scale of S/s, thereby giving a value for V/v on the next scale to the right—in this example, a value of about 2.5:1. From this point on the V/v scale, a third leg of the line runs through the mark of 25:1 on the second scale of A/a, thus effecting the calculation $Q/q=VA/va$. Q/q on the final scale on the extreme right appears as 60:1.

This answer accords with calculations from wetted perimeter alone—specifically, from the general relation

$$p \propto q^{0.5} \quad (26)$$

recommended by Lacey (1930, 1934, 1938), who gave the numerical connection

$$p=2.671q^{0.5}. \quad (27)$$

The previously reported values of wetted perimeter for a group of underfit streams in England (Dury, 1954, table 1) give an average close to 8:1 for P/p and an average of precisely 60:1 for Q/q. The additional data for the Itchen permit comparison between the bedwidth ratio W/w and the ratio of wetted perimeters P/p (fig. 14); P/p is the smaller ratio because of the greater depth and width ratio on the present than on the former channel. Whereas the use of the relation $w \propto q^{0.5}$ gives Q/q for the Itchen as approximately 100:1, the relation $Q/q=(P/p)^2$ gives approximately 60:1.

TABLE 4.—*Examples of lengthening of stream trace on account of manifest underfitness*

[Map sheet: U.S. Geological Survey Topographic Maps; Ordnance Survey of Great Britain. Measurements for Black Earth Creek taken on large-scale field maps]

River	Locality	Map sheet	Factor b which tra lengthen
Pembina	North Dakota	Cavalier (1:62,500)	[illegible]
Coln	Gloucestershire, England	SP/10 (1:25,000)	[illegible]
Pecatonica	Wisconsin	Mineral Point (1:24,000)	[illegible]
Pecatonica, East Branch	do	South Wayne (1:62,500)	[illegible]
Little Eau Pleine	do	Marshfield (1:62,500)	[illegible]
Cherwell	Oxfordshire, England	SP/43 (1:25,000)	[illegible]
Windrush	do	SP/21 (1:25,000)	[illegible]
French Broad	North Carolina	Brevard (1:24,000) Rosman (1:24,000)	[illegible]
Little Platte	Wisconsin	Rewey (1:24,000)	[illegible]
Black Earth Creek	do	Cross Plains (1:62,500)	[illegible]
Pecatonica	do	Mifflin (1:24,000)	[illegible]
Do	do	South Wayne (1:62,500)	[illegible]
Galena, Shullsburg Branch	do	New Diggings (1:24,000)	[illegible]
Pecatonica Mineral Point Branch	do	Mineral Point (1:24,000)	[illegible]
Great Ouse	Buckinghamshire, England	SP/62 (1:25,000)	[illegible]
Pecatonica	Wisconsin	Calamine (1:24,000)	[illegible]
Mounds Creek, East Branch	do	Blue Mounds (1:62,500)	[illegible]
East Pecatonica	do	Mifflin (1:24,000)	[illegible]
Elk Fork Salt	Missouri	Florida (1:62,500)	[illegible]
Avon	Warwickshire, England	SP/57 (1:25,000)	[illegible]
Platte	Wisconsin	Potosi (1:24,000)	[illegible]
Wye	Herefordshire, England	142 (1:63,360)	[illegible]
Lugg	do	142 (1:63,360)	[illegible]
Meramec	Missouri	Pacific (1:24,000)	[illegible]
Monnow	Herefordshire, England	142 (1:63,630)	[illegible]
Gasconade	Missouri	Morrison (1:62,500)	[illegible]
Nene	Northamptonshire, England	SP/65 (1:25,000)	[illegible]
Sowe	Warwickshire, England	SP/37 (1:25,000)	[illegible]
Galena	Wisconsin	New Diggings (1:24,000)	[illegible]
Eau Pleine	do	Stratford (1:24,000)	[illegible]
Meramec (part: see below)	Missouri	St. Clair (1:62,500)	[illegible]
Big	do	Cedar Hill (1:24,000)	[illegible]
Meramec (part: see above)	do	St. Clair (1:62,500)	[illegible]
Dry Fork	do	Meramec Spring (1:62,500)	[illegible]
Gasconade	do	Bland (1:62,500)	[illegible]
Middle Fork Salt	do	Florida (1:62,500)	[illegible]
Gasconade	do	Linn (1:62,500)	[illegible]
Bourbeuse	do	St. Clair (1:62,500)	[illegible]

The second example in figure 13 supposes a cross-sectional area reduced to one-tenth of its former value, as by a reduction to one-fifth in width and one-half in mean depth. Because the ratio S/s often appears to run rather low on rivers that are not the most greatly underfit, the value 1.125:1 is used here (see table 4). As shown, this value combines with the other ratios to give V/v as about 1.7:1 and Q/q as about 17:1. This example is thought to provide a likely indication of the change in bankfull discharge required to reduce wavelength to one-fifth of its former value.

Because the former channels were large in comparison with present channels, some increase in roughness with reduction in size ought to be conceded. But this increase is likely to have been moderated by changes in channel form. The preceding examples of the Pecatonica and Itchen combine with the generality of cross sections through former channels (Dury, 1954, figs. 4–11; 1958, figs. 1–6, fig. 10; 1961; 1964b) to show that the width-and-depth ratio on former channels is greater than that on their present-day successors, even when allowance is made for point bars on some lines of cross section. This circumstance agrees with the analysis of hydraulic geometry by Leopold and Maddock (1953), who found that

$$q \propto w^b \propto d^c \propto v^d$$

where b, c, and d are numerical constants such th $b+c+d=1.0$. If b is taken at the general (averag value of 0.5, then $c+d=0.5$. The foregoing applic tion of Manning's equation requires that, in the prese context, $d>0.0$; hence, $c<0.5$. Therefore, where and D are former bed width and former mean dep respectively, it follows from $q \propto w^b \propto d^c$ that

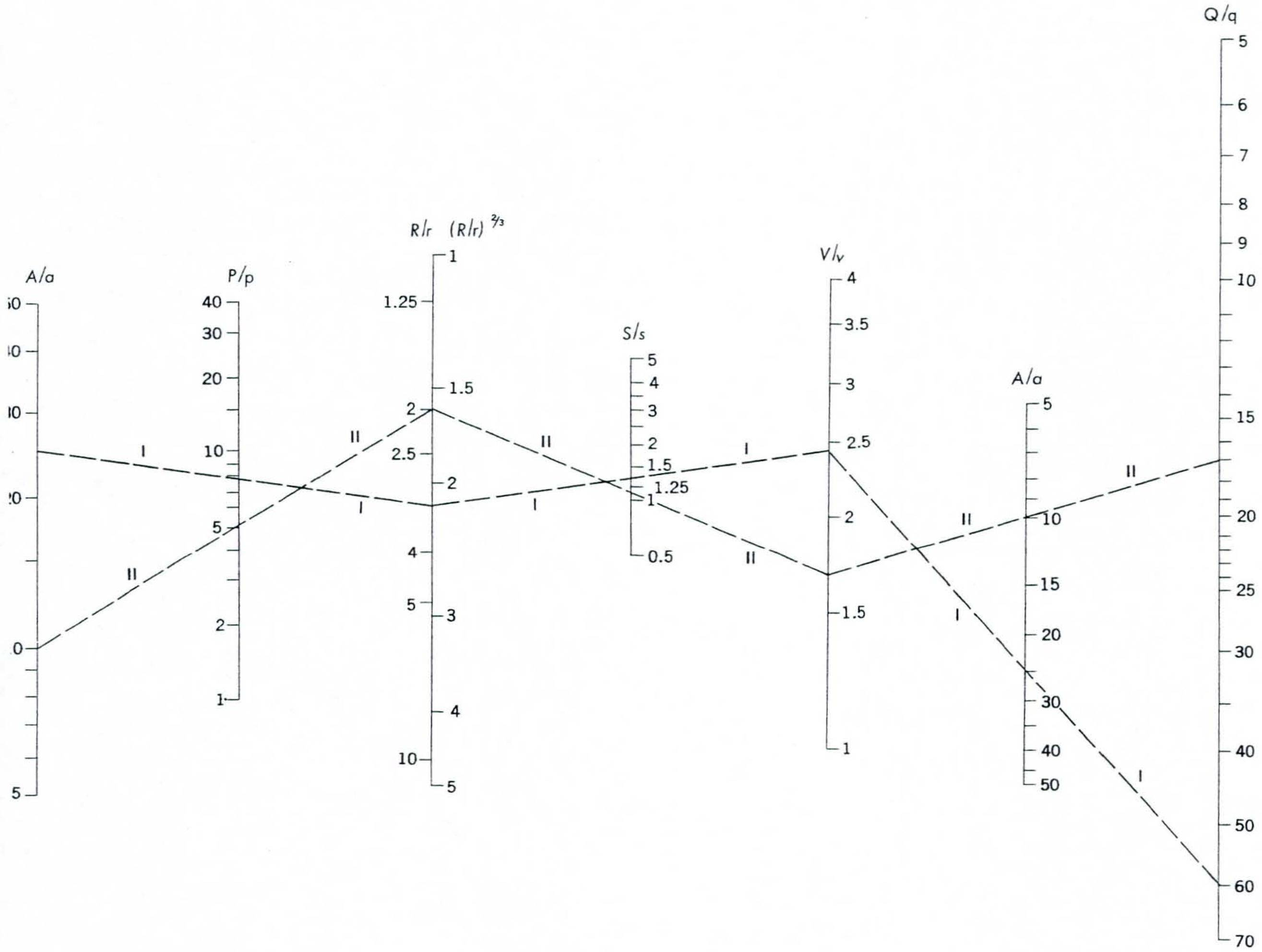
$$Q/q=(W/w)^{1/b}=(D/d)^{1/c}, \quad (\text{:}$$

so that

$$W/w=(D/d)^{b/c}, \quad (\text{:}$$

where $b/c>1.0$. That is, the width-and-depth ra should be greater in the former than in the prese channels, just as it is typically greater today on la rivers than on small ones.

Although specific values cannot be given for both t former and the present coefficients of roughness, range can be suggested for the ratio n/N. If both t former and the present channels are taken as clean b as winding through pools and shallows, their roughn coefficients should lie between 0.033 and 0.045 (Horto revisions, quoted by Linsley, Kohler, and Paulhus, 19 p. 472). The higher the value of n/N—the greater t postulated increase in roughness with reduction in s of channel—the greater is the computed reduction velocity and thus also in discharge. If N and n a identical, the ratio V/v computed from equation



[FIG]URE 13.—Nomogram for Q/q, by Manning's equation, when cross-sectional area (A/a), wetted perimeter (P/p), and comparative downstream slope (S/s), are known for present and for former channels.

[rem]ains unchanged, and Q/q is unaffected. If n is [take]n as 0.045, at the top of the stated range, and N as [0.02]3, at the bottom, then computed values of V/v and [Q/q] are increased by a factor of about 1.36. The dis[char]ge ratio 60:1 then becomes about 80:1, and 17:1 [beco]mes 23:1. The reductions in discharge ratio below [the] values computed from wavelength alone are not [fu]lly restored. In figure 15, where discharge ratios [are] graphed against wavelength ratios, a band of [prob]ability is marked by stippling. Its lower limit [corr]esponds to values of Q/q determined from equation [29, w]hich admits no change in roughness, and the upper [limit] corresponds to an increase in roughness, from large [to sm]all channels, by a factor of 1.36.

[Si]mons and Albertson (1960) inferred, from com[pari]sons of channels in varying bed and bank materials, [that] differences in caliber and cohesiveness produce [mar]ked effects on channel form at a given discharge. [Cha]nge of channel form with change of material is also [inco]rporated in the findings of Blench (1951) and and Schumm (1960), who agreed in distinguishing between the influence on, or of, the bed, and that on, or of, the banks. The graphs of Simons and Albertson show a range in the connection between wetted perimeter and discharge, in the general form

$$p=kq^{0.51}, \tag{30}$$

from

$$p=1.71q^{0.51} \tag{31}$$

for coarse noncohesive material and for sand bed and cohesive banks to

$$p=3.35q^{0.51} \tag{32}$$

for bed and banks of sand throughout. Roughly speaking, that is to say, wetted perimeter of a stable channel, for a stated discharge, can be doubled or halved in response to a change in the material of bed and banks. But, even if these findings are accepted unreservedly, they offer little help in reducing the ratio Q/q below the values already obtained from Manning's

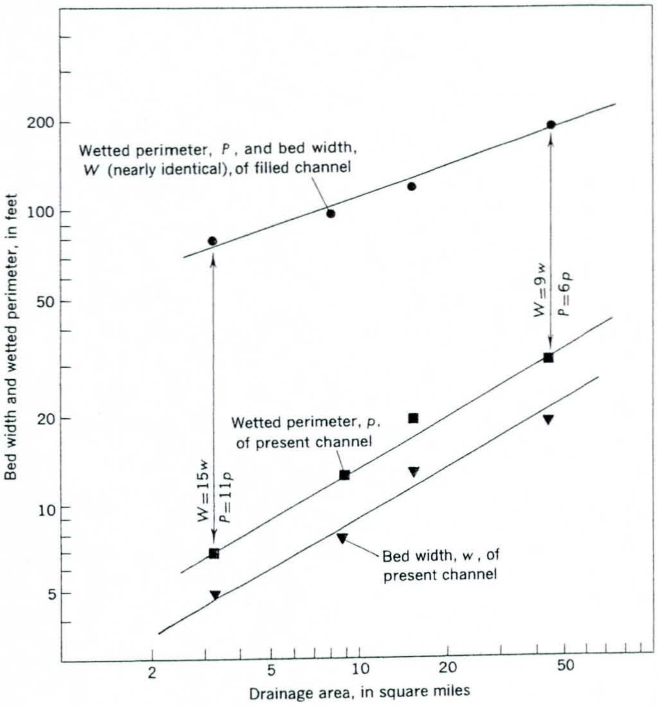


FIGURE 14.—Comparative bed widths and wetted perimeters on the River Itchen, Warwickshire, England.

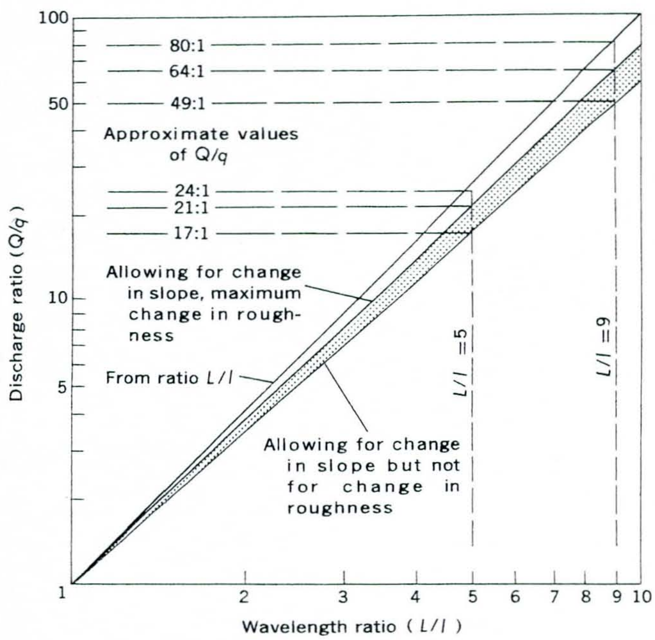


FIGURE 15.—Nomogram for relations of discharge ratio to wavelength ratio.

equation, the slope ratio, and interrelations of wetted perimeter and bed width.

The highest values of k in equation 30 and, in consequence, the highest values of p, together with the highest values of P and W for former channels, are expectable when bed and banks consist of sand. Increasing coarseness and cohesion which reduce the value of k, consequently reduce computed values p, w, P, and W. Several of the former channels, cluding all those described from England and from Driftless Area of Wisconsin, are cut in bedrock, wh ranges from compact shale through sandstone to w cemented limestone. As a group, these chan cannot have undergone unusual widening on acco of low cohesiveness of the bed and banks. Tl wetted perimeters and their widths are likely to h been minimal rather than maximal. It is scar conceivable that materials of the present channels more cohesive than materials of former channels bedrock. If, therefore, $p=kq^b$ and $P=k'Q^b$, k' is n unlikely to be greater than k.

Furthermore, since the fills of at least some of former channels vary little in caliber from top bottom, almost all the reduction in channel size been accompanied by negligible change in part size. As soon as the streams began to shrink, they contact with bedrock through all, or much, of t wetted perimeters. That the change from workin bedrock to working in alluvium, peat, or sapropel accomplished early is largely irrelevant in any ca sense to reduction of channel size.

If some of the former streams transported bed l coarser than that of the successor streams of to then, according to Simons and Albertson, the for coarseness should have tended to reduce the we perimeters of the time. On this count also, observed values of P and W appear minimal ra than maximal, and the computed values of Q/q s not to be excessive.

Support of another kind for this last inference co from the findings of Brush (1959) and of the Waterways Experiment Station (1956). (See also pold and others, 1960.) Brush (1959) concluded roughness increases markedly with sinuosity, and suggested that the concepts of grain roughness bed roughness be elaborated to include also l roughness. The Technical Memorandum of the Ex ment Station offered the conclusion that increas sinuosity reduces channel discharge by about percent when the sinuosity rises from 1.20 to 1.4 from 1.40 to 1.57. Since the present channels of s manifestly underfit streams are distinctly more sin than the former channels, in which existing flood p top the infill, former discharges computed from w length, slope, and generalized values of roughness be somewhat low. Even where the present cha are not sinuous in plan, pool-and riffle sequence the bed amount, in effect, to sinuosity. (See Leo and Wolman 1957, p. 53–55.)

ACCOUNTING FOR FORMER DISCHARGES

Since a wavelength ratio of about 5:1 seems widespread in nature, the hydrologic problem of underfit streams is reduced, by the reasoning presented in the foregoing section, to one of providing bankfull discharges about 20 times as great as the bankfull discharges of today. The exceptionally high wavelength ratios of about 9:1 call for multiplication of bankfull discharge by 50 or 60 times. (See fig. 15.) General arguments cannot go far to explain discharges of these orders. In what follows, numerical values will be freely employed, even though some are merely approximate, in order eventually to establish certain limits of physical possibility.

Bankfull discharges computed from present wavelengths, in tables 2 and 3, cannot be regarded as precise despite the general applicability of equations 1 and 9 to a wide range of data. Nevertheless, these computed values are broadly consistent both with suggestions about the return period of discharge at bankfull and with observations in other regions of the ratio $q_{2.33}/q_{bf}$ between discharge at mean annual flood and discharge at bankfull. This ratio, in turn, can be used to show that something more than a change in discharge frequency is needed within the limit set by present mean annual flood.

Reservations about assigning a fixed return period to discharge at the bankfull stage apply only to entire basins. There can be no objection to specifying a return period for bankfull discharge at a station. Moreover, the interrelation of regional graphs of the 2.33-year flood and graphs of discharge at bankfull stage leaves open the possibility that, in the downstream direction, bankfull discharge may come to assume a fixed return period in the lower parts of some drainage areas. Suggestions that the return period is about 1.1 or 1.2 years on the annual series (Wolman and Leopold, 1957, p. 88–91; Dury, 1961) should be construed as applying to single stations, or to parts of basins; but within these limits and with the usual forms of regional discharge and return periods graphs taken into account, the return period of 1.1 to 1.2 years ensures a modest value for the ratio $q_{2.33}/q_{bf}$.

The last column of table 13 lists values of $q_{2.33}/q_{bf}$. The data for q_{bf} are taken mainly from preceding tables or from regional graphs, but computed data for streams in Georgia and recorded data for the British rivers Great Ouse and Wye have been added for comparison: the Wye drains a particularly rainy basin where high runoffs are typical of small areas. Where apparent values of q_{bf}, in cubic feet per second per square mile (cfs per sq mi), increase in the downstream direction, a single regional value is used. Data for $q_{2.33}$ have been obtained by rigorous analysis (Nene and Great Ouse), adapted from preliminary State reports on floods (Green River (McCabe, 1958), New England (Bigwood and Thomas, 1955)), read from the regional graphs (Wabash and White Rivers, Northeast Ozarks, Alabama River system), or arithmetically determined from the compilations of annual peaks (Driftless Area, Green Bay area, Humboldt and Owyhee Rivers). The results from the last method are probably by far the least reliable.

The ratio $q_{2.33}/q_{bf}$ ranges so widely that comparison and generalization are alike difficult. However, there is nothing to suggest that values relying on computed values of q_{bf} are grossly in error. If any generalization can be offered, it is that $q_{2.33}/q_{bf}$ can descend from about 15:1 at 10 square miles to about 2.5:1 or 1.5:1 at 1,000 square miles. No means of causing the present 2.33-year flood to produce the effects of channel-forming discharge therefore suffice to explain the former channels or to supply the discharges computed for them, even if considerable allowance is made for the effects of change in slope and in roughness. Indeed, on the Red and the Sheyenne Rivers, where $q_{2.33}$ and q_{bf} seem to be identical, the problem of large meanders remains outstanding. However, coincidence of the two discharges appears exceptional, particularly in headward reaches: bankfull discharge usually runs below discharge at the 2.33-year flood. Much is unknown about the mechanics of discharge at bankfull stage and about the cause of its statistical relation to the 2.33-year discharge, but there seems no reason to suppose that the two discharges failed to differ from one another on many rivers when the large meanders were being formed.

The former discharges presumably represent the sum effect of a combination of causes. Since underfit streams are distributed on a continent-wide scale, they require changes in climate. The separate factors most likely to have operated in former times to promote high discharges are:

1. Reduced air temperature
2. Increased total precipitation
3. Changed regimen of precipitation
4. Increased extent of frozen ground
5. Changed regimen of runoff
6. Increased size of individual rains
7. Increased frequency of storms
8. Increased wetness of soil
9. Changed vegetation cover.

Some of these possibilities clearly overlap with, or involve, others. Changes in vegetation are implied in reductions of air temperature. Wetness of soil must be affected by changes in amount and frequency of precipitation. Changes in regimen of precipitation lead to changes in both regimen and total of runoff. In the immediately following paragraphs, changes in air temperature and in precipitation will first be considered with a view to demonstrating their possible combined effect. It will be argued that temperature reductions of entirely probable order, linked with modest and by no means improbable increases in precipitation, are capable of producing marked increases in annual runoff. Throughout a wide range of climates, the estimated increase is by a factor between 5 and 10. To some extent, the effects of the other possible kinds of change will later be presented as revealing parts of the mechanism of changes in temperature and in precipitation.

CHANGES IN AIR TEMPERATURE AND IN PRECIPITATION

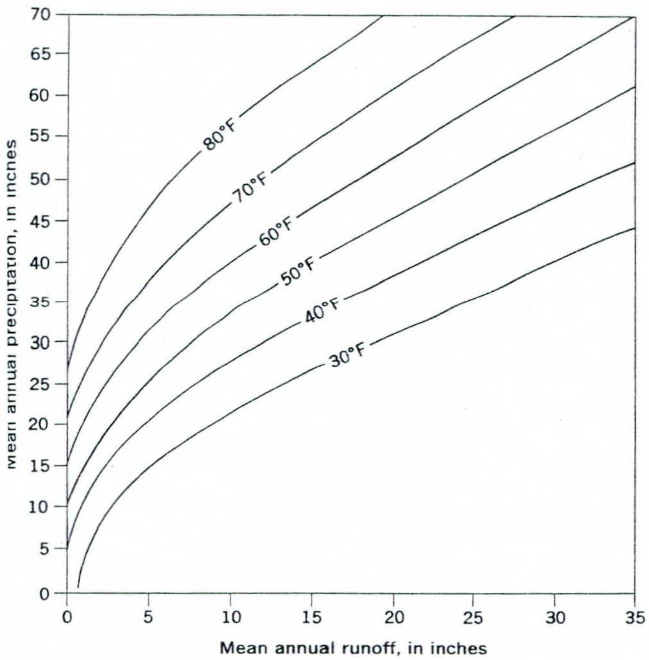
The one fundamental change in climate which is well attested by field evidence and which may reliably be taken to have affected the whole of the conterminous United States is a lowering of air temperature at times of glacial maximum. Previous discussions of underfit streams (Dury, 1954, 1958, 1960) emphasized the importance of dating the cutting of former channels and former meanders because of the relevance of dates not merely to the general inquiry but also to the climatic conditions in which those features were produced. Until very recently, evidence was insufficient to associate the highest former discharges at all firmly with glacial maximums or with early deglacial conditions. Now, however, observations converge to show that maximal former discharges are correlatable with wide former extensions of land ice or with rigorous intervals in the deglacial sequence (Dury, 1964b).

For reasons given in Dury (1964a, b), no general hypothesis can be entertained that the former high discharges were supplied throughout whole regions by streams of melt water from ice fronts or by spillwater from proglacial lakes. Snowmelt is perhaps another matter, although it can scarcely be considered for Puerto Rico, where manifestly underfit streams occur well within the Tropics. Lowering of temperature at glacial maximum is well demonstrated by paleobotanical and paleontological evidence and by the distribution of formerly frozen ground. Reconstructions depending in part on the heat balance of ice sheets permit quantities to be stated for glacial and early deglacial temperatures, so that the extent of temperature reduction below present values can be expressed numerically.

Since the immediate point at issue is the extent of maximum reduction, recapitulation of deglacial floristic sequences is unnecessary. It suffices to recall that many workers have reported great latitudinal displacements, both of ecological regions and of the range of frost action (see, for example, on vegetational change, Deevey, 1949, 1951; Firbas, 1949a, b; Godwin, 1956; Martin, 1958; Wright, 1957, 1961, and references therein; Zumberge and Potzger, 1956; on the former southward extension of frozen ground in the United States, Ashley, 1933; Black, 1954; Denny, 1951; Smith, 1949a, b, 1953; Smith and Smith, 1945; Wolfe, 1953). The reduced temperatures implied in findings of this description cannot fail to have influenced runoff. Their hydrologic effects were certainly not produced in isolation; but if temperature change is first examined for its own sake and its likely efficacy established, then the extent of any additional changes which may be required can be indicated.

Manley (1955), in a survey of the retreat of the Laurentide ice sheet, suggested that mean annual temperature at the maximum of Wisconsin glaciation was 13°F lower than that of today near the Gulf of Mexico, 16°F lower at lat 35° N., 20°F lower in the Ohio valley and 27°F lower in the vicinity of New York City. Dillon (1956) inferred reductions in mean annual temperature of 5°F at the equator, 10°F at lat 35°–40° N. and 25°F at the edge of the ice sheet. Manley and Dillon agreed on a southward temperature gradient steeper than that of the present time. The two sets of inferences accord reasonably well with one another and also with specific conclusions related to single areas. Thus, Antevs (cited by Dillon) found a decrease of 10°F in mean annual temperature on the 105th meridian in Colorado and New Mexico. Flint (1957) suggested a lowering of about 13°F in the Great Basin at the maximum of Wisconsin glaciation. Wright (1961), reviewing evidence for late Pleistocene climates in Europe, observed that frost features and fossil plants in lowland areas suggest temperatures 18°–22°F below those of today. Manley (1951) accepted reductions of 16°F for full-glacial conditions in southern England. Leopold (1951a) and Broecker and Orr (1958) treated certain pluvial lakes in terms of changes both in temperature and in precipitation: Leopold used temperature reductions in New Mexico in the range 10°–15°F, and Broecker and Orr postulated a reduction of 9°F in the Great Basin. In the two publications just cited, these reductions, in combination with increases in precipitation, were held capable of accounting for the former pluvial lakes Estancia, Lahontan, and Bonneville. Broecker, Ewing, and Heezen (1960) relied on counts of plankton and on oxygen-isotope analysis to demonstrate a rapid increase in the surface temperature of the Atlantic, close to the end of Wisconsin glacial times by some 15°–20°F. Flint and Brandtner (1961), reviewing evidence from widely separated regions, signalized changes of the same order of magnitude. They cited the findings of Andersen and others (1960), that July temperatures in Denmark underwent a net increase of some 25°F in the last 15,000 years, and those of va ler Hammen and Gonzalez (1960), that in the same ›eriod mean annual temperature at Bogotá, Colombia, ose about 14°F.

The present interrelation of temperature, precipita- ion, and runoff in the conterminous United States was stablished in generalized form by Langbein and others 1949) and presented as a graph on which figure 16 of



GURE 16.—Interrelation of precipitation, weighted mean annual temperature, and runoff. After Langbein and others (1949).

ıis report is based. This diagram can be used to show ıe likely effects on runoff of the reductions in temper- ;ure postulated for glacial maximum.

Suppose, for instance, that a certain drainage basin now ıs a mean annual temperature of 60° F and a mean annual recipitation of 25 inches, and let the reduction in mperature at glacial maximum be 20° F; then the dicated mean annual runoff for the present day is 2.5 ches, whereas for glacial maximum, with a mean ınual temperature of 40° F, it is 7.9 inches. Without ıange in precipitation, annual runoff has increased ore than three times.

The mean annual temperatures used in figure 16 are ›t, however, mean temperatures in the ordinary sense. hey are weighted to allow for seasonal distribution of ·ecipitation, in the form

$$\overline{T}_w=\sum\frac{t_m p_m}{\overline{P}} \qquad (33)$$

here $\overline{T}_w$ is weighted mean annual temperature, t_m ıd p_m are mean temperature and mean precipitation for single months throughout the year, and $\overline{P}$ is mean annual precipitation.

If precipitation is concentrated in the warmer months, the weighted mean annual temperature is above the ordinary mean. Winter concentration of precipitation depresses the weighted mean below the ordinary mean. In 21 pairs of readings, the departure of weighted from ordinary means, regardless of sign, averages about 5°F (Langbein and others, 1949, table 4).

If weighted means are used for annual temperature, precipitation, as well as temperature, becomes involved in the argument. The possibilities to be considered, in addition to a reduction in ordinary temperatures, then become:

1. No change either in total or in regimen of precipitation
2. Change in regimen of precipitation, but no change in total precipitation
3. Change both in total and in regimen of precipitation

Since determinations of weighted mean annual temperature require monthly means of ordinary temperature, these must be reconstructed for times of glacial maximum. Reconstructions can be made graphically, if ordinary temperatures are available for warmest and coldest months, by comparison with graphs of present temperatures (fig. 17). But, unless the reduction for the warmest month differs markedly from that for the coldest, a uniform change can be applied throughout the year.

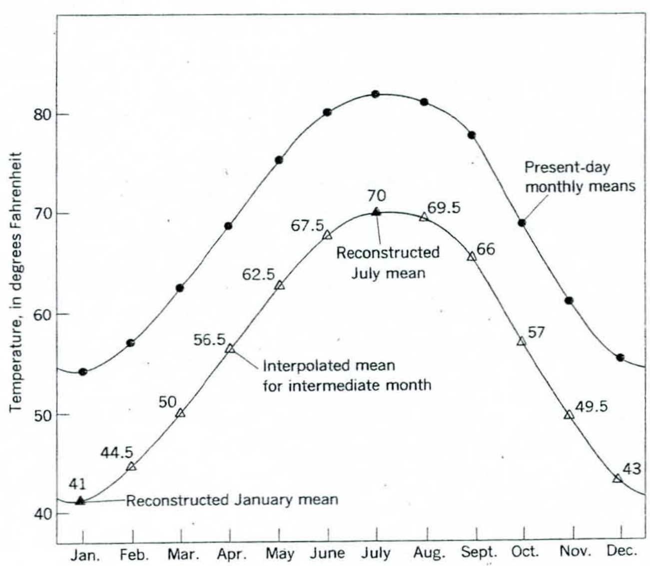


FIGURE 17.—Reconstruction of regimen of mean monthly temperature at the site of New Orleans during full-glacial conditions.

REDUCED TEMPERATURES, BUT NO CHANGE IN PRECIPITATION

Ordinary and weighted mean annual temperatures for New York City are 51.8° and 52.8°F, respectively. With ordinary mean temperatures reduced by 27°F, in accordance with Manley's findings, these become 24.8° and 26°F. The increase in runoff indicated by figure 16 is from 15 to 37 inches—an increase by a factor of about 2.5. For New Orleans, with the more modest reduction of 13°F in ordinary mean temperature, runoff should increase from 16.5 to 27 inches—that is, by a factor of about 1.6.

By a simple process of transformation, the graphs in figure 16 can be used to generalize the ratio F_q, of former to present annual runoff, for selected reductions in mean annual temperature. The generalized values constitute the bases of figures 18, 19, and 20, where the effects on mean annual runoff of temperature reductions of 10°, 15°, and 20°F are shown by nomograms.

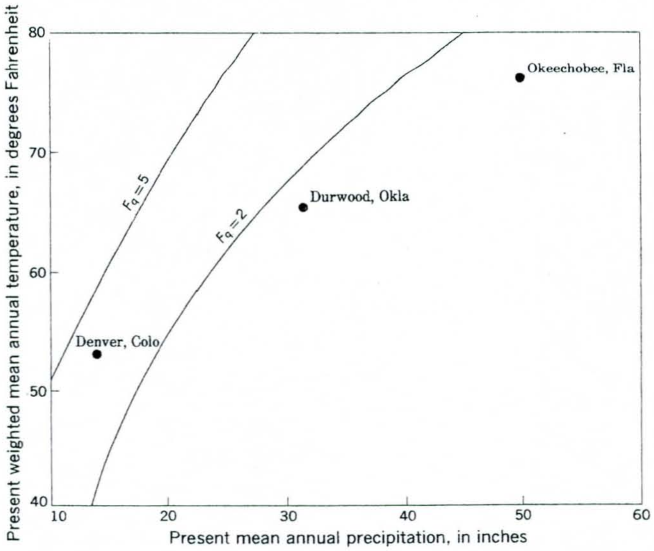


FIGURE 18.—Nomogram for results of a reduction of 10°F in weighted mean annual temperature with precipitation unchanged.

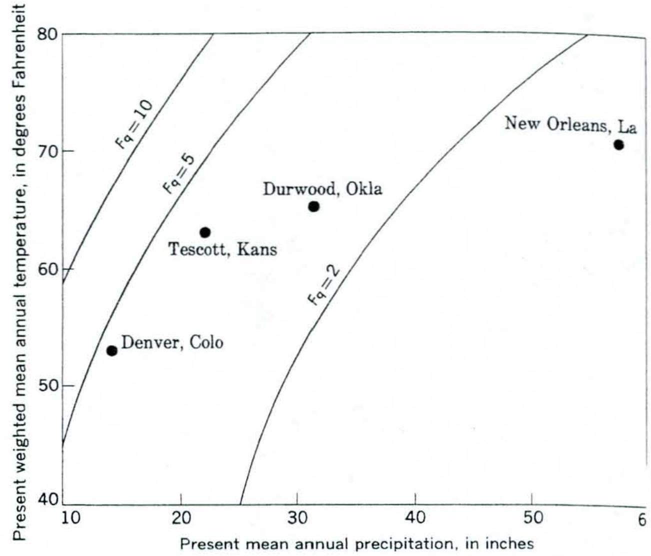


FIGURE 19.—Nomogram for results of a reduction of 15°F in weighted mea annual temperature with precipitation unchanged.

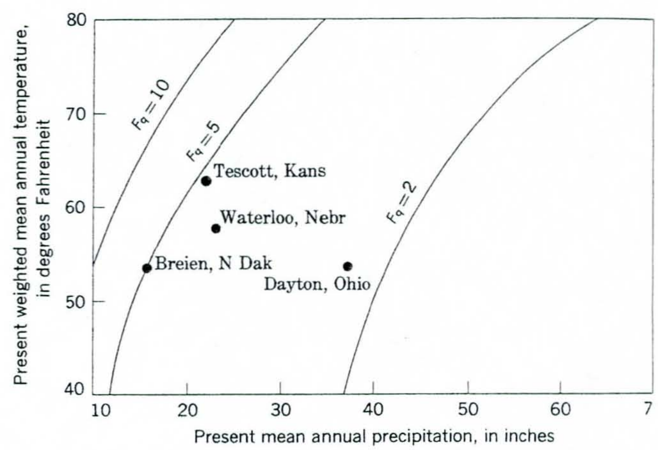


FIGURE 20.—Nomogram for results of a reduction of 20°F in weighted mea annual temperature with precipitation unchanged.

For obvious reasons, the proportional effect of a fixed reduction in temperature increases directly with the mean temperature assumed at the outset and inversely with precipitation. Points for selected stations have been inserted in the nomograms for reference and comparison. As shown, a reduction in temperature of 20°F could increase total runoff in much of the Great Plains by a factor of 5 or even 10 without the aid of increased precipitation.

CHANGE IN REGIMEN OF PRECIPITATION WITHOUT CHANGE IN TOTAL

Although a seasonal concentration of precipitation causes weighted mean temperature to differ from ordinary mean temperature, the effects of the differen upon indicated runoff are mainly small in comparis with the effects of reductions in ordinary temperatu by the range 13°–27°F. The extent to which change in seasonal regimen could affect the ratio F between former and present total runoff, is illustrat by the hypothetical, and quite extreme, exampl in tables 5 and 6.

The highest notional values of F_q are expectabl those associated with marked winter concentrations precipitation at former times of low temperatu Reconstructed patterns of weather for times of glaci maximum, however, usually include summer co centrations of precipitation (Manley, 1955; Dill 1956). Regions close to the former ice fronts may w have been overspread by the cold air of high-pressu

TABLE 5.—*Influence of precipitation regimen on weighted mean annual temperature*

[Hypothetical monthly data throughout]

	Mean temperature, in °F	Precipitation, in inches			Weighted mean temperature, in °F, for indicated precipitation			Mean temperature, in °F, reduced by 20°F	Reduced weighted mean temperature, in °F, for indicated precipitation		
		Evenly distributed	Summer concentrated	Winter concentrated	Evenly distributed (cols. 1×2)	Summer concentrated (cols. 1×3)	Winter concentrated (cols. 1×4)		Evenly distributed (cols. 8×2)	Summer concentrated (cols. 8×3)	Winter concentrated (cols. 8×4)
	1	2	3	4	5	6	7	8	9	10	11
uary	40	2	1	3	80	40	120	20	40	20	60
ruary	45	2	1	3	90	45	135	25	50	25	75
rch	50	2	1.5	2.5	100	75	125	30	60	45	75
ril	55	2	2	2	110	110	110	35	70	70	70
y	60	2	2.5	1.5	120	150	90	40	80	100	60
e	65	2	3	1	130	195	65	45	90	135	45
y	70	2	3	1	140	210	70	50	100	150	50
gust	65	2	3	1	130	195	65	45	90	135	45
tember	60	2	2.5	1.5	120	150	90	40	80	100	60
ober	55	2	2	2	110	110	110	35	70	70	70
vember	50	2	1.5	2.5	100	75	125	30	60	45	75
cember	45	2	1	3	90	45	135	25	50	25	75
Year	55	24	24	24	55	59	52	35	35	39	32

stems during much of the winter. (See also Wright, 961, and references therein.) Consequently, the gh values of F_q derived from hypothetical former inter concentration of precipitation can probably disregarded. This is not to deny that contrasts ay occur between region and region, either in the pe of former precipitation regimen or in the mode of lation of former to present regimen. Generally wever, those moderate values of F_q associated with mmer concentration of precipitation at times of acial maximum are to be preferred.

TABLE 6.—*Influence of temperature reduction on mean annual runoff*

Total precipitation (from table 5) 24 inches

Weighted mean annual temperature, in °F (from table 5)	Mean annual runoff, in inches (from fig. 16)
55	3.15
59	2.5
52	3.75
35	9.5
39	7.9
32	11.25

Seasonal regimen of precipitation		F_q, approximately, for reduction of 20°F in ordinary temperature
Present	Former	
Evenly distributed	Evenly distributed	3.0
Do	Summer concentrated	2.5
Do	Winter concentrated	3.6
Summer concentrated	Evenly distributed	3.8
Do	Summer concentrated	3.2
Do	Winter concentrated	4.5
Winter concentrated	Evenly distributed	2.5
Do	Summer concentrated	2.1
Do	Winter concentrated	3.0

CHANGES BOTH IN TOTAL AND IN REGIMEN OF PRECIPITATION

Although the several changes now under discuson were stipulated to involve reductions in temperaire, the effects of increased precipitation will be onsidered, initially, as if temperature remained untered.

A special case is the increase of precipitation by a xed proportion throughout the year. Such an inease would leave weighted mean annual temperature nchanged, but it would of course affect total runoff. y a process of transformation similar to that used r temperature, nomograms have been contructed show the generalized effects of uniform proporonal increases in precipitation (figs. 21, 22). The ctors of increase employed—namely, 1.5 and 2.0— re not wholly arbitrary.

Some increase in precipitation is required by the ypothesis of increased cyclonic strength at glacial aximum (Willett, 1950; Dury, 1954). Dillon (1956) egarded increased cyclonic activity as having coniderably enhanced the moistness of southeastern Jnited States and as having greatly enhanced that of the Southwest; he also postulated increased precipitation in the North as necessary to the growth of ice. Manley (1955), dealing with the Laurentide ice sheet east of long 90° W., called for heavy precipitation, mainly in late summer, from widespread and persistent clouds near the ice margins; he also pointed out, on the basis of calculable effects of ablation, that temperature reduction alone is insufficient to have maintained the ice at its known limits.

An increase in total precipitation seems to be indicated, again, by the large former meanders of Black

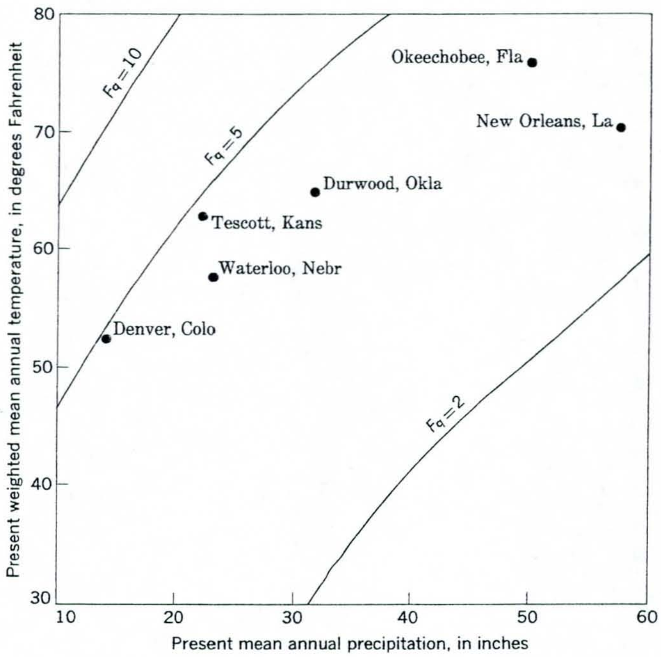


FIGURE 21.—Nomogram for results of an increase of 50 percent in precipitation with temperature unchanged.

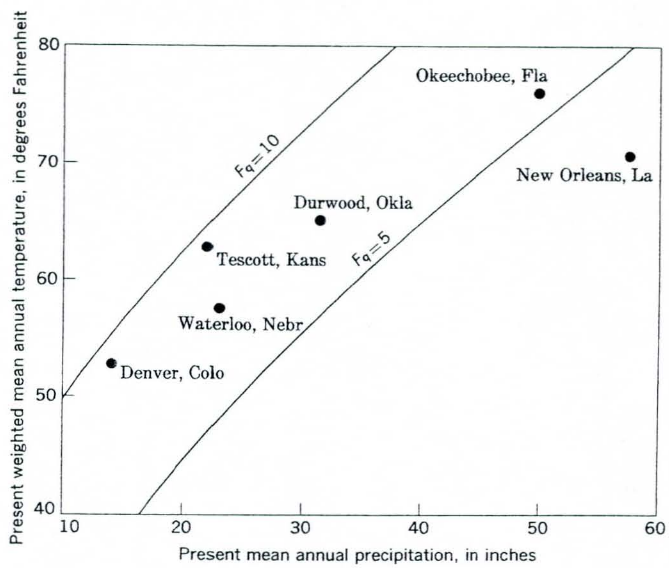


FIGURE 22.—Nomogram for results of an increase of 100 percent in precipitation with temperature unchanged.

Earth Creek, Wis. At the present day, this river discharges about one-third of the precipitation falling on its basin (Cline, 1960). Although matters are complicated by highly permeable outwash in the valley bottom, there is in fact very little net interchange of water between the channel and the ground-water table. Thus, whatever effects are supposed to result—for example, from severe frost (discussed below)—the total possible increase in annual discharge, without increas in precipitation, is by a factor of three. But Blacl Earth Creek, with ratios L/l and W/w in the neighbor hood of 10:1, ranks high on the scale of underfitness The ratio Q/q, between former and present bankful discharges, seems unlikely to be less than 50:1; evei allowing a threefold increase in total runoff, a furthe sixteenfold increase in bankfull discharge has still t be accounted for. This increase seems difficult t explain without appeal to increased precipitation Similar arguments apply to highly underfit stream which drain basins cut in rocks that are largely im permeable.

Leopold (1951a) admitted an increase of precipita tion in New Mexico by a factor of about 1.5. Broecke and Orr (1958) suggested a factor of about 2.0, but the associated this figure with a quite modest reduction o 9° F in temperature for the Great Basin. A greate reduction in temperature, such as Manley's result imply, would reduce the precipitation factor. A Flint (1957) observed, biogeographic evidence fror east Africa requires an increase in precipitation, fo pluvial phases, of about 50 percent over the presen figures. In certain dry regions, that is to say, and a times of glacial maximum, precipitation seems to hav been 50 percent greater than it is today.

These regions, with their former pluvial lakes an their climate-sensitive ecology, are well suited to provid numerical estimates of precipitation change. Humi regions offer no comparable data, although they provid signs of humidity greater than present humidity. Ruh and Scholtes (1956) inferred such humidity from mas sive late Wisconsin gley horizons in Iowa. (See als Ruhe, Rubin, and Scholtes, 1957; Lane, 1931.) Al though gleying is equivocal, as pluvial lakes are equivo cal, in that its requirements of humidity can be me partly by reduced temperatures, the dating of the gle soils of Iowa to the Cary-Mankato stades of the Wis consin Glaciation is instructive. The soils forme shortly before the Two Creeks Interstade. It was pre cisely at this time, with deglacial temperatures alread rising, that Lake Lahontan rose to levels well abov those recorded at glacial maximum at about 20,00 years B.P. The very high stand, therefore, is referabl to increased precipitation rather than to reduced ten peratures, as observed in Professional Paper 452–I Despite the cold fluctuation of Zone Ic (see table 15 the gleying in Iowa may also reasonably be ascribe in part, to increased precipitation. In addition, a ver general argument is possible, to the effect that climati rasts between dry and humid regions may have ı at least as marked at times of glacial maximum in early deglacial times as they now are. If so, ıcrease of 100 percent in the precipitation of humid ›ns is by no means incompatible with an increase of ercent in dry regions.

onditions during the Atlantic phase (Zone VII, sithermal maximum) of deglacial time (see table 15) ;est that increases in the precipitation for humid ›ns by about 50 percent are not unlikely. The ial reexcavation of large channels at this time and paleontological evidence of increased soil moisture t be reconciled with mean temperatures as much as ' higher than those of today. Transformations of re 16 can readily be made to show that, merely to ntain present levels of mean annual runoff against ı a temperature increase, precipitation would need ıcrease 10–15 percent. Since little is known of the ensions of channels cut during Zone VII, bankfull harges cannot be computed for them. Something, ever, can be made of wavelengths of former rivers he English fenland (Dury, 1964b). These wave-;ths, in comparison with those of existing rivers and alley meanders on upstream reaches, suggest that kfull discharge in Zone VII times was locally ⅙ or ½ ;reat as that responsible for the main series of valley ınders and about 20 times as great as that of the sent time. The indicated increase in precipitation ıerefore much greater than the 10–15 percent capable giving present-day values of runoff. The probably dest assumption that mean annual runoff during ıe VII was double that of today entails increases in an annual precipitation by 33–50 percent—that is, reases of the order postulated for high-glacial times.

ls with changes in temperature, so with changes in cipitation: the indicated proportional effects in-ase directly with temperature and inversely with al precipitation. In part of the range of possible nates, the computed results of a doubling of precipi-ion are at least as great as those of a temperature re-ction of 20°F. (See figs. 18–22.)

Marked variations in seasonal concentration of cipitation have little effect on the values of F_q for assumed doubling of precipitation. For the tem-ratures and precipitation listed in table 5, doubling cipitation would increase the values about six times ble 7). A change from the summer concentration, own in table 5, to the winter concentration, shown in e same table, but with monthly totals doubled in dition, would raise F_q only to about 8.5. The objec-›ns to a former winter concentration, which have been ıted above, pertain here also. In any event, the influence of the change in annual total is far more significant than any reasonable change in annual regimen.

TABLE 7.—*Influence of double the amount of precipitation on runoff*

[Precipitation values listed in table 5]

Total precipitation .... 48 inches
Ordinary mean temperature .... 55° F (unchanged)

Seasonal regimen of precipitation	Weighted mean annual temperature, in °F	Runoff from 48 inches of precipitation	Ratio to corresponding runoff from 24 inches of precipitation
Evenly distributed	55	19	6.0
Summer concentrated	60.5	15.25	6.1
Winter concentated	52	21.25	5.7

CONCURRENT REDUCTION OF TEMPERATURE AND INCREASE OF PRECIPITATION

When figure 16 is used to supply F_q for reduced temperatures and for increased precipitation at the same time, some very high values begin to appear. In figures 23–26, F_q is generalized for reductions of 10° and 20°F and for increases of 50 and 100 percent in precipitation. Toward the extreme of heat and aridity, increases in total runoff by a factor of 50 become possible, mainly because present absolute runoff

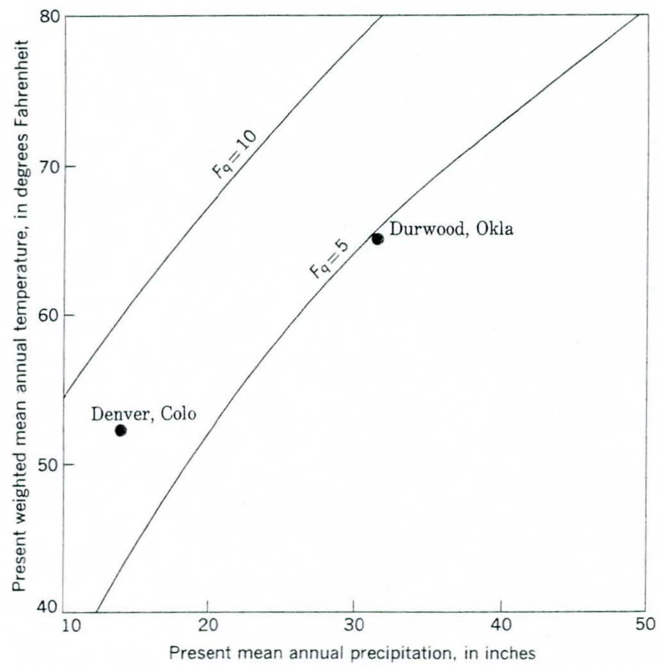


FIGURE 23.—Nomogram for results of an increase of 50 percent in precipitation plus a reduction of 10°F in weighted mean annual temperature.

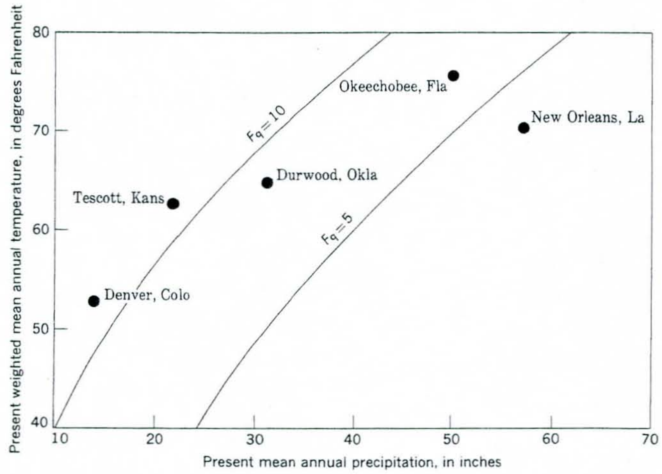


FIGURE 24.—Nomogram for results of an increase of 100 percent in precipitation plus a reduction of 10°F in weighted mean annual temperature.

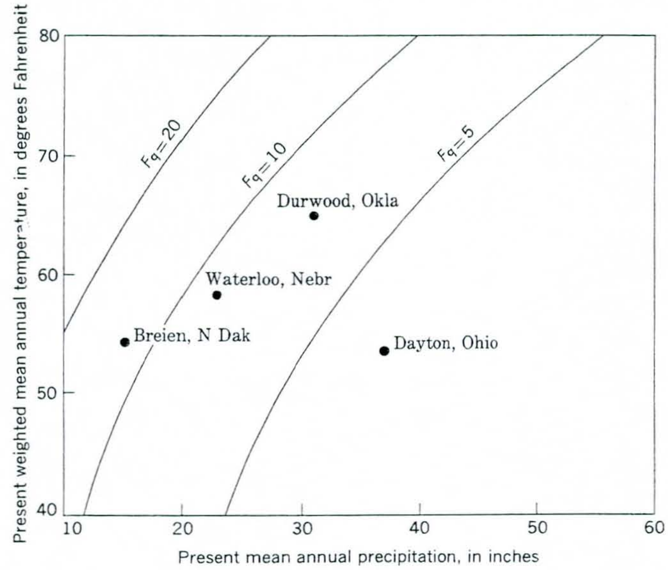


FIGURE 25.—Nomogram for results of an increase of 50 percent in precipitation plus a reduction of 20°F in weighted mean annual temperature.

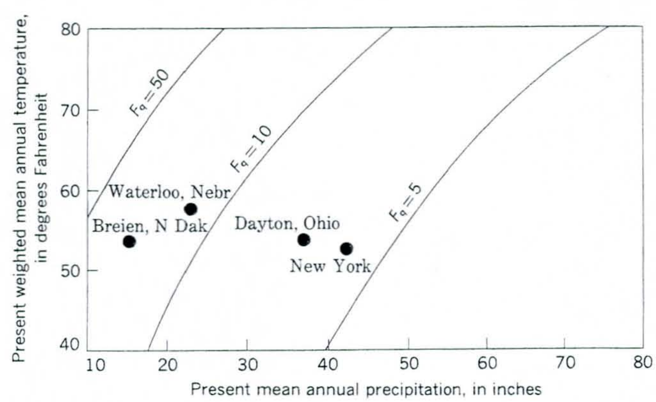


FIGURE 26.—Nomogram for results of an increase of 100 percent in precipitation plus a reduction of 20°F in weighted mean annual temperature.

is very low. Indeed, the logical implication of graphs used in constructing the nomograms is F_q can be infinite in regions which today are too for runoff to occur but which formerly suppo surface drainage.

Although the nomograms in general agree with precipitation-temperature-runoff relations of figur and thus accord with empirical findings, certain of t implications are not especially welcome. In the hu eastern half of the United States, a doubling of pre tation and a reduction in temperature by 13° F the Gulf of Mexico and by twice that amount in North would increase annual runoff some 5–10 ti the respective values of F_q for New Orleans and York City, determined from the data given above, 4.8 and 5.5. But, whether F_q is determined for cha in temperature, changes in precipitation, or cha in both combined, it tends to rise westward across interior and to ascend sharply across the Great Pl No comparably systematic increase in the ratio as determined from L/l, has been detected. Admit ly, too little is known for systematic cross-cou changes in L/l to be specified, even if they oc On the other hand, L/l in the semiarid country dra by the upper members of the Humboldt and Ow systems is rather less than in the distinctly hu Driftless Area of Wisconsin and in southern Engl The ratio L/l on the Mission River of Texas and ratio Q/q computed from it fall well below expectat based on figures 16 and 23.

Part of the immediate difficulty is due to the fo of the individual graphs in figure 16, which show runoff for certain high temperatures and low value precipitation. The general map, which is in La bein and others (1949, pl. 1), allows little scope no mean annual runoff. Therefore, since the acc panying nomograms are based on figure 16, albeit w some adjustments in the range of very high temp tures and very low precipitation, the very highest va of F_q which they indicate should be taken with rese

For all that, any hypothesis that required the eff of greatly reduced temperatures in dry regions to offset by compensatory changes in precipitation wo be of questionable merit. Far more needs to be kn of the regional variation in the ratio L/l before matter can be further examined. Particularly is so because of the apparently irregular occurrence high values of L/l in humid regions. Dry regions r present distinctive problems of their own, incluc reductions in stream density such as can scarcely h influenced regions which were, and remain, hun Pending additional work, the fivefold to tenfold crease in mean annual runoff in humid regions will b be used as the starting point of further inquiry.

CHANGES OTHER THAN CHANGES IN ANNUAL TEMPERATURE AND IN ANNUAL PRECIPITATION

The empirically obtained graphs of figure 16 subsume the influences on annual runoff of geology, relief, slope, soil, vegetation, climate, and weather. Geology, relief, and slope may be regarded as constants for given areas, which combine to cause departures in either direction from the runoffs generalized in figure 16. Variations in regimen of precipitation, causing weighted mean annual temperature to depart from ordinary mean, are averaged by the graphs of figure 16 and have, moreover, been found small in their effects by comparison with changes in ordinary mean temperature or in total precipitation. Nevertheless, changes in seasonality of precipitation are certain to have affected momentary or other short-term discharges more strongly than they affected annual runoff. Some reasonable combination of changes in climate, weather, soil conditions, and vegetation seems quite capable of producing a twentyfold increase in bankfull discharge, in the context of a fivefold or tenfold increase in the annual total. Certain leading possibilities are explored in the following sections.

FROZEN GROUND

In a discussion of the possible effects of frost in promoting high discharges, the point at issue is not the reduced evapotranspiration associated with temperatures low enough to involve significant increases in the duration and severity of frost, but the action of frost itself in sealing permeable rock and in changing the hydrologic characteristics of the topsoil.

It would be desirable, were it possible, to separate the effects of permafrost from those of frost that was merely seasonal. Some of the evidence for past frost action is, however, ambiguous in this connection (Wright, 1961; Black, 1954). Nevertheless, something can be accomplished by reference to areas which were certainly, or at least very probably, subjected to permafrost in former times and to other contrasting areas where permafrost cannot establish itself. Hypothetical translocation of present-day frosty climates make suggestions possible about the effect of seasonal frost. The following discussion produces doubt of the relevance of frost to the bankfull discharges formerly delivered by streams that are now underfit.

The underfit streams first studied in detail lie in southern England, where cryoturbation was formerly severe. Among the Mesozoic rocks, the Chalk is deeply shattered, and Jurassic outcrops possess superficial structures over a wide area (Hollingworth, Taylor, and Kellaway, 1944; Kellaway and Taylor, 1952). Each set of effects is attributed to freezing of the ground in depth. Associated features include masses of head (solifluction earth), fossil cracks of vanished ice wedges, patterned ground, involutions, and patches of loess. While not all of these associated features require more than severe seasonal frost, the depth of frost penetration—measurable in tens of feet—surely indicates permafrost. Let it then be conceded that, at times of maximum cold, the permeable rocks of the English Plain were sealed by ice and that they remained sealed as late as 10,000 years B.P. Sealing would still not account for the regional development of underfit streams, since numbers of them drain rocks of very low permeability. Frost sealing is not wholly adequate even to explain systems of dry valleys.

The Warwickshire Itchen drains a basin floored by rocks that are scarcely permeable. The outcrop is notorious for its poor yields of well water and for low or no base flows in dry seasons. But the Itchen, reduced to about one-tenth of its former width at bankfull stage, is among the most highly underfit of streams unaffected either by derangement or by discharge of melt water. Additional basins where all outcrops are impermeable are few. However, the ground-water table in many parts of the English Plain is restricted in total area, fragmented in distribution, and perched high above the impermeable clays which line the valleys. Such is the case with the upper Cherwell, which, opposing the Itchen across a common divide, is equally underfit.

Even where a given basin includes some outcrops of permeable rock, these may have little influence in those conditions which result in bankfull discharge. Ground-water tables serve to maintain base flow rather than to contribute largely to discharge at high stages. And unless highly permeable bedrock lies bare or is mantled by highly permeable topsoil, it cannot instantly absorb intense precipitation or, for that matter, snowmelt. The necessary conditions are too restricted in distribution to bear on the general problem of underfit streams.

Black Earth Creek, Wis., is once again relevant. Rising on the line of the former ice front, this stream, whose valley meanders were abandoned as early as about 12,000 years B.P., may reasonably be taken as underlain by permafrost during early deglaciation. But, as noted above, its present total runoff is equivalent to one-third of precipitation; since little net exchange now occurs between channel and ground water, perennial sealing of the underlying outwash by frost would be unlikely to increase total runoff by more than three times, even in the most favorable conditions. Since bankfull discharges are required to increase by a factor of about 50, frost seems unlikely to be a prime cause of increase, whatever its contributory effects. If frost has little significance in the context of this highly underfit stream, it need scarcely be considered in regions where underfitness is less marked and where ice fronts were formerly remote.

The speculation that permafrost may be typical of waxing rather than of waning glaciation (Wright, 1961) will not be pursued. A region where the effects of permafrost can be ruled out is the gulf coast of the United States, for which Manley (1955) reconstructed mean temperatures for Wisconsin glacial maximum of slightly above 32°F in January and of about 80°F in July. Even if the resultant annual mean of about 50°F is considerably too high, no possible adjustment will bring it down to the level required for permafrost. As Black (1954) observed, permafrost results when the net heat balance of the surface of the earth, over a period of years, produces temperatures continuously below 32°F. Such a value is far too low for the gulf coast. Indeed, Frye and Willman (1958) concluded that, in parts of the Central and Eastern United States, permafrost extended little if at all beyond the limits of the glaciers. Thus, although the lines of ice fronts cannot be used in reconstructing the former distribution of permafrost (Flint, 1957), the gulf coastland may safely be assumed to have lain well outside the farthest possible limit of perennially frozen ground. Nonetheless, meandering valleys are well in evidence there.

If it is supposed that the discharges which cut these valleys were promoted by seasonal frost, it must also be supposed that those discharges occurred during the cold season and that they ceased to occur as the glaciers receded and as frost in the south became less frequent and less severe. In the present state of knowledge, the question of seasonality of former peaks may be suspected to raise difficulties. The evidence from that part of the coast where the Mission River reaches the sea (Dury, 1964b) is not opposed to the view that here, as in regions further north, the former high discharges ceased in early deglacial times—precisely when a climatic jump is likely to have reduced the incidence of cold waves and consequently, of frost. But the coincidence is one of time and not of cause, as will next be shown.

Although former patterns of weather are even more difficult to reconstruct than are those of climate, it is possible to translocate to the gulf coast a present-day climate likely to be comparable to the coastal climate at glacial maximum. Mean January temperatures in the Driftless Area of Wisconsin and near the head of Green Bay are about 20°F; those of July are about 68°F. Mean annual rainfall in these two areas is 30 inches or somewhat more. Since present mean annual rainfall in the Mission River basin is about 20 inches, a transfer of the Wisconsin climate to the Texas coast would involve an increase of 50 percent in precipitatio in addition to a reduction in temperature greater tha that inferred by Manley for glacial maximum. Th extra reduction in temperature may be accepted, how ever, in compensation for the possible extra severity southward bursts of cold air at glacial maximum, whe weather gradients were unusually steep. If, accord ingly, the Mission River basin was formerly penetrate by frost to the depth of 40 inches now recorded i Wisconsin and if frost was responsible for promotin the discharges which cut the valley meanders of tl Mission River, then these meanders should be simil in wavelength, area for area, to the stream meanders Wisconsin.

Although the valley meanders of the Mission Riv are smaller, area for area, than those either of tl Driftless Area or of the Green Bay country, they a distinctly larger than the present meanders on tl rivers of Wisconsin (fig. 27). At 10 square miles tl

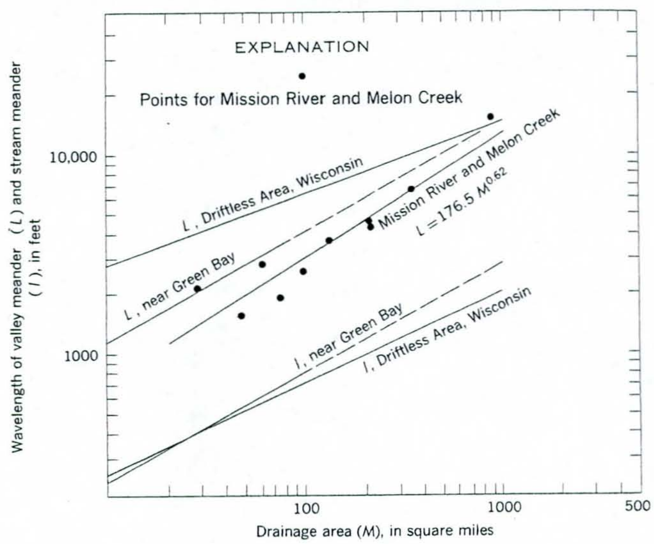


FIGURE 27.—Relation of wavelength of valley meanders to drainage area of Missio River, Tex., in comparison with other drainage areas.

valley meanders of the Mission River are three tim as long as the stream meanders of the Driftless Are and at 1,000 square miles they are six times as lon Whereas the valley meanders of the Driftless Area a nearly four times as long as those of the Mission Riv at 10 square miles, those of the Mission River becor the longer, by extrapolation, at 1,650 square mil Extrapolation of the graphs for the Green Bay count indicates that, at 1,000 square miles, the valley meanders the Mission River are 4.5 times as long as the stream mea ders near Green Bay, whereas at 1,650 square mil the valley meanders of the Green Bay country are b 15 percent longer than those of the Mission River. T valley meanders of the Mission River obviously belo to the group which includes valley meanders of oth regions. They are far too large to be grouped with the stream meanders now developing in a climate as rainy as, and more frosty than, that likely to have affected the gulf coastland at glacial maximum. Once again, frost appears irrelevant to the basic problem of underfit streams.

The Interior Plateau country of Kentucky shows that, in any event, frost was not invariably capable of sealing bedrock while valley meanders were being cut. In the usual manner, the distribution of manifestly underfit streams in the plateau is limited or obscured by the drowning of reservoirs, by the wide opening of some valleys, or by the narrowness of other valleys at the base. Rolling Fork (Howardston quadrangle, Kentucky, 1:24,000) is manifestly underfit where it trenches resistant members of the Mississippian and Silurian Systems (fig. 28), but it loses the definition of its valley meanders where it enters the shaly Devonian rocks farther downstream (Jillson, 1929; McFarlan, 1943). On the Green River system, bold meander trains are uncommon on the present channels (compare fig. 29), but stream meanders are nevertheless sufficiently numerous to permit a comparative plot of wavelength between stream and valley (fig. 30). The indicated wavelength ratio is almost exactly 5:1 up to 500 square miles; this ratio is similar to those obtained in several other regions and to the 4.6:1 for Rolling Fork. Above the 500-square-mile mark, the rate of increase in wavelength of valley meanders, with increasing size of drainage area, lessens rather suddenly.

Since the trains of valley meanders are deeply incised, they are necessarily inherited from levels higher than the present valley floors. The inflection in the wavelength–drainage-area graph therefore results from conditions of a former time. About at the point where the rate of increase in wavelength becomes less, the Green River crosses the boundary between the Osage and Meramec Series of the Mississippian and passes on downstream to rocks which, in bulk, are distinctly permeable. Although the crossing of the boundary seems not to be associated with a reduction of discharge per square mile at the present day, either at the 2.33-year flood or at mean discharge (fig. 30; U.S. Geol. Survey, 1957a; McCabe, 1958), this circumstance probably implies nothing more than a ground-water table not lower than the streambed. In places, indeed, the ground-water table is higher than the streambed, as in the area of the Horse Cave quadrangle (Kentucky, 1:24,000); there river level at bankfull stage ranges from about 470 feet above sea level, upstream, to about 460 feet above sea level, downstream, whereas the floors of nearby sinks are nearly always flooded to at least 550 feet above sea level. Actually, the ancestral river could have reached limestone at heights of 800 feet or more above sea level—and may well have been suffering considerable losses to percolation at the bankfull stage—when the wavelength of its valley meanders was fixed by incision. If this view is correct, then the Green River demonstrates one possible effect of percolation: a reduction in wavelength, not of stream meanders but of valley meanders, independent of the reduction which makes streams underfit on the regional scale. If frost action in the past has affected the discharge of the Green River system, it has certainly not contributed to the present state of underfitness downstream from the 500-square-mile mark.

References to Puerto Rico and Alaska further advance discussion of the possible effects of frost. Any attempt to make frost generally responsible for the former high channel-forming discharges of underfit streams must surely fail because such streams are widely observed in Puerto Rico. Although the traces of present streams may well have been generalized on the topographic maps (U.S. Geol. Survey, 1:30,000) and although some trains of valley meanders are so deep and narrow that they leave no room for stream meanders, bends of the two orders are combined often enough to show that the rivers of Puerto Rico are typically underfit (figs. 31, 32). At lat 18°–18°30′ N. and in a very maritime situation, Puerto Rico is well outside the main limits of frost at the present day. It is most unlikely that, at glacial maximum, its temperatures were lower than the present temperatures 10° of latitude to the north. A possible comparison is with central Florida, where present-day runoff reaches a maximum not in the coldest season but in September–October (Harbeck and Langbein, 1949).

Scanty observations (table 8) give a value of 3.7 for the ratio of wavelength between valley meanders and stream meanders in Puerto Rico, and thus suggest that the discharges required to cut the large bends were about 10 times those which shape the present channels. The 3.7:1 ratio of wavelength agrees precisely with that cited in Professional Paper 452–B for the Orontes River in Syria, whereas the modest suggested discharge ratio of 10:1 indicates for the southerly latitude of

TABLE 8.—*Concurrent values of meander wavelengths for Puerto Rico*

River	Valley meanders		Stream meanders	
	Number of wavelengths	Average wavelength, in feet	Number of wavelengths	Average wavelength, in feet
Rio Grande de Arecibo	4	8,125	4	2,500
Do	3	3,975	3	875
Rio Grande de Manati	3	4,335	5	1,230
Do	3	4,000	5	1,115
Rio Camuy	3	2,100	5	490
Do	3	2,100	5	680
Rio Gurabo	5	2,640	6	750

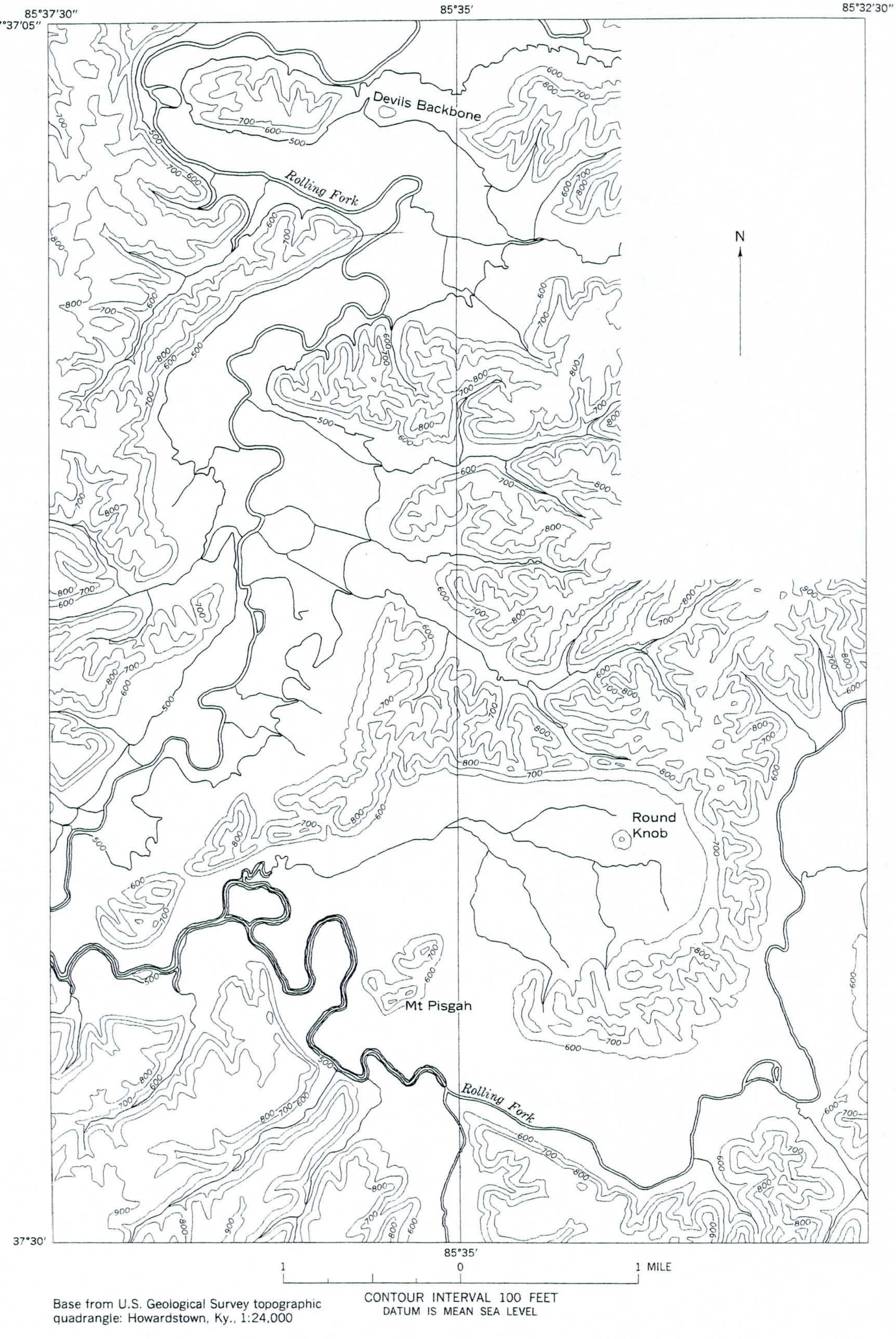
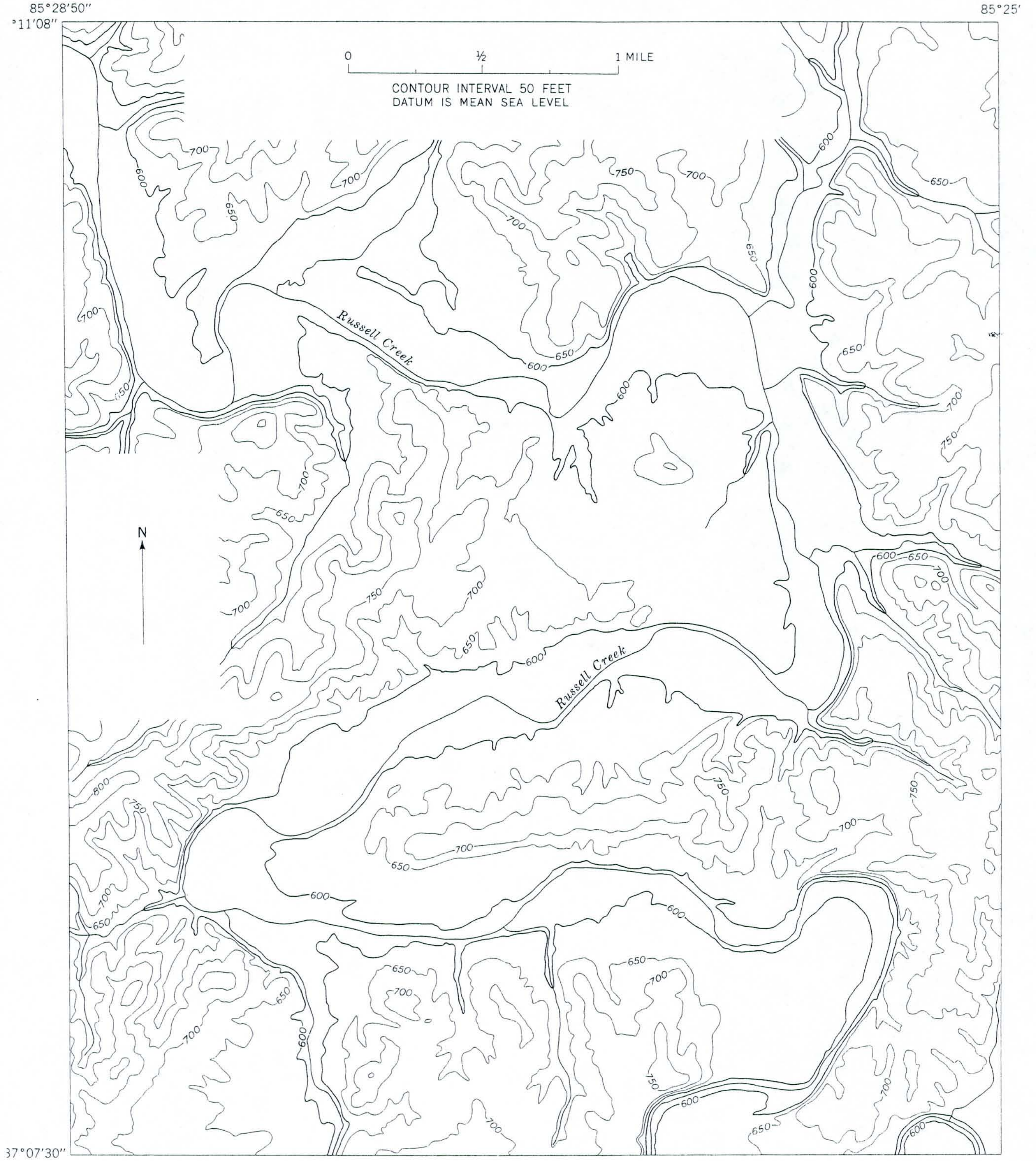


FIGURE 28.—Rolling Fork, Ky., an example of a manifestly underfit stream.



Base from U.S. Geological Survey topographic quadrangle: Gresham, Ky., 1:24,000

FIGURE 29.—Russell Creek, Ky., showing poor development of stream meanders.

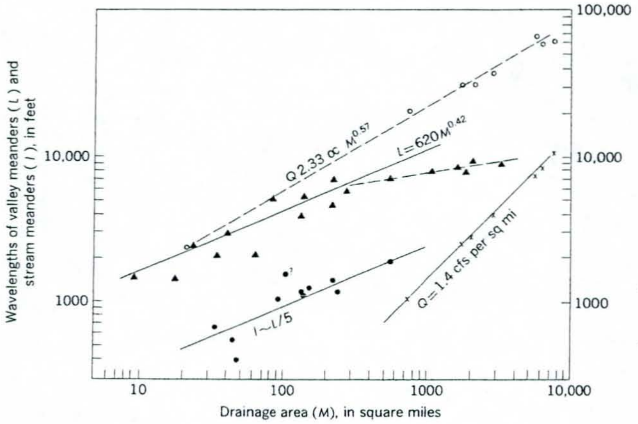


FIGURE 30.—Relation of wavelengths of stream and valley meanders to drainage area and relation of former discharge to drainage area, Green River system, Kentucky.

Puerto Rico a change far slighter than in, say, the Driftless Area or even in the mid-West generally. Both the reduced extent of change and the inference that a change did occur in Puerto Rico accord with hypotheses of displacement of storm belts and pressure cells. Furthermore, they agree with findings that variations of ocean-surface temperature during the late Pleistocene were less marked in low latitudes than in high (Emiliani, 1955; Ericson and Wollin, 1956).

Although streamflow records from Alaska are understandably few in relation to the size of the State, they can nevertheless supply comparisons with possible conditions in more southerly regions during glacial maximum.

Regional contrasts of stream regimen within Alaska separate the southeast coastland from the interior and separate the central-interior districts from the coastland of Seward Peninsula (U.S. Geol. Survey, 1957b, p. 10). In southeast Alaska, where annual precipitation ranges from 20 inches near the summits to 150 inches or more near the shore (U.S. Weather Bureau, 1943), runo copious and base flow is high. Many stations re two maximums of runoff: one promoted by rainfal October and November and a second in May and J that in part reflects melting. At glacial maxim the climatic and hydrologic conditions of this area presumably translocated to southern British Colun and the Pacific Northwest; but it may not be unrea able to suggest a comparison with New England du the early waning of glaciation, since New England, the Alaskan Panhandle, is strongly influenced cyclones and was probably so influenced when receding ice front still lay close to its shore.

Fish Creek, near Ketchikan, in the very south of Alaskan Panhandle, discharges the equivalent of s 150–200 inches of precipitation a year. In the 24 y during the period 1916–45 for which peak records available, peak momentary annual discharge aver about 90 cfs per sq mi (table 9). At this rat discharge—that is, at the rate of the 2.33-year floo 200 inches of precipitation could be run off in 60 d But high base flow ensures a mean of 7.8 cfs per s for March, the month of lowest average runoff, and mean for October, the month of highest runof correspondingly brought down to 21.5 cfs per sq the highest mean monthly discharge is but three ti as great as the lowest.

If the comparison of southeastern Alaska with] glacial New England has any force, it suggests possible reason for the modest wavelength ratio betv valley meanders and stream meanders in New Engl even if the regimen of precipitation was modifie this region at glacial maximum, seasonal variatio runoff may not have been extreme.

In the interior of Alaska, as exemplified by Tanana River at Big Delta (table 9), runoff rises maximum in late summer, when precipitation coinc

TABLE 9.—*Some runoff data for three Alaskan stations*

Creek or river, drainage area, and location of measuring site	Period of record, in years	Mean discharge, in cubic feet per second per square mile [1]											
		Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	S
Fish Creek (32.1 sq mi) near Ketchikan: lat 55°23′30″ N., long 131°11′40″ W.	32 (1915–32, 1938–50).	21.5	19.9	12.1	11.8	9.0	7.8	10.0	15.6	15.0	11.2	10.9	
Tanana (13,500 sq mi) at Big Delta: lat 64°09′20″ N., long 141°51′00″ W.	8 (1949–52, 1954–57).	.70	.42	.38	.36	.36	.35	.42	1.27	1.91	2.77	2.67	
Kruzgamepa (84.0 sq mi) near Iron Creek: lat 64°55′00″ N., long 164°57′20″ W.	5 water years (1906–11).	1.67	1.01	.87	.76	.63	.55	.50	5.58	12.96	6.36	4.05	

[1] Arithmetic mean annual flood: Fish Creek (24 peaks recorded), 90 cfs per sq mi; Tanana River (8 peaks), 3.66 cfs per sq mi; Kruzgamepa River (2 peaks reco averages 3.8 times average discharge for months involved.

ith melting. Annual precipitation of about 17.5 ches over the basin is associated with discharges of ore than 2.5 cfs per sq mi in July and August and ith a mean annual flood of 3.66 cfs per sq mi. Because ie basin of the Tanana above Big Delta is very much rger than that of Fish Creek above Ketchikan, direct ımparisons between their respective momentary and onthly peak discharges are inadmissible. The fact ıat the Tanana, at the rate of 3.66 cfs per sq mi, ould require 120 days to discharge all the precipitation ı its basin may reflect the results of channel storage ıd basin storage rather than the results of other ctors.

Mean monthly midwinter temperatures in the Tanana ısin above Big Delta fall considerably below those constructed for the north-central conterminous United tates at glacial maximum, but midsummer temperaıres perhaps run a little high. The concentration of ınoff in the summer season that is indicated, for ıstance, by a hypothetical translocation of the climate : the Tanana basin to the Driftless Area may therefore e slightly exaggerated. But somewhat more general ›nsiderations support the hypothesis that a summer ›ncentration of runoff did in fact occur close to the ›rmer ice fronts of the interior conterminous United tates. The Alaskan plateau is a source region for ›ld polar continental air in winter, whereas in summer is affected by an extension of the polar front from the ering Sea and by single traveling lows. At glacial ıaximum the central part of the conterminous United tates may equally have been overlain by cold connental air in winter, receiving, however, lows during ımmer, especially in view of the steep energy gradient etween the ice fronts and the Gulf of Mexico. On limatic grounds, therefore, a marked summer conentration of runoff can be postulated for the interior t glacial maximum.

Permafrost brings its own special influences to bear n regimen of runoff, the effects being exemplified by ıe Kruzgamepa River on Seward Peninsula (table 9). 'here, with a mean annual precipitation of about 15 ıches and with perennially frozen ground seeming to ffset the effects of nearness to the ocean, runoff is very trongly concentrated in the open-water season. Mean ischarge for the Kruzgamepa near Iron Creek in the eriod 1906–11 rose to nearly 13 cfs per sq mi in June— ıore than three times as high as the 2.33-year momenıry peak on the Tanana at Big Delta. At 13 cfs per q mi, the Kruzgamepa could discharge all the precipiation on its basin in about 30 days. Unfortunately, o more than two momentary annual peaks are on ecord. For what they are worth, they suggest a ıomentary peak rate of flow about four times as great

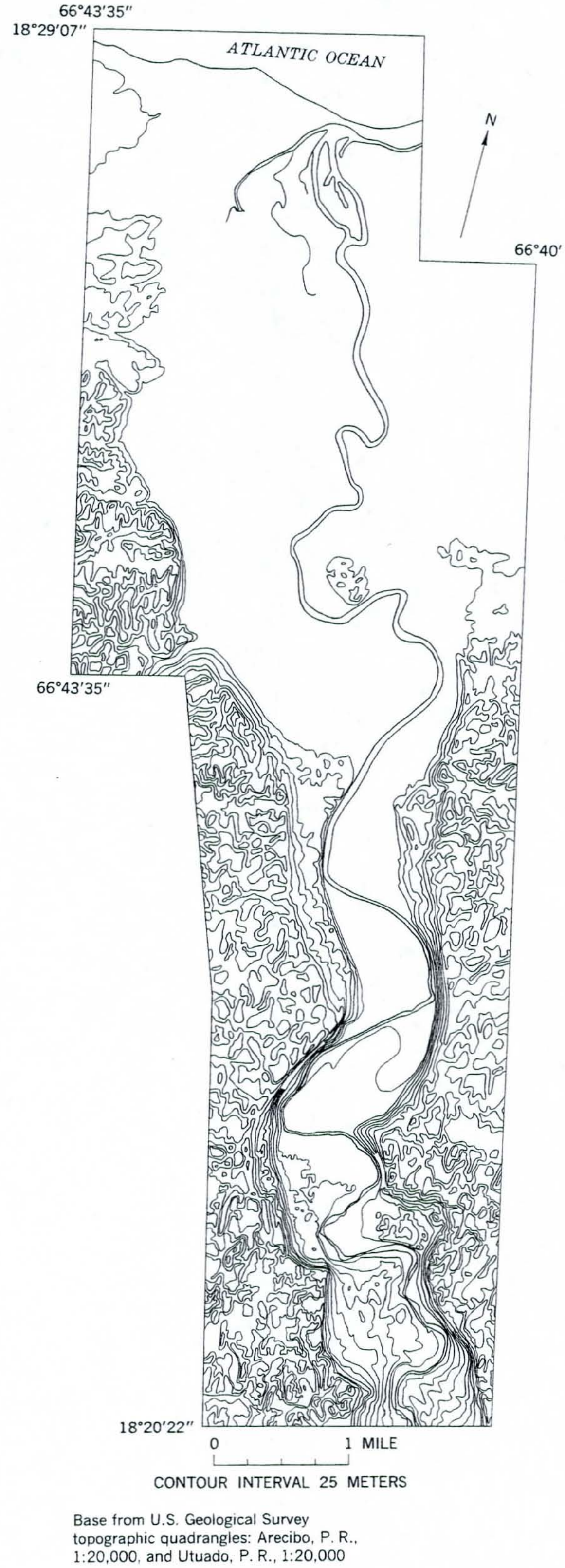
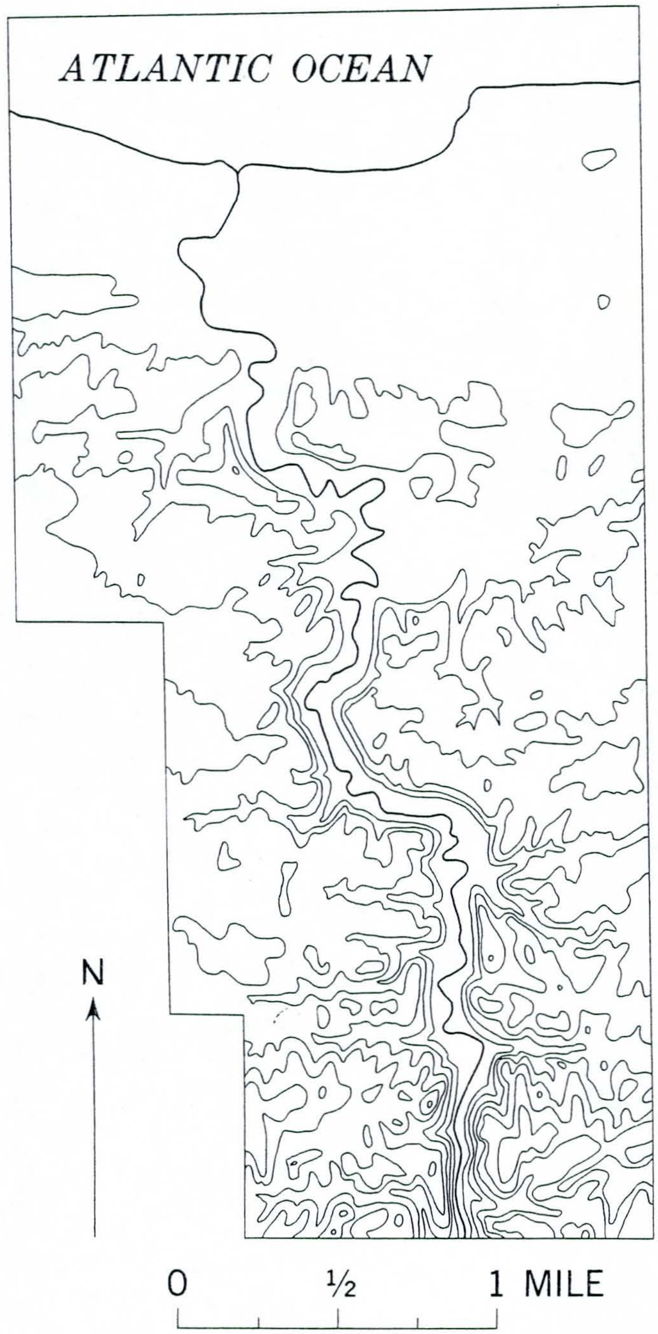


FIGURE 31.—Rio Grande Arecibo, P. R., showing contrast between valley and stream meanders. An example of manifest underfitness.



CONTOUR INTERVAL 25 METERS

Base from U.S. Geological Survey topographic quadrangle: Camuy, P. R., 1:20,000

FIGURE 32.—Rio Camuy, P. R., an example of manifest underfitness.

as the average for June—that is, a rate that could ensure runoff of all precipitation in little more than a week.

Although the seasonal concentration of runoff is very strong on the Kruzgamepa, it is, however, equaled in the conterminous United States, in regions well outsi the present limits of permafrost. Similarly, the se sonal concentration on the Tanana is well below th observed at many inland stations south of the 49 parallel (table 10; see also Harbeck and Langbei

TABLE 10.—*Selected data on seasonal concentration of runoff*

Creek or river	Percentage of annual runoff in peak month	Peak month
Kruzgamepa, near Iron Creek, Alaska	32.6	June.
Tanana, near Big Delta, Alaska	21.8	July.
Fish Creek, near Ketchikan, Alaska	13.6	October.
Roaring Fork, at Glenwood Springs, Colo	32.0	July.
Yellowstone, at Corwin Springs, Mont	29.0	June.
Red of the North, at Grand Forks, N. Dak	28.0	April.
Republican, near Bloomington, Nebr	25.0	June.
Pecatonica, at Freeport, Ill	16.0	March.

1949). If the necessary information were available, would be desirable to make comparisons not only peak monthly mean discharges but also of momenta peak discharges. However, even without such inf mation, reasoning in general terms is possible. F example, the percentage of total runoff discharg during the peak month is about 32 percent on t Kruzgamepa and 16 percent on the Pecatonica Freeport (table 10). If bankfull discharges were sin larly affected, then translocation of the climate of t Kruzgamepa basin to the basin of the Pecatonica wou insure no more than a twofold increase. If bankf discharge on the Kruzgamepa is taken at the exa gerated value of four times the mean for the pe month and that of the Pecatonica is taken at the undu low value of mean discharge for the peak month, th translocation of climate could still not account for mo than an eightfold increase of bankfull discharge in t Pecatonica basin. Such an increase automatica includes the effects of reduced temperatures. But ev with the aid of grossly exaggerated assumptions, fro fails signally to provide the required increase of least 20 and possibly more than 50 times in bankf discharge. Despite the first attraction of its possib ities, frost once again fails to supply an explanation the reconstructed discharges of meandering valleys.

CHANGED REGIMEN OF RUNOFF

Discussion of certain types of change in regimen h largely been anticipated above. Summer concentr tion of precipitation has been shown to reduce annu runoff, and winter concentration of precipitation increase it. But the proportional effects on total runo obtainable even from a very marked winter conce tration are small. Again, whereas increased wint concentration could perhaps be imagined for low la tudes during high-glacial times, it is difficult to app to regions which formerly experienced frequent ou spreading of cold air during the winter. Neither t ailable reconstructions of weather nor reference to ıld climates of the present day suggests anything but rmer summer concentration both of precipitation and runoff for many regions now typified by underfit reams. The effects of summer concentration of pre- pitation could have been largely offset by reduced mperatures, but the necessary allowance has already een made in the transformations of figure 16. And, s has been seen, the summer concentration of runoff ı Alaska, powerfully influenced by severe winters and y summer melt, does not necessarily exceed that in arts of the conterminous United States.

Too little is known of the interconnections of bankfull ischarge, discharge at mean annual flood, and mean ischarge for peak months to sustain a general argu- ıent for (or against) disproportionate increases in ankfull discharge over discharge of the other two inds. Even if former mean annual floods or peak ıonthly discharges were reconstructed, by means of ypothetical translocations of climate, they would not ıdicate former discharges at bankfull stage. Changes ı frequency relations can readily be imagined which ould increase bankfull discharge proportionally more han discharge at mean annual flood—and, indeed, hich would increase bankfull discharge while leaving ıean annual flood unchanged. But, quite apart from he close approach of q_{bf} to $q_{2.33}$ on some rivers at the resent time, the magnitudes of discharge involved ıave already been shown to demand something more han a readjustment of frequency relations in the ange from 2.33 years downward.

Changes in seasonal regimen of runoff, without change n total runoff, are then regarded as capable of making ıo more than a small contribution to the required ormer discharges—a contribution insignificant in com- parison with the fivefold or greater increase in total unoff inferred to result from reduced temperature and ncreased precipitation or in comparison with the postulated increase in bankfull discharge by a factor f commonly about 20 and somewhat exceptionally ıbout 50–60.

This conclusion is substantiated by a comparison of ıumerical values of former discharges with extreme values now observed and with return periods obtained for discharges of the computed former order. Approxi- mate calculations from the regional discharge-frequency graphs in flood reports for Illinois, Indiana, Kentucky, Missouri, Montana, and North Carolina suggest that 10-year floods are (with considerable range, however) about six times as great as 1.1-year to 1.2-year floods. Return periods of at least 200 years and, in some regions, of more than 10,000 years are needed to specify dis- charges 20 times as great as the present 1.1-year to 1.2- year floods. Where the rate of increase in discharge falls off with increasing return period—as, notably, on the Nene and Great Ouse (Dury, 1959)—discharges 50 times as great as present discharge at bankfull stage may simply not lie on any extension of the present frequency graph.

Computed former discharges for some other regions fall mainly within the scope of possible floods, as is shown by figures 33–34. Recorded extremes are plotted

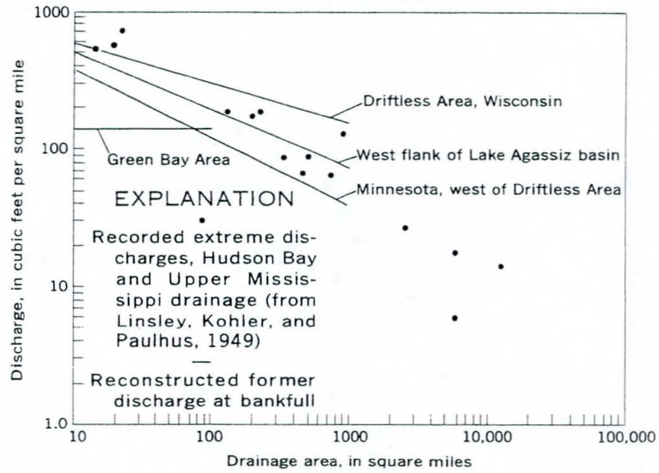


FIGURE 33.—Recorded extreme discharges, contrasted with reconstructed former discharge at bankfull stage. (Example 1.)

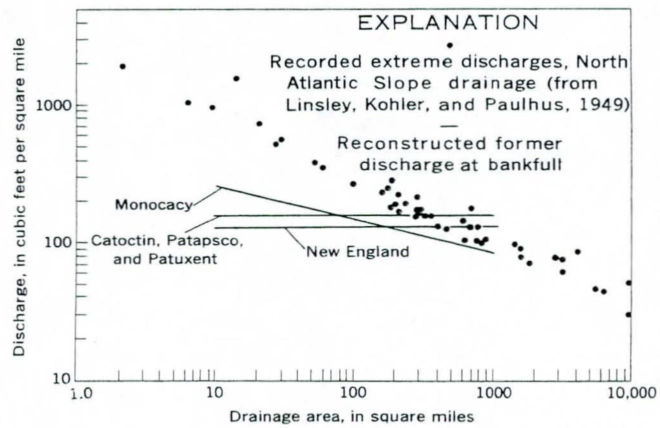


FIGURE 34.—Recorded extreme discharges, contrasted with reconstructed former discharge at bankfull stage. (Example 2.)

in these diagrams alongside reconstructed discharges, the discharges being somewhat arbitrarily reduced in accordance with figure 15. Apparent downstream reductions in former bankfull discharge, in terms of cubic feet per second per square mile, result directly from the form of the equations for wavelength used in computing table 2. Nothing presented here should be taken to imply either that bankfull discharge per square mile does or does not change with area of drainage. The sole point to be made is that former discharges appear to range well up into the present extreme values.

CHANGES IN STORMINESS, IN STATE OF SOIL, AND IN VEGETATION

The complex of possible changes now to be examined overlaps to some extent with the postulate of increased total precipitation. In the following paragraphs, however, emphasis is on changes in runoff from single rainfalls. Because storm rainfall is usefully specified not merely by quantity but also by intensity (or duration), the possible effects of changes in rainfall intensity will be examined. As with annual data, transformations of available generalized material will be used; but the nature of this material prevents a constant separation of the various factors from one another.

In a study of episodic erosion in New Mexico, Leopold (1951b) observed that an increase in annual precipitation does not necessarily accompany changes in the frequency of storms of given magnitude or changes in the intensity of rainfall in storms of given frequency. But the geomorphic effects now in question are of a different order from those described by Leopold, and increased total precipitation has already been inferred both for high-glacial times and for the Atlantic phase of postglacial time. Accordingly, changes in single storms too great to be accommodated in an unchanged annual total may fittingly be envisaged.

In effect, changes in runoff from single storms amount to changes in retardation of overland flow and in rate and capacity of infiltration. Studies of infiltration are now so numerous that the most summary review of their results is impracticable. However, data assembled and generalized by the U.S. Bureau of Reclamation (1960) can be made, by transformation, to predict the results of changes in infiltration capacity.

Figure 35 (curves II and III) shows the results of assuming that infiltration capacity for hydrologically medium forest, not at present having continuously moist soil, is reduced by one-half and by three-quarters. With the reduction by one-half, predicted factors of increase in runoff vary from 2 for rainfalls of 12 inches to 10 for rainfalls of 4 inches. With the reduction by three-quarters, they vary from 3 for rainfalls of 10.5 inches to 20 for rainfalls of 4 inches. However, unless it can be shown that bankfull discharges today are normally supplied when infiltration capacity is high, whereas former bankfull discharges were associated with low infiltration capacity, this particular set of results is not greatly relevant to the present problem. Changes in infiltration, throughout the year as a whole, have automatically been allowed for in transformations of annual data. Moreover, this part of the discussion remains entirely hypothetical so long as no means is suggested for effecting a significant change in infiltration.

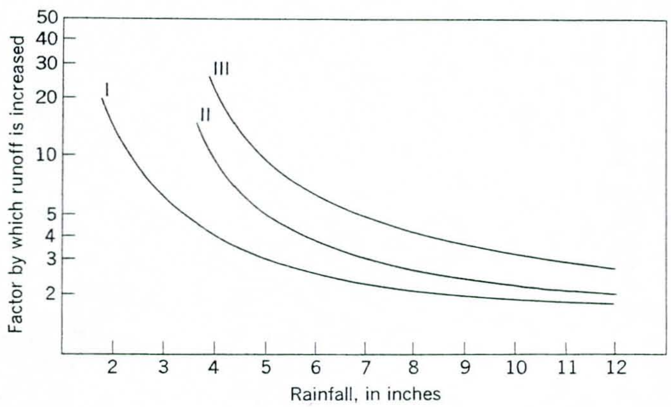


FIGURE 35.—Nomogram for effects on runoff of vegetational change and of change in infiltration capacity. Curve I, effect of conversion from retentively poor herbaceous vegetation to retentively fair oak-aspen forest; effect of reduction of infiltration capacity by one-half (curve II) and by three-quarters (curve III) for hydrologically below-medium forest, not at present having continuously moist soil.

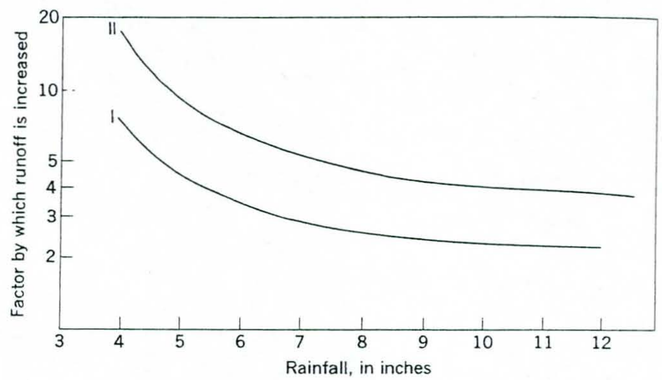
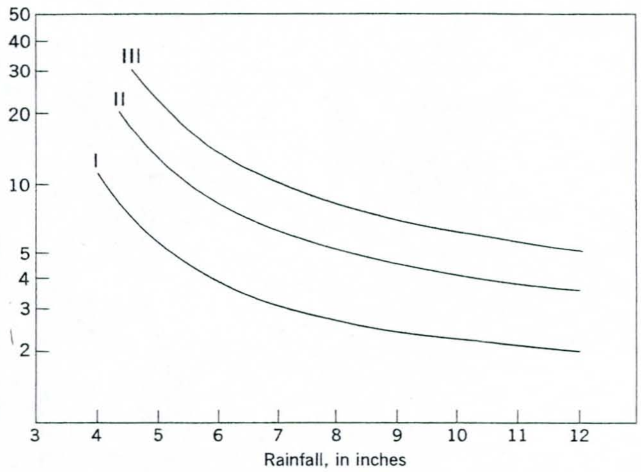


FIGURE 36.—Nomogram for effects on runoff of increase in amount of individual rainstorm. Curve I, 50 percent increase for hydrologically medium forest not having continuously wet soil; curve II, 100 percent increase for same forest.

The most obvious changes in infiltration are those due to changes in vegetation cover and in antecedent soil conditions. Their effects are implicit in further transformations of the data supplied by the U.S. Department of Agriculture. Let the initial conditions be taken as those of hydrologically medium forest, not having continuously most soil: the effects of increases of 50 percent and 100 percent in the amount of individual rainfalls are then those illustrated in figure 36. An increase from 8.4 to 12.6 inches in the amount of rainfall gives a predicted factor of 2.5 for increase in runoff, whereas a factor of 5.0 accompanies an increase of rainfall from 4.8 to 7.2 inches. The predicted runoff from a 20-inch rainfall is 4 times that from a 10-inch rainfall, whereas that from a rainfall of 9.6 inches is 10 times that from a rainfall of 4.8 inches. Extrapolation of the curves in figure 36 suggests that, in the existing range of 3–4 inches, doubling the amount of individual rainfalls could effect a twentyfold increase in individual runoffs.



URE 37.—Nomogram for effects on runoff of changes in antecedent wetness of il, hydrologically medium forest. Effect of conversion from dry to wet soil ıd, in addition, of no increase in amount of rain (curve I); of an increase of 50 ırcent (curve II); and of an increase of 100 percent (curve III).

Figure 37 shows the results expectable in a region of drologically medium forest when initially dry soil converted to a state of continuous moistness and .en (a) amount of rain is unchanged, (b) amount of n is increased by one-half, and (c) amount of rain is ubled. Mere conversion of soil condition doubles the ıoff from a rainfall of 12 inches and increases tenfold ıt from a rainfall of 4 inches. When amount of nfall increases, increases of runoff by factors of 20 more are predicted in the range of 4–5 inches for ›sent rainstorms. This set of comparisons is not .olly just, since, like the assumptions made about ınges in infiltration capacity, it implies a contrast ;ween present-day conditions which are adverse to ;h runoff and former conditions which were favorable. The lowest curve (curve I) in figure 35 was obtained transforming data for vegetation cover alone. ıssumes the replacement of retentively fair oak-aspen est by retentively poor herbaceous vegetation. th amount of rainfall unchanged, runoff predicted individual rainstorms increases by a factor of 2 a rainfall of 8.5 inches and by a factor of 10 for a nfall of 2.3 inches. However, if changes in vege-ion are considered as partial explanation of underfit eams, they must be supposed to have operated in ingle direction throughout vast areas—an unlikely ›position, especially as some regions which contain lerfit streams support, at the present time, vege-ion which is retentively poor. And it seems most probable that the increased runoff of the Atlantic ıse in northwest Europe was associated with a ;ened retention by the plant cover. Vegetation ınges are therefore regarded as complicating the lrologic history of late-glacial and postglacial times, but they are not regarded as a prime cause of fundamental change in peak discharges.

Despite the reservations made necessary by the types of conditions assumed, one valid principle emerges from the transformations carried out: for a given assumption of change, proportional increases in runoff rise toward the lower end of the scale of rainfalls. This does not necessarily mean that peak discharges would undergo comparable increases, but it does show how the proportional rise in total runoff becomes greater than that in total precipitation (figs. 21–22). Since, moreover, large proportional increases in runoffs from single rainstorms of moderate amount can accompany lesser increases in total runoff, it will presently be necessary to inquire whether moderate rainstorms are capable of having promoted former discharges at the bankfull stage and whether they are likely to have done so.

When duration (or intensity) of rainfall is introduced, the inquiry becomes still further complicated. On previous occasions, the present writer has expressed calculated values of former bankfull discharge in terms of rainfall intensity with a view to demonstrating that former discharges are beyond the expectable limits of precipitation at the present day (Dury, 1954, p. 215–218; 1958, p. 113–114). The facts that the values adopted for rates of discharge now appear rather high and that the streams discussed are highly underfit constitute no obstacle to the type of reckoning employed. At the time, however, the signal difference in magnitude between present-day bankfull discharges and expectable intensities of rainfall was not clear, so that comment was deficient in a respect now seen to be significant.

The equation used by the writer (Dury, 1954) to convert discharge to its equivalent rainfall was the so-called rational formula,

$$q_{\max}=640\ CiM, \tag{34}$$

where $q_{\max}$ is the expectable peak discharge past a given point, C is a coefficient of runoff, and i is the intensity of rainfall, in inches per hour, on a drainage area, M, in square miles. The rational formula assumes that, for rainfall exceeding the time of concentration, the rate of runoff equals the rate of rainfall reduced by an appropriate runoff factor (Linsley, Kohler, and Paulhus, 1949, p. 575).

When C is taken as unity, the equation can be made to give rainfall intensity equivalent to a given rate of discharge, in the form

$$i=q/640M. \tag{35}$$

Although this equation is valid as a statement of equivalence, it is unsuitable for predicing the intensities actually necessary for given discharges. It ignores the influence of infiltration, retardation, channel storage, and form of hydrograph; on this count, it tends to predict intensities that are too low for large basins. On the other hand, it also ignores the flow already occurring before a storm; on this count, it tends to predict intensities that are unnecessarily high. Actual values of bankful discharge and of discharge at mean annual flood were not available to the writer in 1954. When these are inserted in equation 34, equivalent rainfall intensities run well below intensities in the observed range, as will be described below. For the present, it is enough to note that the rational formula and its derivative, equation 35, have little bearing on the reconstruction of actual former intensities of rainfall—not merely because of discrepancies in magnitude but also because they omit, among other factors, initial flow. And since cross-sectional areas of channels do not increase linearly downstream, no linear equation will serve. It remains true, however, that if rainfall equivalents calculated for former discharges are beyond the reasonable limits of present intensities, some change in intensity relations is required.

For all that, expressions of bankfull discharge in terms of rainfall equivalent are not wholly valueless, nor is it impossible to examine some of the effects, or implications, of changes in rainfall intensity. Coaxial relations of antecedent precipitation, duration of storm, storm precipitation, and storm runoff, as developed by the U.S. Weather Bureau, can be made to predict the effects upon runoff of change in any one of the named factors and also in intensity. For instance, if duration is reduced from 48 to 24 hours, other conditions remaining constant, then the indicated change in total runoff shows the results of a doubling of rainfall intensity.

Linsley, Kohler, and Paulhus (1949, figs. 16–7 and 16–8) reproduced coaxial diagrams for two groups of basins, one in Ohio and one in Kansas. Since the effect of antecedent precipitation varies greatly from season to season, comment will be restricted to those parts of the year when this effect is greatest, on the assumption that these are the times most liable to record discharge at high stages. Calculations of increased runoff with increased antecedent precipitation—that is, with increased wetness of soil—and with increased duration give rather modest values even when quite extreme changes are assumed. Thus, if the index of antecedent precipitation is raised from 0.5 to 2.0 inches and intensity is assumed to increase infinitely with a reduction in duration from 96 to 0 hours, the percentage increase in runoff is no more than 20–50 percent for an 8-inch rainfall and 100–150 percent for a 1-inch rainfall.

The predicted increase in runoff rises if the amount of rainfall also is assumed to increase. But, even then, the results of a hypothetical doubling of storm rainfall in the present range of 1–4 inches, an increase in the index of antecedent precipitation from 0.5 to 2.0 inches, and a reduction in duration from 96 to 0 hours would still not do more than increase total runoff 3–5 times. The conflict between this and the earlier set of results cannot be resolved without further study, although the results of transforming annual data on precipitation and runoff favor the larger rather than the smaller predicted increases of runoff from single storms.

Current intensities of rainfall can be obtained, as averages for storms of stated duration, from the analyses of frequency-intensity regime issued by the U.S. Weather Bureau as technical papers. In these technical papers, storms are specified by frequency and duration; divided by the duration, the total rainfall for a given frequency gives the mean intensity of fall. Intensities decrease with increasing extent of storm and also with increasing duration. Shifts of intensity along the scale of frequency and duration would change the total falls set against given return periods, durations and areas.

The rate of increase in storm rainfall with increasing duration or return period varies considerably from station to station. Regional values may, however, be approximated by averaging data for representative stations. These data suggest that if the intensities now appropriate to 2-year 1-hour rainfalls in the Middle Atlantic region and in the Southeast were extended to the 2-year 6-hour and 24-hour rainfalls, then total falls would increase by about 3.5 times for 6-hour rains and by about 11 times for 24-hour rains.

Shifts in frequency necessary to effect a stated change in total rainfall for given durations—that is, a selected change in intensity—can be read from figure 38, which is based on regional data and which includes 1-year values extrapolated from 2-year and 100-year rains. Tables 11 and 12 list, against duration and present return period, the return periods now associated with

TABLE 11.—*Return periods of rainstorms 50 percent greater than those of today*

Region	Present return period, in years	Return period, in years of rainstorms of indicated duration that are 50 percent greater than those of today	
		1 hour	24 hours
Middle Atlantic	1	4	
	2	10	1
Southeast	1	6	
	2	17	

ɜLE 12.—*Return periods of rainstorms 100 percent greater than those of today*

Region	Present return period, in years	Return period, in years, of rainstorms of indicated duration that are 100 percent greater than those of today	
		1 hour	24 hours
dle Atlantic	1	15	15
	2	50	35
heast	1	25	10
	2	100	35

nfalls 50 and 100 percent greater than present 1-year 1 2-year rainfalls. The shifts in frequency increase :h decreasing duration and with increasing return ·iod. Thus, a given proportional increase in total l could occur with minimal disturbance of frequency ıtions if it were concentrated in rainfalls of long ration but of high frequency.

[f two parallel frequency series could be constructed, ɘ for discharge and one for rainfall, the rainfalls ociated with selected frequencies of discharge could read off. In practice, however, correlation is disıraged by variations in the extent and location of gle storms, in tributary contributions to flood peaks, 1 in seasonal and short-term conditions of infiltration 1 retardation. Rainfall of high intensity can occur the low-water season; also, seasonal concentration of ıoff commonly is greater than that of rainfall. Part of the potential effect on runoff of rainfall of stated frequency is therefore lost.

A rough kind of comparison may nevertheless be attempted. Let it be assumed that the rain which occurs with a frequency of 2.33 years is responsible for discharge at mean annual flood and that the rain responsible for discharge at bankfull stage has a return period of somewhat more than 1 year. As a rough approximation, the 2-year point rainfall may be taken as equivalent to the 2.33-year areal fall, and the 1-year rainfall may be taken as equivalent to the areal fall having a return period of 1.1–1.2 years. When discharges are converted to equivalent rates of precipitation, they run well below expectable intensities of rainfall, frequency by frequency.

In table 13, miscellaneous data for discharge at bankfull stage and at the 2.33-year flood are listed against their equivalents in rainfall intensity. Although the discharge data relate to drainage areas varying widely in climate, it is probably fair to contrast the discharge equivalents of streams in Georgia and Alabama with rainfall intensities in the Southeast, to contrast those of streams in New England with rainfall intensities for the Middle Atlantic region, and to infer that, in other regions also, rates of discharge are likely to be far less than rates of rainfall at equal return period. The highest rates of bankfull discharge listed in table 13, for streams in Georgia (computed) and for

BLE 13.—*Discharges at bankfull stage, q_{bf}, and at the 2.33-year flood (mean annual flood), $q_{2.33}$, with equivalents in terms of rainfall intensity*

Region or basin	Area, in square miles	q_{bf} (present discharge at bankfull)		$q_{2.33}$ (present discharge at the 2.33-year flood)		$q_{2.33}/q_{bf}$
		Cubic feet per second per square mile	Equivalent rainfall, in inches per hour	Cubic feet per second per square mile	Equivalent rainfall, in inches per hour	
ːtless Area, Wis	1,000	[1] 4.6	0.007	(?)6.0	0.009	1.3
	10	[1] 7.0	.011			
en Bay area, Wisconsin	100	[2] 7.0	.011	(?)12.0	.019	1.7
	10	[2] 7.0	.011	(?)50.0	.078	7.1
thern New England		[2] 7.8	.012	[2] 22.5	.035	2.9
en River, Ky	500	[1] 6.6	.010	35.0	.055	5.3
	10	[1] 13.4	.021	150.0	.235	11.2
nboldt River, Nev., and Owyhee River, Oreg	1,000	[1] .4	.0005	(?)0.7	.001	1.75
	10	[1] 1.6	.003	(?)25.0	.039	15.6
bash and White Rivers, Ind	20,000	[3] 2.0	.003	[3] 5.0	.008	2.5
	100	[3] 10.0	.016	[3] 25.0	.039	2.5
rgia streams (mean of five)		[4] 24.8	.039			
theast Ozarks	3,000	[2,5] 7.0	.011	[5] 7.0	.011	1.0
	100	[2,5] 7.0	.011	[5] 52.0	.083	7.4
bama River system	20,000	[2,5] 4.5	.007	[5] 5.0	.008	1.1
	1,000	[2,5] 4.5	.007	[5] 15.0	.024	1.3
l River, N. Dak.-Minn., and Sheyenne River, N. Dak		[2] .63	.001	[2,5] .63	.001	1.0
ıe and Great Ouse, England	1,000	[1] 1.7	.003	[2] 4.4	.007	2.6
	100	[1] 2.9	.005	[2] 4.4	.007	1.5
		[2] 1.8	.003	[2] 4.4	.007	2.4
at Ouse River, England	[6] 1,200	[7] 1.96	.003	[2] 4.4	.007	2.25
	[6] 550	[7] 4.8	.008	[2] 4.4	.007	(?).9
e River, England and Wales	[6] 65	33.6	.054			

[1] From preceding tables.
[2] Regional values.
[3] Dury (1961).
[4] U.S. Geol. Survey, Water Resources Div., Research Seminar Paper, May 1958.
[5] From preceding figures.
[6] Approximately.
[7] Nixon (1959).

the upper Wye in Great Britain (observed), are both below the equivalent rainfall rate of 0.06 inch per hour, and most of the remaining rates of bankfull discharge equal less than 0.01 inch per hour.

The contrast between discharge and its equivalent in rainfall is due in part to the slow variation of runoff in comparison to rainfall. In part it reflects loss or retention of water. But the magnitude of the contrast depends also upon the duration selected for a given return period. Thus, whereas 1-year 1-hour rainfalls in the Middle Atlantic region are equivalent to discharges of 700 cfs per sq mi and those of the Southeast are equivalent to discharges of 1,150 cfs per sq mi, the 1-year 24-hour rainfalls are equivalent but to 60 and 100 cfs per sq mi, respectively.

If water could be delivered to rivers at the present rates of rainfall in 1-year 1-hour storms, the discharges produced would be capable of cutting valley meanders. Delivery at the present rate of 1-year 24-hour storms could also suffice with the aid of high initial flow. Although analyses of rainfall probability for some regions where streams are highly underfit give present intensities of rather low order (Dury, 1954), these analyses are defective in overlooking duration. The quantities involved in the preceding discussion suggest that increases in rainfalls of long duration and high frequency, themselves promoting high indices of antecedent precipitation and high rates of initial flow, could result in disproportionate increases in discharge at stated return period. The shift in frequency of discharge of given magnitude could exceed the shift in frequency of rainfall of given intensity and duration. Indeed, it would have to do so if rainfalls on the present range of frequency are to be capable of promoting discharges of the former order, because some of the former discharges, as has been said previously, lie well beyond the uppermost limits of present discharges.

For a given frequency and duration, any increase in intensity of rainfall during high-glacial times must be reconciled with greatly reduced air temperatures. No such qualification, of course, applies to the Atlantic phase of postglacial time, when increased temperatures can be assumed to have increased the water-bearing capacity of the air. But temperature reductions postulated for high-glacial times are so great, even if high rainfall intensity is referred to waning rather than to maximum glaciation, that their possible effect on totals of rainfall cannot be ignored.

Generalized estimates of maximum possible precipitation, published by the U.S. Weather Bureau (1947) for the United States east of the 105th meridian, permit a rough judgment of the potential results of change in temperature. The maximum height of cumulonimbus tops at the present time ranges from 28,000 feet (300 millibars) in winter to 53,000 feet (100 millibars) summer. If 34,000 feet is assumed as the depth column, on the grounds that maximum height formerly less than it now is but that precipitation formerly concentrated in summer, then it can be sho that a reduction of 20°F in surface temperature wo reduce the depth of precipitable water by two-thi Specifically, in the Ohio valley, where mean July te perature is about 75°F and where high-glacial tempe tures are held to have been some 20°F lower than th of today, a reduction in surface temperature from 7 to 55°F would reduce the depth of precipitable wa from about 3 inches to about 1 inch. A greater prop tional reduction is obtained when allowance is made differences in depth of column between present former times.

Maximum possible precipitation does not, howe depend merely on depth of precipitable moisture also on transport of air. Extent and duration of sto are both involved. The generalized estimates sh that, at many stations, possible maximums incre rather slowly with increase of duration above 6 ho and that they decrease rather slowly with increas area. Thus, they are commonly not more than 50 cent greater for duration of 24 hours than for durat of 6 hours; the maximum possible fall on an area of square miles is commonly less than twice that on area of 500 square miles at 6-hour duration and less t 1½ times as great at 24-hour duration.

For the sake of inquiry, let the area of 200 squ miles and the duration of 24 hours be adopted. Ta 14 lists, against selected stations, expectable 2-y 24-hour rainfalls that have been converted from po

TABLE 14.—*Some values of 24-hour 200-square-mile rainfa*

Station	2-year rain [1]	Maximum possible rai	
		Actual [2]	×⅓
Mobile, Ala	5.3	31.9	
Savannah, Ga	4.2	28.2	
Pensacola, Fla	5.5	31.5	
Macon, Ga	3.5	28.2	
Montgomery, Ala	4.1	30.5	
Birmingham, Ala	4.1	30.0	
Richmond, Va	3.3	24.0	
Charleston, N.C	4.1	28.2	
Pittsburgh, Pa	2.2	23.0	
Harrisburg, Pa	2.8	22.0	
New York, N.Y	3.2	23.6	
Baltimore, Md	3.2	22.2	

[1] U.S. Weather Bur. Tech. Paper 29, "Rainfall Intensity-Frequency Regime.
[2] U.S. Weather Bur. (1947).

rainfall to areal rainfall for 200 square miles by factor 0.92. Also listed are maximum possible cipitations—for 200 square miles and 24-hour h duration—and these same precipitations multiplied 0.33. The intention is to inquire whether rainf of the given extent and duration could possibly oc in conditions of greatly reduced temperature.

sumptions made are that maximum possible precipitation is reduced in the same proportion as depth of precipitable water (an assumption which may not be justified) and that the temperature reduction of 20°F applies to all regions (an assumption which is certainly at variance with reconstructed latitudinal gradients of temperature). Let these assumptions stand, however, as rough hypotheses. Then the 2-year 24-hour fall for 200 square miles is still only ⅓ or ½ as great as reduced values of maximum possible 24-hour precipitation. Pittsburgh and Richmond are perhaps fairly representative of areas which formerly experienced a reduction of 20°F in surface temperature: their reduced maximum possible precipitation for 24 hours is 2½-3½ times as great as the present 2-year 24-hour fall. Reference to figure 38 suggests that the reduced

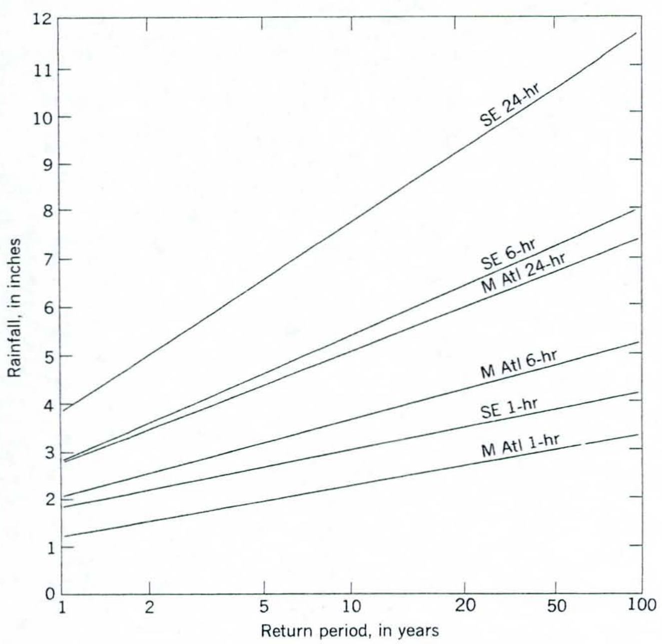


FIGURE 38.—Generalized rainfalls for varying return periods and duration, Middle Atlantic region and Southeast United States.

possible maximums are about 4 times the value of 24-hour falls at a return period of 1 year.

Since maximum possible rainfalls increase with increasing duration in about the same proportion as do recorded rainfalls of given frequency and extent, a fixed proportional reduction in possible maximums would still leave room for increases in actual rainfalls throughout the range of frequency and duration. The reduced possible 24-hour maximum for an area including Richmond, Charleston, Pittsburgh, and Harrisburg, allowing for a reduction of 20°F in surface temperature, appears to be about 8 inches, which corresponds to a present return period of about 200 years (fig. 38). If smaller temperature reductions are assumed to reduce possible maximums in the South by about one-half, then these maximums run at about 15 inches. A 15-inch 24-hour rainfall in the Southeast has at present a return period of about 700 years.

Great reductions of temperature in high-glacial times therefore do not preclude either shifts in frequency relations of the order shown in table 12 or twofold increases in total precipitation. The necessary water could have been carried in the air. But unless a great allowance is made for increased transport, the fourfold increase in rainfalls of short return period and long duration seems to mark the practicable limit. Since this increase would demand maximum possible precipitation in former times, it can be taken as outside the limits of probability.

HYDROLOGIC CHANGES IN THE PERSPECTIVE OF LATE-GLACIAL AND DEGLACIAL TIME

The foregoing discussion of the hydrologic effects of low temperatures in high-glacial times assumes that the maximal former discharges of streams that are now underfit were associated with times of maximum cold. Although there is evidence that increased rates of deep-sea sedimentation coincided with glacial maximums (Broecker, Turekian, and Heezen, 1958; Broecker Ewing, and Heezen, 1960; Emiliani, 1955; Ericson and Wollin, 1956; Ewing and Donn, 1956, 1958; Hough, 1953), it still does not follow that the large former streams existed in strict contemporaneity with maximum glaciation rather than with early deglaciation.

Three points arise here:

1. Troll (1954) and Kremer (1954) showed that certain European rivers, which now flow in single channels within meandering valleys, assumed a braided habit at times of greatest cold, thus tending to infill long reaches of their valleys instead of deepening the valley meanders. If such conditions were at all common, then the incision of valley meanders ought not to be referred precisely to times of maximum glaciation.
2. The very high stands of Lake Lahontan in Zones Ic and III have already been referred to increased pluviation rather than to reduced temperatures (Dury, 1964b; foregoing text; table 15).
3. Reconstructions of climate for glacial maximum frequently embody high-pressure systems in regions now typified by manifestly underfit streams (Büdel, 1949, 1953; Flohn, 1953; Klute, 1951; Poser, 1948; Wright, 1957, 1961). If such highs were capable of persistent blocking action, then precipitation of the order required by former channels and valley meanders seems unlikely while the highs endured.

On all these counts, the reconstructed former discharges seem to belong to waning rather than to maximum glaciation. Such a view accords with the appearance of large meanders on Black Earth Creek, Wis., during waning glaciation, with the considerably later initiation of valley meanders in emerging parts of the Lake Borders, and with the scouring to bedrock of large channels in southern England as late as Zone III times. Although the chronological evidence does no more, in one sense, than fix limits for the conditions appropriate to the cutting of valley meanders, the dates involved require that already rising deglacial temperatures should have been fully compensated by increasing precipitation. Full compensation between 20,000 and 10,000 years B.P. in northwest Europe means compensation for a rise in July temperatures by 18°F—more than half the total rise between glacial maximum and hypsithermal maximum (Andersen and others, 1960; Flint and Brandtner, 1961, fig. 1).

TABLE 15.—*Outline of chronology*

Years B.P.	Zonal No.	Zonal name
	X	
100		
	IX	Sub-Atlantic.
2, 500		
	VIII	Sub-Boreal.
4, 500		
	VII	Main Atlantic.
6, 200		
	VI	Early Atlantic (Transitional).
7, 750		
	V	Boreal.
9, 000		
	IV	Pre-Boreal.
10, 000		
	III	Younger Dryas.
10, 600		
	II	Allerød (Two Creeks).
12, 000		
	Ic	Older Dryas.
12, 700		
	Ib	Bölling.
13, 300		
	Ia	Oldest Dryas.
16, 000		

Nevertheless, the last recorded scourings of large channels occurred in conditions which, if they were distinctly wetter than those of today, were also distinctly colder. Temperature and precipitation worked together in favoring high discharge. Lesser episodes of channeling were associated with climatic fluctuations in which changes in temperature and in precipitation did not necessarily work in the same direction.

If stream shrinkage occurred in southern England during Zone II, similar to that inferred for Wisconsin at the corresponding time, it may have been due either to reduced precipitation, to a rise in temperature to within 8° of existing values for July (Manley, 1951), or to both combined. Manley's estimate of temperature closely resembles that obtained for New England by E. B. Leopold (1958), who considered that July temperatures in New England during the Two Creeks interval (=Zone II) were 5½°–9°F below current values. Both writers agreed on a modest reductio July means subsequent to Zone II amounting to s 4°F. This scarcely seems enough to account for renewed channeling in southern England in the r Zones III–IV, so that increased precipitation is requ to accompany temperature reduction. Although Ma (1959) inferred reductions in summer mean temp tures for northwest England by 7°–16°F for the fluctuation after Zone II, he also called for a consider increase in precipitation at the same time to exp extensions of glaciers. Since increased precipitati is needed to explain the renewed rise of Lake Laho to very high levels, a cold-humid fluctuation ca inferred for Zone III (approximately), both in hu and in arid regions.

Infilling of large channels in southern England du Zone V coincided, as the floristic record shows, temperatures reduced anew. During this interv reduction of precipitation must be inferred capabl counteracting the effects of reduced temperatures in addition, of causing runoff to decrease. In di contrast, as previously observed, increased tem atures during Zone VII were associated with incre runoff, which considerably increased precipitat supplied runoff.

The deglacial succession affords room for episode minor cutting—in addition to those identified here in Professional Paper 452–B—such as would be expe to accompany the minor fluctuations of degla sea level reported by Fairbridge (1961) or would expected from the climatic evidence reviewed by C way (1948, and references therein). In the longer te the tally of about 15 complete temperature cy thought to relate to the Pleistocene (Emiliani, 1 suggests considerable fluctuations of runoff during last half-million years or more. The last main onse underfitness, though its location on the scale of degla time is somewhat imprecise, may have been the such episode of many.

Broad climatic implications of the findings descri in this Professional Paper series lie, for the most p outside the limits of present inquiry. No attempt be made to reconstruct the meteorological pattern relevant intervals of time, whether those of channel or those of filling. In a general way, of course, the h precipitation inferred for times of incision of va meanders and for times of complete scouring of larg channels relates to steep meteorological gradients ac belts of latitude. Willett (1950) postulated intensifi tion of the general circulation at the last glacial m mum, and Ewing and Donn (1958) called for very gr storminess in middle latitudes as a result of frequ encounters between glacial air from the icecaps moist equatorial masses. In the shorter term, Wins

(1955) described rapid cyclogenesis—including the rejuvenation of old secondaries—in association with steep thermal gradients in Alaska. However, it is difficult to extrapolate to times of maximum glaciation the fluctuations of lengths measurable in days or weeks, such as are coming to be well understood for the present-day Arctic (Namias, 1958a, b, and 1960, and references therein). Still further complications ensue from the probability that the last main episode of cutting of valley meanders was early deglacial rather than maximum glacial in date, although a time lag between the insolation cycle and the temperature cycle (see Emiliani, 1955) opens interesting possibilities in this connection.

It may well emerge that underfitness is by no means confined, as at one time seemed likely, to the belt of midlatitude westerlies, even in the widest possible sense of this term; for, quite apart from the former pluvial conditions in some regions inside the tropics, there are signs that Alaska also has been effected by reductions of discharge powerful enough to make rivers manifestly underfit (fig. 39). In this way, the general theory of

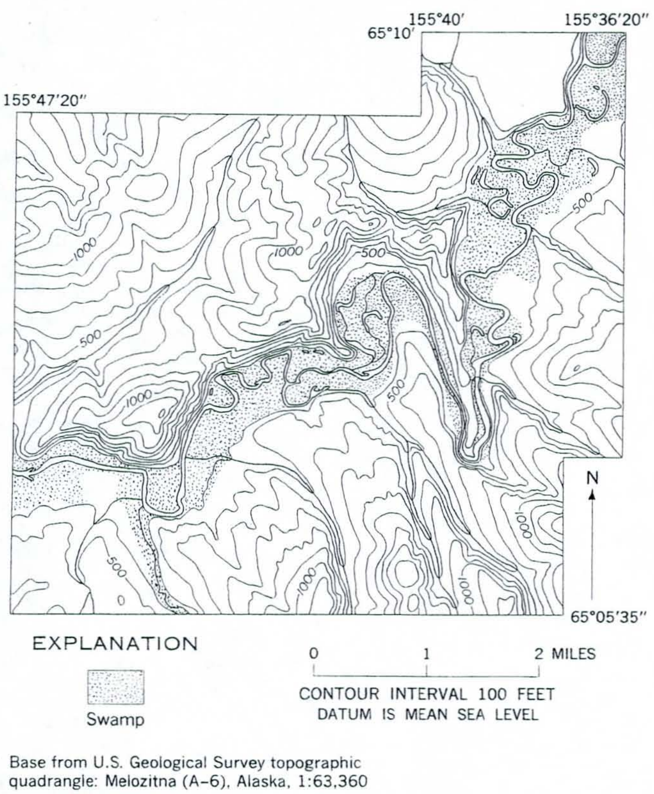


FIGURE 39.—Part of the Dulbi River, a manifestly underfit stream in Alaska.

underfit streams reinforces such interpretations as those of Büdel (1949), in which the general atmospheric circulation was held to be strengthened at times of extension of land ice. (See also Charlesworth, 1957, p. 1131–1145.) Meanwhile, the necessary changes in total precipitation that are required to account for the former discharges of streams now underfit are not greater than the season-to-season variations recorded in some regions at the present time.

SUMMARY

1. As previous workers concluded, meander wavelength varies with bankfull discharge in the form $l \propto q^b$: specifically, the amplified data presented here indicate that $l \approx 30q^{0.5}$, where l is in feet and q is in cubic feet per second.

2. The wavelength ratio L/l between valley meanders and stream meanders can be used to compute discharge ratios Q/q from $(L/l)^2$; these ratios range above 100:1 in some instances.

3. Allowance for change in downstream slope and in channel form gives discharge ratios of about 60:1 or 50:1 for streams showing marked underfitness, where wavelength ratios approximate 9:1, and of about 20:1 in the more common circumstance where wavelength ratios approximate 5:1.

4. Reductions in air temperature of the order reconstructed for glacial maximum, combined with increases in precipitation above present-day values by 50–100 percent, are capable of increasing mean annual runoff by factors of 5–10 within a wide range of existing climates.

5. Change of temperature is not alone sufficient to explain the observed effects, particularly since cutting of valley meanders and scouring of large channels persisted into early deglacial times.

6. Frozen ground cannot provide a general explanation of former high discharges.

7. The most likely cause of increased momentary peak discharge, in addition to the increase in mean annual runoff, seems to be increase in single rainfalls, particularly in rainfalls of long duration and high frequency.

8. The required increases in precipitation lie within the limits of physical possibility set by air temperatures.

9. Varying combinations of trends in temperature and in precipitation, at certain intervals of the deglacial succession, can be associated with episodes of minor channeling or filling.

10. The last main episode of channeling and of deepening and enlargement of valley meanders agrees with inferences of increased storminess during times of glaciation (in the broad sense).

11. Precise regional contrasts in degree of underfitness are difficult to make. However, high degrees of underfitness characterize the Driftless Area of Wisconsin, the upper reaches of streams in the Ozarks, and the English Plain—areas that were subject to very rigorous climates, but were not ice covered, at relevant glacial maximums. Elsewhere, a wavelength ratio of

about 5:1 is widespread. Lower ratios appear in areas, such as Syria and Puerto Rico, well distant from the former ice fronts. The identifiable contrasts agree with postulates of displacement of storm tracks and of locales of frontal activity, and with hypotheses of increased strength in the general atmospheric circulation, at or near times of glacial maximum.

TRANSFORMATION OF PRECIPITATION-TEMPERATURE VALUES TO GIVE THE FACTOR F_q OF INCREASE IN ANNUAL RUNOFF

Whether the factor F_q is computed or is obtained mainly by graphical means, its values depend on comparison of runoff for stated precipitation-temperature values with runoff for other precipitation-temperature values. For graphical determination, let, for example, present runoffs be listed against amounts of precipitation, for the isotherm 70°F in figure 16. A second listing, against the same precipitations but for the isotherm of 50°F, shows the amounts to which runoffs should rise for a reduction of 20°F below the weighted mean of 70°F and through a range of precipitation. Division of items in the second list by corresponding items in the first list gives values of F_q against amounts of precipitation for the present isotherm of 70°F and for a postulated reduction of 20°F in weighted mean annual temperature. The following sample exemplifies the results obtained:

Present precipitation, in inches	Present runoff, in inches for weighted mean annual temperature		F_q for reduction from 70°F to 50°F
	70°F	50°F	
50.0	12.2	24.0	1.97
47.5	10.6	21.25	2.00
45.0	9.1	19.25	2.12
42.5	7.7	17.0	2.21

In practice, values obtained in this manner for F_q are likely to need smoothing. Smoothing can itself be effected graphically, for example, with the aid of a plot of values of F_q against present annual precipitation for a given isotherm (in the foregoing example, 70°F) and for a stated reduction in temperature (in the example, 20°F). Similar graphs for other isotherms and for other extents of temperature reduction provide integral values of F_q for use in constructing nomograms of the type given in figures 18–20.

A second set of integral values for F_q, obtained when changes in precipitation are envisaged for conditions of unchanged temperature, supply the basis for nomograms such as figures 21–22. Values of F_q for changes both in precipitation and temperature are obtained by listing the present runoff from present rainfalls at the isotherm of 70°F against runoff from 1½ times the present rainfall at the isotherm of 60° in this example, the values of F_q correspond to th used in constructing figure 23. Appropriate adju ments supply values for figures 24–26.

TRANSFORMATION OF RUNOFF VALUES FOR SING STORMS

The procedure here is essentially similar to that j outlined for values of annual runoff. It makes use the information presented by the U.S. Bureau Reclamation (1960, Appendix A, p. 412–431, "Es mating rainfall runoff from soil and cover data"; secti A–5 of that appendix, containing nomograms for run equations supplied by the U.S. Soil Conservati Service, is especially relevant).

For example, two comparative series of values c be derived from the runoff equation $q=\frac{(p-0.2}{p+0.8}$ —where q is direct runoff, in inches; p is storm rainf in inches; and S is maximum potential difference l tween p and q, in inches, at the time of storm's beg ning—by taking a range of values for p and a seco range in which each entry is twice as large as the c responding entry in the first range. The two results ranges of values for q will then indicate the effects o doubling in the amount of rainfall. Results of chan in vegetation cover are obtainable by use of the runc curve numbers listed by the U.S. Bureau of Reclamati (1960), table A–2, p. 426. As with data for ann runoff, smoothing by means of graphs is usually nec sary, and in any event it is useful in obtaining integ values for increase in storm runoff. The followi sample exemplifies the results obtained:

Storm rainfall, in inches	Storm runoff, in inches, for hydrologically medium forest, antecedent condition II, Soil Conservation Service runoff curve 40	Storm runoff, in inches, same vegetation, infiltration capacity reduced by one-half, Soil Conservation Service runoff curve 60	Factor of incre in storm runo
5.0	0.24	1.30	5.
6.0	.50	1.92	3.
7.0	.85	2.60	3.

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