that viscous losses due to the flow of water into and out of the well are negligible. From the seismological point of view, the assumption of simple harmonic earth-motion is also objectionable. The importance of these factors can only be determined by further work which is necessary before the method may be considered applicable.

A disadvantage of the formula should be pointed out. For very large values of V/A, the phreatic magnification becomes sensibly independent of the volume of the aquifer, the response of the well is no longer a good index of its volume, and ultimately the formula reduces to

$$M = H/E = 2\pi b/\rho g\lambda \tag{3}$$

which is independent of the volume. The result of this is that, for example, with a 6-inch well, penetrating an aquifer containing 30 acre-feet of water, an accuracy in M within one per cent would be required to evaluate V within ten per cent. Thus the determination of very large volumes is impractical. On the other hand, an 18-inch well would permit the determination of a volume of nearly 300 acre-feet with the same accuracy. An accuracy within one per cent in the determination of M would be hard to attain. Hence it may be doubted whether the method will find a practical use, and it is here presented primarily for its theoretical interest.

A rough verification of formula (3) was made in the following way. For the Lower California Earthquake of December 30, 1934, recorded at the well, the first well-water motion in the P-phase was computed by the formula, using for the earth-motion at Lodi the first ground-motion of the first preliminary waves at Berkeley as computed from the records of the Bosch-Omori and Wiechert seismographs. This computed amplitude by formula was 0.035 cm as compared with the observed amplitude recorded by the instrument of 0.040 cm. This agreement tends to substantiate the formula, although it is recognized that the neglect of transients in both cases (earth-motion at Berkeley and water-motion at Lodi) is questionable.

For a well equipped with a pressure recording-device, such as the artesian wells reported by Leggette and Taylor (Bull. Seis. Soc. Amer., v. 25, pp. 169-175, 1935), the value of A/V is practically zero. Hence we have $M = H/E = 2\pi b/\rho g\lambda$. Since the recording is in pressure-units, $H/E = p/\rho gE = 2\pi b/\rho g\lambda$ and $p = (2\pi b/\lambda)E$ where p = amplitude of pressure-change.

During large earthquakes it is well known that underground waters are frequently forced up in the form of earthquake-fountains. These flow for a short time. In the earthquake of 1906, sand brought up from such fountains near Salinas came from a depth of 80 feet. It is conceivable that the effects of earthquake-waves in the vicinity of the epicenter of a large earthquake might cause some spurting of water in this fashion. However, the continued flow frequently observed is probably due to an actual displacement of part of the aquifer.

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NATURAL STREAM CHANNEL-STORAGE

Robert E. Horton

It is assumed that the reader has some familiarity with the author's infiltration-theory of runoff[1] and also with Sherman's unit-graph method of calculating runoff[2]. [References: (1) Robert E. Horton, The role of infiltration in the hydrologic cycle, Trans. Amer. Geophys. Union, 14th annual meeting, pp. 446-460, 1933, and Surface-runoff phenomena: Part I--Analysis of the hydrograph, Pub. 101, Horton Hydrologic Laboratory, Edwards Bros., Inc., Ann Arbor, Mich., 1935; (2) Water-Supply Paper 772, U. S. Geol. Surv., 1936.]

Figure 1 is an example of a channel-outflow graph for Wabash River above Logansport, Indiana, August 1929, from Water-Supply Paper 772, United States Geological Survey, 1936 (the original graph as plotted from daily average discharges has been corrected to allow for the fact that the peak is higher than the crest-day average). In a previous paper (op. cit., Pub. 101) the author showed: (a) Direct surface-runoff ends soon after rainfall ends--sometimes a little earlier; the time at which the direct surface-runoff ends is usually at or near the point of inflection a (Fig. 1) on the hydrograph. (b) After direct surface-runoff ends on a drainage-basin without surface-storage in reservoirs, lakes, swamps, and marshes, the entire flow at a given point, gaging-station, or outlet, is derived either from ground-water flow or channel-storage until the effects of subsequent rainfall begin to appear in the stream.

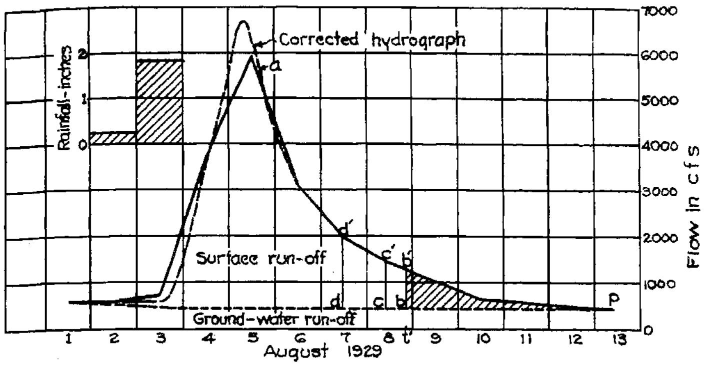


FIG. 1- WABASH RIVER ABOVE LOGANSPORT, IND.
STREAM-RISE OF AUG. 2-12, 1929

It is generally possible by different methods to draw a line separating the ground-water or base-flow from the surface-runoff, such as the dotted line of Figure 1.

It can be shown analytically that, subject to one exceptional case, if there is a constant or nearly constant rate of ground-water flow entering a stream, the channel-storage will drain out in a finite time, ending at a point p (Fig. 1), where there is, as a rule, a more or less abrupt flattening of the slope of the recession-side of the graph.

It is easy to see that at any time t', measured backward from the end of the channel-storage outflow, the shaded area on Figure 1 represents the total volume of channel-storage which remained in the stream-channels comprising the drainage-net of the stream-system at the time t'. Also the ordinate bb' represents the rate of channel-storage outflow at a time when the total volume of channel-storage is as represented by the shaded area. If, now, the area of the graph above the base-flow line and between the points p and a is determined for each of a series of ordinates bb', cc', dd', etc., these data when plotted will show the law of relationship between the rate of channel-storage outflow and the total volume of channel-storage. It can be shown analytically that if the width of the streams is a simple function of the stage, the volume of channel-storage will be a simple power-function of the stage above a suitable reference-plane. If there was no ground-water flow, this would be the plane of zero-discharge at the given gaging-station or outlet.

A stream-channel usually consists of a succession of deeps and shoals, and the fact that a given gaging-station may be located in a deep reach where there is a quiescent pool of water remaining in the channel-bottom at zero-discharge makes no difference since the stage-outflow relation at such a station is determined by the plane of zero-discharge or by the bottom of the channel at a point downstream where there is a rapid forming a low-water control. Ground-water flow can be considered as raising the plane of zero-outflow of channel-storage. Many natural stream-channels have approximately parabolic cross-sections, particularly at control-points. It can easily be shown that if the slope is constant and the width is a simple power-function of the stream-stage, the stage-discharge relation can always be represented by an equation of the form

$$q = K_c\, h^M \qquad (1)$$

where h is the stage above the plane of zero-discharge. Similar equations can be applied, but with different values of the exponent M, where the slope varies as a simple function of the stage. If the stage-discharge relation above the plane of zero-channel storage-outflow is given by the equation

$$q_s = K_c'\, h^R \qquad (2)$$

and the volume of channel-storage is given by the equation

$$S_c = K_s'\, h^S \qquad (3)$$

then the channel storage-outflow rate will be a simple power-function of the volume of remaining channel-storage, expressed by the equation

$$h = \sqrt[R]{q_s/K_c'} = \sqrt[S]{S_c/K_s} \qquad (4)$$

from which

$$q_s = K_c'(S_c/K_s)^{(R/S)} = K_c'(S_c/K_s)^{(R-S)} \qquad (5)$$

or, letting $M = (R - S)$ and $K_s = (K_c'/K_s^{R/S})$

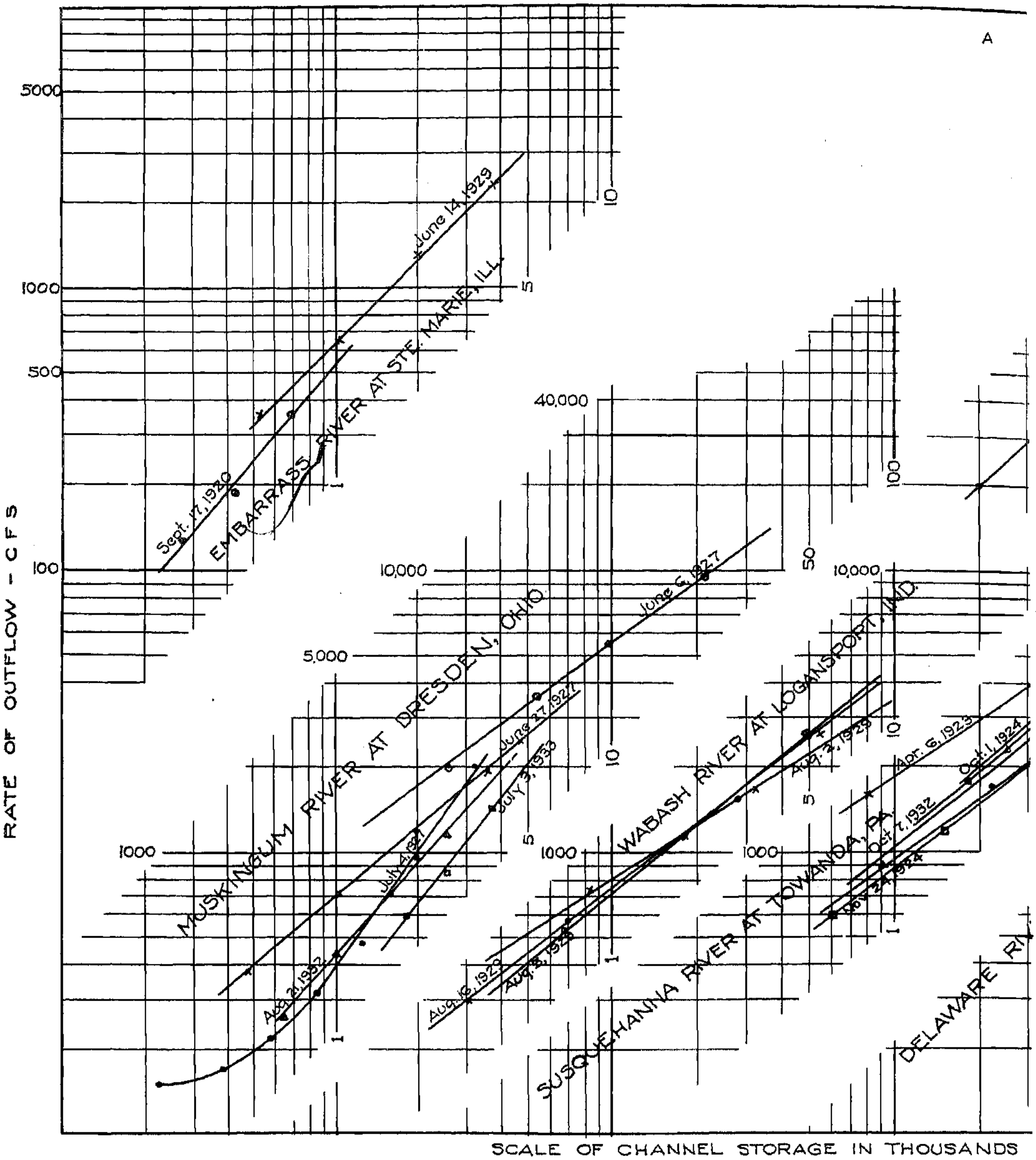
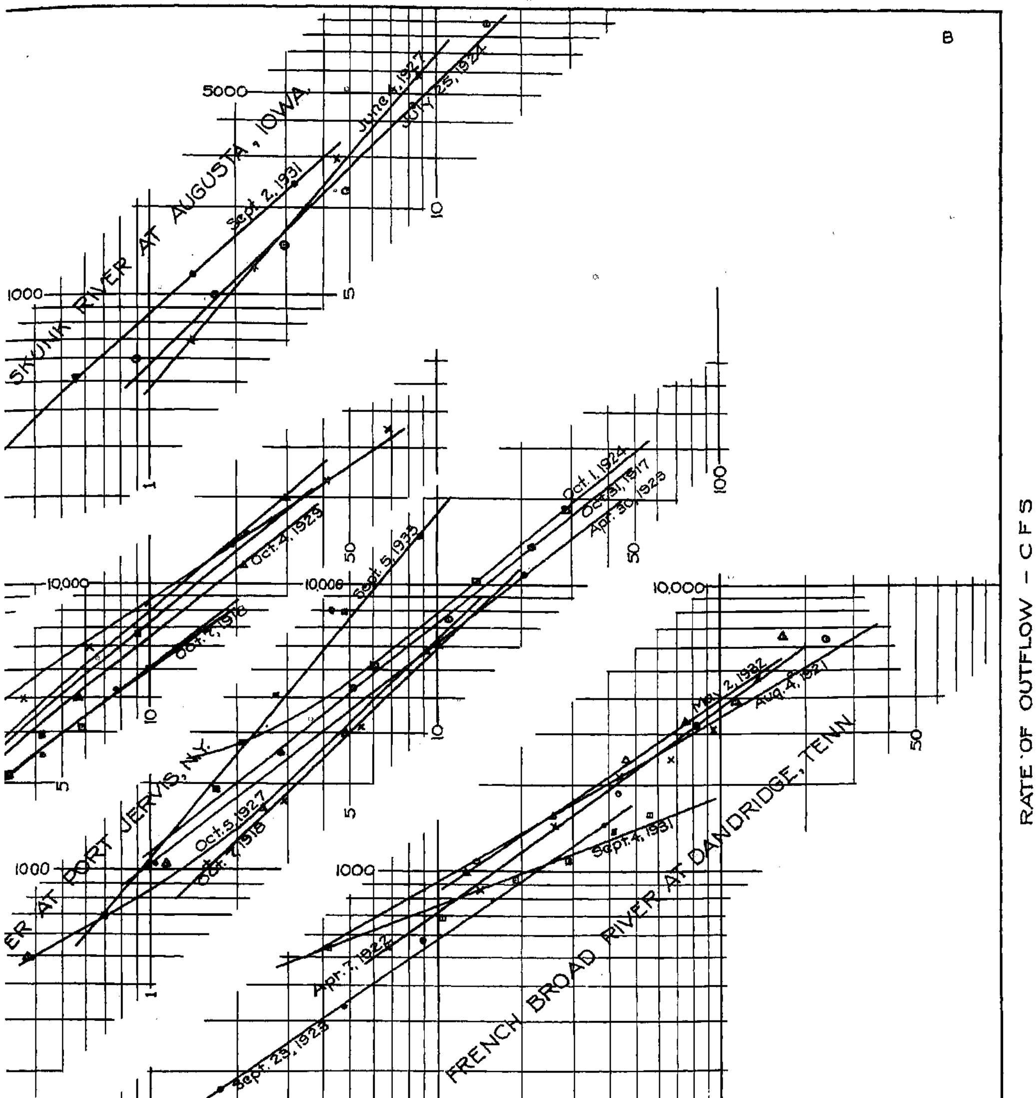


FIG.2-CHANNEL-STORAGE OUTFLOW-VOLUME RELATION LINES

$$q_s = K_s S_c^M \qquad (6)$$

Equation (6) expresses the rate of channel-storage outflow in terms of the volume of remaining channel-storage. If this simple power-function relationship holds true for channel-storage in the stream-system of a given drainage-net, then the values of q_s plotted in terms of the corresponding values of S_c, determined from the recession side of the hydrograph of a stream-rise in the manner described, will plot as straight lines on logarithmic paper and the appurtenant values of K_s and the exponent M can easily be determined.

Water-Supply Paper 772 (Hoyt and others, Studies of relations of rainfall and runoff in the United States, U. S. Geol. Surv., 1936) contains a large number of examples of graphs of stream-rises. These graphs were plotted for use in the study of Sherman's unit-graph method. For this purpose it was deemed sufficient to make a roughly approximate separation of ground-water from



FROM GRAPHS OF STREAM RISES GIVEN IN W.S.P. 772

surface-runoff. Subject to errors from this cause, as subsequently discussed, these graphs furnish an excellent basis for determining if the channel storage-outflow-volume relation is, in fact, a simple power-function in most cases, and also for determining the values and variations of the constants K_s and M in equation (6).

Selecting those graphs in which the recession-side of the graph was reasonably free from surface-runoff from residual rainfall, the volumes of channel-storage were determined and plotted on logarithmic paper and the resulting lines are shown on Figure 2. [Caution is required in plotting such lines not to extend the computation to the left of the point of inflection on the graph, as surface-runoff is then entering the stream and the volume of channel-storage is changing more or less independent of the outflow-rate.] It will be noted that the channel storage-outflow relations are nearly all straight lines. Furthermore, note that they have as a rule nearly the same slope for different graphs on the same stream at the same gaging-station.

TABLE 1

VALUE OF K_s AND M IN EQUATION $q_s = K_s S_c^M$ FROM STREAM-FLOW GRAPHS IN W.S.P. 772

Stream	Drainage Area Sq. Mi.	Date of flood	K_s	M	$\sqrt[M]{K_s}$	$\frac{1}{\sqrt[M]{K_s}}$
(1)	(2)	(3)	(4)	(5)	(6)	(7)
French Broad River at Dandridge, Tenn.	4,450	Aug. 4, 1921	17.8	0.573		
		April 7, 1922	5.1	0.716		
		Sept. 23, 1923	4.9	0.692		
		Sept. 4, 1931	60.0	0.366		
		May 2, 1932	9.1	0.660		
		Average	19.38	0.601	5.93	0.167
Red River at Denison, Texas.	39,400	June 18, 1930	39.1	0.510		
		Dec. 7, 1930	17.6	0.573		
		Jan. 7, 1932	12.3	0.660		
		Sept. 5, 1932	2.54	0.773		
		Average	17.88	0.629	6.12	0.164
Susquehanna River at Towanda, Pa.	7,770	Oct. 7, 1918	6.02	0.730		
		Apr. 6, 1923	1.92	0.666		
		Oct. 1, 1924	2.91	0.860		
		Nov. 24, 1924	4.63	0.760		
		Oct. 4, 1929	4.61	0.786		
		Oct. 7, 1932	4.19	0.810		
		Average	4.05	0.769	2.93	0.341
Delaware River at Port Jervis, N.Y.	3,070	Oct. 31, 1917	4.8	0.793		
		Oct. 7, 1918	0.984	0.945		
		Apr. 30, 1923	4.93	0.773		
		Oct. 1, 1924	5.00	0.800		
		Oct. 5, 1927	4.22	0.786		
		Sept. 5, 1933	0.55	1.185		
		Average	3.41	0.880	2.94	0.340
Embarrass River at Ste. Marie, Ill.	1,540	Sept. 17, 1920	0.22	1.13		
		June 14, 1929	0.78	0.965		
		Average	0.50	1.048	0.484	2.065
Skunk River at Augusta, Iowa	4,290	July 25, 1924	0.772	0.963		
		June 4, 1927	0.18	1.152		
		Sept. 2, 1931	1.79	0.893		
		Average	0.914	1.003	0.912	1.098
Muskingum River at Dresden	5,980	June 6, 1927	7.0	0.726		
		June 27, 1927	2.45	0.820		
		July 4, 1927	.048	1.316		
		Aug. 21, 1932	.196	1.098		
		July 3, 1933	.05	1.25		
		Average	1.95	1.042	2.005	0.500
Wabash River at Logansport, Ind.	3,830	Aug. 5, 1929	2.61	0.816		
		Aug. 16, 1929	9.10	0.650		
		Aug. 2, 1929	3.71	0.770		
		Average	5.14	0.745	3.390	0.295

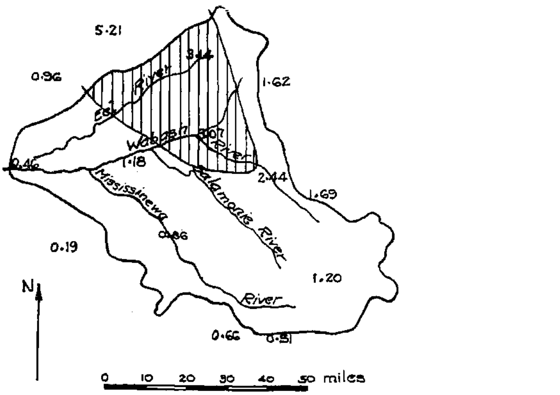
Table 1 shows the values of K_s and M determined from these graphs. It will be noted that the value of the exponent M is relatively constant for different graphs on the same stream. So far as these studies go, the value of the exponent M appears to hover around two values. For many streams M is from 2/3 to 3/4 or not far from (5/3 - 1), the value which it would have if the channel had a constant width within the range of the observations, the slope was constant, the discharge was that corresponding directly to the Manning formula, $q = (1.486/n)W\ h^{5/3}\ \sqrt{S}$, and the channel-storage was that corresponding to a channel of uniform width, that is, with the volume of storage increasing directly in proportion to the stage h. For some other streams the exponent M has a value close to but usually a little more than unity.

The values of the coefficient K_s for different graphs taken at the same gaging-station are much more variable than the values of M. It can be shown analytically that the values of both K_s and M are sensitive to the effect of errors in the elimination of ground-water flow. This probably accounts for some of the variation in M and K_s. The major reason for variation in K_s, which represents the rate of outflow from channel-storage when the channel-storage is unity, is that many of the stream-rises used in this analysis did not represent runoff from the entire drainage-basin. It is possible by the application of the author's infiltration-theory of the analysis of the hydrograph (op. cit., Pub. 101) to determine in a given case the part of a drainage-basin from which surface-runoff takes place. For example, in case of the Wabash River, the surface-runoff which produced the graph shown on Figure 1 was confined to the shaded area shown on Figure 3, that is, to the northerly portion of the drainage-basin, entirely excluding the Mississinewa River. A stream-rise or flood where direct surface-runoff takes place from only part of the drainage-basin may be designated a "partial-area rise" or flood. It is undoubtedly true that a large proportion of the runoff graphs given in Water-Supply Paper 772 represent partial-area rises.

The unit-graphs given in Water-Supply Paper 772 are nearly all for relatively large drainage-basins. It is generally true that as the size of the drainage-basin increases, the frequency of occurrence of stream-rises resulting from surface-runoff over the whole basin decreases. If the stream-rise results from runoff from the lower portion only of the drainage-basin, the entire channel-system will not contain channel-storage. The rate of outflow will therefore be increased and the coefficient K_s increases relative to that for a full-area rise on the same stream. The situation where surface-runoff takes place only from the upper portion of the drainage-basin is more complicated and will not be discussed here but it appears that, in general, the coefficient K_s will have its minimum value for a full-area rise. This fact is important since the minimum value of K_s is apparently fairly fixed and definite, and can be applied in computations relating to the most severe floods, which are, in general, full-area floods.

Channel storage-outflow-volume relations were also worked out for a large number of stream-rises on Ralston Creek at Iowa City, Iowa. [The author is indebted to the late Dr. Floyd A. Nagler for data regarding Ralston Creek stream-rises used in the preparation of Figure 4 and Table 2.]

The results are shown on Figure 4 and the values of the coefficient K_s and exponent M for the different rises are shown on Table 2. This drainage-



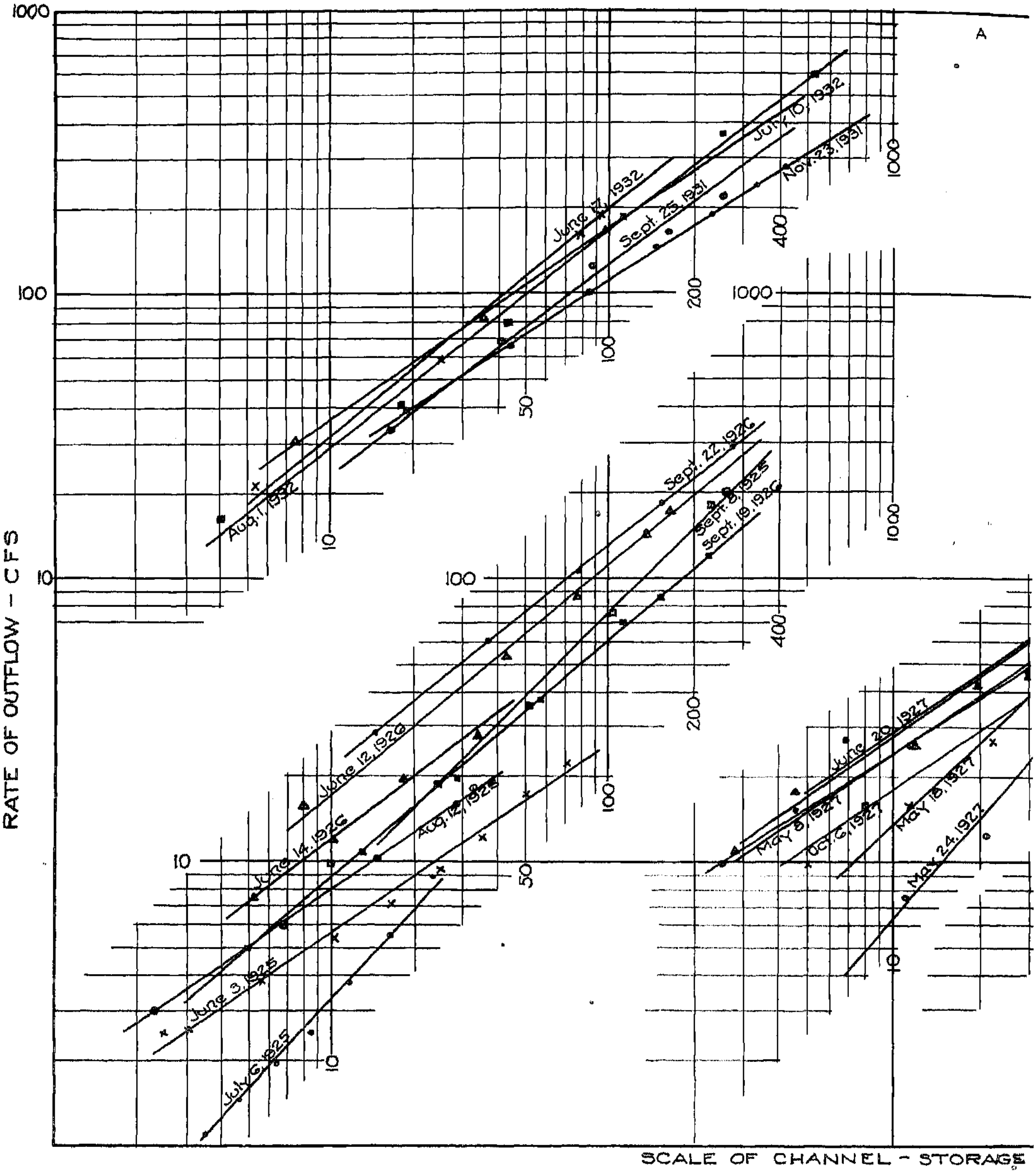
WABASH RIVER BASIN ABOVE LOGANSPORT, IND.
DRAINAGE-AREA 3830 SQ.MI.

FIG 3- TOTAL STORM RAINFALL AND AREA (SHADED) WHICH PRODUCED RISE OF AUG. 3-12, 1929
THE INFILTRATION-CAPACITY WAS 0.30 INCH PER HOUR

TABLE 2

VALUE OF K_s AND M IN EQUATION $q_s = K_s S_c^M$
RALSTON CREEK, IOWA CITY, IOWA

Date	Maximum Rate of Discharge c.f.s.	K_s	M	$\sqrt[M]{K_s}$	$\frac{1}{\sqrt[M]{K_s}}$
(1)	(2)	(3)	(4)	(5)	(6)
June 3, 1925	43	1.205	0.667	1.132	0.883
July 6, 1925	12	0.322	1.050	0.304	3.290
Aug. 12, 1925	21	1.66	0.683	1.414	0.707
Sept. 8, 1925	267	0.78	1.00	0.780	1.283
June 12, 1926	272.	2.70	0.81	2.238	0.447
June 14, 1926	45	2.13	0.75	1.763	0.567
Sept. 19, 1926	146	1.33	0.833	1.268	0.787
Sept. 22, 1926	380	3.42	0.793	2.65	0.377
May 8, 1927	295	5.02	0.680	2.99	0.334
May 18, 1927	360	1.565	0.946	1.528	0.654
May 23, 1927	95	5.50	0.640	2.98	0.336
May 24, 1927	123	0.434	1.163	0.379	2.640
June 20, 1927	650	5.83	0.689	3.37	0.297
Oct. 2, 1927	385	5.49	0.703	3.31	0.302
Oct. 6, 1927	236	3.70	0.686	2.455	0.407
May 18, 1928	70	3.48	0.759	2.572	0.389
June 24, 1928	220	5.25	0.650	2.94	0.340
July 4, 1928	75	3.87	0.600	2.41	0.415
July 19, 1928	185	4.25	0.651	2.565	0.390
Aug. 23, 1928	353	3.30	0.769	2.500	0.400
June 11, 1929	395	3.38	0.829	2.742	0.365
June 13, 1929	108	2.75	0.883	2.44	0.410
July 1, 1929	200	3.32	0.776	2.535	0.394
Aug. 2, 1929	372	3.12	0.777	2.42	0.413
June 22, 1930	150	5.54	0.670	3.15	0.317
June 14, 1930	735	5.29	0.700	3.21	0.311
July 19, 1931	68	4.55	0.680	2.805	0.357
Sept. 21, 1931	132	5.35	0.658	3.02	0.331
Sept. 25, 1931	250	4.00	0.746	2.815	0.355
Nov. 23, 1931	338	5.79	0.639	3.075	0.325
June 17, 1932	340	4.87	0.800	3.545	0.282
July 10, 1932	270	7.50	0.675	3.90	0.256
Aug. 1, 1932	850	4.85	0.763	3.340	0.299
Average		3.68	0.761	2.441	0.596



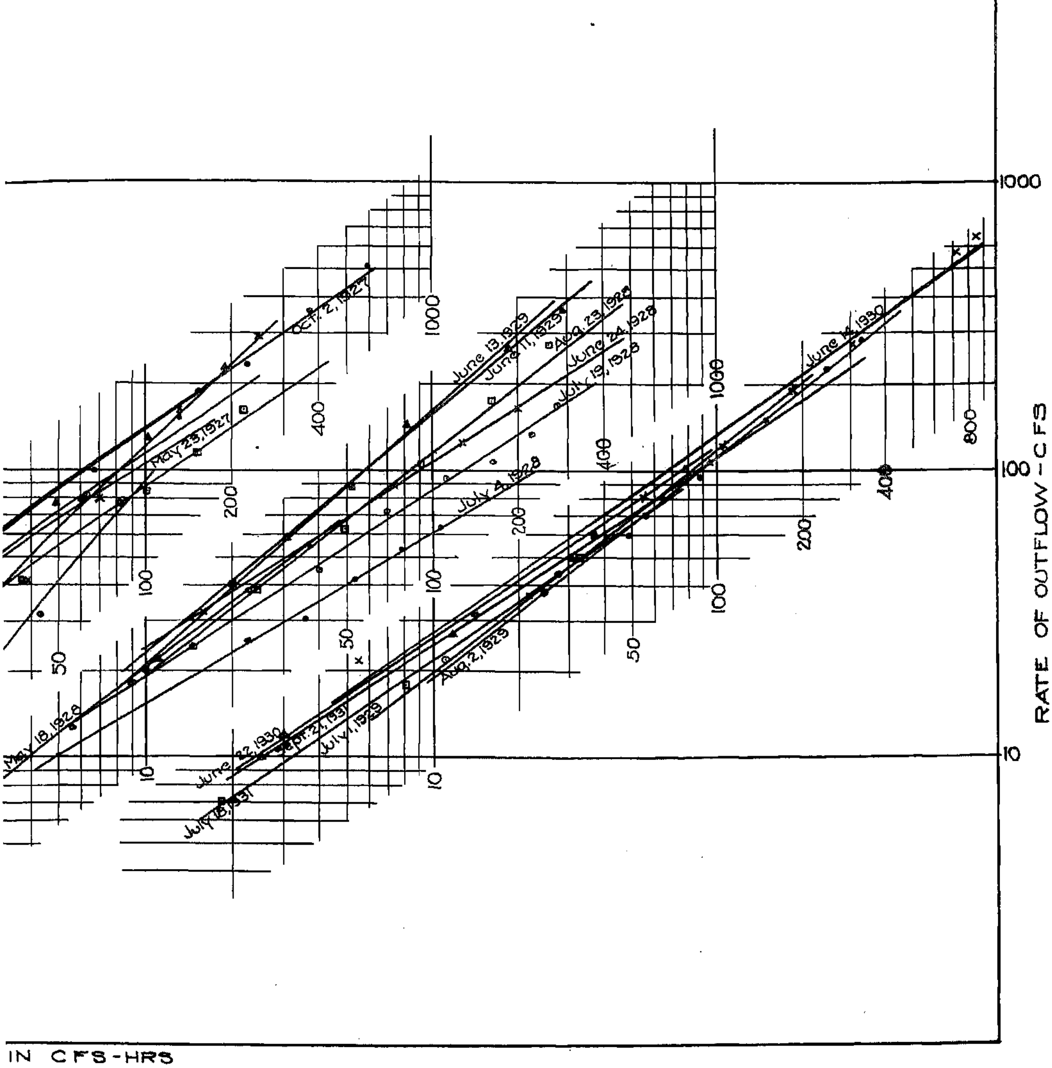
basin comprises only three square miles but, owing to high infiltration-capacity of part of the area during the summer season, it happens that here again many of the rises are partial-area rises, since full-area rises result only from rains having intensities exceeding the highest infiltration-capacity on any part of the drainage-basin. [In preparing the examples of channel storage-outflow-volume relation lines for Ralston Creek, given at p. 39 of Pub. 101, it was assumed that there was no appreciable ground-water flow. Some of the resulting lines show considerable curvature as the stage- and outflow-rates decrease. Later recomputation of these lines, making corrections for ground-water flow, reduces them substantially to straight lines.]

The slopes of the graphs for Ralston Creek (Fig. 4) and the values of the exponent M are of the same order of magnitude and show about the same dispersion as for the larger drainage-basins studied (Fig. 2). In Table 2, as in Table 1, most of the values of M are between 0.60 and 0.80.

Writing equation (6) in the form

$$S_c = \sqrt[M]{q_s/K_s} \qquad (7)$$

FIG. 4—CHANNEL STORAGE OUTFLOW-VOLUME RELATION LINES, RALSTON CREEK, IOWA CITY, IOWA



and letting $q_s = 1$, gives

$$S_1 = \sqrt[M]{1/K_s} \tag{8}$$

Equation (8) gives the volume of channel-storage when the channel storage-outflow rate is unity. This quantity will have a maximum value for a full-area rise. Its value for different rises is given in column (7) of Table 1 and column (6) of Table 2.

In the preceding discussion q_s is in cubic feet per second and S_c in c.f.s.-days in Table 1, and in c.f.s.-hours in Table 2. Since one c.f.s.-day equals, very nearly, two acre-feet, channel-storage given in c.f.s.-days can easily be expressed in acre-feet by multiplying these formulae by 2.

Also, since one c.f.s.-day = 0.03719-inch depth on one square mile, channel-storage can easily be expressed in terms of inches depth on the drainage-area. For example, the data for

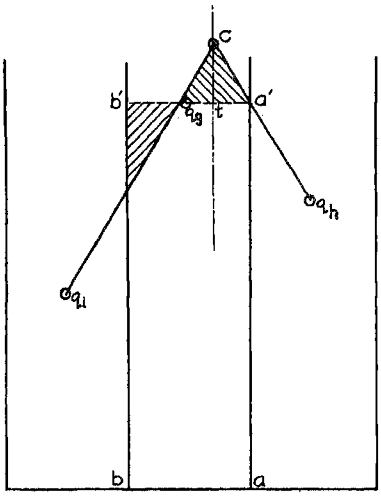


FIG. 5 - CORRECTION OF CREST OF GRAPH PLOTTED FROM DAILY AVERAGE DISCHARGES

Delaware River, given on Figure 2, show that for large floods on that stream the channel-storage may be equivalent to 100,000 acre-feet or 1/2 inch depth on the drainage-basin.

The channel-storage for unit channel-outflow rate as given by equation (8) is apparently a physical characteristic of the storage-capacity of the drainage-net of a given stream-system. Perhaps a better characteristic would be the same quantity reduced to the basis of channel-storage in acre-feet per square mile corresponding to a channel outflow-rate of one c.s.m.

Let S_A be this quantity and A the drainage-area in square miles. Then from equation (8)

$$S_A = 2(S_1/A) \tag{9}$$

Combining equations (6) and (9)

$$S_A = 2A^{(M-1)}/\sqrt[M]{K_s} \tag{10}$$

If applied to full-area rises, this equation should give the maximum channel-storage per square mile of drainage-area corresponding to a channel outflow-rate of one c.s.m.

The data in Tables 1 and 2 afford some indication that K_s and M are inversely related. Owing to the fact that an undetermined portion of the stream-rises included in this study were partial-area rises, no attempt has been made to work out the relation between these quantities or the true values of S_A for full-area rises. The data thus far collected have been presented in the hope that others may pursue similar investigations and publish the results so that these quantities and relations may be further studied and evaluated.

Some studies have been made which indicate that the channel-storage relation once determined for a given stream-system from the recession-side of a graph of a stream-rise can also be applied, approximately at least, throughout the entire stream-rise in case of full-area rises.

Summarized briefly, the above analysis of channel-storage seems to offer the following possibilities of usefulness:

(1) It supplies the missing link in the problem of routing a graph of surface-runoff down a system of stream-channels, approximately at least.
(2) It affords a simple, direct means of determining the volume of channel-storage and the consequent regulating effect of channel-storage on floods.
(3) It appears that the values of the quantities S_1 and S_A are definite physical characteristics of a given drainage-basin.

[Note: The purpose of the present paper being to present facts derived directly from observation, theoretical treatment of the subject has been omitted. Full analytical study of the basis of the equations given has, however, been made and will be published in a subsequent paper.]

Appendix--Correction of crest of runoff-graph plotted from daily discharges

Let q_g = average discharge for crest-day, q_h = average discharge on the higher, and q_ℓ that on the lower, of the two days, one preceding, the other following the crest-day. Then

$$t = 12[(q_g - q_h)/(q_g - q_\ell)]$$

where t is the time of the crest, measured from the dividing line between the crest-day and the higher of the adjacent days. Referring to Figure 5, the crest is on a vertical line through t and is so located that the shaded areas between aa' and bb' and above and below a'b', respectively, are equal. The method is very sensitive and the point can be so located by one or two trials that the shaded areas balance.

A somewhat similar correction is often required at the beginning of an abrupt rise. A smooth curve drawn through the corrected points and through the points corresponding to average discharge for other days will give nearly a correct graph of a rise and permits the inflection-point on the recession-side to be approximately determined and with it the limit of the values of q which may be used in computing storage outflow-rate-volume relations.

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THE CHANNEL-STORAGE METHOD OF DETERMINING EFFLUENT SEEPAGE

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(Published with the approval of the Director, U. S. Geological Survey)

Some years ago the senior author, in collaboration with Norah Dowell Stearns, undertook to make a monthly inventory of the water-supply of the Pomperaug River Basin, in Connecticut, from a study of data obtained by A. J. Ellis from 1913 to 1916. For this purpose approximate determinations or estimates were made of the ground-water runoff, that is, of the part of the daily discharge of the river that consisted of water derived from the zone of saturation by effluent seepage. The determinations or estimates were essentially based on the assumption that after a week of fair weather the storm-water was all discharged and that until the next rain the runoff was essentially all ground-water (U. S. Geol. Surv. Water-Supply Paper 597, pp. 107-116 and pl. 19, 1929). In the discussion of the methods and results of this investigation, the following statement was made (Water-Supply Paper 597, pp. 145-146):

> "In this investigation the estimates of ground-water runoff were based on the discharge of the Pomperaug at Bennetts Bridge--that is, on the discharge during the periods between rains, when there was virtually no direct runoff left in the stream-system. Much better results could be obtained by basing the estimates on periods beginning as soon after rains as all of the direct runoff has reached the streams. With this method the ground-water runoff during any particular day would be the total runoff minus the decrease in stream-storage. The decrease in stream-storage could be estimated by maintaining gages at several points on the trunk-stream and on selected tributaries and making surveys of the stream-system showing the approximate water-areas of different parts of the system at different gage-heights."

Recently a critical study has been made by W. G. Hoyt, L. L. Harrold, F. F. Snyder, and R. C. Cady of several possible methods for determining effluent seepage or ground-water runoff, the results of which have been published in Water-Supply Paper 772 (pp. 111-119). The subject is consequently much better understood now than it was at the time when the Pomperaug study was made. Nevertheless, all these other methods involve intangible assumptions and therefore they should at least be checked by the more tangible channel-storage method. The channel-storage method is based on the simple law that in any period when there is no overland runoff, that is, when no storm- or snow-water is running into the stream-system from the surface, the quantity of water derived by effluent seepage into a given stream-system equals the quantity of water discharged from the stream-system minus the decrease or plus the increase in channel-storage during the same interval. The law thus stated does not take into account the quantities of stream-water that are lost during the period by evaporation from the stream-system. If any part of the stream-system is influent, the law gives the net effluent seepage. It includes the seepage derived from bank-storage and perched ground-water.

Last year a beginning was made on a modest scale in the investigation of the channel-storage method, the stream-basin selected being that of Difficult Run, in Virginia, near Washington, D. C. It has not been possible to advance the investigation very far but certain definite results have been obtained.

Difficult Run rises near Fairfax, Virginia, and flows toward the northeast into the Potomac River. It has a drainage-area of about 58 square miles. Near its mouth the stream has cut a narrow gorge into the granite bed-rock, through which it descends from the plateau to the entrenched Potomac River below Great Falls. Back from the gorge the stream occupies a mature valley with graded sides and a fairly wide flood-plain. A substantial gaging-station with an automatic recorder has been installed near the mouth of the stream and one automatic rain-gage near the head. Three observation-wells with automatic recorders have been maintained in the basin since 1931 or 1932. Several additional shallow wells have been put down recently to study the relation of bank-storage and recharge on the flood-plain to stream-flow.

One channel-storage experiment has thus far been begun, and this is on a headwater-area of about one square mile near Fairfax. The stream-system of this area includes four small branches