

## DOWNSTREAM CHANGE OF VELOCITY IN RIVERS

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**ABSTRACT.** Because river slope generally decreases in a downstream direction, it is generally supposed that velocity of flow also decreases downstream. Analysis of some of the large number of velocity measurements made at stream-gaging stations demonstrates that mean velocity generally tends to increase downstream. Although there are many reaches in nearly all rivers where mean velocity decreases downstream, the general tendency for conservation or for downstream increase was found in all data studied.

Computations of bed velocity indicate that this parameter also tends to increase downstream.

Near the streambed, shear in the vertical profile of velocity (rate of decrease of velocity with depth) tends to decrease downstream. This down-valley decrease of shear implies decreasing competence downstream.

### GENERAL STATEMENT

**B**ECAUSE rivers tend to be steep in their headwaters and flow at flat gradients near their mouths, it appears logical to suppose that the stream velocities are great in the headwaters and decrease downstream.

Yet, the engineer who actually measures stream velocities for purposes of computing the water discharge uses the same current meter for a measurement in a small brook as in the Mississippi River. The velocities that he measures are not sufficiently different in small headwaters and downstream reaches to furnish any obvious indication of a definite pattern of organization in the data.

The purpose of the present paper is to organize some of the large amount of available data on river velocities so that the downstream change may be evaluated.

### DATA ON RIVER VELOCITY

There are more than 6000 gaging stations in the United States. Records on many streams are more than 50 years in length. At each station periodic measurements of discharge are made by use of a current meter. The discharge on other days is determined indirectly from the recording chart or daily readings of stage (water surface elevation).

When a river measurement is made by current meter, the water speed is measured at about 20 to 30 verticals across the section and usually at two depths in each vertical. From these direct measurements the form of the river cross section may be plotted, as well as an approximate pattern of the vertical and horizontal distribution of velocity.

In stream-gaging practice, the mean velocity is not a simple average of the velocity measurements made at various positions across the stream. The current meter placed in a given position in the cross section measures the downstream movement of the water in a given interval of time. The metering thereby defines a prism of water whose width and depth are known and whose downstream dimension is represented by the measured movement. The volume of the prism that passed the meter during the measured time interval represents the discharge in units of volume per unit time as denoted by the identity,

$$Q \equiv aV$$

where  $Q$  is discharge in cubic feet per second,  $a$  is cross-sectional area in square feet, and  $V$  is velocity in feet per second. The total discharge past the cross section is the sum of the discharges represented by the individual prisms.

The total discharge having been obtained in this way, the mean velocity used in hydraulic practice is an expression of the velocity of a single rectangular prism having the same cross-sectional area as the stream. This mean velocity is defined as the quotient of the discharge in the cross section divided by the cross-sectional area of flowing water. Arithmetically the mean velocity so defined is equivalent to a weighted mean in which the velocities of the individual prisms are weighted by the cross-sectional area of the respective prisms.

Owing to the distribution of the velocity with depth, the mean velocity is closely equal to that existing at a position 0.6 of the distance from the water surface to the streambed.

It is recognized (Rubey, 1938, p. 129) that the mean velocity in the cross section is not the most meaningful velocity with respect to sediment transport. Direct measurements of the bed velocity, however it may be defined, are not possible at present. On the other hand, a large fund of measurements of mean velocity exists. It appears instructive to examine the downstream change of that velocity parameter for which meas-

urements exist, recognizing that the parameter, mean velocity, is not ideal for discussing geomorphic progresses. In a later section, downstream changes of a computed bed velocity will be presented.

As indicated previously, velocity in a stream is a conservative quantity. Even the maximum point velocity (maximum at any point in the cross section) does not ordinarily greatly exceed the mean velocity for the cross section. An unpublished study by the Geological Survey of 2950 measurements of maximum point velocity values from a variety of rivers showed a median value of 4.11 ft per sec., mean of 4.84 ft per sec. and less than one per cent of the total exceeded 13 ft per sec.

The largest value of maximum point velocity in a natural river channel ever measured by stream-gaging personnel of the U. S. Geological Survey was about 22 ft per sec. Approximately this value was measured on several occasions: Brazos River at Waco, Texas, Sept. 27, 1936; Canadian River near Newcastle, Okla., May 4, 1941; New River at Caperton, W. Va., Aug. 1940; Potomac River at Washington, D. C., May 14, 1932.

#### FLOW DURATION AND FLOOD FREQUENCY

In order to compare the mean velocity in a small headwater tributary with that in a large trunk stream it is necessary to establish a basis of comparability. It would be meaningless to compare the velocity during a flood flow at one point on a river with that of a low flow at some other point. The two points to be compared must both be experiencing flood flow or must both be experiencing low flow. The relative magnitude of flow is discussed in hydrologic practice by the use of the flood-frequency concept. If the mean flow for each day of record at a station is tallied in categories of size, the number of days on which each category of size is equaled or exceeded may be counted. The number of days in each category may be accumulated from the largest category to the smallest, and divided by the total number of days. The resulting quotients represent the percentage of the total days each given discharge is equaled or exceeded. When the discharge equaled or exceeded is plotted against per cent of time the resulting curve, called the "flow-duration curve," is directly analogous to the plot used in sedimentation work to describe the size-frequency distribution of a sediment sample.

The velocity at two stations along a river system may be logically compared when the discharge at each represents the same percentage of time equaled or exceeded.

The flow-duration curve adequately describes the per cent of time a given mean daily flow is exceeded. A flood peak, however, may last only a few hours and therefore would not be represented by the mean flow for the day on which it occurs. The frequency of flood peaks, therefore, is usually determined by the use of a series of data representing the momentary peak-discharge values for each year of record. This series is arranged in order of magnitude, and plotted in terms of the recurrence interval in years versus the peak discharge.

There are minor variations in methods of preparing flow-duration and flood-frequency curves, and procedures for analysis are adequately described in the literature (Jarvis, 1936, pp. 68-85; Linsley, Kohler, and Paulhus, 1949, pp. 182-198), but the general principles underlying all methods are essentially similar. In the analysis contained herein, the procedures for obtaining flow-duration and flood-frequency curves were those in current use in the U. S. Geological Survey.

#### MEAN VELOCITY AT A GIVEN RIVER CROSS SECTION

The mean velocity in a river cross section tends to increase with discharge. This becomes evident from a consideration of a relation widely used in hydraulic practice, the Manning equation, expressing the balance of frictional losses (which are dependent on velocity, hydraulic radius, and roughness) and the slope of the energy grade line.

$$V = 1.49 R^{2/3} \frac{s^{1/2}}{n} \quad (1)$$

where  $V$  = mean velocity

$$R = \frac{\text{area of cross section}}{\text{wetted perimeter}} = \text{hydraulic radius}$$

$s$  = slope of the energy grade line, equal to slope of water surface for steady, uniform flow

$n$  = roughness parameter

For most natural river channels the width is great compared with depth and the hydraulic radius,  $R$ , is approximately equal

to the mean depth,  $d$ . The discharge,  $Q$ , is expressed as the product of mean velocity and cross-sectional area,

$$Q \equiv aV$$

With increasing discharge in most natural river cross sections, the water-surface slope remains nearly constant and roughness tends to decrease. Thus velocity tends to increase with increasing depth and therefore with increasing discharge.

Figure 1 presents a typical relation of mean velocity to discharge in a river cross section. The data are representative of the large volume of observational material collected at stream-gaging stations by the use of current meter. Such data are available in the files of the Water Resources Division, U. S. Geological Survey. For the station pictured, the mean annual discharge during a 9-year period of record was 881 cfs, at which the corresponding velocity is 3.2 ft per sec. At any given discharge a corresponding mean velocity may be read off the graph.

#### CHANGE OF MEAN VELOCITY DOWNSTREAM

It has been demonstrated (Leopold and Maddock, 1953) that the velocity corresponding to mean discharge tends to increase with increasing discharge downstream.

The reason for the downstream increase of velocity with increasing discharge has been previously discussed in detail (Leo-

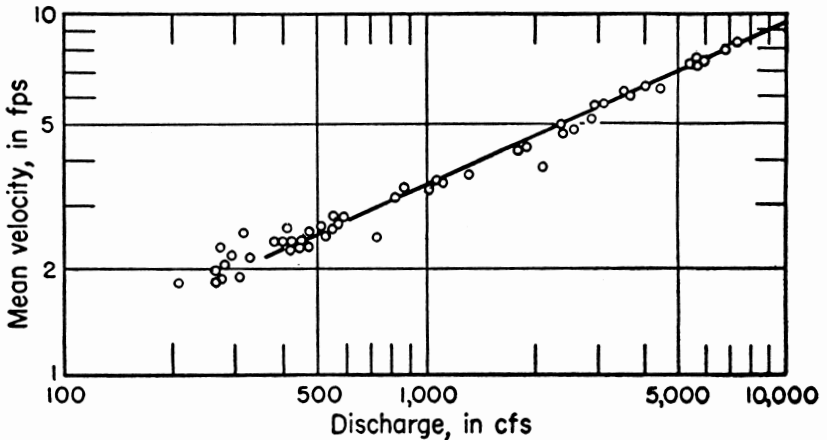


Fig. 1. Relation of mean velocity to discharge in a given river cross section, Stillwater River near Absarokee, Montana.

pold and Maddock, 1953, pp. 15, 42, and 43). The explanation can be summarized by saying that the downstream increase of depth overcompensates for the decrease in slope.

In that study, the river data considered in analyzing the downstream change of velocity did not include enough high mountain streams of very steep gradient and in which caliber of load rapidly decreased downstream. Moreover, downstream changes in bed velocity were not discussed. The present paper reviews the evidence previously discussed, presents data for streams representing a more stringent test of the generalization, and discusses similar relations involving bed velocity.

The mean annual discharge at a given station is the arithmetic mean of all the daily discharge values during the period of record. In most river systems, the mean discharge increases downstream owing to the increase in drainage area in the downstream direction. As a general rule, mean discharge increases directly with some power (near unity) of drainage area in a given basin. Though there are many basins, particularly in the western United States, in which mean discharge decreases owing to irrigation diversion and to channel losses, in most basins there is an increase downstream.

The mean discharge at most points along a river system occurs somewhat less often than half the time. The per cent of time the mean discharge is equaled or exceeded is not constant but is sufficiently conservative to allow a comparison of upstream and downstream points. Because it is a parameter readily available for a large number of gaging stations, its use allows the comparison of a considerable number of stations. The computation of complete flow-duration and flood-frequency curves has been completed for a smaller number of selected stations chosen for a more critical evaluation.

The mean velocity at a number of stations along a river length will be compared at discharges corresponding to their respective mean annual values. Figure 2a presents a sample of this relation for a small portion of the upper Missouri River basin and a few of stations on the Missouri and Mississippi rivers.

Each point in figure 2a represents a particular gaging station and its abscissa value corresponds to its mean annual discharge. Within each tributary basin, the points are numbered in sequence from downstream toward the headwaters. In nearly all

basins the mean annual discharge increases downstream but there are examples, such as the Greybull River, in which the mean discharge at a downstream station, Basin, Wyoming (point 23), is less than at Meeteetsee farther upstream (point 24), owing to irrigation diversion and channel losses.

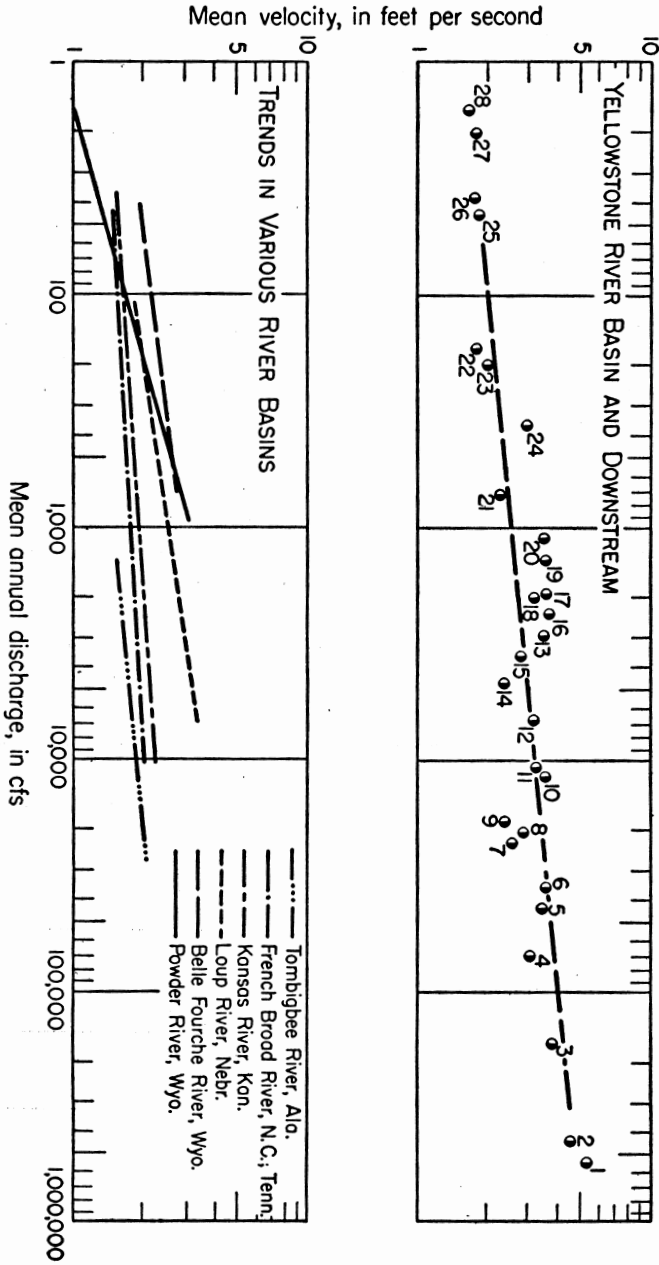
Stations in the Yellowstone and Bighorn River basins chosen to represent a headwater area have velocities at mean annual discharge of 2 to 4 ft per sec., whereas the lower Missouri and Mississippi rivers have corresponding velocities between 4 and 6 ft per sec. The mean line in the graph of figure 2a has been fitted by inspection and is intended merely to call attention to the definite tendency of the points to progress toward higher velocity values downstream.

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Fig. 2. Change of mean velocity as mean annual discharge increases downstream. In 2a above, each point represents a gaging station on Yellowstone River system and downstream. In 2b, mean trend lines for various river basins are drawn.

Identification of Stream-gaging Stations  
in Figure 2A

1. Mississippi R.	at	Vicksburg, Miss.
2. "	"	Memphis, Tenn.
3. "	"	St. Louis, Mo.
4. Missouri R.	at	Hermann, Mo.
5. "	"	Kansas City, Mo.
6. "	"	St. Joseph, Mo.
7. "	"	Pierre, S. D.
8. "	"	Bismark, N. D.
9. "	"	Williston, N. D.
10. Yellowstone R.	nr.	Sidney, Mont.
11. "	"	Miles City, Mont.
12. "	"	Billings, Mont.
13. "	"	Corwin Springs, Mont.
14. Bighorn R.	at	Custer, Mont.
15. "	"	St. Xavier, Mont.
16. "	"	Kane, Wyo.
17. "	"	Thermopolis, Wyo.
18. "	"	Manderson, Wyo.
19. Wind R.	nr.	Crowheart, Wyo.
20. "	"	Riverton, Wyo.
21. "	"	Burriss, Wyo.
22. "	"	Dubois, Wyo.
23. Greybull R.	at	Basin, Wyo.
24. "	"	Meeteetsee, Wyo.
25. Owl Creek	nr.	Thermopolis, Wyo.
26. Medicine Lodge Cr.	nr.	Hyattville, Wyo.
27. Gooseberry Cr.	nr.	Grass Creek, Wyo.
28. N. Fk. Owl Cr.	nr.	Anchor, Wyo.



In any given reach of river the velocity may decrease downstream, as can be seen by a study of figure 2a. Also there is a considerable scatter of points in the graph. But for present purposes attention is directed to the general trend which is toward increasing velocity downstream.

In figure 2b the trend lines of velocity versus mean annual discharge are plotted for a number of rivers chosen to represent a variety of physiographic provinces. The general tendency for increasing velocity as mean annual discharge increases downstream is well exemplified.

The mean annual discharge in many streams is equaled or exceeded about one out of every 4 days in a long period. Inasmuch as the greatest total sediment load is carried by a stream during high or flood flows, it is of interest to examine the trend of velocity at discharge rates less frequent than the mean annual rate. For this purpose figure 3 presents the trend lines for discharges equaled or exceeded various percentages of time from 50 per cent to 1 per cent, in the Maumee and Scioto River basins in Ohio. The two basins were considered together in order to be able to utilize a sufficient number of points to define the downstream trends. The Maumee and Scioto basins were chosen because complete flow-duration curves for stations in Ohio are available in the study by Cross and Bernhagen (1949). Figure 3 is taken from Leopold and Maddock (1953).

The scatter of points about the lines represented in figure 3 was again great, but the general trend of increasing velocity

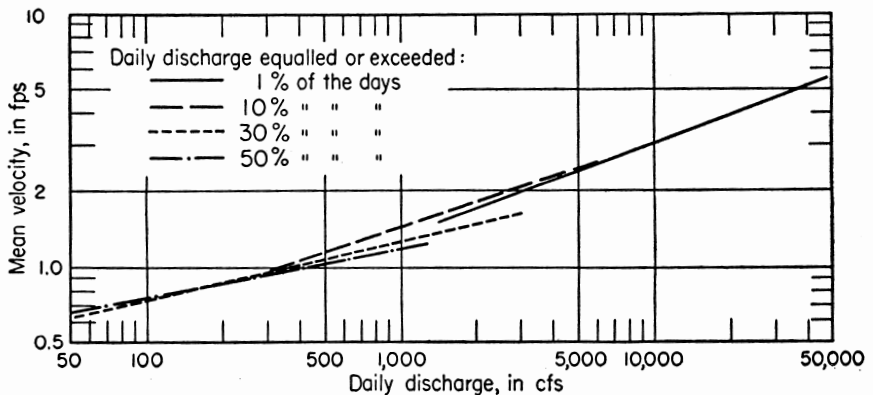


Fig. 3. Relation of mean velocity to daily discharge equaled or exceeded various percentages of time, Maumee and Scioto river basins, Ohio.

downstream in these basins appears to characterize each of the duration percentages analyzed.

But the Maumee and Scioto basins still do not furnish a critical test of how mean velocities vary downstream because the topographic relief in those drainage basins is not great. Moreover, the materials composing the beds of those rivers are relatively fine grained, consisting for the most part of clay, silt, and sand. It is logical to suppose that if in any stream the velocity tends to decrease downstream, such a stream would be one that is very steep in the headwaters and flat downstream. Moreover, it would be a stream the bed of which consists of coarse gravel in the headwaters, decreasing in grain size to sand downstream. Furthermore, the graphs presented earlier were for discharges exceeded no less often than 1 per cent of the days, or one day in 100. It might be supposed that the more extreme floods, which are the most effective in moving coarse gravel in the headwaters, would be the flow condition most likely to show a tendency for decreasing velocity downstream.

As an example of these stringent conditions a series of tributaries of the upper Yellowstone River in Wyoming and Montana were selected. The streams chosen include the Greybull River, Rock Creek, the Clarks Fork, and some of their headwater tributaries. These streams drain into the Bighorn River, a tributary to the Yellowstone River, or directly into the Yellowstone. These headwater tributaries carry coarse gravel which decreases in caliber rapidly downstream. The slopes of the uppermost tributaries are up to 100 feet per mile. More than one tributary to the Yellowstone was selected in order that a sufficient number of gaging stations would be included to provide an adequate test.

In accordance with good hydrologic practice, the flood-frequency curves for all gaging stations were developed for an identical period of record, 1923-1950, inclusive, and the records shorter than this were adjusted to the long record.

The flood data utilized at each station consisted of the peak or momentary discharge for each year of available record. Flood-frequency curves were developed for recurrence intervals of 5, 10, 25, and 50 years. By definition, a flood of 50-year recurrence interval means the momentary discharge equaled or exceeded on the average once in 50 years over a long period.

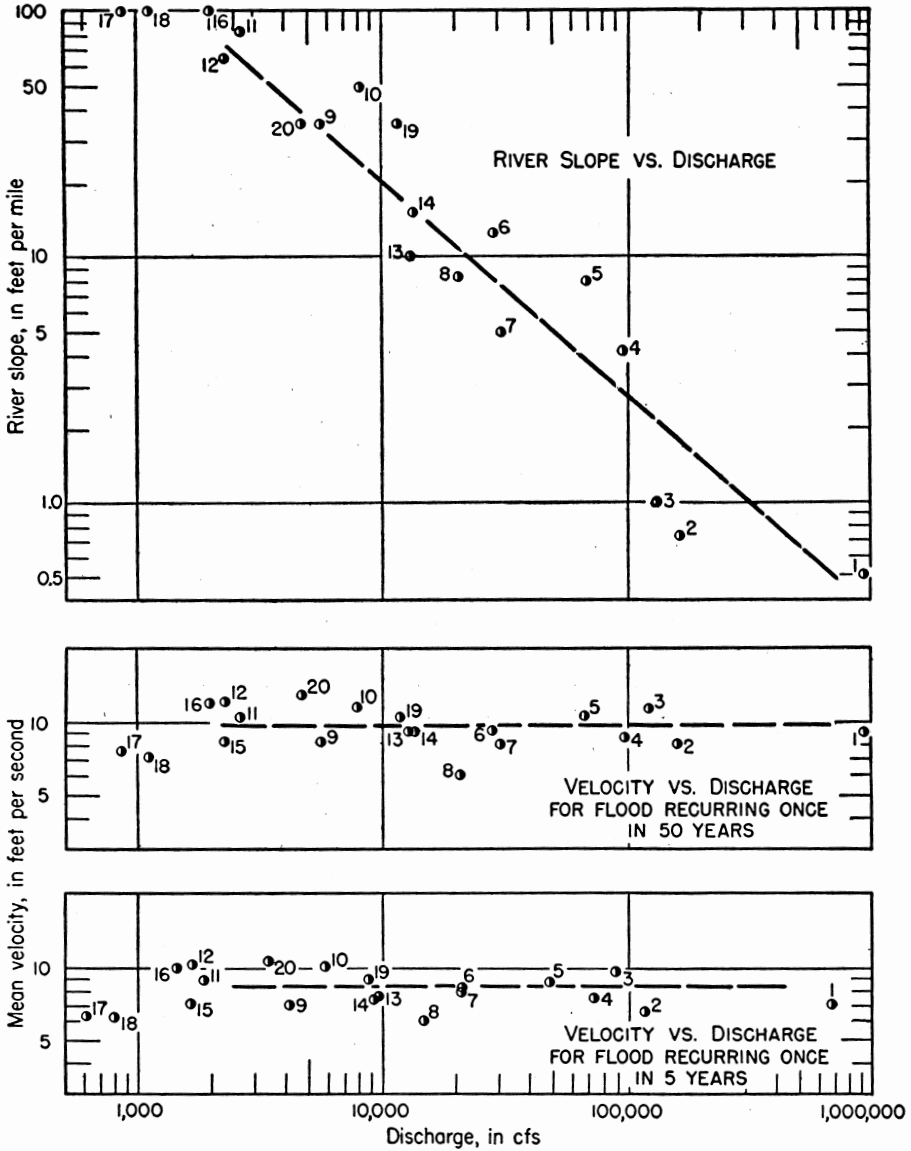


Fig. 4. Change of mean velocity with increase of discharge downstream for flood flows, Yellowstone River Basin and downstream. Upper figure shows how river slope decreases downstream. Lower figures show downstream change of velocity for 50 year and 5 year floods.

The relation of mean velocity to discharge was plotted for each station as in the example of figure 1. After the discharge for each recurrence interval chosen had been determined from the flood-frequency curves, the velocity corresponding to that discharge was read from the velocity-discharge curves.

The curves for recurrence intervals of 5 years and 50 years are reproduced in figure 4 for the stations in the upper Yellowstone basin. For comparison, a few stations on the upper Missouri River were included in the study, and one station on the main stem of the Mississippi River.

Dashed lines have been fitted by eye through the points in the velocity-discharge graphs of figure 4 merely to indicate the trend. Though there is scatter to the points, this indicates a remarkable conservation of the mean velocity downstream, and there is no definite tendency to decrease downstream in these rivers. A valid interpretation of these data is that in these streams mean velocity is essentially constant downstream.

The graph of river slope in feet per mile plotted against discharge (50-year recurrence interval), which is presented for comparison with the velocity-discharge graphs, shows the marked decrease of gradient downstream from 100 feet per mile to about half a foot per mile at the stations studied. That the velocity should remain essentially constant despite

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Identification of Stream-gaging Stations Shown  
by Points in Figure 4

1. Mississippi R. at St. Louis, Mo.
2. Missouri R. at Williston, N. D.
3. Yellowstone R. nr. Sidney, Mont.
4. " at Miles City, Mont.
5. " at Billings, Mont.
6. " at Corwin Springs, Mont.
7. Bighorn R. at Kane, Wyo.
8. " at Thermopolis, Wyo.
9. Greybull R. nr. Basin, Wyo.
10. " at Meeteetsee, Wyo.
11. " nr. Pitchfork, Wyo.
12. Wood R. at Sunshine, Wyo.
13. Clarks Fork at Edgar, Mont.
14. " at Chance, Mont.
15. Rock Cr. at Joliet, Mont.
16. " nr. Red Lodge, Mont.
17. W. Fk. Rock Cr. below Basin nr. Red Lodge, Mont.
18. Red Lodge Cr. above Reservoir nr. Boyd, Mont.
19. Stillwater R. nr. Absarokee, Mont.
20. Rosebud Cr. nr. Absarokee, Mont.

such a decrease in slope, and under the conditions of extreme flood in streams of rapidly decreasing bed-material size, is of interest.

Whether further study of additional data will discover streams in which the mean velocity rapidly decreases downstream is unknown, but no such trend has been found in any of a large number of streams studied by the author.

#### CHANGES DOWNSTREAM IN THE VELOCITY NEAR THE STREAMBED

Though the velocity near the streambed cannot at present be measured directly, the theory of vertical velocity distribution in pipes and open channels is sufficiently well developed that computations of velocity near the bed are possible. Though these computed velocities are approximate, they allow, nevertheless, an indication of the downstream trends which are under discussion here.

The velocity distribution in the vertical tends to follow a logarithmic law down to a position very close to the bed itself where the flow is laminar, and within the laminar sublayer the vertical changes are linear.

Bed velocity has been defined in a number of ways. Rubey (1938) considered it to be the velocity at the laminar boundary. Einstein (1950) discussed the movement of bed load in terms of the velocity in the "bed layer" which he found to be about 2 grain diameters in thickness.

The theory of turbulent flow derived from the study of pipe flow was applied to wide rivers by O'Brien (1937). Langbein (1940) further developed this concept and modified the O'Brien equations to a form which relates the velocity at a given distance above the bed (but outside the laminar sublayer) to the mean velocity. As O'Brien had previously done, Langbein showed that actual measurements of velocity distribution gave results similar to those from the O'Brien equation. On the basis of measurement data, Langbein presents a dimensionless graph of the velocity distribution in the vertical which related  $F(z/d)$  to  $(z/d)$  where

$$F(z/d) = \frac{V - V_m}{\sqrt{gds}} \quad (2)$$

$V$  is the velocity at distance  $z$  from the bed,  $d$  is the depth,  $s$  is energy slope,  $V_m$  is the mean velocity in the vertical, and

$F(z/d)$  signifies a function of  $(z/d)$ . Values of the function  $F(z/d)$  may be read off Langbein's graph for given values of  $z$  and  $d$ . For concurrent values of  $V_m$ ,  $d$ , and  $s$  representing a given cross section, the vertical velocity distribution may be computed.

Leopold and Maddock (1953) have studied the changes in depth, mean velocity, and water-surface slope in the downstream direction for numerous rivers. The downstream changes have been expressed as functions of discharge of a given frequency. From their data, which were in a form similar to those presented in figure 2b of the present paper, representative curves of the downstream changes of  $V_m$ ,  $d$ , and  $s$  as functions of mean annual discharge have been chosen and are defined by the mutually consistent equations below. These relations may be considered to describe the characteristics of a typical midwestern river, though no single river is exactly like the composite represented by the equations.

$$d = 0.16 Q^{0.40} \quad (3)$$

$$V_m = 1.3 Q^{0.10} \quad (4)$$

$$s = 0.021 Q^{-0.49} \quad (5)$$

where  $d$  is depth in feet,  $V_m$  is mean velocity in feet per second,  $s$  is water-surface slope (approximately equal to energy slope), and  $Q$  is mean annual discharge in cfs.

For a given value of  $Q$ , values of  $V_m$ ,  $d$ , and  $s$  provided by equations 3-5 were substituted in equation 2, and using Langbein's graph, the velocity  $V$  at different distances  $z$  above the bed was computed. The computed velocities at three distances above the bed are shown in table 1 for several discharge values representing progressive downstream changes. Complete vertical-velocity curves for four of these assumed discharges are plotted in figure 5.

The four diagrams in figure 5 represent plots of velocity against stream depth at four places along the length of the assumed composite river. The logarithmic decrease of velocity with depth in a given cross section is apparent from the shape of the curves. The Langbein empirical relation between  $F(z/d)$  extended only to  $z/d$  value of 0.1 and had to be extrapolated somewhat to compute the values of velocity near the bed for the deeper sections, but because his relation was a straight line

TABLE 1  
 Computed Values of Velocity and Shear near Streambed  
 Representing Changes in Downstream Direction

Depth $d$ —ft—	Discharge $Q$ —cfs—	Slope $s$	Mean velocity $V_m$ —ft/sec—	Velocity in ft/sec at various distances above streambed —0.02 ft      0.05 ft      0.1 ft—	Shear between 0.05 and 0.1 ft —ft/sec/ft—
0.4	10	0.0070	1.6	1.1      1.34      1.52	3.6
1.0	100	.0023	2.0	1.3      1.54      1.71	3.4
2.0	580	.00098	2.4	1.6      1.82      1.97	3.0
3.5	2400	.00048	2.7	1.9      2.10      2.24	2.8
5.0	5800	.00030	3.0	2.1      2.32      2.45	2.6

on logarithmic paper and parallel to the O'Brien theoretical relation, the extrapolation appeared to be justified.

The measured velocities in the canal data used by Langbein gave the average position of the mean velocity at 0.67 the distance from the water surface to the streambed, whereas in most rivers it usually occurs at about 0.60 the depth. This slight difference does not materially affect the shape of the vertical-velocity curves.

It can be seen in both figure 5 and table 1 that, progressing downstream, depth increases faster than does mean velocity. Moreover, the velocity at any specific distance above the bed also increases downstream, as can be seen in the table. In the case of the particular river characteristics used here as typical, the downstream rate of increase of velocity 0.02, 0.05 and 0.10 foot above the bed is about the same as the downstream rate of increase of mean velocity.

On figure 5 dashed lines have been drawn between the points representing the velocity at 0.05 feet above the bed and the mean velocity. The slope of the dashed line then represents the shear  $\frac{\Delta V}{\Delta d}$  between these two depths. The changing slope

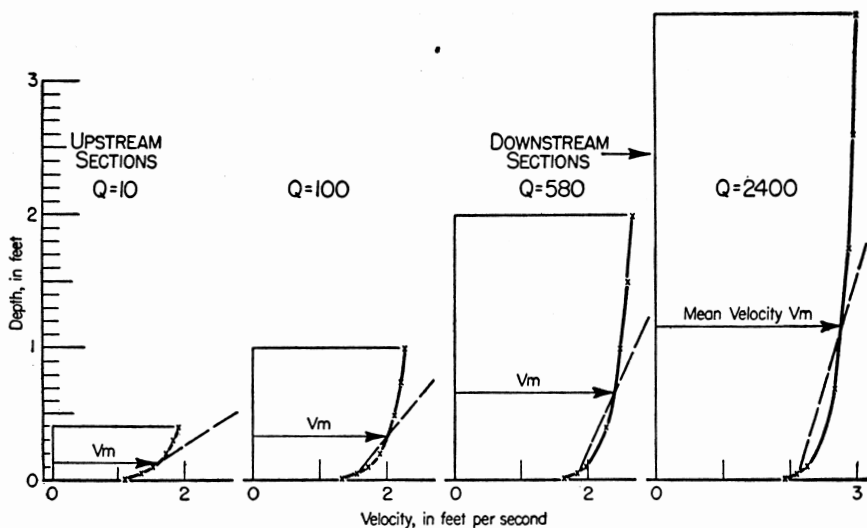


Fig. 5. Vertical velocity distribution computed for four sections representing progressive positions downstream in a river. Note that velocity near the bed as well as mean velocity increases downstream. Dashed lines connect velocity vectors at 0.05 feet above bed and the mean velocity.

of these dashed lines from the upstream section to the downstream section shows that this shear decreases downstream. In table 1, values of the shear between points 0.05 and 0.1 feet above the bed are listed and it can be seen that these shear values also decrease downstream.

The finite values of shear indicated by the righthand column in table 1 are presented only to provide a better visual picture of the physical nature of shear stress in the vertical profile of velocity. They represent merely examples of shear quantities and it is not implied that these are particularly important depths through which to measure shear.

The total shear on the bed,  $\tau$ , is proportional to the product of slope and hydraulic radius; for wide channels where hydraulic radius is approximately equal to mean depth,  $\tau \propto ds$ . As indicated by the equations (3) and (5) for the composite river discussed earlier and for many of the individual rivers studied, slope decreases downstream slightly faster than depth increases. Thus the product depth  $\times$  slope, to which total bed shear is proportional, decreases slightly downstream.

It appears, then, that in numerous rivers and in a synthetic composite river described here, total shear on the bed and shear between different levels near the bed decrease in the downstream direction. This decrease may obtain despite a downvalley increase of mean velocity and of velocity near the bed.

#### RELATION TO COMPETENCE AND LOAD

It is well known that the particle size of material composing the bed of a river decreases downstream owing to abrasion and sorting, and that there is a concomitant decrease in slope. It has long been supposed, therefore, that the velocity of a stream decreases downstream.

The reasoning on which this supposition is based may be summarized as follows: The downstream decrease in caliber of load caused by abrasion and sorting means that a smaller velocity is required to carry the small particles near the stream mouth than the large particles in headwater reaches. The smaller velocity required downstream is attained by progressive flattening of the slope downstream. If, in any reach, the contributed load is of such large caliber that the existing velocity is insufficient to move it, deposition occurs until the slope through the reach of deposition is increased sufficiently to produce a velocity great enough to move the particles.

The present analysis indicates that the velocity near the bed, as well as the mean velocity, tends to increase downstream. It has been shown, furthermore, that even under conditions of extreme flood in steep mountain streams the mean velocity does not decrease downstream. It is necessary, then, to revise our concept of how the mutual adjustment of slope, roughness, velocity and caliber of load are interrelated in the downstream direction.

In the present state of knowledge it is not possible to substitute as simple and readily understandable a concept as the one to which we have been accustomed. A few relations which may be necessary and useful for further understanding can be stated now, but formulation of a complete picture must await further work.

In streams carrying primarily coarse gravel, it is safe to assume that bed roughness is governed primarily by particle size. Downstream, where the bed is composed of fine sand, silt, or clay, bed roughness is materially affected by the configuration of the bed in the form of transitory riffles, waves, and dunes.

The vertical-velocity profile is dependent in part on the bed roughness. Roughness affects the rate at which velocity decreases as one approaches the bed—that is, the shear in the moving water close to the bed. This shear is intimately related to the production of turbulence, which in turn is responsible for the movement of bed particles.

Movement of bed particles is accomplished by the energy of turbulent eddies immediately adjacent to the particles. A similar principle applies to the quarrying of large blocks of bedrock exposed in the stream bed and to the movement of large boulders or coarse gravel on the bed. On the basis of keen observation, Matthes (1947) described how kolk, a form of macroturbulence consisting of strong intermittent vortices, is a primary mechanism for plucking and for movement of large bed material. Vortical motion is probably, in most cases, a phenomenon arising from shear stress.

Therefore, a change in roughness, through its effect on shear, would alter the competence even without any concurrent change in mean velocity. Similarly, competence may be changed without necessarily altering slope.

If roughness decreased downstream, shear would tend to decrease downstream. This implies a downvalley decrease in competence.

The relation between shear and bed roughness and between these same factors and competence and load are beyond the scope of this paper. The importance of these relationships has been brought out by the many studies of bed load movement in which the grain size and shear factors appear prominently. The present paper merely calls attention to a tendency for certain shear functions to decrease downstream. The tendency for particle size to decrease downstream is well known. There is implied a downstream decrease of competence but any definitive discussion of this aspect requires further detailed study.

From the foregoing analysis it is clear that we may no longer consider stream adjustment in terms of a downstream decrease in velocity. It is equally important to recognize, however, that the introduction of the concept of decreasing shear downstream does not mean that we may simply substitute decreasing shear or roughness for decreasing velocity in our old concept of stream adjustment. As can be seen in the Manning equation, a change of roughness,  $n$ , requires a change in the interrelations of velocity, depth, and slope in order that the equation be satisfied. Moreover, if stream adjustment is considered in terms of the eight variables, discharge, width, depth, velocity, slope, roughness, load, and caliber of load, as was done by Leopold and Maddock (1953), then there are not one but eight simultaneous equations involved in the description of the interrelations.

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