Theoretical Implications of Underfit Streams

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By G. H. DURY

GENERAL THEORY OF MEANDERING VALLEYS

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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

A technique of defining discharge at the natural bankfull stage, upon rivers subjected to artificial banking and dredging, is applied to extending the data available on the relation between amount of discharge and wavelength of meanders. The new data confirm that wavelength varies with the square root of discharge and support the contention that bankfull discharge of manifestly underfit streams has been reduced by the square of the reduction shown by meander wavelengths.

Allowances, however, are necessary for additional changes, notably those in channel form and in downstream slope. The large channels occupied by former streams are thought to have a higher width-and-depth ratio than the present channels, thus giving, for instance, a 25:1 ratio of cross-sectional area where the bed-width ratio is 9:1 or 10:1. If velocity through the former cross section at bankfull stage were identical with present-day velocity at the corresponding stage, then the discharge ratio for highly underfit streams would be about 25:1 instead of 80:1 or 100:1 as previously suggested by the writer. But reduction to manifest underfitness involves a reduction of downstream slope on account of lengthened trace, for which infilling of headward valleys or excavation of downstream reaches do not compensate. The reduced slope, in turn, involves reduction in velocity, which considerably offsets the change in channel form. The net outcome of revised calculations is that a discharge ratio of about 50:1 or 60:1 is required where streams are highly underfit—that is, where the wavelength ratio is 9:1 or 10:1—and a discharge ratio of about 20:1 is required where the wavelength ratio is about 5:1. This last ratio is widely represented in nature.

Inquiry into the possible hydrologic effects of climatic change takes into account the temperatures reconstructed for full-glacial times and applies them in transformations of the empirically determined interrelation of temperature, precipitation, and annual runoff. In conjunction with increases in total precipitation by a factor of 1.5–2.0, the inferred temperature changes are capable of increasing annual runoff by factors in the approximate range of 5.0–10.0 in a wide range of existing climates. Computed proportional increases in runoff rise toward increasingly dry and increasingly warm climates.

Temperature change alone is not sufficient to explain the observed morphological effects, especially in view of the dating of certain episodes of channeling and of the assigning of initiation of valley meanders to parts of the deglacial succession when the trough of low full-glacial conditions had already been passed and when air temperatures were distinctly rising.

Frozen ground, whether seasonally frozen or permafrost itself, cannot 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 framework 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 location 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 sedimentation.

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 of former discharges refer throughout to discharge at bankfull stage.

EMPIRICAL CONNECTION BETWEEN WAVELENGTH AND DISCHARGE

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

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

Formidable difficulties surround the collection of concurrent data on meander wavelength and on bankfull discharge. Because many streams have been artificially embanked, the natural bankfull stage is hard to define. However, a possible technique of definition, based on minimal discharge and drainage-area values, has been developed in a study of the White and Wabash Rivers (Dury, 1961). The technique uses discharge and area values for flow at reported bankfull, flood, or flood-damage stage and also uses estimates of channel capacity. When the various discharge and area values are plotted, they form irregular clouds; lines drawn for the bases of the clouds are taken as regional graphs of discharge at natural bankfull stage.

The relevant diagram for the White and Wabash Rivers is here reproduced as figure 1. Corresponding graphs for the Red River of the North and for the Sheyenne River, for a group of rivers in the northeast Ozarks and for the Salt River of Missouri, and for members of the Alabama River system constitute figures 2–4. Bankfull discharges, read off against areas for which meander wavelength is known, are listed in table 1 against the associated wavelengths.

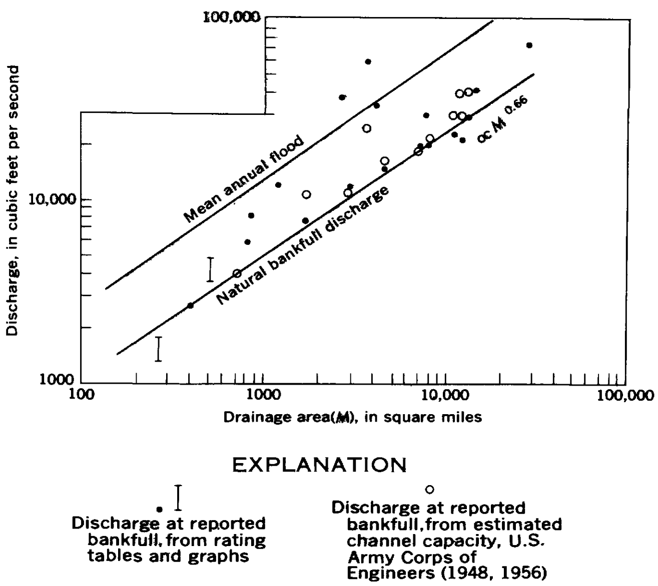


FIGURE 1.—Determination of discharge at natural bankfull stage, Wabash and White Rivers.

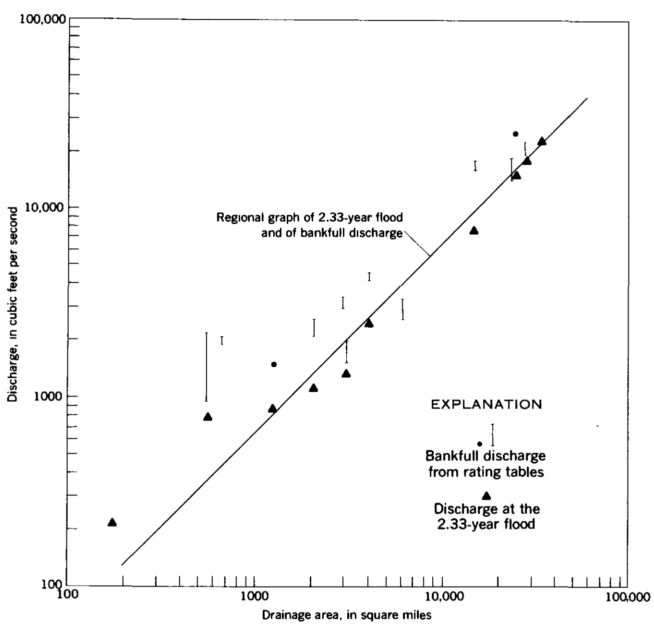


FIGURE 2.—Determination of discharge at natural bankfull stage, Red River of the North and the Sheyenne River.

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

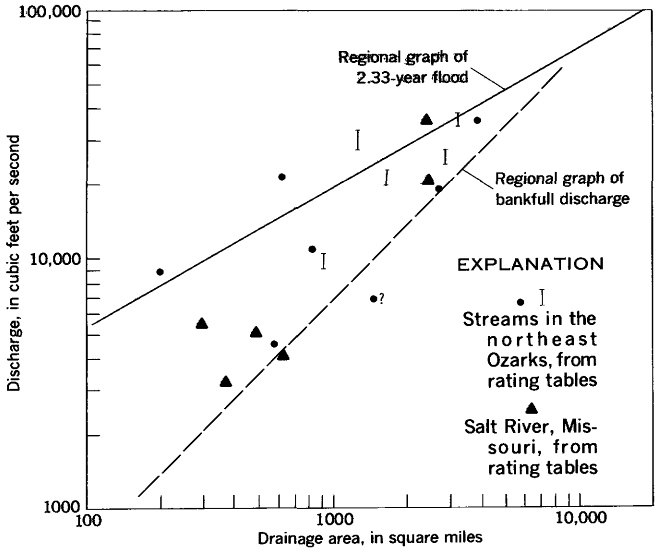


FIGURE 3.—Determination of disacharge at natural bankfull stage, rivers in the northeast Ozarks and the Salt River of Missouri.

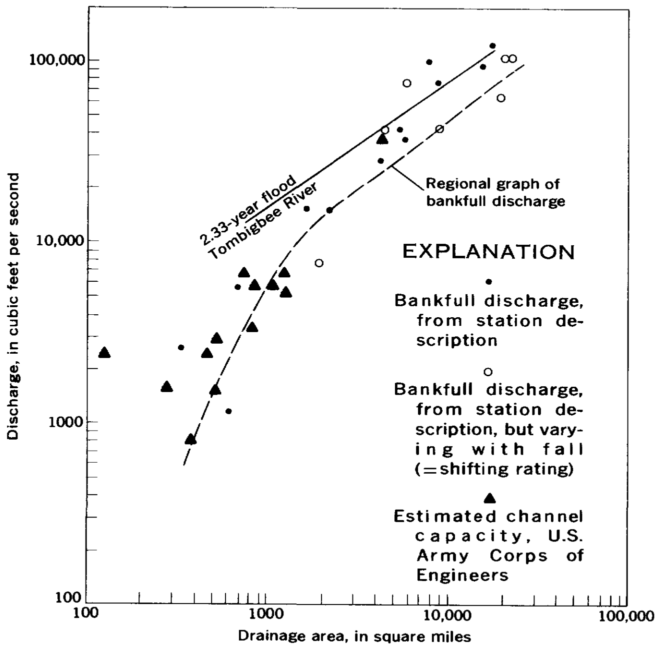


FIGURE 4.—Determinations of discharge at natural bankfull stage, members of the Alabama River system.

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 annual flood has a return period of 2.33 years; the mean annual and the 2.33-year floods are indeed taken as identical in many reports. Not uncommonly, floods of other return periods are expressed as multiples or fractions of the 2.33-year flood, so that their graphs on Gumbel paper run parallel to the graph of that flood. Downstream convergence of the pairs of graphs in figures 3 and 4 means a downstream increase in the return period of bankfull discharge; this downstream increase denies the convenient possibility that discharge at bankfull stage might be expressible as a fraction of discharge at the 2.33-year flood.

The circumstance is not surprising, for many headstreams are notoriously flashy, whereas channel storage tends to suppress peaks progressively in the downstream direction. Nor does a downstream increase in return period conflict with a downstream increase in total duration (Dury, 1961). Again, parallel graphs may appear for the middle and lower reaches of a given stream despite marked lack of parallelism for the headwaters. The curved graph of discharge and area, drawn in figure 4 for bankfull discharge on members of the Alabama River system, might well become parallel to the graph of the 2.33-year flood, if it could be extended downstream. In that event, bankfull discharge would have a fixed return period for part of the drainage. The rectilinear graphs drawn in figure 3 for streams in Missouri would, if projected downstream, eventually cross. But this would give bankfull discharge—quite anomalously for a humid region—a return period greater than 2.33 years. Here, also the regional graph of bankfull discharge ought somewhere to be inflected.

The type of relation here envisaged between graphs of discharge at fixed return period and graphs of discharge at bankfull stage obviously raises problems—of duration—that need further study. Meanwhile, the values of bankfull discharge obtained from the accompanying diagrams are both consistent among themselves and in agreement with values obtained by other workers. Figure 5 shows the 70 wavelength and discharge readings of table 1 plotted against the corresponding values assembled by Leopold and Wolman (1957, fig. 45 and Appendix E). The data of Leopold and Wolman, relating to observations on rivers and flumes by the two authors and by Friedkin (1945), Inglis (1940, 1949), Qraishy (1944), and Brooks and Eakin (unpub. data), ranged from a discharge of 0.021 cfs (cubic feet per second) for a meander model to 1 million cfs for the Mississippi River. Though the data from the present writer's observations are more closely grouped, they still range through more than three log

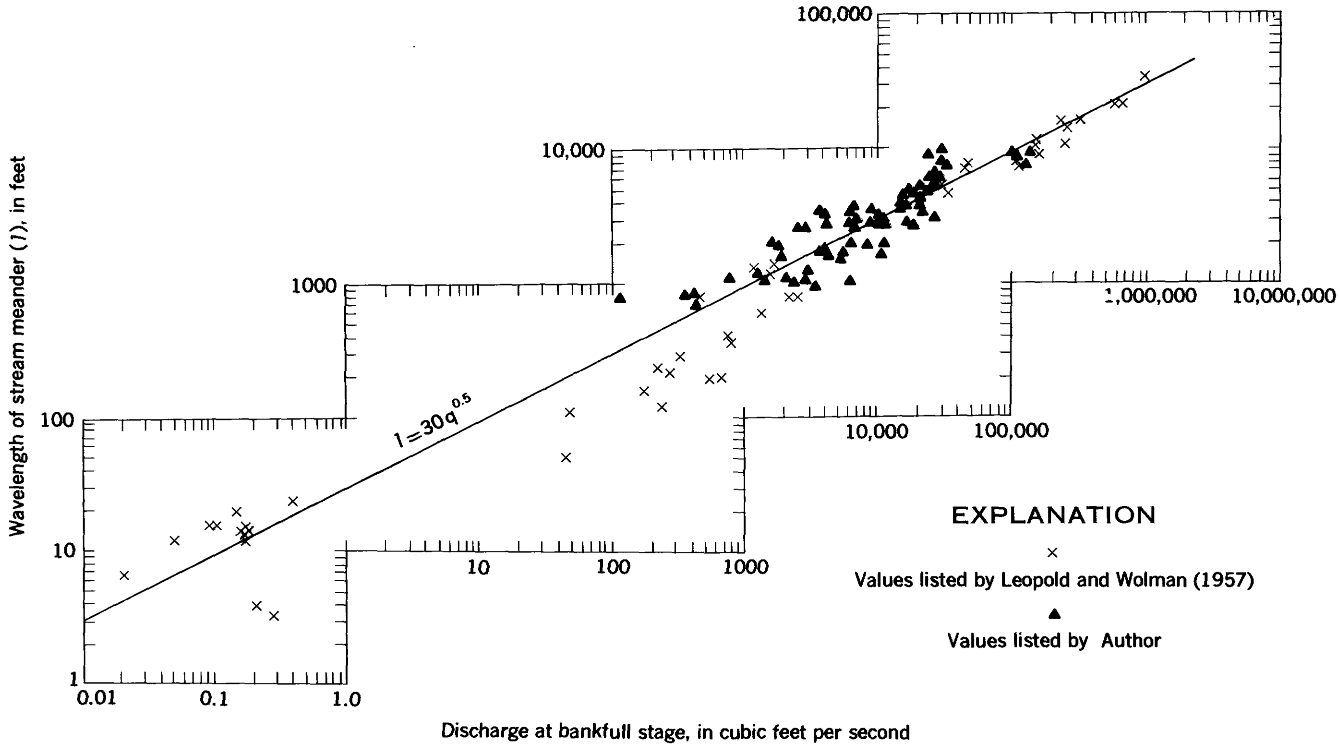


FIGURE 5.—Relation between meander wavelength and discharge at bankfull stage.

cycles on the scale of discharge and lie mainly in a gap left by the data of Leopold and Wolman. The two series of data accord well with one another—a particularly encouraging circumstance, in view of their lack of reference to sediment load, roughness, channel form, slope, and velocity.

Variations in load, roughness, and channel form are perhaps mainly responsible for the irregularity of scatter in figure 5, although the present writer's use of regional discharge and area values alongside specific values of wavelength may also contribute. Simons and Albertson (1960) suggested that regime equations generally may depend on the conditions on which they are based and that they may thus be valid only within the range of observed data. Specifically, these authors claimed to identify variations in wetted perimeter, hydraulic radius, and cross-sectional area according to variations in the material of bed and banks. (See also Schumm, 1960.) But if a general equation for wavelength and discharge is to be obtained, variations of this kind must be provisionally neglected. Slope and velocity may not, in one sense, be immediately relevant. Mean velocity through the cross section at the bankfull stage promises to vary little, if at all, in the downstream direction. If that is so, then changes of slope in the same direction may be omitted from the argument for the 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

$$l=kq^b. \tag{1}$$

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

$$l=36.1q^{0.47} \tag{2}$$

or

$$l=26.8q^{0.5} \tag{3}$$

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

$$l=6.06w^{0.99} \tag{4}$$

or

$$l=6.06w \tag{5}$$

and bed width to discharge in the form

$$w=4.88q^{0.5}. \tag{6}$$

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

$$l=29.6q^{0.5}, \tag{7}$$

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

$$l=36q^{0.5}, \tag{8}$$

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

$$l=30q^{0.5}; \tag{9}$$

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 wavelength alone. But the reductions are pronounced only if roughness is left out of account. Discharges computed from wavelength ratio appear to be of the correct order of magnitude, even though leaving considerable room for uncertainty.

The range of bankfull discharges and of meander wavelengths through which the relation $l \propto q^{0.5}$ holds is large enough to justify the assumption that $L \propto Q^{0.5}$, where L is the wavelength of former (valley) meanders and Q is the associated discharge at bankfull stage. Therefore

$$L/l = (Q/q)^{0.5} \tag{10}$$

and

$$Q/q = (L/l)^2; \tag{11}$$

in more general terms,

$$L/l = (Q/q)^b \tag{12}$$

and

$$Q/q \doteq (L/l)^{1/b}. \tag{13}$$

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

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

$$l = eM^f. \tag{14}$$

Where the discharge and area relation is known or can be defined, it usually takes the form $q_{rp} \propto M^g$, where q_{rp} is momentary peak discharge at fixed return period. Consequently, if bankfull discharge is assumed to have a fixed return period, then $l \propto q^{gf}$. The relation $l \propto q^{0.5}$ can still hold where f is less than 0.5, for g is commonly less than unity. For instance, if $q \propto M^{0.8}$ and $l \propto M^{0.4}$, then still $l \propto q^{0.5}$. However, analysis of wavelength and area and discharge and area relations on the English rivers Nene and Great Ouse (Dury, 1958, 1959) suggest that something more is involved. On these rivers, $q_{rp} \propto M$, but $l \propto M^{0.44}$. Therefore, if bankfull discharge has a fixed return period, then

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

Equation 11 then gives

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

since

$$L/l=9,$$

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

If, however,

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

then

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

The higher the value of b, the lower is the calculated ratio Q/q and the simpler is the problem of accounting for the discharges computed for the former meanders. A mere wish to simplify, however, carries no logical force.

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

$$l=eM^{f} \tag{14}$$

and

$$L=e'M^{f}, \tag{16}$$

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

$$L=e'M^{f'}, \tag{17}$$

where $f'<f$.

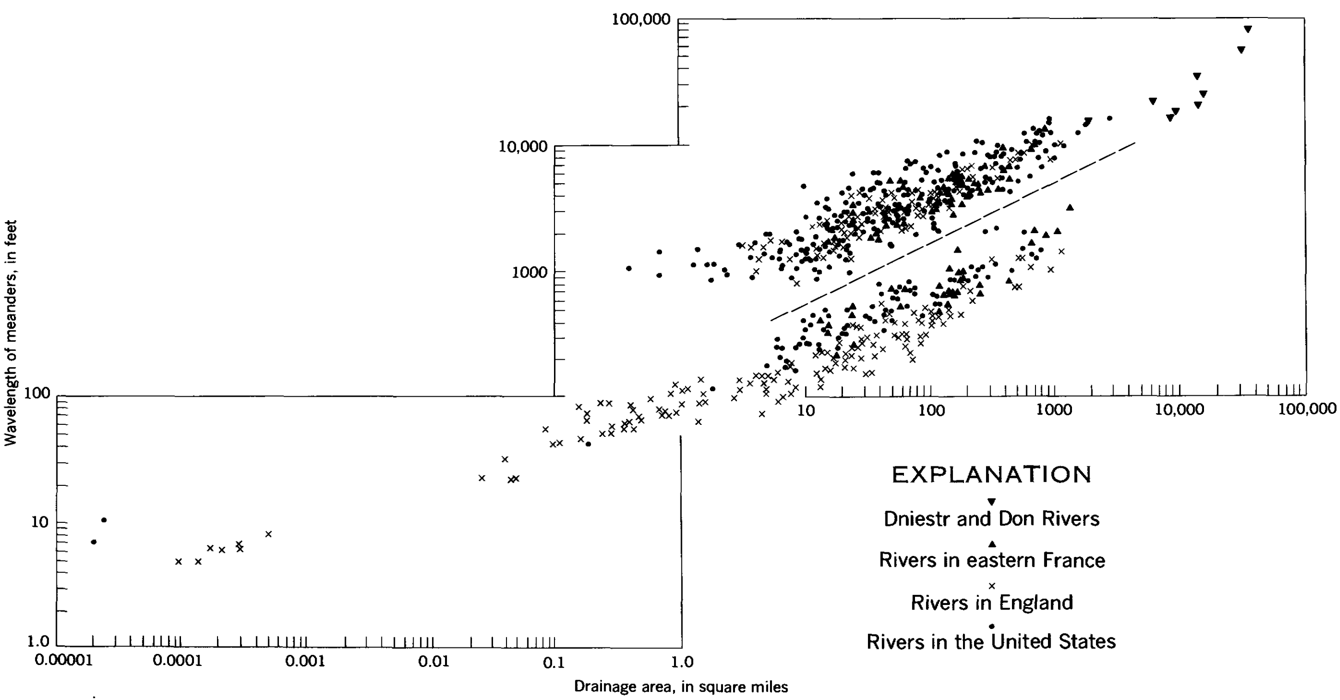


FIGURE 6.—Massed plots of wavelength against drainage area. Upper band, wavelength of valley meanders; lower band, wavelength of stream meanders.

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

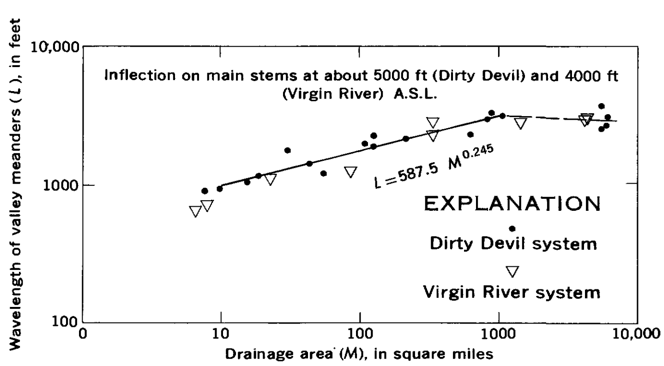


FIGURE 7.—Relation of wavelength of valley meanders to drainage area, Dirty Devil and Virgin Rivers, Utah.

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 above sea level. Increasing aridity, with change toward existing climates, probably shifted the zone of counterbalance vertically upward. Available maps for the two river systems unfortunately do not permit a comparison between former and present wavelengths. On the Humboldt River, Nev., and Owyhee River, Oreg. (fig. 8), the scanty evidence collected is not sufficient

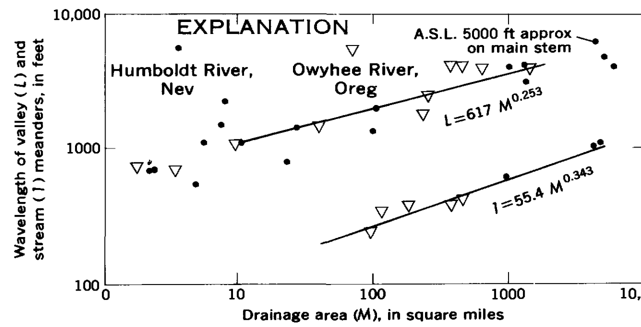


FIGURE 8.—Relation of wavelengths of valley and stream meanders to drainage area, Humboldt River, Nev., and Owyhee River, Oreg.

to show whether or not the graph of stream meanders becomes inflected at a level different from that of any inflection in the graph of former meanders. However, it seems highly likely that the change toward aridity would reduce the total area of contributing drainage, so that the effective basins are now smaller than they once were.

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

Where the whole of a given basin was formerly contributing, but parts of it now fail to contribute, the plot of former wavelengths against drainage area is correct. But the plot of present wavelengths should be adjusted by reductions in the values of area against which wavelengths are set, and the slope of the plot would consequently steepen. Even if the two wavelength graphs, as first drawn, ran parallel, they would then converge: the adjusted plot of present wavelenghts would ensure a downstream decrease in the ratio L/l.

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

$$l=85.2M^{0.46} \tag{18}$$

and

$$L=1,232M^{0.36} \tag{19}$$

that is, there is upward convergence (fig. 9). But, on the likely view that the former meanders were cut at

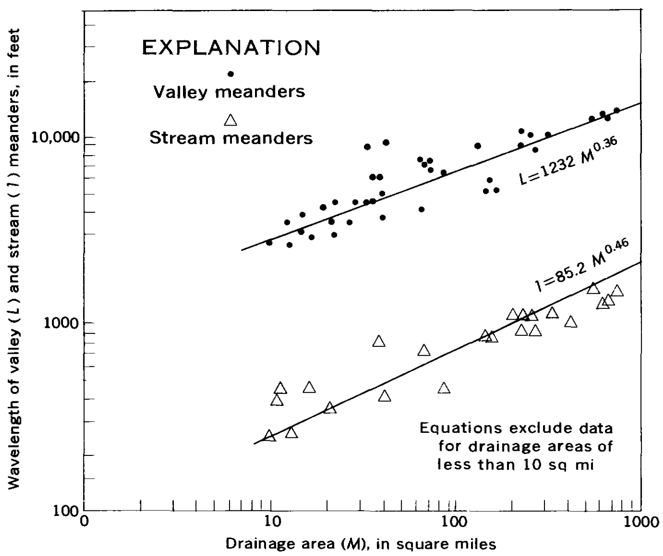


Figure 9.—Relation of wavelength of valley and stream meanders to drainage area, rivers in the Driftless Area of Wisconsin (Kickapoo, Platte, Galena, and Pecatonica Rivers).

times of great cold, there seems little scope for increasing the steepness of the graph of former meanders. Any allowance made with respect to present conditions, for noncontribution in narrow bands along the divides or for percolation, would steepen the graph of present meanders and would thus increase the downstream reduction in the ratio L/l. It is difficult to imagine that hydrologic contrasts between headwater and other parts of drainage areas were greater in former times than they are today, and hypothetical climatic contrasts—for example, heavy snow on high ground—are problematic, especially if they are required to account for downstream changes in stream behavior greater than the changes which now occur.

It therefore seems necessary to conclude that, in certain basins, a downstream reduction in the ratio L/l does occur, for whatever reason. The ratio Q/q, calculated from L/l, then also decreases downstream. Bankfull discharge can no longer be specified as having been reduced to a set fraction of its former value throughout an area, and interregional comparisons can no longer be made unless numerical values of area are given. Computed values of Q/q, derived from a variable ratio L/l, are increased above the regional value in the upper basin and reduced below it in the lower basin.

Discharge ratios between former and present streams may therefore be computed in one of two ways, each of which is open to some objection. The regional values of wavelength, as specified by best-fit equations, may be accepted, and discharges may be computed for them by means of the general wavelength and discharge equation 9. In this event, discharges must be referred to specific areas—that is, near the upper and lower ends 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

$$q = (l/30)^2 \quad (20)$$

and

$$Q = (L/30)^2 \quad (21)$$

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.

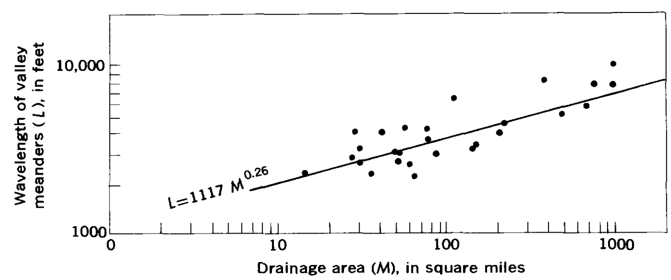


Figure 10.—Relation of wavelength of valley meanders to drainage area, in part of Minnesota.

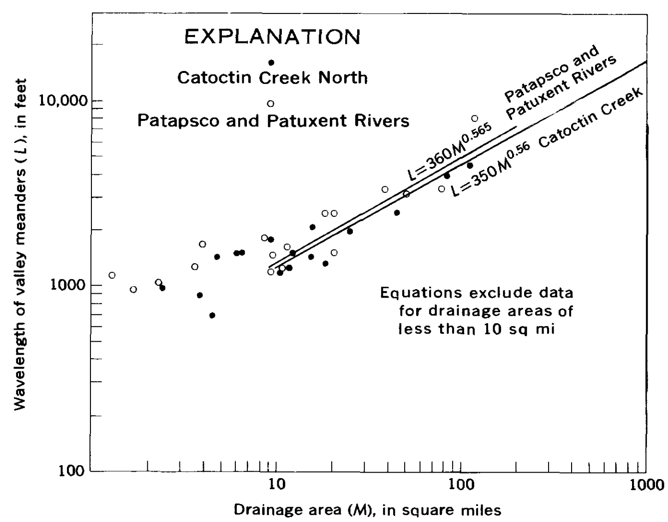


Figure 11.—Relation of wavelength of valley meanders to drainage area, Patapsco and Patuxent Rivers and Catoctin Creek, Md.

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,

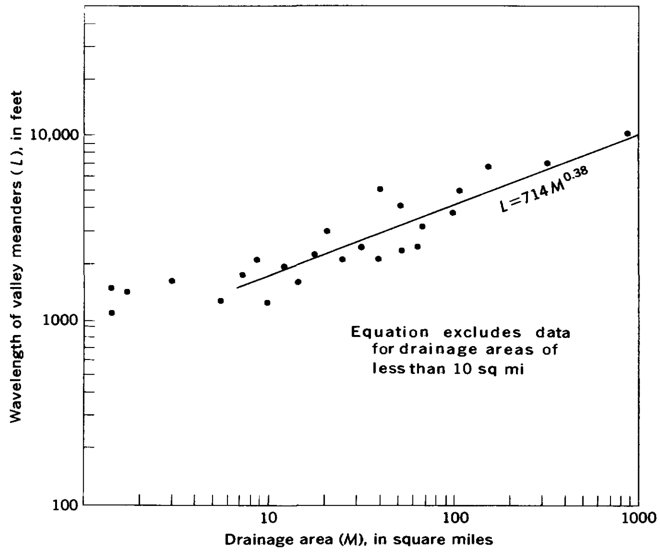


FIGURE 12.—Relation of wavelength of valley meanders to drainage area, Monocacy River, Md.

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 the relation of bankfull discharge to discharge at fixed return period. At the same time, the difference between f' and 0.5, in the expression $L \propto M^{f'}$, is so well marked that a wholesale conversion of best-fit graphs for former meanders to the general form $L \propto M^{0.5}$ is difficult to justify. The values obtained from such conversion are here required merely for discussion.

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

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: e	Steam meanders: f	Steam meanders: q, in cubic feet per second per square mile	Valley meanders: e'	Valley meanders: f'	Valley meanders: Q, in cubic feet per second per square mile	Q/q
Nene and Great Ouse Rivers, England	1,000	59	0.44	1.7	708	0.38	106	63
	10			2.9			320	110
Driftless Area, Wis	1,000	85	.46	4.6	1,232	.36	243	53.0
	10			7.0			927	132.0
Green Bay area, Wisconsin	100	68	.54	7.4	338.5	.54	184	25.0
	10			6.1			153	25.0
Southern New England	1,000	58	.56	8.5	260	.56	172	20.0
	10			4.9			99	20.0
Green River, Ky	500	135	.41	6.6	663	.41	160	24.0
	10			13.4			325	24.0
Eastern France	1,000	49	.53	4.0	400	.50	178	44.5
	10			3.8			178	47.0
Humboldt River, Nev., and Owyhee River, Oreg	1,000	55	.34	.4	617	.25	13.4	33.5
	10			1.6			13.4	84.0
Cotswolds, England	100	-----	-----	-----	680	.37	155	-----
	10						282	
West side of glacial Lake Agassiz area	1,000	-----	-----	-----	1,155	.30	94	-----
	10						618	
Minnesota, west of Driftless Area	1,000	-----	-----	-----	1,117	.26	50	-----
	10						460	
Mission River, Texas	1,000	-----	-----	-----	176.5	.62	181	-----
	10						60	
Catoctin Creek and Patapsco and Patuxent Rivers, Md	100	-----	-----	-----	355	.56	243	-----
	10						185	
Monocacy River, Md	1,000	-----	-----	-----	714	.38	108	-----
	10						326	
Dirty Devil and Virgin Rivers, Utah	1,000	-----	-----	-----	587.5	.245	18	-----
	10						119	
Deep Spring Valley, Calif	100	-----	-----	-----	373	.25	16	-----
	10						49	

thought to be much more nearly representative of widespread conditions. As will be now demonstrated, moreover, the whole series of discharge ratios obtained from wavelength ratios is susceptible of reduction.

TABLE 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	
Nene and Great Ouse Rivers, England	41	1.8	355	125	68.5
Driftless Area, Wis	66	4.9	515	295	60
GreenBay area, Wisconsin	79	7.0	397	175	25
Southern New England	84	7.8	377	158.5	20
Green River, Ky	82	7.5	403	180	24
Eastern 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 channels, calculation can involve cross-sectional area, wetted perimeter, and downstream slope. In Manning's equation, let v, n, r, s, and a be, respectively, the mean velocity (at bankfull stage), the roughness, the hydraulic radius, the downstream slope, and the cross-sectional area of the present stream, and let V, N, R, S, and A be the corresponding values for the former stream. Then, since,

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

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

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

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

But since $Q=VA$, and $q=va$, then

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

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

Professional Paper 452-B notes that, on becoming manifestly underfit, a stream lengthens its trace; simple calculation shows that, on many streams, infilling on the headwaters could not possibly compensate for the reduction of slope involved in a lengthened trace, and field observation reveals that the tendency 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

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

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

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

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 bed-width 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 by which trace lengthened
Pembina	North Dakota	Cavalier (1:62,500)	2.19
Coln	Gloucestershire, England	SP/10 (1:25,000)	1.45
Pecatonica	Wisconsin	Mineral Point (1:24,000)	1.44
Pecatonica, East Branch	do	South Wayne (1:62,500)	1.42
Little Eau Pleine	do	Marshfield (1:62,500)	1.33
Cherwell	Oxfordshire, England	SP/43 (1:25,000)	1.33
Windrush	do	SP/21 (1:25,000)	1.33
French Broad	North Carolina	Brevard (1:24,000) Rosman (1:24,000)	1.32
Little Platte	Wisconsin	Rewey (1:24,000)	1.32
Black Earth Creek	do	Cross Plains (1:62,500)	1.31
Pecatonica	do	Mifflin (1:24,000)	1.30
Do	do	South Wayne (1:62,500)	1.29
Galena, Shullsburg Branch	do	New Diggings (1:24,000)	1.29
Pecatonica Mineral Point Branch	do	Mineral Point (1:24,000)	1.28
Great Ouse	Buckinghamshire, England	SP/62 (1:25,000)	1.28
Pecatonica	Wisconsin	Calamine (1:24,000)	1.25
Mounds Creek, East Branch	do	Blue Mounds (1:62,500)	1.23
East Pecatonica	do	Mifflin (1:24,000)	1.22
Elk Fork Salt	Missouri	Florida (1:62,500)	1.22
Avon	Warwickshire, England	SP/57 (1:25,000)	1.22
Platte	Wisconsin	Potosi (1:24,000)	1.21
Wye	Herefordshire, England	142 (1:63,360)	1.21
Lugg	do	142 (1:63,360)	1.20
Meramec	Missouri	Pacific (1:24,000)	1.17
Monnow	Herefordshire, England	142 (1:63,630)	1.16
Gasconade	Missouri	Morrison (1:62,500)	1.15
Nene	Northamptonshire, England	SP/65 (1:25,000)	1.15
Sowe	Warwickshire, England	SP/37 (1:25,000)	1.14
Galena	Wisconsin	New Diggings (1:24,000)	1.13
Eau Pleine	do	Stratford (1:24,000)	1.12
Meramec (part: see below)	Missouri	St. Clair (1:62,500)	1.10
Big	do	Cedar Hill (1:24,000)	1.10
Meramec (part: see above)	do	St. Clair (1:62,500)	1.07
Dry Fork	do	Meramec Spring (1:62,500)	1.06
Gasconade	do	Bland (1:62,500)	1.06
Middle Fork Salt	do	Florida (1:62,500)	1.06
Gasconade	do	Linn (1:62,500)	1.05
Bourbeuse	do	St. Clair (1:62,500)	1.03

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 that $b+c+d=1.0$. If b is taken at the general (average) value of 0.5, then $c+d=0.5$. The foregoing application of Manning's equation requires that, in the present context, $d>0.0$; hence, $c<0.5$. Therefore, where W and D are former bed width and former mean depth, respectively, it follows from $q \propto w^b \propto d^c$ that

$$Q/q=(W/w)^{1/b}=(D/d)^{1/c}, \tag{28}$$

so that

$$W/w=(D/d)^{b/c}, \tag{29}$$

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

Although specific values cannot be given for both the former and the present coefficients of roughness, a range can be suggested for the ratio n/N. If both the former and the present channels are taken as clean but as winding through pools and shallows, their roughness coefficients should lie between 0.033 and 0.045 (Horton's revisions, quoted by Linsley, Kohler, and Paulhus, 1949, p. 472). The higher the value of n/N—the greater the postulated increase in roughness with reduction in size of channel—the greater is the computed reduction in velocity and thus also in discharge. If N and n are identical, the ratio V/v computed from equation 24

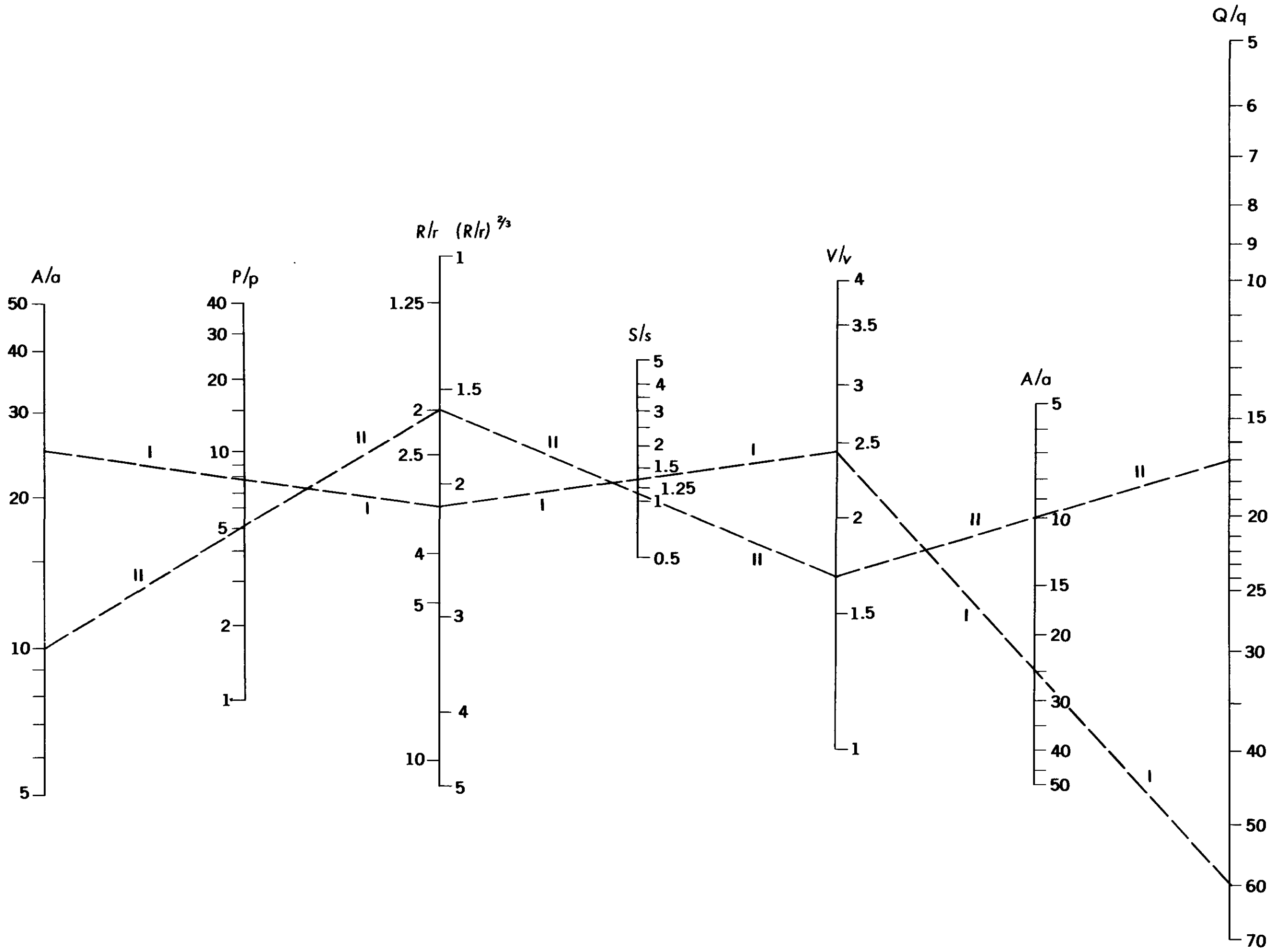


FIGURE 13.—Nomogram for Q/q, by Manning's equation, when cross-sectional area (A/a), wetted perimeter (P/p), and comparative downstream slope (S/s), are known for present and for former channels.

remains unchanged, and Q/q is unaffected. If n is taken as 0.045, at the top of the stated range, and N as 0.033, at the bottom, then computed values of V/v and Q/q are increased by a factor of about 1.36. The discharge ratio 60:1 then becomes about 80:1, and 17:1 becomes 23:1. The reductions in discharge ratio below the values computed from wavelength alone are not wholly restored. In figure 15, where discharge ratios are graphed against wavelength ratios, a band of probability is marked by stippling. Its lower limit corresponds to values of Q/q determined from equation 25, which admits no change in roughness, and the upper limit corresponds to an increase in roughness, from large to small channels, by a factor of 1.36.

Simons and Albertson (1960) inferred, from comparisons of channels in varying bed and bank materials, that differences in caliber and cohesiveness produce marked effects on channel form at a given discharge. Change of channel form with change of material is also incorporated 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

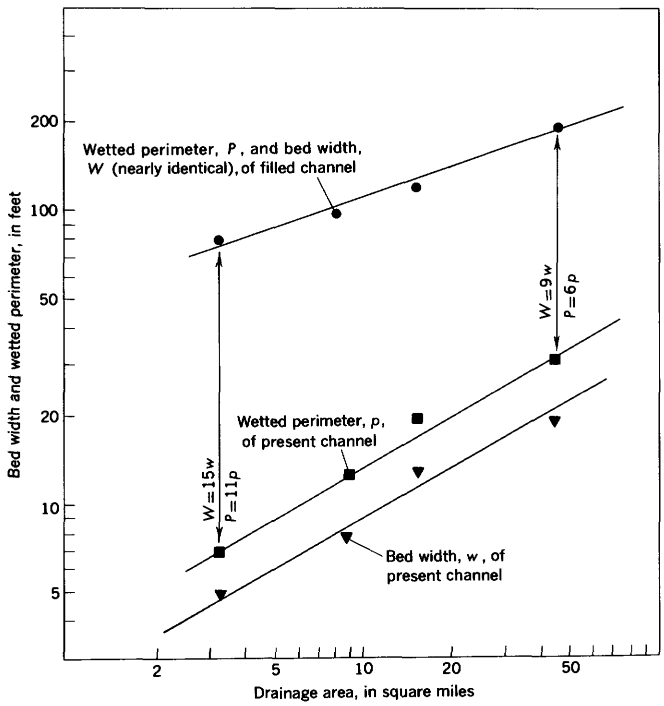


Figure 14.—Comparative bed widths and wetted perimeters on the River Itchen, Warwickshire, England.

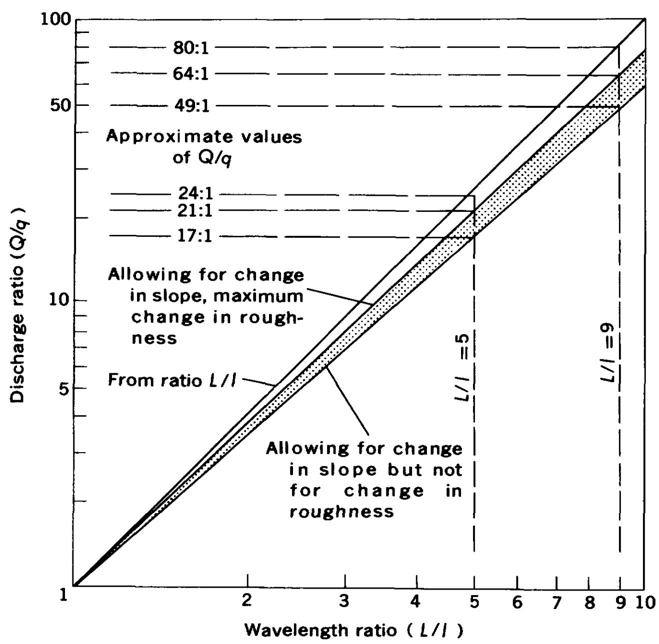


Figure 15.—Nomogram for relations of discharge ratio to wavelength ratio.

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 of p, w, P, and W. Several of the former channels, including all those described from England and from the Driftless Area of Wisconsin, are cut in bedrock, which ranges from compact shale through sandstone to well-cemented limestone. As a group, these channels cannot have undergone unusual widening on account of low cohesiveness of the bed and banks. Their wetted perimeters and their widths are likely to have been minimal rather than maximal. It is scarcely conceivable that materials of the present channels are more cohesive than materials of former channels in bedrock. If, therefore, $p=kq^b$ and $P=k'Q^b$, k' is most unlikely to be greater than k.

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

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

Support of another kind for this last inference comes from the findings of Brush (1959) and of the U.S. Waterways Experiment Station (1956). (See also Leopold and others, 1960.) Brush (1959) concluded that roughness increases markedly with sinuosity, and he suggested that the concepts of grain roughness and bed roughness be elaborated to include also bend roughness. The Technical Memorandum of the Experiment Station offered the conclusion that increase of sinuosity reduces channel discharge by about 8–10 percent when the sinuosity rises from 1.20 to 1.40 or from 1.40 to 1.57. Since the present channels of some manifestly underfit streams are distinctly more sinuous than the former channels, in which existing flood plains top the infill, former discharges computed from wavelength, slope, and generalized values of roughness can be somewhat low. Even where the present channels are not sinuous in plan, pool-and riffle sequences on the bed amount, in effect, to sinuosity. (See Leopold 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 van der Hammen and Gonzalez (1960), that in the same period mean annual temperature at Bogotá, Colombia, rose about 14°F.

The present interrelation of temperature, precipitation, and runoff in the conterminous United States was established in generalized form by Langbein and others (1949) and presented as a graph on which figure 16 of this report is based. This diagram can be used to show the likely effects on runoff of the reductions in temperature postulated for glacial maximum.

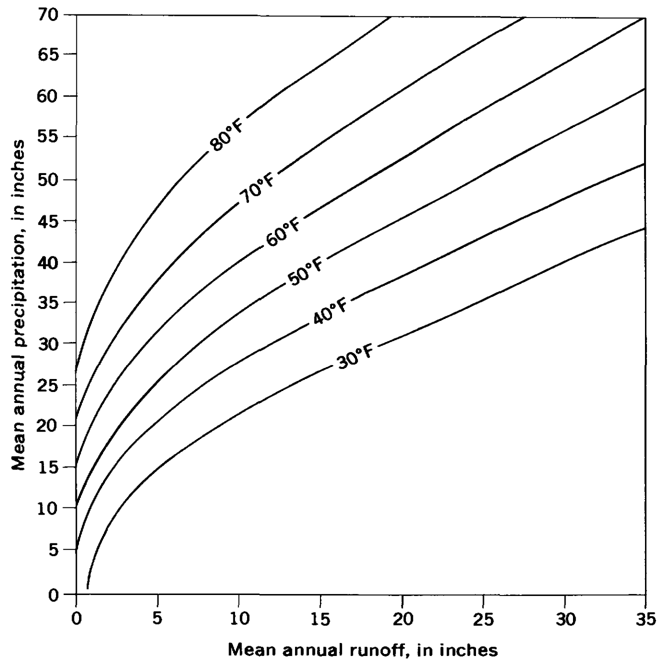


FIGURE 16.—Interrelation of precipitation, weighted mean annual temperature, and runoff. After Langbein and others (1949).

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

The mean annual temperatures used in figure 16 are not, however, mean temperatures in the ordinary sense. They are weighted to allow for seasonal distribution of precipitation, in the form

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

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

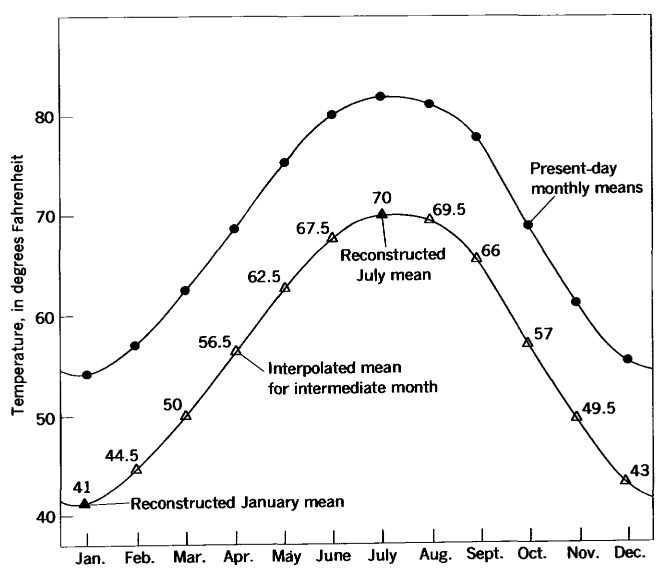


FIGURE 17.—Reconstruction of regimen of mean monthly temperature at the site of New Orleans during full-glacial conditions.

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,

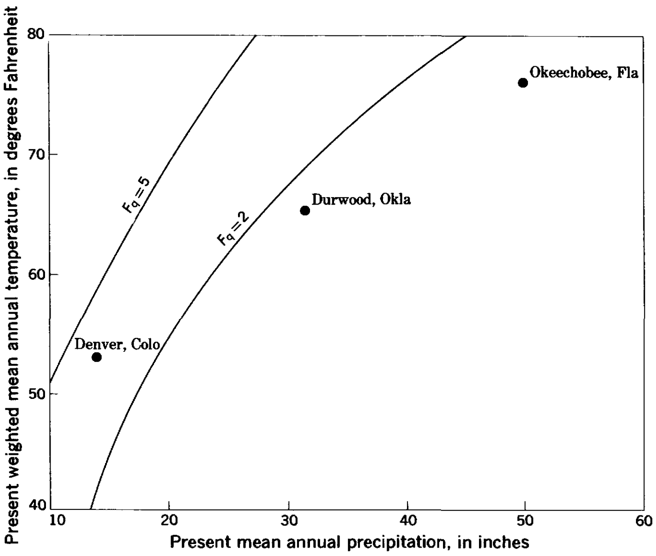


FIGURE 18.—Nomogram for results of a reduction of 10°F in weighted mean annual temperature with precipitation unchanged.

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.

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

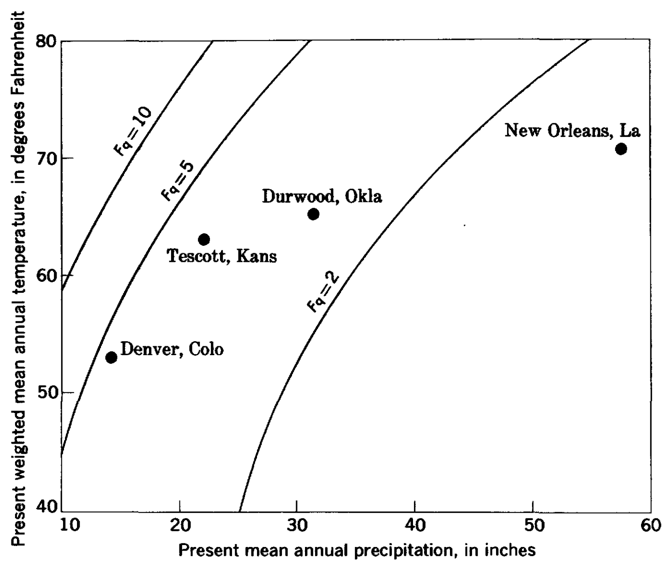


FIGURE 19.—Nomogram for results of a reduction of 15°F in weighted mean annual temperature with precipitation unchanged.

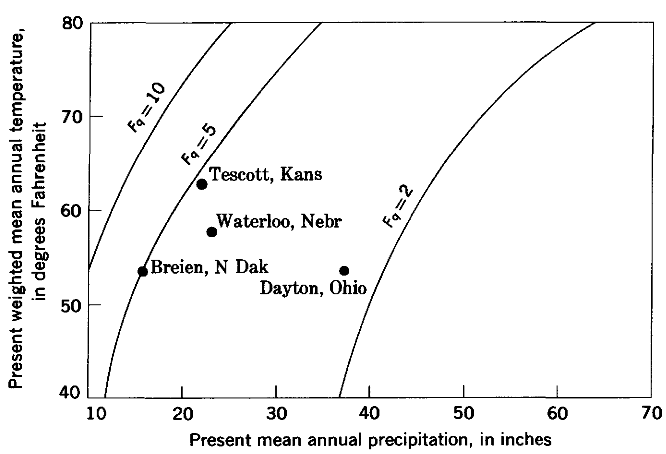


FIGURE 20.—Nomogram for results of a reduction of 20°F in weighted mean annual temperature with precipitation unchanged.

ordinary mean temperature, the effects of the difference upon indicated runoff are mainly small in comparison with the effects of reductions in ordinary temperature by the range 13°–27°F. The extent to which a change in seasonal regimen could affect the ratio F_q, between former and present total runoff, is illustrated by the hypothetical, and quite extreme, examples in tables 5 and 6.

The highest notional values of F_q are expectably, those associated with marked winter concentrations of precipitation at former times of low temperature. Reconstructed patterns of weather for times of glacial maximum, however, usually include summer concentrations of precipitation (Manley, 1955; Dillon 1956). Regions close to the former ice fronts may well have been overspread by the cold air of high-pressure

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
January	40	2	1	3	80	40	120	20	40	20	60
February	45	2	1	3	90	45	135	25	50	25	75
March	50	2	1.5	2.5	100	75	125	30	60	45	75
April	55	2	2	2	110	110	110	35	70	70	70
May	60	2	2.5	1.5	120	150	90	40	80	100	60
June	65	2	3	1	130	195	65	45	90	135	45
July	70	2	3	1	140	210	70	50	100	150	50
August	65	2	3	1	130	195	65	45	90	135	45
September	60	2	2.5	1.5	120	150	90	40	80	100	60
October	55	2	2	2	110	110	110	35	70	70	70
November	50	2	1.5	2.5	100	75	125	30	60	45	75
December	45	2	1	3	90	45	135	25	50	25	75
Year	55	24	24	24	55	59	52	35	35	39	32

systems during much of the winter. (See also Wright, 1961, and references therein.) Consequently, the high values of F_q derived from hypothetical former winter concentration of precipitation can probably be disregarded. This is not to deny that contrasts may occur between region and region, either in the type of former precipitation regimen or in the mode of relation of former to present regimen. Generally however, those moderate values of F_q associated with summer concentration of precipitation at times of glacial maximum are to be preferred.

CHANGES BOTH IN TOTAL AND IN REGIMEN OF PRECIPITATION

Although the several changes now under discussion were stipulated to involve reductions in temperature, the effects of increased precipitation will be considered, initially, as if temperature remained unaltered.

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

Some increase in precipitation is required by the hypothesis of increased cyclonic strength at glacial maximum (Willett, 1950; Dury, 1954). Dillon (1956) regarded increased cyclonic activity as having considerably enhanced the moistness of southeastern United States and as having greatly enhanced that of

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

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

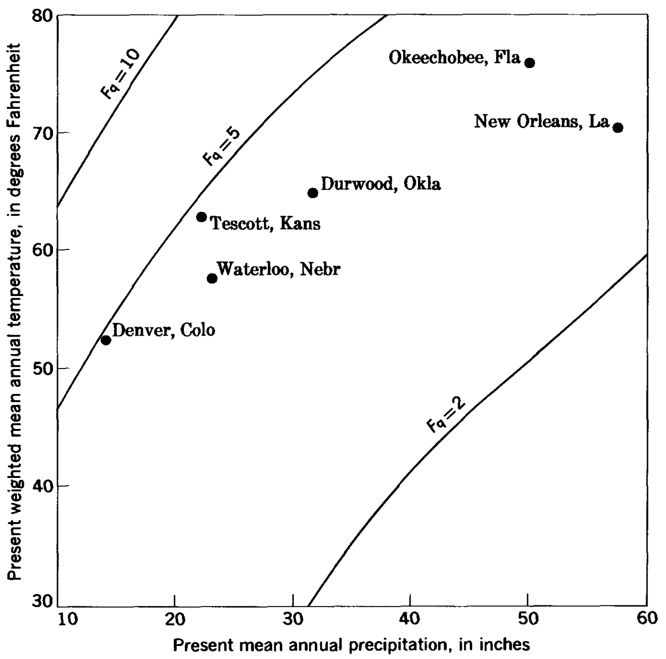


FIGURE 21.—Nomogram for results of an increase of 50 percent in precipitation with temperature unchanged.

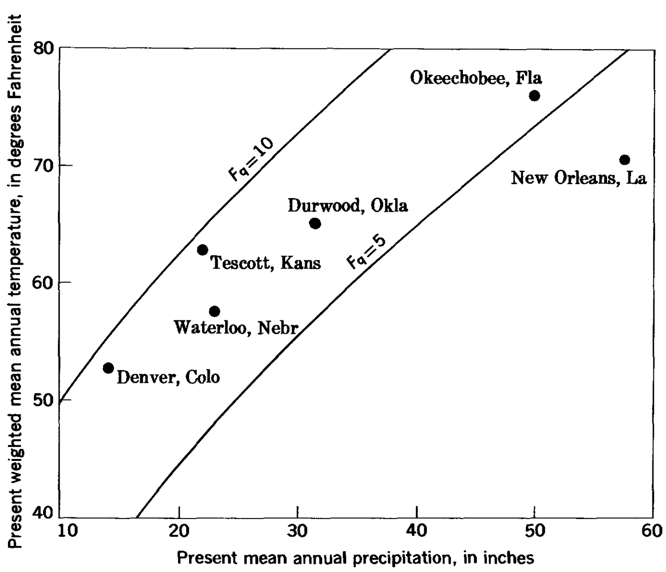


FIGURE 22.—Nomogram for results of an increase of 100 percent in precipitation with temperature unchanged.

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 increase in precipitation, is by a factor of three. But Black Earth Creek, with ratios L/l and W/w in the neighborhood of 10:1, ranks high on the scale of underfitness. The ratio Q/q, between former and present bankfull discharges, seems unlikely to be less than 50:1; even allowing a threefold increase in total runoff, a further sixteenfold increase in bankfull discharge has still to be accounted for. This increase seems difficult to explain without appeal to increased precipitation. Similar arguments apply to highly underfit streams which drain basins cut in rocks that are largely impermeable.

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

These regions, with their former pluvial lakes and their climate-sensitive ecology, are well suited to provide numerical estimates of precipitation change. Humid regions offer no comparable data, although they provide signs of humidity greater than present humidity. Ruhe and Scholtes (1956) inferred such humidity from massive late Wisconsin gley horizons in Iowa. (See also Ruhe, Rubin, and Scholtes, 1957; Lane, 1931.) Although gleying is equivocal, as pluvial lakes are equivocal, in that its requirements of humidity can be met partly by reduced temperatures, the dating of the gley soils of Iowa to the Cary-Mankato stades of the Wisconsin Glaciation is instructive. The soils formed shortly before the Two Creeks Interstade. It was precisely at this time, with deglacial temperatures already rising, that Lake Lahontan rose to levels well above those recorded at glacial maximum at about 20,000 years B.P. The very high stand, therefore, is referable to increased precipitation rather than to reduced temperatures, as observed in Professional Paper 452–B. Despite the cold fluctuation of Zone Ic (see table 15), the gleying in Iowa may also reasonably be ascribed, in part, to increased precipitation. In addition, a very general argument is possible, to the effect that climatic contrasts between dry and humid regions may have been at least as marked at times of glacial maximum and in early deglacial times as they now are. If so, an increase of 100 percent in the precipitation of humid regions is by no means incompatible with an increase of 50 percent in dry regions.

Conditions during the Atlantic phase (Zone VII, hypsithermal maximum) of deglacial time (see table 15) suggest that increases in the precipitation for humid regions by about 50 percent are not unlikely. The partial reexcavation of large channels at this time and the paleontological evidence of increased soil moisture must be reconciled with mean temperatures as much as 7° F higher than those of today. Transformations of figure 16 can readily be made to show that, merely to maintain present levels of mean annual runoff against such a temperature increase, precipitation would need to increase 10–15 percent. Since little is known of the dimensions of channels cut during Zone VII, bankfull discharges cannot be computed for them. Something, however, can be made of wavelengths of former rivers in the English fenland (Dury, 1964b). These wavelengths, in comparison with those of existing rivers and of valley meanders on upstream reaches, suggest that bankfull discharge in Zone VII times was locally ⅓ or ½ as great as that responsible for the main series of valley meanders and about 20 times as great as that of the present time. The indicated increase in precipitation is therefore much greater than the 10–15 percent capable of giving present-day values of runoff. The probably modest assumption that mean annual runoff during Zone VII was double that of today entails increases in mean annual precipitation by 33–50 percent—that is, increases of the order postulated for high-glacial times.

As with changes in temperature, so with changes in precipitation: the indicated proportional effects increase directly with temperature and inversely with total precipitation. In part of the range of possible climates, the computed results of a doubling of precipitation are at least as great as those of a temperature reduction of 20°F. (See figs. 18–22.)

Marked variations in seasonal concentration of precipitation have little effect on the values of F_q for an assumed doubling of precipitation. For the temperatures and precipitation listed in table 5, doubling precipitation would increase the values about six times (table 7). A change from the summer concentration, shown in table 5, to the winter concentration, shown in the same table, but with monthly totals doubled in addition, would raise F_q only to about 8.5. The objections to a former winter concentration, which have been stated above, pertain here also. In any event, the

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

influence of the change in annual total is far more significant than any reasonable change in annual regimen.

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

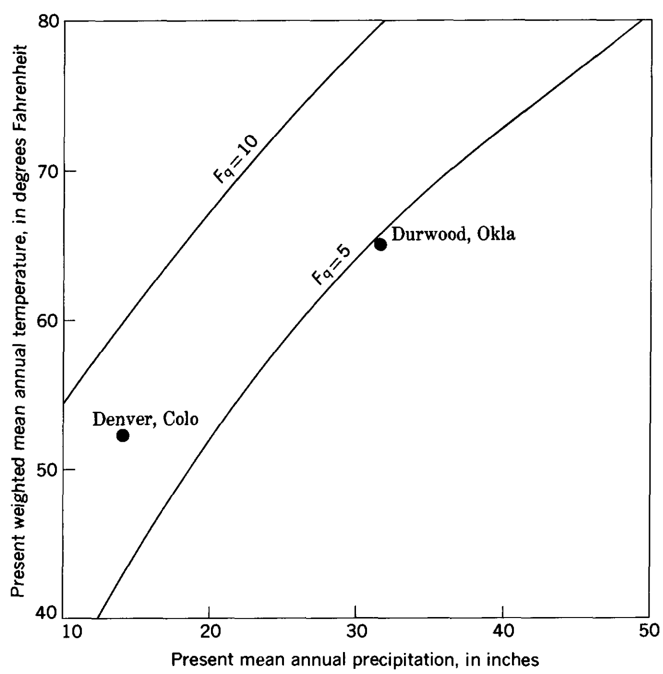


FIGURE 23.—Nomogram for results of an increase of 50 percent in precipitation plus a reduction of 10°F in weighted mean annual temperature.

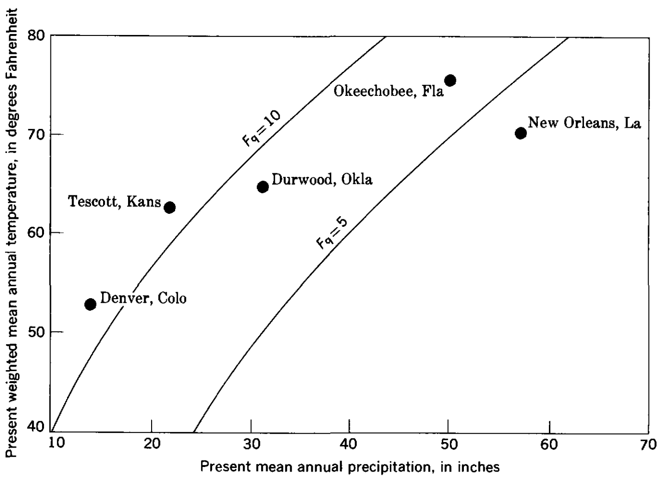


FIGURE 24.—Nomogram for results of an increase of 100 percent in precipitation plus a reduction of 10°F in weighted mean annual temperature.

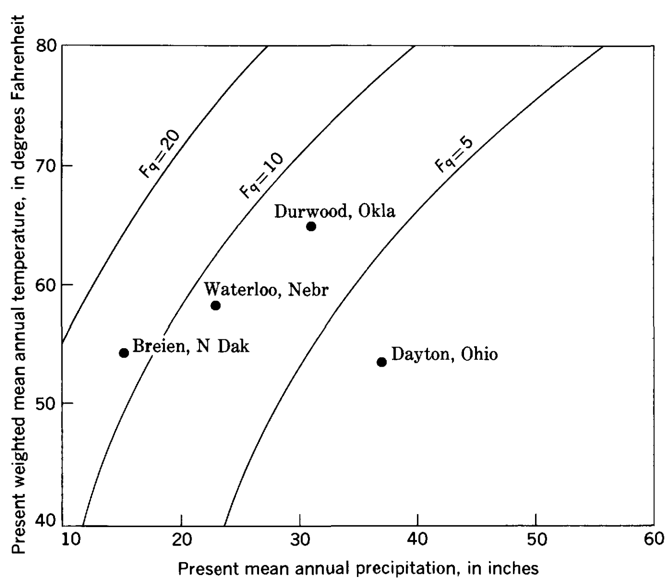


FIGURE 25.—Nomogram for results of an increase of 50 percent in precipitation plus a reduction of 20°F in weighted mean annual temperature.

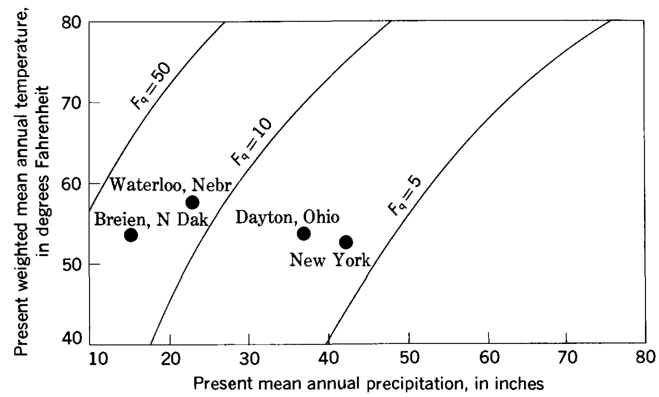


FIGURE 26.—Nomogram for results of an increase of 100 percent in precipitation plus a reduction of 20°F in weighted mean annual temperature.

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

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

Part of the immediate difficulty is due to the forms of the individual graphs in figure 16, which show no runoff for certain high temperatures and low values of precipitation. The general map, which is in Langbein and others (1949, pl. 1), allows little scope for no mean annual runoff. Therefore, since the accompanying nomograms are based on figure 16, albeit with some adjustments in the range of very high temperatures and very low precipitation, the very highest values of F_q which they indicate should be taken with reserve.

For all that, any hypothesis that required the effects of greatly reduced temperatures in dry regions to be offset by compensatory changes in precipitation would be of questionable merit. Far more needs to be known of the regional variation in the ratio L/l before the matter can be further examined. Particularly is this so because of the apparently irregular occurrence of high values of L/l in humid regions. Dry regions may present distinctive problems of their own, including reductions in stream density such as can scarcely have influenced regions which were, and remain, humid. Pending additional work, the fivefold to tenfold increase in mean annual runoff in humid regions will here 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 precipitation, in addition to a reduction in temperature greater than that inferred by Manley for glacial maximum. The extra reduction in temperature may be accepted, however, in compensation for the possible extra severity of southward bursts of cold air at glacial maximum, when weather gradients were unusually steep. If, accordingly, the Mission River basin was formerly penetrated by frost to the depth of 40 inches now recorded in Wisconsin and if frost was responsible for promoting the discharges which cut the valley meanders of the Mission River, then these meanders should be similar in wavelength, area for area, to the stream meanders of Wisconsin.

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

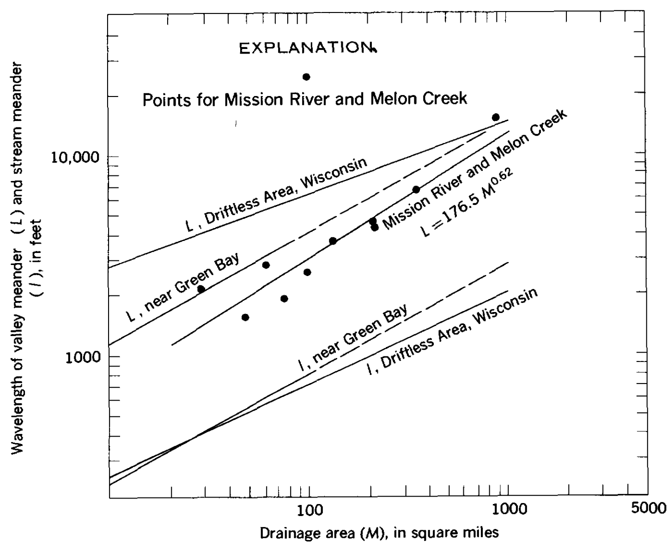


FIGURE 27.—Relation of wavelength of valley meanders to drainage area of Mission River, Tex., in comparison with other drainage areas.

valley meanders of the Mission River are three times as long as the stream meanders of the Driftless Area, and at 1,000 square miles they are six times as long. Whereas the valley meanders of the Driftless Area are nearly four times as long as those of the Mission River at 10 square miles, those of the Mission River become the longer, by extrapolation, at 1,650 square miles. Extrapolation of the graphs for the Green Bay country indicates that, at 1,000 square miles, the valley meanders of the Mission River are 4.5 times as long as the stream meanders near Green Bay, whereas at 1,650 square miles, the valley meanders of the Green Bay country are but 15 percent longer than those of the Mission River. The valley meanders of the Mission River obviously belong to the group which includes valley meanders of other 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

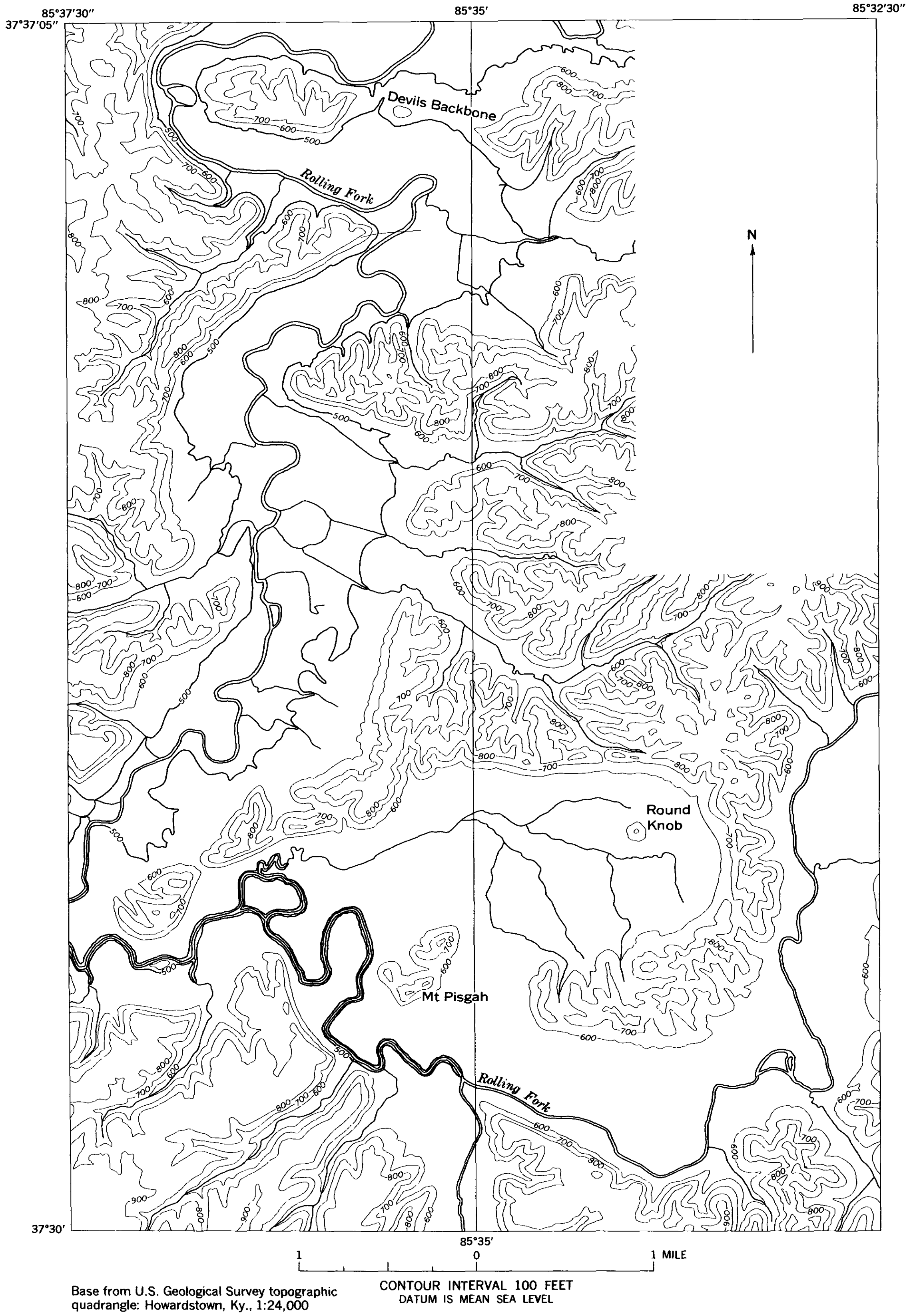
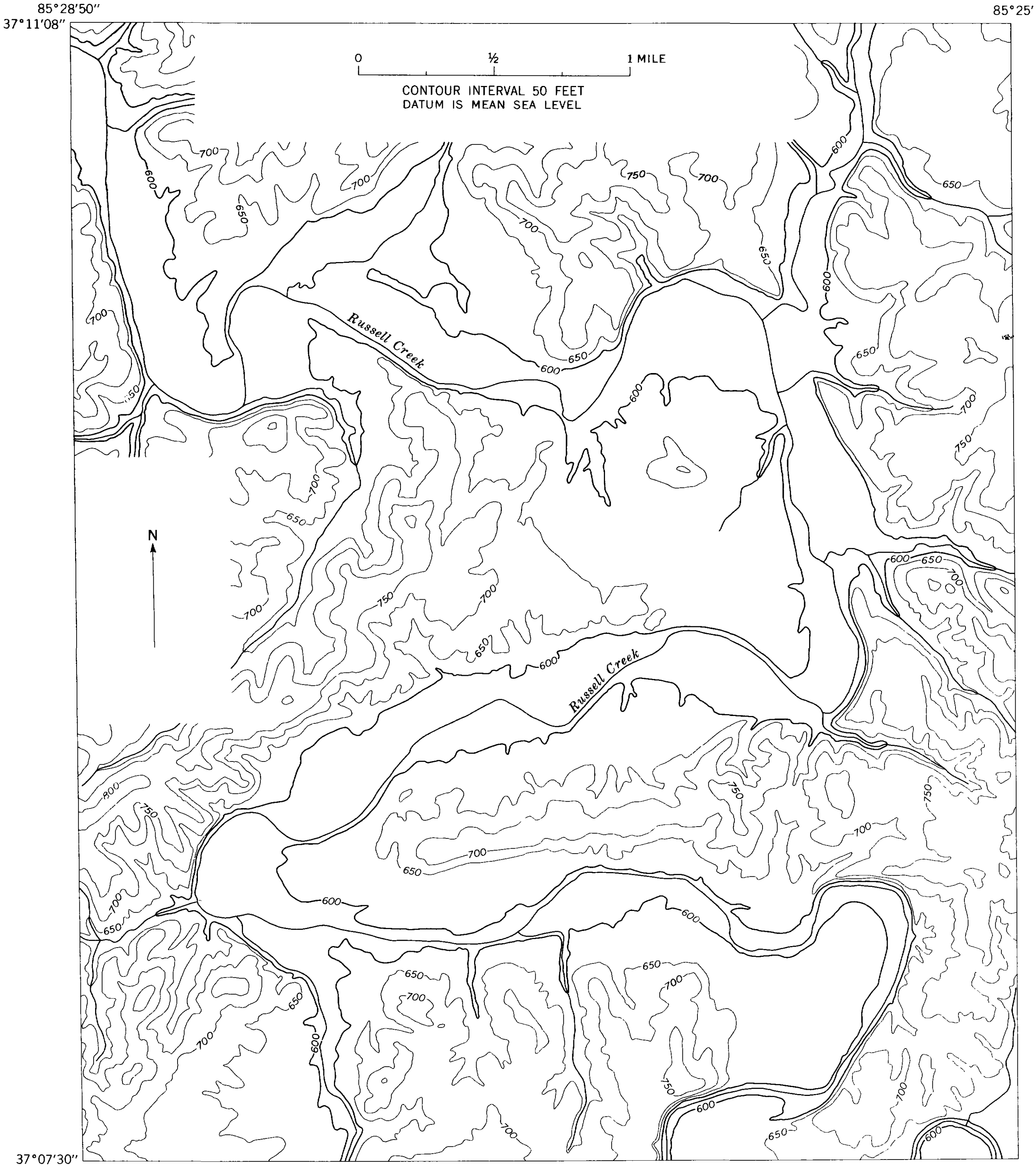


FIGURE 28.—Rolling Fork, Ky., an example of a manifestly underfit stream.



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

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

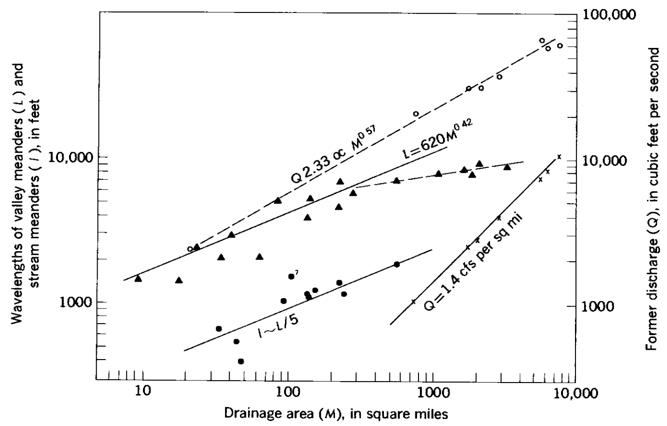


FIGURE 30.—Relation of wavelengths of stream and valley meanders to drainage area and relation of former discharge to drainage area, Green River system, Kentucky.

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), runoff is copious and base flow is high. Many stations record two maximums of runoff: one promoted by rainfall in October and November and a second in May and June that in part reflects melting. At glacial maximum, the climatic and hydrologic conditions of this area were presumably translocated to southern British Columbia and the Pacific Northwest; but it may not be unreasonable to suggest a comparison with New England during the early waning of glaciation, since New England, like the Alaskan Panhandle, is strongly influenced by cyclones and was probably so influenced when the receding ice front still lay close to its shore.

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

If the comparison of southeastern Alaska with late-glacial New England has any force, it suggests one possible reason for the modest wavelength ratio between valley meanders and stream meanders in New England: even if the regimen of precipitation was modified in this region at glacial maximum, seasonal variation in runoff may not have been extreme.

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

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.	Sept.
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	13. 1
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	1. 63
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	4. 80

[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 recorded), averages 3.8 times average discharge for months involved.

with melting. Annual precipitation of about 17.5 inches over the basin is associated with discharges of more than 2.5 cfs per sq mi in July and August and with a mean annual flood of 3.66 cfs per sq mi. Because the basin of the Tanana above Big Delta is very much larger than that of Fish Creek above Ketchikan, direct comparisons between their respective momentary and monthly peak discharges are inadmissible. The fact that the Tanana, at the rate of 3.66 cfs per sq mi, would require 120 days to discharge all the precipitation on its basin may reflect the results of channel storage and basin storage rather than the results of other factors.

Mean monthly midwinter temperatures in the Tanana basin above Big Delta fall considerably below those reconstructed for the north-central conterminous United States at glacial maximum, but midsummer temperatures perhaps run a little high. The concentration of runoff in the summer season that is indicated, for instance, by a hypothetical translocation of the climate of the Tanana basin to the Driftless Area may therefore be slightly exaggerated. But somewhat more general considerations support the hypothesis that a summer concentration of runoff did in fact occur close to the former ice fronts of the interior conterminous United States. The Alaskan plateau is a source region for cold polar continental air in winter, whereas in summer it is affected by an extension of the polar front from the Bering Sea and by single traveling lows. At glacial maximum the central part of the conterminous United States may equally have been overlain by cold continental air in winter, receiving, however, lows during summer, especially in view of the steep energy gradient between the ice fronts and the Gulf of Mexico. On climatic grounds, therefore, a marked summer concentration of runoff can be postulated for the interior at glacial maximum.

Permafrost brings its own special influences to bear on regimen of runoff, the effects being exemplified by the Kruzgamepa River on Seward Peninsula (table 9). There, with a mean annual precipitation of about 15 inches and with perennially frozen ground seeming to offset the effects of nearness to the ocean, runoff is very strongly concentrated in the open-water season. Mean discharge for the Kruzgamepa near Iron Creek in the period 1906–11 rose to nearly 13 cfs per sq mi in June—more than three times as high as the 2.33-year momentary peak on the Tanana at Big Delta. At 13 cfs per sq mi, the Kruzgamepa could discharge all the precipitation on its basin in about 30 days. Unfortunately, no more than two momentary annual peaks are on record. For what they are worth, they suggest a momentary peak rate of flow about four times as great

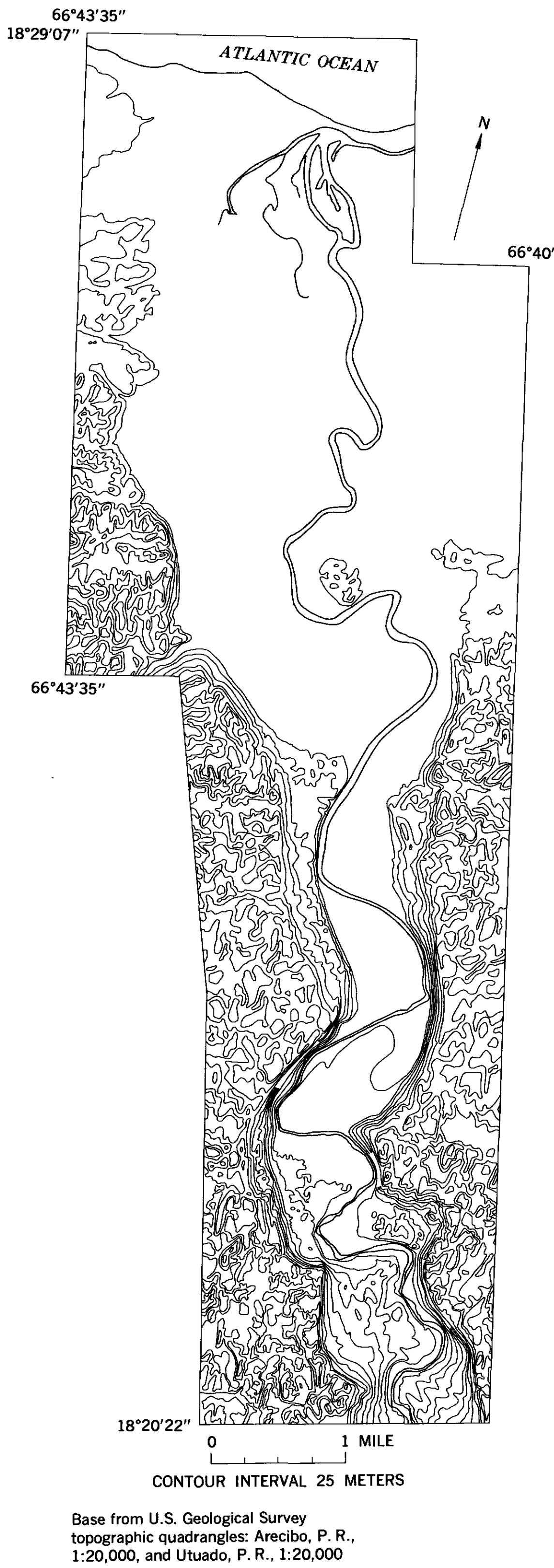


FIGURE 31.—Rio Grande Arecibo, P. R., showing contrast between valley and stream meanders. An example of manifest underfitness.

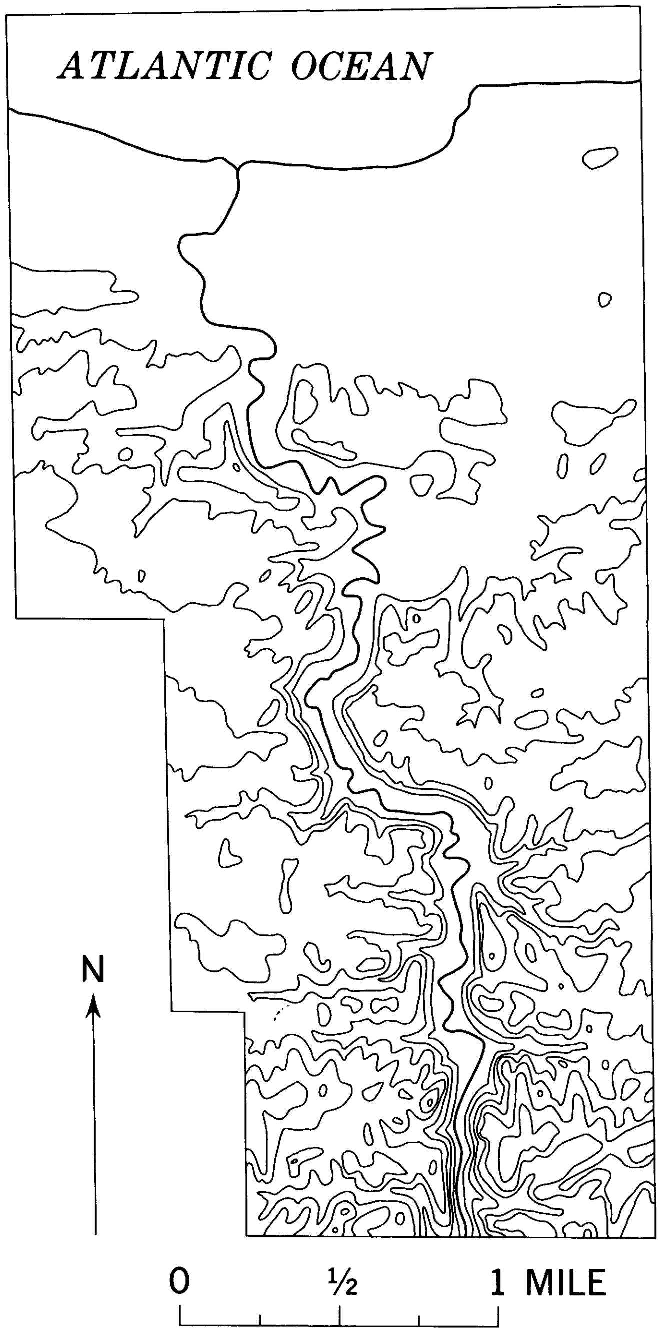


FIGURE 32.—Rio Camuy, P. R., an example of manifest underfitness.

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 outside the present limits of permafrost. Similarly, the seasonal concentration on the Tanana is well below that observed at many inland stations south of the 49th parallel (table 10; see also Harbeck and Langbein, 1949). If the necessary information were available, it would be desirable to make comparisons not only of peak monthly mean discharges but also of momentary peak discharges. However, even without such information, reasoning in general terms is possible. For example, the percentage of total runoff discharged during the peak month is about 32 percent on the Kruzgamepa and 16 percent on the Pecatonica at Freeport (table 10). If bankfull discharges were similarly affected, then translocation of the climate of the Kruzgamepa basin to the basin of the Pecatonica would insure no more than a twofold increase. If bankfull discharge on the Kruzgamepa is taken at the exaggerated value of four times the mean for the peak month and that of the Pecatonica is taken at the unduly low value of mean discharge for the peak month, then translocation of climate could still not account for more than an eightfold increase of bankfull discharge in the Pecatonica basin. Such an increase automatically includes the effects of reduced temperatures. But even with the aid of grossly exaggerated assumptions, frost fails signally to provide the required increase of at least 20 and possibly more than 50 times in bankfull discharge. Despite the first attraction of its possibilities, frost once again fails to supply an explanation of the reconstructed discharges of meandering valleys.

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.

CHANGED REGIMEN OF RUNOFF

Discussion of certain types of change in regimen has largely been anticipated above. Summer concentration of precipitation has been shown to reduce annual runoff, and winter concentration of precipitation to increase it. But the proportional effects on total runoff obtainable even from a very marked winter concentration are small. Again, whereas increased winter concentration could perhaps be imagined for low latitudes during high-glacial times, it is difficult to apply to regions which formerly experienced frequent outspreading of cold air during the winter. Neither the available reconstructions of weather nor reference to cold climates of the present day suggests anything but former summer concentration both of precipitation and of runoff for many regions now typified by underfit streams. The effects of summer concentration of precipitation could have been largely offset by reduced temperatures, but the necessary allowance has already been made in the transformations of figure 16. And, as has been seen, the summer concentration of runoff in Alaska, powerfully influenced by severe winters and by summer melt, does not necessarily exceed that in parts of the conterminous United States.

Too little is known of the interconnections of bankfull discharge, discharge at mean annual flood, and mean discharge for peak months to sustain a general argument for (or against) disproportionate increases in bankfull discharge over discharge of the other two kinds. Even if former mean annual floods or peak monthly discharges were reconstructed, by means of hypothetical translocations of climate, they would not indicate former discharges at bankfull stage. Changes in frequency relations can readily be imagined which would increase bankfull discharge proportionally more than discharge at mean annual flood—and, indeed, which would increase bankfull discharge while leaving mean annual flood unchanged. But, quite apart from the close approach of q_{bf} to $q_{2.33}$ on some rivers at the present time, the magnitudes of discharge involved have already been shown to demand something more than a readjustment of frequency relations in the range from 2.33 years downward.

Changes in seasonal regimen of runoff, without change in total runoff, are then regarded as capable of making no more than a small contribution to the required former discharges—a contribution insignificant in comparison with the fivefold or greater increase in total runoff inferred to result from reduced temperature and increased precipitation or in comparison with the postulated increase in bankfull discharge by a factor of commonly about 20 and somewhat exceptionally about 50–60.

This conclusion is substantiated by a comparison of numerical values of former discharges with extreme values now observed and with return periods obtained for discharges of the computed former order. Approximate 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 discharges 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

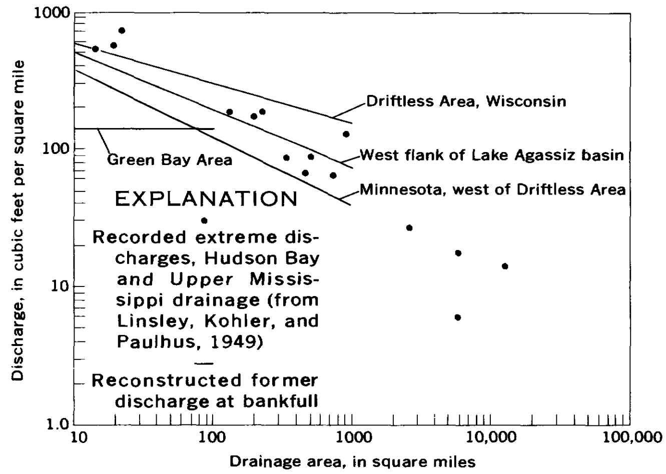


FIGURE 33.—Recorded extreme discharges, contrasted with reconstructed former discharge at bankfull stage. (Example 1.)

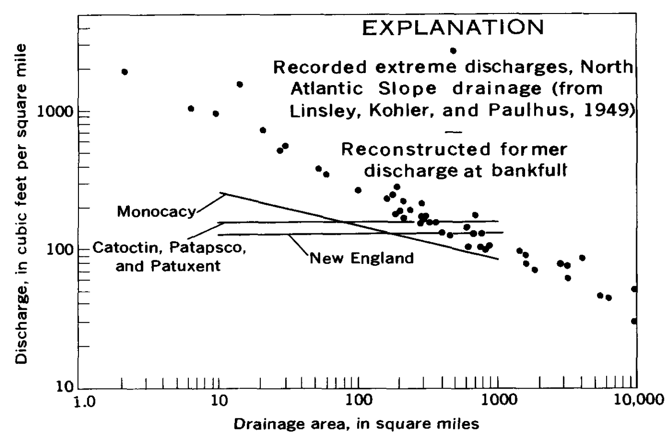


FIGURE 34.—Recorded extreme discharges, contrasted with reconstructed former discharge at bankfull stage. (Example 2.)

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.

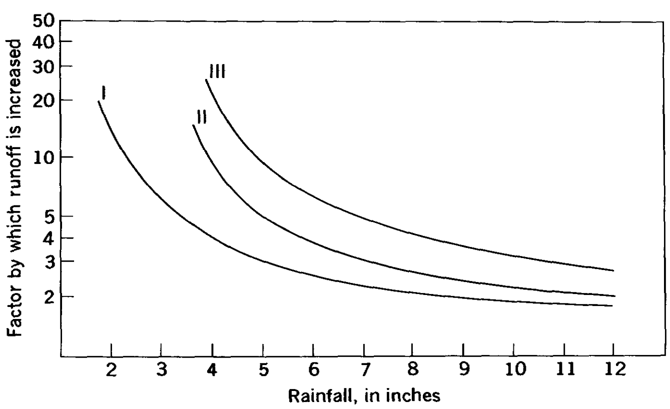


FIGURE 35.—Nomogram for effects on runoff of vegetational change and of change in infiltration capacity. Curve I, effect of conversion from retentively poor herbaceous vegetation to retentively fair oak-aspen forest; effect of reduction of infiltration capacity by one-half (curve II) and by three-quarters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

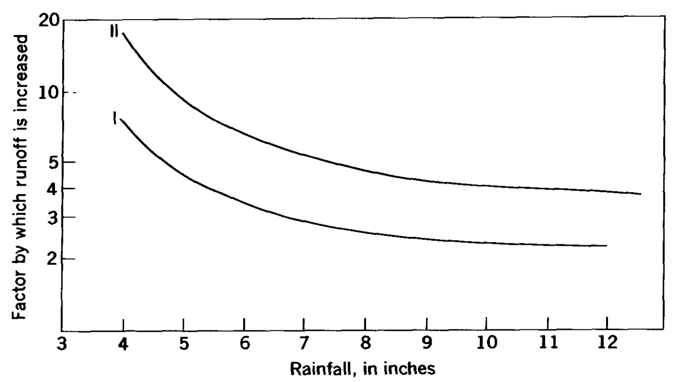


FIGURE 36.—Nomogram for effects on runoff of increase in amount of individual rainstorm. Curve I, 50 percent increase for hydrologically medium forest not having continuously wet soil; curve II, 100 percent increase for same forest.

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.

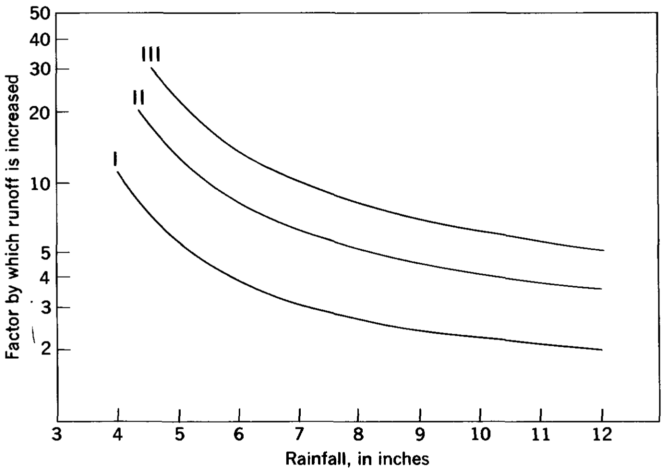


FIGURE 37.—Nomogram for effects on runoff of changes in antecedent wetness of soil, hydrologically medium forest. Effect of conversion from dry to wet soil and, in addition, of no increase in amount of rain (curve I); of an increase of 50 percent (curve II); and of an increase of 100 percent (curve III).

Figure 37 shows the results expectable in a region of hydrologically medium forest when initially dry soil is converted to a state of continuous moistness and when (a) amount of rain is unchanged, (b) amount of rain is increased by one-half, and (c) amount of rain is doubled. Mere conversion of soil condition doubles the runoff from a rainfall of 12 inches and increases tenfold that from a rainfall of 4 inches. When amount of rainfall increases, increases of runoff by factors of 20 or more are predicted in the range of 4–5 inches for present rainstorms. This set of comparisons is not wholly just, since, like the assumptions made about changes in infiltration capacity, it implies a contrast between present-day conditions which are adverse to high runoff and former conditions which were favorable.

The lowest curve (curve I) in figure 35 was obtained by transforming data for vegetation cover alone. It assumes the replacement of retentively fair oak-aspen forest by retentively poor herbaceous vegetation. With amount of rainfall unchanged, runoff predicted for individual rainstorms increases by a factor of 2 for a rainfall of 8.5 inches and by a factor of 10 for a rainfall of 2.3 inches. However, if changes in vegetation are considered as partial explanation of underfit streams, they must be supposed to have operated in a single direction throughout vast areas—an unlikely proposition, especially as some regions which contain underfit streams support, at the present time, vegetation which is retentively poor. And it seems most improbable that the increased runoff of the Atlantic phase in northwest Europe was associated with a lessened retention by the plant cover. Vegetation changes are therefore regarded as complicating the hydrologic 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	4
	2	10	10
Southeast	1	6	3
	2	17	9

TABLE 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
Middle Atlantic	1	15	15
	2	50	35
Southeast	1	25	10
	2	100	35

rainfalls 50 and 100 percent greater than present 1-year and 2-year rainfalls. The shifts in frequency increase with decreasing duration and with increasing return period. Thus, a given proportional increase in total fall could occur with minimal disturbance of frequency relations if it were concentrated in rainfalls of long duration but of high frequency.

If two parallel frequency series could be constructed, one for discharge and one for rainfall, the rainfalls associated with selected frequencies of discharge could be read off. In practice, however, correlation is discouraged by variations in the extent and location of single storms, in tributary contributions to flood peaks, and in seasonal and short-term conditions of infiltration and retardation. Rainfall of high intensity can occur in the low-water season; also, seasonal concentration of runoff 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

TABLE 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	
Driftless Area, Wis	1,000	[1] 4.6	0.007	(?)6.0	0.009	1.3
	10	[1] 7.0	.011			
Green Bay area, Wisconsin	100	[2] 7.0	.011	(?)12.0	.019	1.7
	10	[2] 7.0	.011	(?)50.0	.078	7.1
Southern New England		[2] 7.8	.012	[2] 22.5	.035	2.9
Green River, Ky	500	[1] 6.6	.010	35.0	.055	5.3
	10	[1] 13.4	.021	150.0	.235	11.2
Humboldt 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
Wabash 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
Georgia streams (mean of five)		[4] 24.8	.039			
Northeast 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
Alabama 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
Red River, N. Dak.-Minn., and Sheyenne River, N. Dak		[2] .63	.001	[2, 5] .63	.001	1.0
Nene 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
Great 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
Wye 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) in summer. If 34,000 feet is assumed as the depth of column, on the grounds that maximum height was formerly less than it now is but that precipitation was formerly concentrated in summer, then it can be shown that a reduction of 20°F in surface temperature would reduce the depth of precipitable water by two-thirds. Specifically, in the Ohio valley, where mean July temperature is about 75°F and where high-glacial temperatures are held to have been some 20°F lower than those of today, a reduction in surface temperature from 75°F to 55°F would reduce the depth of precipitable water from about 3 inches to about 1 inch. A greater proportional reduction is obtained when allowance is made for differences in depth of column between present and former times.

Maximum possible precipitation does not, however, depend merely on depth of precipitable moisture but also on transport of air. Extent and duration of storm are both involved. The generalized estimates show that, at many stations, possible maximums increase rather slowly with increase of duration above 6 hours and that they decrease rather slowly with increasing area. Thus, they are commonly not more than 50 percent greater for duration of 24 hours than for duration of 6 hours; the maximum possible fall on an area of 10 square miles is commonly less than twice that on an area of 500 square miles at 6-hour duration and less than 1½ times as great at 24-hour duration.

For the sake of inquiry, let the area of 200 square miles and the duration of 24 hours be adopted. Table 14 lists, against selected stations, expectable 2-year 24-hour rainfalls that have been converted from point rainfall to areal rainfall for 200 square miles by the factor 0.92. Also listed are maximum possible precipitations—for 200 square miles and 24-hour hour duration—and these same precipitations multiplied by 0.33. The intention is to inquire whether rainfalls of the given extent and duration could possibly occur in conditions of greatly reduced temperature. As-

TABLE 14.—*Some values of 24-hour 200-square-mile rainfall*

Station	2-year rain [1]	Maximum possible rainfall	
		Actual [2]	×⅓
Mobile, Ala	5.3	31.9	10.6
Savannah, Ga	4.2	28.2	9.4
Pensacola, Fla	5.5	31.5	10.5
Macon, Ga	3.5	28.2	9.4
Montgomery, Ala	4.1	30.5	10.2
Birmingham, Ala	4.1	30.0	10.0
Richmond, Va	3.3	24.0	8.0
Charleston, N.C	4.1	28.2	9.4
Pittsburgh, Pa	2.2	23.0	7.7
Harrisburg, Pa	2.8	22.0	7.3
New York, N.Y	3.2	23.6	7.9
Baltimore, Md	3.2	22.2	7.4

[1] U.S. Weather Bur. Tech. Paper 29, "Rainfall Intensity-Frequency Regime."
[2] U.S. Weather Bur. (1947).

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 possible maximums are about 4 times the value of 24-hour falls at a return period of 1 year.

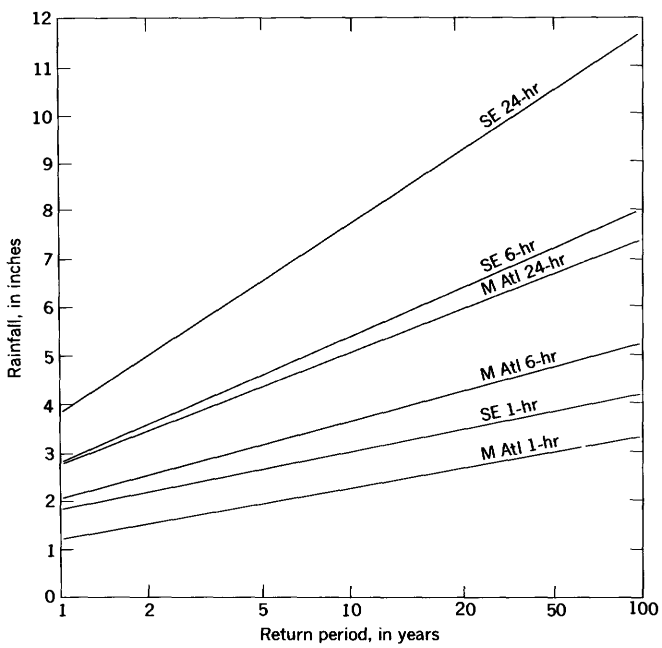


FIGURE 38.—Generalized rainfalls for varying return periods and duration, Middle Atlantic region and Southeast United States.

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 ot 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
100	X	
2,500	IX	Sub-Atlantic.
4,500	VIII	Sub-Boreal.
6,200	VII	Main Atlantic.
7,750	VI	Early Atlantic (Transitional).
9,000	V	Boreal.
10,000	IV	Pre-Boreal.
10,600	III	Younger Dryas.
12,000	II	Allerød (Two Creeks).
12,700	Ic	Older Dryas.
13,300	Ib	Bölling.
16,000	Ia	Oldest Dryas.

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 reduction in July means subsequent to Zone II amounting to some 4°F. This scarcely seems enough to account for the renewed channeling in southern England in the range Zones III–IV, so that increased precipitation is required to accompany temperature reduction. Although Manley (1959) inferred reductions in summer mean temperatures for northwest England by 7°–16°F for the cold fluctuation after Zone II, he also called for a considerable increase in precipitation at the same time to explain extensions of glaciers. Since increased precipitation is is needed to explain the renewed rise of Lake Lahontan to very high levels, a cold-humid fluctuation can be inferred for Zone III (approximately), both in humid and in arid regions.

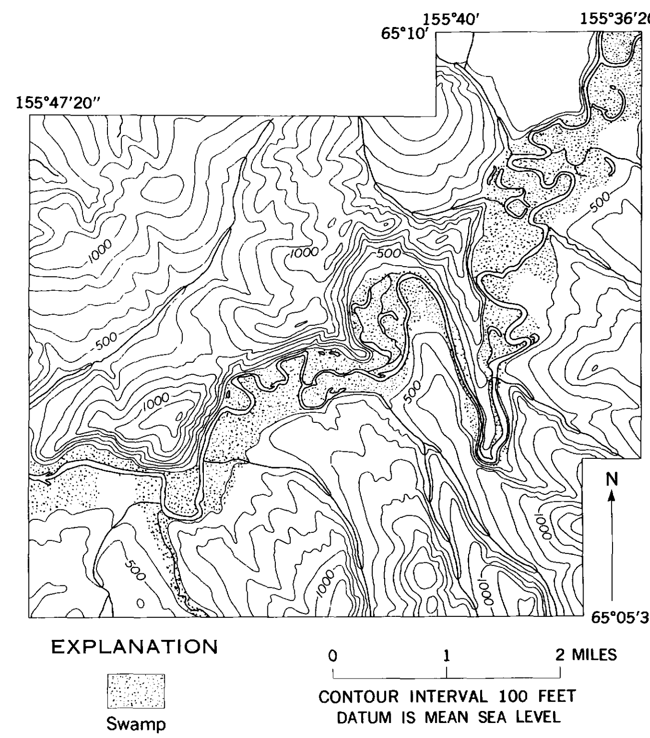
Infilling of large channels in southern England during Zone V coincided, as the floristic record shows, with temperatures reduced anew. During this interval, a reduction of precipitation must be inferred capable of counteracting the effects of reduced temperatures and, in addition, of causing runoff to decrease. In direct contrast, as previously observed, increased temperatures during Zone VII were associated with increased runoff, which considerably increased precipitation-supplied runoff.

The deglacial succession affords room for episodes of minor cutting—in addition to those identified here and in Professional Paper 452–B—such as would be expected to accompany the minor fluctuations of deglacial sea level reported by Fairbridge (1961) or would be expected from the climatic evidence reviewed by Conway (1948, and references therein). In the longer term, the tally of about 15 complete temperature cycles thought to relate to the Pleistocene (Emiliani, 1955) suggests considerable fluctuations of runoff during the last half-million years or more. The last main onset of underfitness, though its location on the scale of deglacial time is somewhat imprecise, may have been the last such episode of many.

Broad climatic implications of the findings described in this Professional Paper series lie, for the most part, outside the limits of present inquiry. No attempt will be made to reconstruct the meteorological patterns of relevant intervals of time, whether those of channeling or those of filling. In a general way, of course, the high precipitation inferred for times of incision of valley meanders and for times of complete scouring of largest channels relates to steep meteorological gradients across belts of latitude. Willett (1950) postulated intensification of the general circulation at the last glacial maximum, and Ewing and Donn (1958) called for very great storminess in middle latitudes as a result of frequent encounters between glacial air from the icecaps and moist equatorial masses. In the shorter term, Winston

(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



Base from U.S. Geological Survey topographic quadrangle: Melozitna (A-6), Alaska, 1:63,360

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°F; in this example, the values of F_q correspond to those used in constructing figure 23. Appropriate adjustments supply values for figures 24–26.

TRANSFORMATION OF RUNOFF VALUES FOR SINGLE STORMS

The procedure here is essentially similar to that just outlined for values of annual runoff. It makes use of the information presented by the U.S. Bureau of Reclamation (1960, Appendix A, p. 412–431, "Estimating rainfall runoff from soil and cover data"; section A–5 of that appendix, containing nomograms for runoff equations supplied by the U.S. Soil Conservation Service, is especially relevant).

For example, two comparative series of values can be derived from the runoff equation $q=\frac{(p-0.2S)^2}{p+0.8S}$ —where q is direct runoff, in inches; p is storm rainfall, in inches; and S is maximum potential difference between p and q, in inches, at the time of storm's beginning—by taking a range of values for p and a second range in which each entry is twice as large as the corresponding entry in the first range. The two resultant ranges of values for q will then indicate the effects of a doubling in the amount of rainfall. Results of changes in vegetation cover are obtainable by use of the runoff-curve numbers listed by the U.S. Bureau of Reclamation (1960), table A–2, p. 426. As with data for annual runoff, smoothing by means of graphs is usually necessary, and in any event it is useful in obtaining integral values for increase in storm runoff. The following 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 increase in storm runoff
5.0	0.24	1.30	5.42
6.0	.50	1.92	3.84
7.0	.85	2.60	3.06

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