Relation Between Ground Water and Surface Water in Brandywine Creek Basin Pennsylvania

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By F. H. OLMSTED *and* A. G. HELY

CONTRIBUTIONS TO STREAM-BASIN HYDROLOGY

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RELATION BETWEEN GROUND WATER AND SURFACE WATER IN BRANDYWINE CREEK BASIN, PENNSYLVANIA

By F. H. Olmsted and A. G. Hely

ABSTRACT

The relation between ground water and surface water was studied in Brandywine Creek basin, an area of 287 square miles in the Piedmont physiographic province in southeastern Pennsylvania. Most of the basin is underlain by crystalline rocks that yield only small to moderate supplies of water to wells, but the creek has an unusually well-sustained base flow. Streamflow records for the Chadds Ford, Pa., gaging station were analyzed; base flow recession curves and hydrographs of base flow were defined for the calendar years 1928–31 and 1952–53. Water budgets calculated for these two periods indicate that about two-thirds of the runoff of Brandywine Creek is base flow—a significantly higher proportion of base flow than in streams draining most other types of consolidated rocks in the region and almost as high as in streams in sandy parts of the Coastal Plain province in New Jersey and Delaware.

Ground-water levels in 16 observation wells were compared with the base flow of the creek for 1952–53. The wells are assumed to provide a reasonably good sample of average fluctuations of the water table and its depth below the land surface.

Three of the wells having the most suitable records were selected as index wells to use in a more detailed analysis. A direct, linear relation between the monthly average ground-water stage in the index wells and the base flow of the creek in winter months was found.

The average ground-water discharge in the basin for 1952–53 was 489 cfs (316 mgd), of which slightly less than one-fourth was estimated to be loss by evapotranspiration. However, the estimated evapotranspiration from ground water, and consequently the estimated total ground-water discharge, may be somewhat high.

The average gravity yield (short-term coefficient of storage) of the zone of water-table fluctuation was calculated by two methods. The first method, based on the ratio of change in ground-water storage as calculated from a winter base-flow recession curve to seasonal change in ground-water stage in the observation wells, gave values of about 7 percent (using 16 wells) and 7½ percent (using 3 index wells). The second method, in which the change in ground water storage is based on a hypothetical base-flow recession curve (derived from the observed linear relation between ground-water stage in the index wells and base flow), gave a value of about 10½ percent. The most probable value of gravity yield is between 7½ and 10 percent, but this estimate may require modification when more information on the average magnitude of water-table fluctuation and the sources of base flow of the creek becomes available.

Rough estimates were made of the average coefficient of transmissibility of the rocks in the basin by use of the estimated total ground-water discharge for the period 1952–53, approximate values of length of discharge areas, and average water-table gradients adjacent to the discharge areas. The estimated average coefficient of transmissibility for 1952–53 is roughly 1,000 gpd per foot. The transmissibility is variable, decreasing with decreasing ground-water stage.

The seeming inconsistency between the small to moderate ground-water yield to wells and the high yield to streams is explained in terms of the deep permeable soils, the relatively high gravity yield of the zone of water-table fluctuation, the steep water-table gradients toward the streams, the relatively low transmissibility of the rocks, and the rapid decrease in gravity yield below the lower limit of water-table fluctuation. It is concluded that no simple relation exists between the amount of natural ground-water discharge in an area and the proportion of this discharge that can be diverted to wells.

INTRODUCTION

In the Piedmont Upland of the Delaware River region the dense crystalline rocks yield only small to moderate supplies of water to wells, but ground-water discharge sustains unusually high base flow in streams. To explain this seeming inconsistency, the Brandywine Creek basin, which lies wholly within the Piedmont Upland section of the Piedmont physiographic province (Fenneman and others, 1930) in southeastern Pennsylvania, was selected as an area in which to study the relation between ground water and surface water.

The part of the basin studied is an area of 287 square miles above the former stream-gaging station at Chadds Ford, Pa. The hydrologic and meteorologic records that were used in the study included daily discharge of Brandywine Creek at Chadds Ford, precipitation at 6 stations, air temperature at 2 stations, and water levels in 16 observation wells (fig. 1). Though not abundant, the water-level data are more numerous in this area than in other parts of the Delaware River region underlain by consolidated rocks. Fieldwork was restricted to a brief reconnaissance of the area and visits to the observation wells. The most detailed part of the study was made for the calendar years 1952–53, because that was the only period for which concurrent records of ground-water levels and streamflow were available, but a monthly water budget was computed for the period 1928–31.

The chief purposes of the study were to define the base flow of Brandywine Creek, to calculate water budgets for the basin and compare these budgets with those of other basins in northeastern United States, to compare the fluctuations and stages of the water table with the base flow of the creek, and, by using both the ground-water and the base-flow data, to estimate the average gravity yield and coefficient of transmissibility of the rocks in the basin. Many of the results are rough estimates at best, and the figures given must be interpreted with great caution. Nevertheless, important conclusions are reached regarding the overall hydrology and especially the ground-water hydrology of the area. Such conclusions add significant insight into the occurrence and movement of water in the weathered and fractured crystalline rocks of the Piedmont Upland of Pennsylvania and help explain the apparent inconsistency that the crystalline rocks discharge large amounts of water to streams but only small to moderate amounts to wells.

HYDROLOGY OF THE BASIN

PHYSICAL CHARACTERISTICS

The channel characteristics of Brandywine Creek, and the general features of the basin, are described by Wolman (1955). Bascom and Stose (1932, 1938) have described the areal geology, which is generalized in figure 1.

The basin is part of a dissected upland which is underlain largely by metamorphic and igneous rocks of Precambrian to early Paleozoic age. Chester Valley, a long, narrow lowland underlain by dolomite and limestone, crosses the middle of the basin in a roughly east-west direction. Gneiss and granitic to ultramafic rocks of Precambrian age predominate north of Chester Valley; schist of early Paleozoic(?) age underlies much of the south half of the basin. Although several types of rocks occur within the basin, the hydrologic characteristics of the rocks, with the possible exception of the dolomite and limestone, are believed to be comparatively uniform for a basin of this size.

A mantle of weathered material of variable thickness has formed on all these rocks. The zone of water-table fluctuation probably lies within the lower part of the weathered material or, locally, within the immediately underlying fractured rock. At most places and at most times the gradient of the water table is toward the streams, which therefore act as ground-water drains.

Most of the soils are permeable and well drained. According to a classification by the U.S. Department of Agriculture (1952) for the basin above Wilmington, Del. (an area 27 square miles larger than that of the present study), 56 percent of the area is underlain by deep, well-drained soils, 21 percent by shallow, well-drained soils, and 23 percent by imperfectly and poorly drained soils. Many of the imperfectly and poorly drained soils are in swampy areas where ground-water discharge occurs.

Land use in the basin is similar to that in most other areas in the Piedmont province. About 51 percent of the area above Wilmington, Del., is cropland and pastureland, 21 percent is woodland, 21 percent is in miscellaneous use, and 7 percent is occupied by highways, roads and streams (U.S. Department of Agriculture, 1952).

The results of this study probably are applicable to other nearby Piedmont Upland basins having similar rocks, soils, weathered zones and patterns of land use.

PRECIPITATION

The average precipitation in Brandywine Creek basin for 1921–50, 44.1 inches, was computed by the U.S. Weather Bureau from an isohyetal map of the Delaware River basin and New Jersey (Parker and others, written communication; Hely, Nordenson, and others, 1961). The areal range in the average annual precipitation within the basin is less than 2 inches.

Daily precipitation in the basin for 1928–31 was calculated as the weighted average of the precipitation at Coatesville, Pa. (weight 0.7), and West Chester, Pa. (weight 0.3). The monthly and annual values are included in table 2. Daily precipitation for 1952–53 was calculated from the records at the six stations shown in figure 1. Each station record was weighted on the basis of a Thiessen net, as follows: Coatesville, 47.6 percent; Morgantown, 19.1 percent; West Chester, 17.1 percent; Chadds Ford, 6.6 percent; Devault (1 mile west), 5.0 percent; and West Grove (1 mile southeast), 4.6 percent. The daily values thus determined are shown in figure 5, and the monthly and annual values are included in table 3.

RUNOFF

Runoff data were obtained from the annual series of U.S. Geological Survey reports, Surface Water Supply of the United States. The effect of regulation is be-

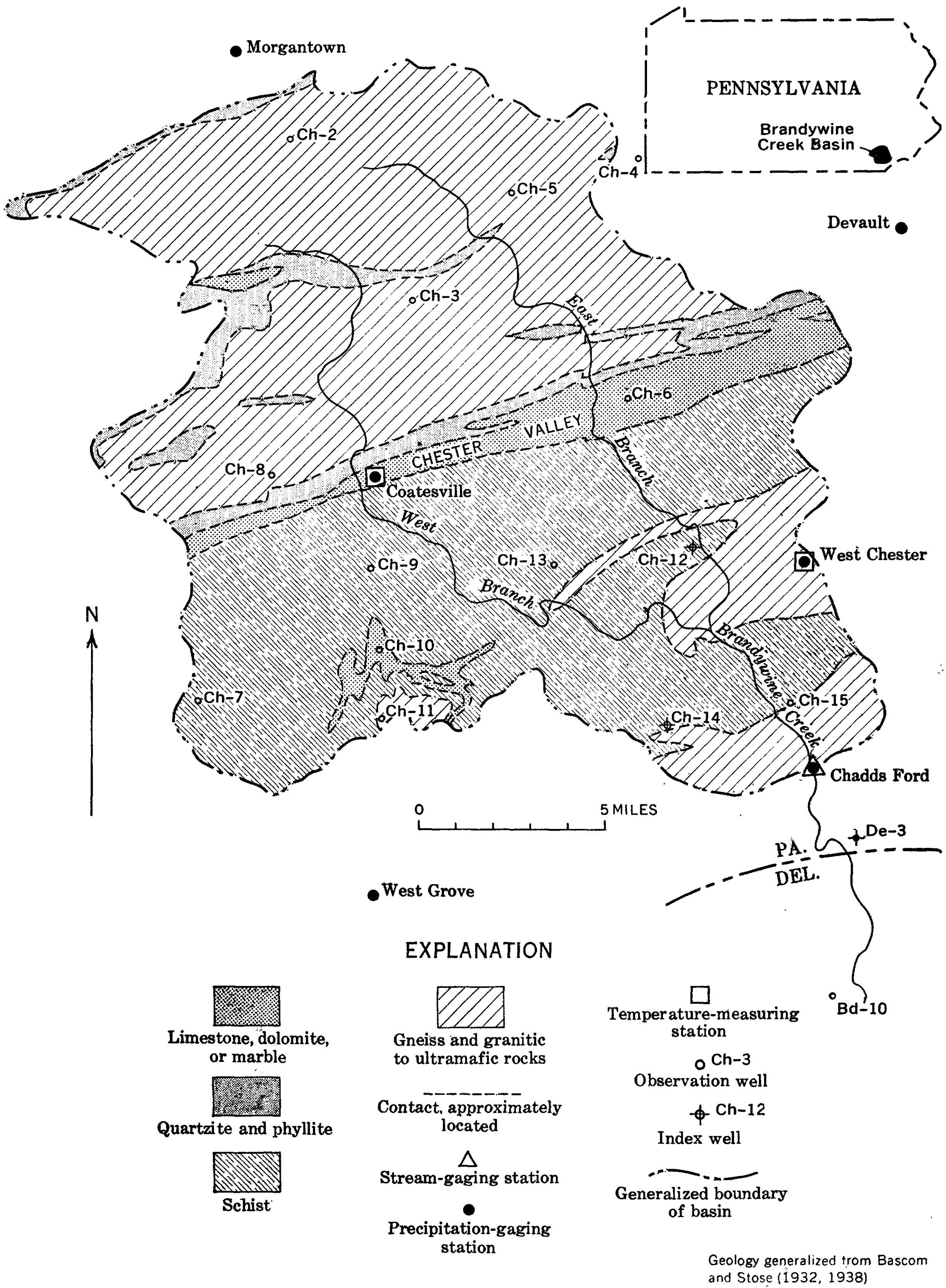

FIGURE 1.—Sketch map of Brandywine Creek basin, showing generalized geology and location of hydrologic and meteorologic stations used in report.

lieved to be slight, although measurable at low flows. The streamflow for any period longer than a week or two is practically equivalent to the runoff (unregulated streamflow). The difference between the two at low flow is shown in figure 2 when the estimated base flow is greater than the streamflow for short periods when there was no evidence of any direct runoff.

Water is diverted from Coatesville Reservoir (capacity 320 million gallons) on Rock Run, a tributary of the West Branch, Brandywine Creek for use by the city of Coatesville. Probably a very small part of this water is used consumptively, however, so that the consequent reduction in streamflow is negligible.

Ground-water withdrawals within the basin probably are very small and are considered negligible.

The average runoff in the basin for the period 1921–50 was 17.5 inches; the areal range in average within the basin was only about 1 inch (Hely, Nordenson, and others, 1961; Parker and others, written communication).

Base-flow recession curves were developed and used as a partial basis for separation of the daily discharge for 1928–31 and 1952–53 into the two components, base flow and direct runoff. Base flow is considered to be a good estimate of the ground-water discharge to streams in this basin because changes in surface storage are very small in relation to total flow.

The calendar years 1928–31 were selected for analysis because the period contained both wet and dry years and average runoff was near the long-term average. The base-flow recession curve used for the period was derived from hydrographs of daily discharge by a graphical method (Parker and others, written communication) and represents primarily summer conditions. The base flow and direct runoff are summarized in table 2.

The base-flow recession curves used for 1952–53 were derived by H. C. Riggs (written communication, 1957) from tables of daily discharge. The method used was slightly different from that used in developing the curve used for 1928–31, and separate curves were developed for winter and summer conditions. The summer curve for 1952–53 has a slightly different shape from that of the curve used for 1928–31 but a similar slope; the winter curve is somewhat flatter. However, the differences in slopes of the recession curves generally apply for only short periods of record and the resulting differences in estimated base flow are probably much smaller than the errors involved in estimating the base flow during periods of direct runoff.

In separating the discharge into its components, records of daily precipitation, snowfall, and temperature were used as guides for interpreting slopes of the hydrograph. Thus there are few sharp rises in the base flow hydrograph in winter, as ground-water levels respond to recharge much more slowly when the ground is frozen or when there is a heavy snow cover. A few minor revisions of the base-flow graph were made on the basis of comparison with hydrographs of ground-water stage described in a later section.

The observed daily discharge and the estimated base flow are shown in figure 2; the monthly values are summarized in table 3. Direct runoff was computed as the difference between total runoff and base flow.

GROUND WATER

Ground-water data for the period 1952–53 consist of water-level measurements in 16 observation wells in or near the basin (U.S. Geological Survey: Annual series on ground-water levels in the United States). These wells are described briefly in table 1; their general location is shown in figure 1. They were selected for measurement because they were unused, were presumably unaffected or only slightly affected by pumping of nearby wells, and consequently probably indicated the natural levels and fluctuations of the water table. All the wells are relatively shallow, ranging in depth from 12 to 40 feet, and all but one (Ch–10) are of the dug rather than the drilled type. Water-level measurements were made by voluntary observers, usually the owners or tenants, except those for wells Ch–4, Ch–7, Ch–13, and Bd–10, which were made by the Geological Survey. A water-stage recorder was installed on well De–3 for brief periods.

Not all the 16 observation wells were measured often enough to be of value in interpreting short-period fluctuation; only 6 or 7 were measured at weekly intervals or immediately after storms.

Wells De–3, Ch–13, and Ch–14 were selected as index wells to use in a more detailed analysis. The selection was based on an examination of the water-level records of all the observation wells and a comparison of these records with the base-flow hydrograph of the creek and meteorological data for nearby stations.

In an area of 287 square miles, 16 well records, only 3 of which were analyzed in detail, may not provide a representative sample of water table depths and fluctuations. However, the areal distribution of the 16 wells shown in figure 1 is reasonably good. The major rock types are represented in roughly the proportions of their areal extent in the basin; three wells are in gneiss, five in granitic to ultramafic rocks, six in schist, one in dolomite, and one in marble. The topographic position also is believed to be good; elevations from ridgetops to valley bottoms are represented.

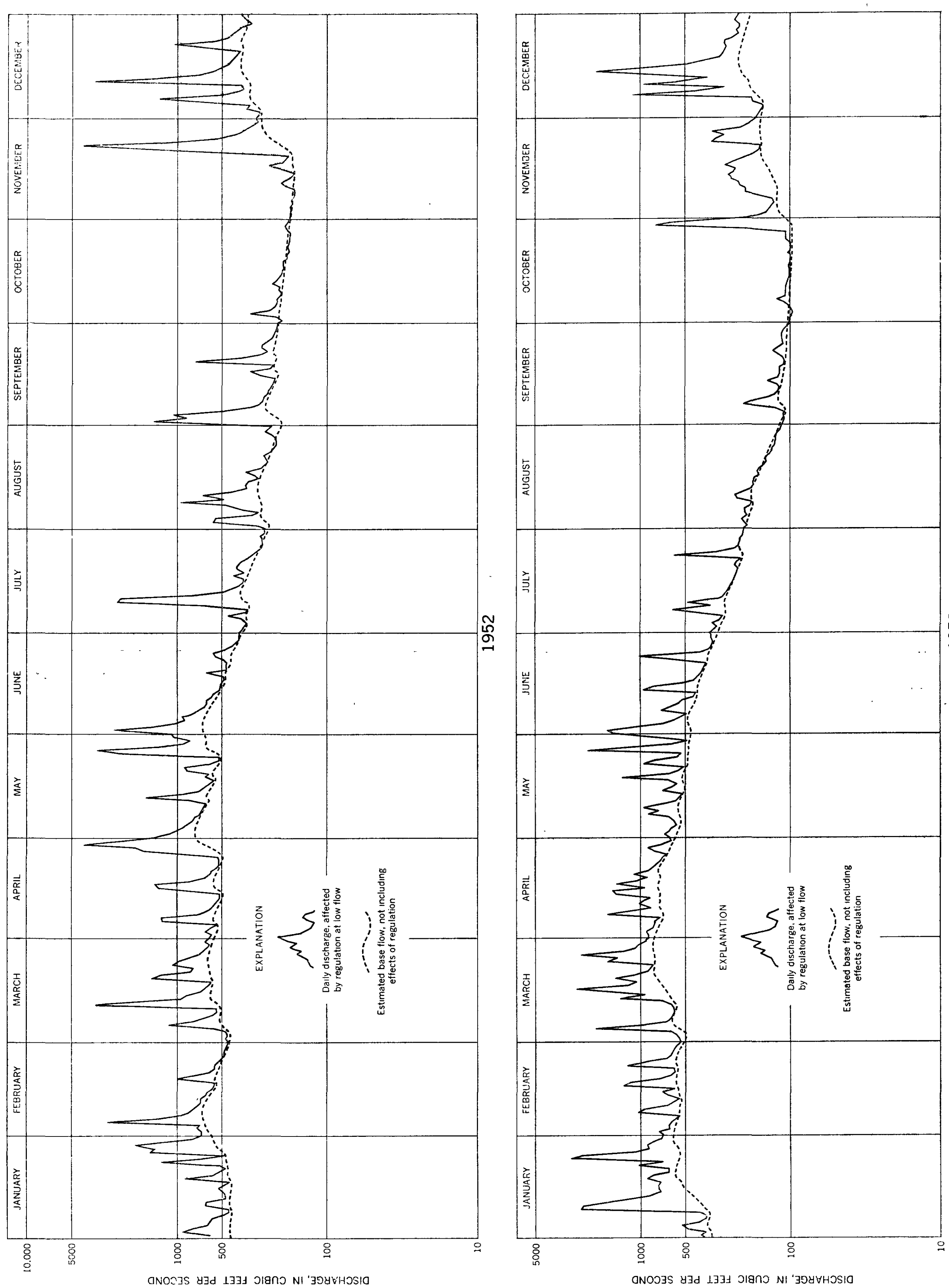

FIGURE 2.—Hydrograph of Brandywine Creek at Chadds Ford, Pa., 1952–53.

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TABLE 1.—*Description of observation wells in or near Brandywine Creek basin measured in 1952–53*

Well	Owner or tenant (1953)	Approximate location (fig. 1)	Latitude (N.)			Longitude (W.)			Type of well	Diameter (inches)	Depth (feet)
			Degrees	Minutes	Seconds	Degrees	Minutes	Seconds			
Bd-10	F. B. Crowninshield	6 mi S. of Chadds Ford	39	47		75	34		Dug	42	23
Ch-2	L. R. Shingle	3 mi SE. of Morgantown	40	06	54	75	51	20	do	36	15
3	F. M. Anderson	5 mi N. of Coatesville	40	03	20	75	48	03	do	36	29.5
4	C. R. Rebmann, Jr	7 mi W. of Devault	40	06	44	75	40	21	do	48	29
5	Richard Cadbury	8 mi NE. of Coatesville	40	05	42	75	44	55	do	30	12
6	J. J. Englerth	6 mi NW. of West Chester	40	00	47	75	41	32	do	60	20.5
7	D. L. Gibbs	7 mi NW. of West Grove	39	53	43	75	54	50	do	48	39.5
8	John Robinson	3 mi W. of Coatesville	39	59	00	75	52	19	do	36	20.5
9	C. R. Young	2 mi S. of Coatesville	39	57	03	75	49	14	do	30	25
10	R. J. Kleberg, Jr	5 mi S. of Coatesville	39	54	50	75	48	48	Drilled	6	33.5
11	J. E. Ryan	5 mi N. of West Grove	39	53	15	75	48	46	Dug	24	20
12	T. P. Harney	3 mi W. of West Chester	39	57	18	75	39	22	do	30	38.5
13	E. S. Barr	5 mi SE. of Coatesville	39	56	57	75	43	38	do	24	18.5
14	J. T. Crossland	4 mi NW. of Chadds Ford	39	53	12	75	40	25	do	36	25.5
15	W. C. Appleton	2 mi N. of Chadds Ford	39	53	41	75	36	41	do	42	38.5
De-3	M. S. Ebert	2 mi SE. of Chadds Ford	39	50	42	75	34	18	do	42	20

Well	Approximate altitude of land surface (feet)	Topography of well site	Water-bearing material	Frequency of water-level measurements	Average depth to water 1952–53 (feet)
Bd-10	250	Near base of slope above pond	Weathered gabbro	Monthly	12.45
Ch-2	635	Gentle slope	Granodiorite or quartz monzonite	Weekly	8.02
3	610	do	Granodiorite	Irregular	23.26
4	560	About midway up hillside	Quartz monzonite	do	23.39
5	570	Broad ravine bottom near seep	do	Weekly	4.12
6	265	Gentle slope in valley	Dolomite	Irregular	15.06
7	575	Summit of ridge	Schist	do	32.47
8	655	Gentle slope	Gneiss	do	12.56
9	510	Near top of broad ridge	Schist	do	15.80
10	313	Small rise in valley bottom	Marble	do	11.22
11	535	Gentle slope	Gneiss	Weekly or monthly	11.71
12	250	Hillside near floodplain of creek	do	Weekly	33.64
13	390	Hillside above small ravine	Schist	Irregular	14.46
14	345	Nose of gentle ridge	do	Weekly	19.99
15	245	Ridge above floodplain of creek	do	Irregular	26.33
De-3	285	Small swale above ravine on slope	do	Weekly	14.61

The three index wells are in the southeastern part of the area (fig. 1), but their records probably represent average water-table fluctuations in the basin fairly accurately. However, some of the shorter period fluctuations, especially those associated with local storms, may not be accurately indicated.

The topographic and hydrologic settings of the index wells are shown in figure 3. Each of the wells is on a southeastward-facing slope several hundred feet from a small perennial stream that presumably is the discharge outlet for the ground water moving past the well. The water table is sufficiently deep at each well, so that ground-water discharge by evapotranspiration probably is negligible, although such discharge certainly occurs in the stream valleys several hundred feet downslope from the wells. The slope of the water table and its highest and lowest positions during 1952–53 are indicated for each well in figure 3. The average slope of the water table toward the discharge outlet was about 140 feet per mile at well Ch–12, 290 feet per mile at Ch–14, and 410 feet per mile at De–3. All three wells are in strongly weathered rock, although well Ch–12, which is 500 feet southwest of a prominent outcrop, may penetrate relatively unweathered rock near the bottom. The parent rock is mica schist or gneiss of the Wissahickon formation at wells De–3 and Ch–14 and the Baltimore gneiss at Ch–12.

Hydrographs for each of the three wells for 1951–55 and daily precipitation at nearby stations were plotted. A ground-water level recession curve was developed for each well by synthesizing water-level recessions for periods of little or no precipitation in the same manner that a base-flow recession curve is derived from a hydrograph of daily discharge. Each well was found to have a characteristic recession curve that showed little or no seasonal variation, although there were some irregular variations.

Average depth to the water table is largely a function of topographic position with respect to discharge outlets. Since the sampling of topographic position probably is good, the average depth to water in the observation wells probably represents approximately the average depth to the water table in the basin. The depth to water, obtained by averaging the seasonal highs and lows listed in table 6, was 17.45 feet in the 16 observation wells and 22.75 feet in the 3 index wells. By comparison, in the crystalline rocks of northern Delaware, just south of Brandywine Creek basin, Ward and Ras-

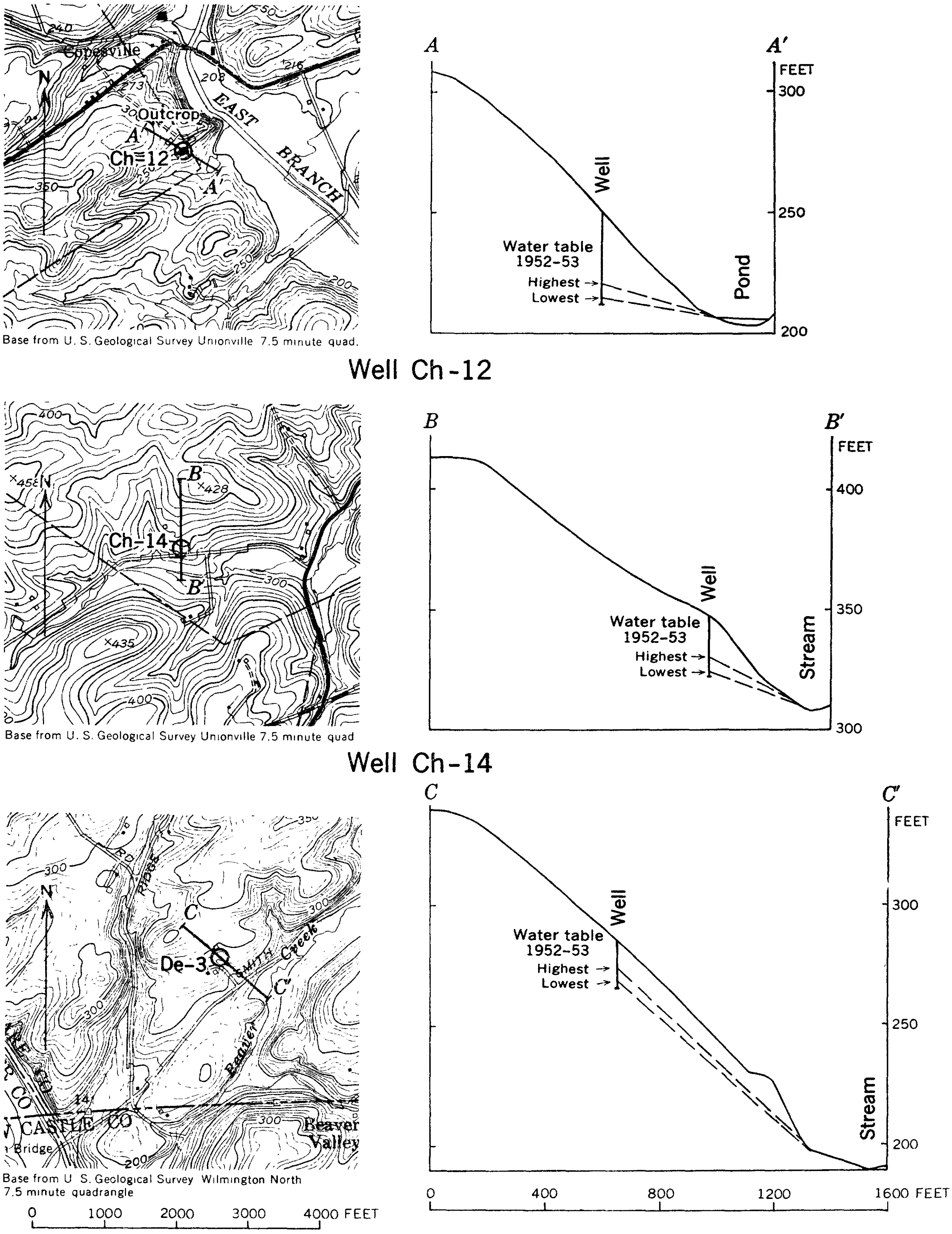

Figure 3.—Topographic and hydrologic settings of index wells De-3, Ch-4, and Ch-12.

mussen (*in* Rasmussen and others, 1957) reported average depths to water as listed below:

Formation	Number of wells	Average depth to water (feet)
Cockeysville marble	18	13.1
Wissahickon formation (schist in fig. 1)	149	23.6
Gabbro	63	18.1
Average, 230 wells		21.7

The magnitude of seasonal water-table fluctuation is determined partly by topographic position but largely by the nature of the rock; the smaller the specific yield (proportion of drainable void space to total volume), the greater the fluctuation. It is difficult to evaluate the accuracy of an estimate of average seasonal fluctuation of the water table in the basin based on records for only 16 wells, and little information in adjacent areas is available for comparison. It is assumed that the average seasonal fluctuation for the basin is indicated by the average of the 16 wells; this average was 5.75 feet for 1952–53 (table 6). The average seasonal fluctuation in the 3 index wells was 5.33 feet for the same period. These estimates of average seasonal fluctuation of the water table may require substantial modification when more water-level data become available for the region.

WATER BUDGETS

The data on precipitation and runoff described in the preceding sections are summarized in tables 2 and 3, which give examples of water budgets for the basin. The general budget equation (inflow equals outflow plus net increase in storage) is simplified by the assumption that net increase in storage for a period of years is negligible. For Brandywine Creek basin above Chadds Ford, inflow is assumed to be equal to the precipitation, and outflow is assumed to be equal to runoff plus evapotranspiration because underground inflow and outflow are probably negligible.

Average annual evapotranspiration, which cannot be measured directly for large areas, may be estimated as the long-term difference (commonly called water loss) between precipitation and runoff. Although 1 year is too short a period for accurate determination of the portion of the water loss that is not appreciably affected by change in storage, values of precipitation minus runoff for each year of the budget period are shown in tables 2 and 3 to indicate the relative magnitude and variability of this quantity. Thus, for the 6 budget years the maximum difference is about 10.4 inches as compared to 22.0 inches for precipitation and 18.3 inches for runoff.

TABLE 2.—*Monthly water budget for Brandywine Creek basin, 1928–31*

Month	Precipitation (inches)	Precipitation minus runoff (inches)	Runoff (inches)	Direct runoff (inches)	Base flow	
					(Inches)	(Percentage of runoff)
1928						
January	3.47		1.82	0.33	1.49	82
February	4.32		3.64	1.65	1.99	55
March	2.92		2.31	.30	2.01	87
April	5.41		2.57	1.01	1.56	61
May	2.62		2.16	.31	1.85	86
June	11.33		3.06	1.29	1.77	58
July	6.39		2.34	.77	1.57	67
August	7.73		2.26	.79	1.47	65
September	4.81		1.77	.33	1.44	81
October	.69		1.27	.04	1.23	97
November	2.19		1.05	.08	.97	92
December	1.56		1.06	.18	.88	83
Year	53.44	28.13	25.31	7.08	18.23	72.0
1929						
January	3.59		1.42	.58	.84	59
February	4.48		2.24	1.46	.78	35
March	2.78		2.70	1.01	1.69	63
April	6.58		2.80	1.13	1.67	60
May	4.78		2.33	.52	1.81	78
June	3.10		1.29	.16	1.13	88
July	2.85		.83	.19	.64	77
August	4.63		.65	.19	.46	71
September	3.84		.43	.08	.35	81
October	5.93		1.25	.63	.62	50
November	3.36		1.32	.46	.86	65
December	2.76		1.14	.26	.88	77
Year	48.68	30.28	18.40	6.67	11.73	63.8
1930						
January	2.99		1.17	.19	.98	84
February	3.03		2.01	.70	1.31	65
March	3.05		2.07	.40	1.67	81
April	2.48		1.61	.29	1.32	82
May	4.11		1.17	.29	.88	75
June	3.52		.79	.15	.64	81
July	2.90		.58	.16	.42	72
August	1.77		.33	.03	.30	91
September	4.58		.42	.14	.28	67
October	.86		.29	.02	.27	93
November	1.94		.39	.09	.30	77
December	2.56		.56	.26	.30	54
Year	33.79	22.40	11.39	2.72	8.67	76.1
1931						
January	2.26		.92	.52	.40	43
February	1.99		.79	.35	.44	56
March	4.99		.99	.43	.56	57
April	2.68		.91	.29	.62	68
May	6.76		1.33	.69	.64	48
June	3.51		1.14	.56	.58	51
July	8.20		1.71	1.07	.64	37
August	5.66		1.15	.59	.56	49
September	1.89		.60	.11	.49	82
October	1.46		.47	.06	.41	87
November	.97		.42	.04	.38	90
December	2.17		.56	.13	.43	77
Year	42.54	31.55	10.99	4.84	6.15	56.0
4-year average	44.61	28.09	16.52	5.32	11.20	67.8

The direct runoff and base flow also indicate the relative magnitude of ground-water discharge to streams. The average base flow for the 6 budget years is about 67 percent of the total runoff. This compares with about 64 percent for North Branch Rancocas Creek in the coastal plain of New Jersey (computed for this study); 74 percent [1] for Beaverdam Creek, a small stream in the coastal plain of Maryland (Rasmussen

[1] Modified slightly in order to account for net changes in storage within the budget periods.

and Andreasen, 1959); 42 percent for Perkiomen Creek (Parker and others, written communication), a relatively flashy stream in the Triassic lowland in Pennsylvania; and 44 percent [1] for the Pomperaug River basin (Meinzer and Stearns, 1929), a small stream in Connecticut.

TABLE 3.—*Monthly water budget for Brandywine Creek basin, 1952–53*

Month	Precipitation (inches)	Precipitation minus runoff (inches)	Runoff (inches)	Direct runoff (inches)	Base flow	
					Inches	Percentage of runoff
1952						
January	5.05	----------	2.94	1.24	1.70	58
February	2.13	----------	2.78	.67	2.11	76
March	5.45	----------	3.49	1.25	2.24	64
April	7.53	----------	3.89	1.77	2.12	54
May	6.39	----------	3.80	1.30	2.50	66
June	2.59	----------	2.59	.60	1.99	77
July	6.29	----------	2.02	.72	1.30	64
August	4.65	----------	1.40	.38	1.02	73
September	5.01	----------	1.41	.53	.88	62
October	.82	----------	.80	.03	.77	96
November	5.51	----------	1.74	1.03	.71	41
December	4.36	----------	2.42	1.08	1.34	55
Year	55.78	26.50	29.28	10.60	18.68	63.8
1953						
January	5.82	----------	3.85	1.97	1.88	49
February	2.64	----------	2.59	.57	2.02	78
March	6.41	----------	4.29	1.59	2.70	63
April	4.34	----------	3.59	.84	2.75	77
May	6.30	----------	3.03	1.00	2.03	67
June	2.78	----------	2.17	.64	1.53	71
July	3.52	----------	1.15	.19	.96	83
August	1.16	----------	.66	.02	.64	97
September	2.24	----------	.47	.04	.43	91
October	4.05	----------	.59	.19	.40	68
November	2.28	----------	.79	.26	.53	67
December	4.46	----------	1.63	.89	.74	45
Year	46.00	21.19	24.81	8.20	16.61	66.9
2-year average	50.89	23.85	27.04	9.40	17.64	65.2

The North Branch Rancocas Creek basin is underlain by predominantly unconsolidated and semiconsolidated sandy sediments; the Beaverdam Creek basin by rather uniformly sandy unconsolidated deposits; the Parkiomen Creek basin in large part by shale, argillite, sandstone, and diabase; and the Pomperaug River basin by crystalline rocks and some sandstone, conglomerate, and shale, all of which are covered by a discontinuous mantle of glacial deposits. The base flow characteristics of Brandywine Creek are more characteristic of streams draining the unconsolidated deposits of the Coastal Plain than of most other streams in areas underlain by consolidated rocks.

Factors other than the geology of a basin affect the amount of base flow. Differences in topography, soils, and vegetation affect infiltration capacity, which in part determines the amount of base flow; some of these factors can be modified considerably by land-management techniques such as those reported extensively in the literature of the Forest Service and Soil Conservation Service of the U.S. Department of Agriculture. A summary and evaluation of most of these techniques is presented by Colman (1953).

However, the writers believe that the great differences in runoff characteristics between the Brandywine Creek basin and the other two basins in consolidated rock must be due chiefly to geological differences.

RELATION BETWEEN GROUND-WATER STAGE AND BASE FLOW OF BRANDYWINE CREEK

FLUCTUATIONS OF WATER TABLE COMPARED TO FLUCTUATIONS OF BASE FLOW

The hydrographs of five wells that were measured weekly or immediately after storms are compared with the hydrograph of base flow of Brandywine Creek in figure 4. The depths to the water table in these wells represent the full range in all 16 observation wells, and the fluctuations in water level in the 5 wells indicate the known variation in fluctuation of the water table throughout the basin.

The general similarity of the fluctuations in both the wells and the creek is apparent. All the hydrographs rise after the first heavy precipitation in the fall and decline during the growing season from late spring to fall. The hydrographs differ in detail, however. Except for the hydrograph of well Ch–5, the short-period fluctuations decrease in number and amplitude with increasing depth to water. In general, this is to be expected because increasing the distance from land surface to water table increases the time of transit and the volume of water in transit (gravity water).

Factors other than depth to water also influence the fluctuations. The permeability of the materials in the zone between the water table and the land surface in part determine the transit time of recharge to the water table; variations in permeability below the water table, and the relation of the site to nearby discharge outlets are also important. For example, well Ch–5, which has the shallowest depth to water of any of the five wells in figure 4, is in the bottom of a small ravine only a few feet upslope from a wet-weather spring, and the hydrograph reflects the local conditions in the zone supplying the spring discharge. The water level in the well remains very nearly at the level of the discharge outlet (the spring) as long as the discharge continues (which was much of the year in both 1952 and 1953), but recedes rapidly after the spring ceases to flow.

The hydrographs of the three index wells were averaged to form a composite hydrograph, which is compared with the base-flow hydrograph of the creek in figure 5. In order to simplify the computations of

[1] See footnote page 8.

gravity yield (discussed on pages A-16–A-18), average ground-water stage is shown in inches above an arbitrary datum 28 feet below the land surface. Scales for ground-water stage and base flow were selected to produce similar amplitudes of seasonal fluctuation of the two curves.

The principal differences between the hydrographs are the relatively greater amplitude and greater number of short-period fluctuations of the base flow, especially during winter and spring storm periods, and the delay of as much as a week between the occurrence of the peaks and troughs of the ground-water stage and the equivalent peaks and troughs of the base flow. A less obvious, but equally important, difference is the slower winter recovery in water level in the wells compared to the more rapid increase in base flow in the creek.

As emphasized by Hurst and Brater (1941), the water table gradient adjacent to a stream is one of the factors that determines the rate of ground-water discharge to the stream (which approximately equals base flow);

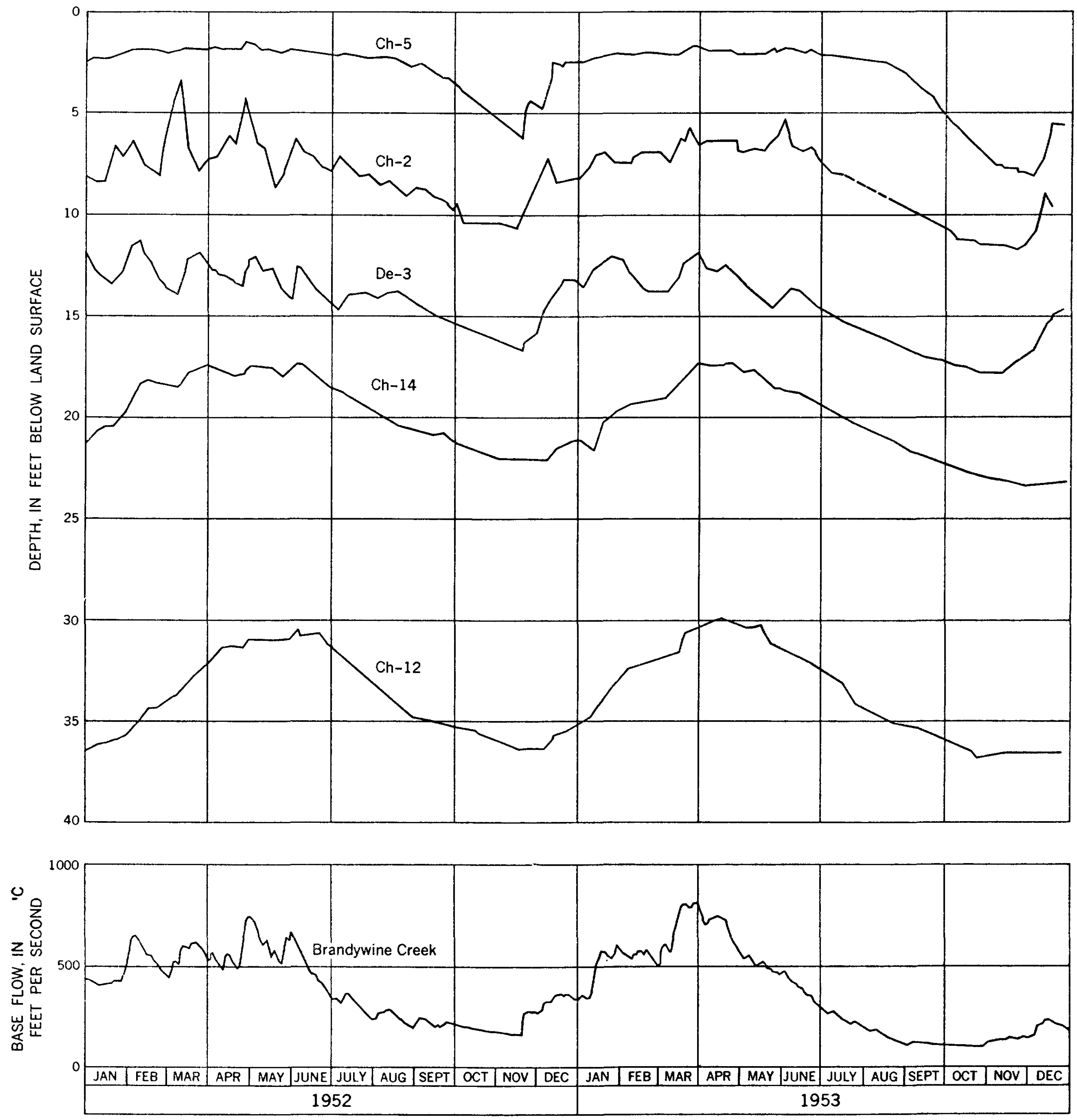

FIGURE 4.—Hydrographs of five wells compared with hydrograph of base flow of Brandywine Creek at Chadds Ford, 1952–53.

consequently, the base-flow hydrograph should most closely resemble the hydrographs of wells adjacent to the creek, where the water table is relatively shallow. However, the index wells are several hundred feet upslope from streams where the water table is much deeper: because of this greater depth, recharge water takes a longer time to reach the water table at the index wells than it does near the streams. Consequently the ground-water stage at the index wells does not respond to precipitation as rapidly as it does near the stream where its effect on base flow is rapid.

Evapotranspiration from ground water affects the base-flow hydrograph but does not materially affect the ground-water stage where the depth to the capillary fringe above the water table is greater than the maximum depth of plant roots. In Brandywine Creek basin evapotranspiration from ground water probably is limited almost entirely to stream valleys. Part of the

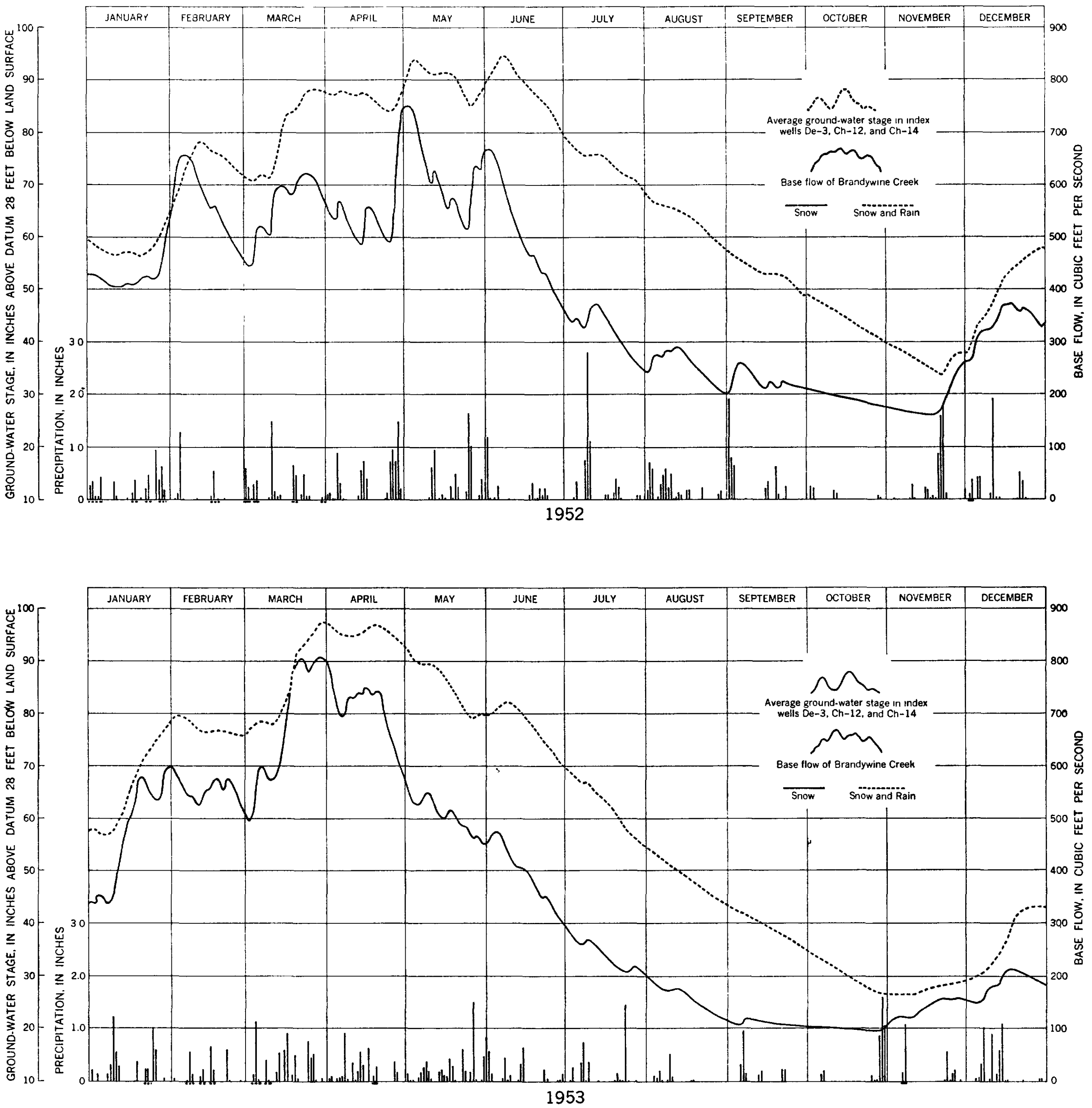

FIGURE 5.—Ground-water stage, base flow, and daily precipitation in Brandywine Creek basin, 1952–53.

ground water moving toward the stream from higher areas is intercepted and transpired by riparian plants. Small summer showers may replenish soil moisture and lessen the demand on ground water, thereby allowing more of the ground-water discharge to reach the stream, even though no recharge to ground water has occurred. This probably accounts for some of the increases in base flow that commonly occur after storms when rainfall is insufficient to produce recharge to ground water.

Some of the differences indicated by the comparison of the hydrographs may not be real because of uncertainities in the delineation of base flow during periods of direct runoff.

GROUND-WATER STAGE IN INDEX WELLS COMPARED TO BASE FLOW

Recession curves of ground-water levels in the three index wells (De—3, Ch—12, Ch—14) were combined to form a composite curve (fig. 6) which is considered representative of the entire basin.

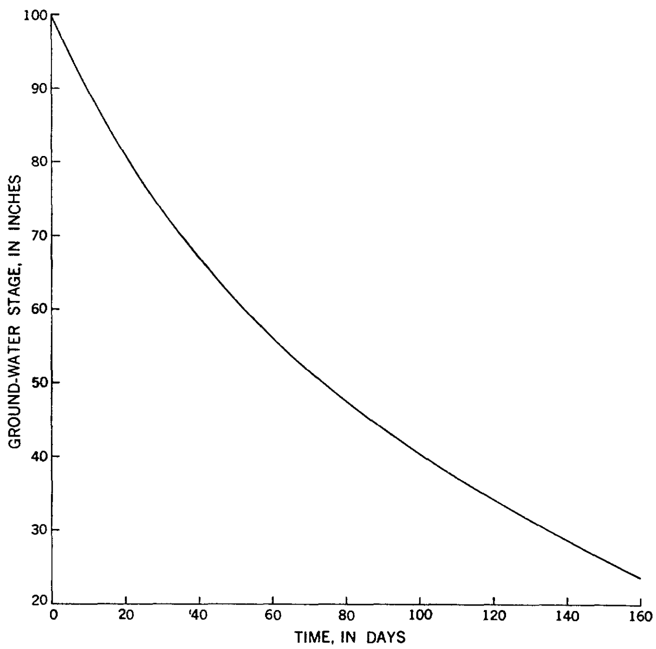

FIGURE 6.—Composite curve of ground-water-level recession for three index wells.

A relation between base flow and ground-water stage is shown in figure 7. This figure shows the monthly base flow (table 3) plotted against monthly average ground-water stage in the three index wells (determined from the hydrographs in fig. 5), and a line representing the relation between them for the condition of no evapotranspiration from ground water. The points for the 24 months represented form a flat loop (actually two similar loops—one for each year) sloping down to the left. The points for the winter months lie very near a straight line marking the lower right side of the loop. The points for the summer months also lie near a straight line having a similar slope but displaced to the left. The fall and spring months lie between these extremes.

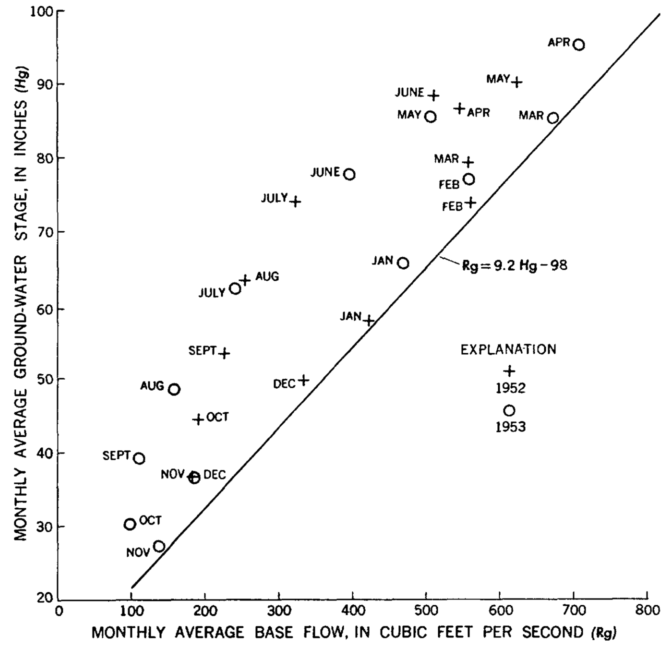

FIGURE 7.—Relation of monthly average base flow to ground-water stage.

Qualitatively this relation suggests not only that the higher ground-water stages are associated with higher base flows, as would be expected, but also that the base flow corresponding to no evapotranspiration from ground water (or total ground-water discharge) is directly proportional to ground-water stage. On the basis of this plot, it is assumed that the total monthly ground-water discharge is a direct linear function of monthly average ground-water stage, and that the displacement of the monthly points to the left of a relation line representing no loss is the measure of loss by evapotranspiration. For example, if the monthly average ground-water stage were the same for July and January, but the base flow in July averaged 250 cfs less than that in January, the evapotranspiration from ground water would be 0.87 cfs per square mile, or 1.00 inch more in July than in January.

The slope of the relation line was determined in the following manner. Pairs of months having about the same average potential evapotranspiration were selected on the basis of data for Seabrook, N.J. (Thornthwaite and Mather, 1955, table 2.1). The ratio of change in base flow to change in ground-water stage for each pair of months is shown on facing page:

Pairs of months	Ratio		
	1952	1953	Average
February–December	9.5	9.2	9.4
March–November	8.8	9.2	9.0
April–October	8.4	9.3	8.8
May–September	10.8	8.5	9.6
June–August	10.3	8.1	9.2
Average	9.6	8.9	9.2

The average ratio, 9.2, is the slope of the relation line.

A period much longer than 2 years, and points differing appreciably in stage and flow for the same months rather than different months would have been preferable, but the computed slope probably is reasonably close to the correct value.

A trial location of the line was determined on the basis of losses for January and July 1952. The equation of the line in this location (fig. 7) is

$$R_g = 9.2\,H_g - 98$$

in which R_g is the average ground-water discharge to streams—assumed equal to base flow in cubic feet per second, and H_g is the average ground-water stage in inches. Table 4 shows the computation of losses for each month based on this line.

TABLE 4.—*Monthly evapotranspiration from ground water in Brandywine Creek basin, 1952–53, as calculated from difference between base flow computed from hydrograph and base flow at zero evapotranspiration from ground water*

Month	Weighted average ground-water stage (inches)	Base flow computed for zero evapotranspiration from ground water (cfs)	Base flow computed from hydrograph (cfs)	Evapotranspiration from ground water (Difference between computed base flows)	
				(cfs)	(inches)
1952					
January	58.0	435	423	12	0.05
February	73.9	582	561	21	.08
March	79.3	632	557	75	.30
April	86.7	700	545	155	.60
May	90.1	731	623	108	.43
June	88.5	716	511	205	.80
July	74.1	584	323	261	1.05
August	63.5	486	254	232	.93
September	53.5	394	226	168	.65
October	44.5	311	191	120	.47
November	36.7	240	183	57	.22
December	49.9	361	334	27	.11
Year					5.69
1953					
January	65.7	506	468	38	.15
February	77.1	611	557	54	.20
March	85.4	688	672	16	.06
April	95.3	779	707	72	.28
May	85.6	690	505	185	.74
June	77.7	617	394	223	.87
July	62.5	477	240	237	.95
August	48.6	349	159	190	.76
September	39.2	263	111	152	.59
October	30.2	180	99	81	.33
November	27.3	153	137	16	.06
December	36.7	240	185	55	.22
Year					5.21
2-year average		489			

TABLE 5.—*Adjusted monthly evapotranspiration from ground water in Brandywine Creek basin, 1952–53*

Month	A Average temperature (°C)	B Percentage of annual daylight	A×B	Percentage of 2-year total A×B	Adjusted evapotranspiration from ground water (inches)	Evapotranspiration from ground water (from table 4) (inches)
1952						
January	1.8	6.69	12	0.45	0.05	0.05
February	2.0	6.93	14	.52	.06	.08
March	3.8	8.26	31	1.15	.12	.30
April	11.8	8.91	105	3.91	.43	.60
May	15.3	9.95	152	5.66	.62	.43
June	22.2	10.05	223	8.31	.91	.80
July	25.0	10.20	255	9.50	1.04	1.05
August	22.7	9.56	217	8.09	.88	.93
September	19.0	8.39	159	5.92	.64	.65
October	10.8	7.79	84	3.13	.34	.47
November	6.5	6.76	44	1.64	.18	.22
December	1.9	6.51	12	.45	.05	.11
Year		100.00	1,308	48.73	5.32	5.69
1953						
January	2.1	6.71	14	.52	.06	.15
February	2.8	6.69	19	.71	.08	.20
March	5.8	8.28	48	1.79	.19	.06
April	10.6	8.93	95	3.54	.39	.28
May	18.0	9.98	180	6.71	.73	.74
June	21.5	10.07	217	8.08	.88	.87
July	24.0	10.23	246	9.17	1.00	.95
August	22.8	9.58	218	8.12	.88	.76
September	19.5	8.41	165	6.15	.67	.59
October	14.0	7.81	109	4.06	.44	.33
November	7.0	6.78	47	1.75	.19	.06
December	2.8	6.53	18	.67	.07	.22
Year		100.00	1,376	51.27	5.58	5.21
2 year period			2,684	100.00	10.90	10.90

Computation of adjustments to the computed losses is shown in table 5. The product of monthly temperature (obtained from publications of the U.S. Weather Bureau) and percentage of annual daylight (computed from standard ephemeris tables) is shown for each month and year and for the 2-year period. The percentage of this 2-year total that occurred during each month is multiplied by the 2-year total loss (from table 4) to obtain the adjusted loss.

Comparison of the adjusted losses (table 5) with those derived from the relation in figure 7 indicates that the trial location of the relation line is satisfactory. This line is used in one method of estimating the gravity yield of the zone of water-table fluctuation, described in a following section. The total ground-water discharge for the 2-year budget period was calculated as the sum of the total loss and the total base flow, and the yearly average for this period was used to calculate the approximate coefficient of transmissibility of the rocks of the basin. The average discharge for the 2-year period was 489 cfs (table 4), or 316 mgd. Slightly less than one-fourth of this total was loss by evapotranspiration; the remainder was base flow.

SIGNIFICANCE OF OBSERVED RELATIONS

Three index wells provide a very small sample of water-table fluctuations in an area of 287 square miles, and water levels in other wells might relate to the flow of the creek quite differently from those in the three index wells. A different period of analysis—particularly one that included relatively dry years—might have altered not only the quantitative relations of the water levels in the index wells to base flow but the qualitative conclusions as well. For example, Harrold (1934) in his study of a small basin in the Piedmont province in Maryland, found a difference in the relations between ground-water stage and base flow in wet and dry seasons. No such difference is apparent in the records for Brandywine Creek basin in 1952–53, but both 1952 and 1953 were relatively wet years. A dry year or even a normal year might have shown a difference. However, at least part of the difference in the stage-flow relations noted by Harrold was due to the difference in evapotranspiration from ground water. This factor is accounted for in the Brandywine data because stage-flow relations are estimated for conditions of no evapotranspiration from ground water.

The linear relation (fig. 7) between ground-water stage in the index wells and base flow could be accidental. If this relation were based on ground-water stage from the composite recession curve in figure 6 and base flow at the corresponding time from the winter recession curve in figure 8, it would not be linear but curved, the slope increasing with decreasing stage.

A hypothetical base-flow recession curve that represents no loss by evapotranspiration from ground water was computed from the ground-water-level recession curve (fig. 6) and the relation (fig. 7) corresponding to no loss. Thus from figure 6, at a time of 80 days the ground-water stage is 47.5 inches, which, when inserted in the equation of the relation line (fig. 7), gives $(9.2)(47.5)-98=338$ cfs as the corresponding base flow. This hypothetical recession is shown in figure 8 and was used in a later section in the computation of gravity yield by method B.

The hypothetical recession curve is flatter than the winter recession curve at the upper end and steeper at the lower end (fig. 8), partly because the winter recession curve includes effects of some evapotranspiration from ground water. Probably the greater part of the discrepancy results from selection of the index wells to indicate average water-table fluctuations throughout the basin rather than the water-table gradients adjacent to the streams (ground-water discharge outlets). When the water table is relatively deep, as at index well Ch–12, the response to precipitation is delayed as much as several weeks. Thus, if wells located nearer streams and having shallower depths to water had been used, the recharge from precipitation might have occurred more quickly, and would have been sustained from increased flow in the aquifer, with the result that the water-table fluctuations would have more nearly resembled those of the creek. In particular, the rise in water table during the winter would have been more rapid, the average stages for the winter months, especially December, January and February, would have been relatively higher than those in figure 7, and consequently the relation line between ground-water stage and base flow for the condition of no evapotranspiration from ground water would be curved and would increase in slope at lower stages. The estimated evapotranspiration from ground water would then be less than that estimated in tables 4 and 5. A base-flow recession curve derived from such a curved relation between ground-water stage and base flow would resemble the winter recession curve more nearly than the hypothetical recession curve in figure 8.

A comparison of local relief and depth to the water table in wells shows that, at most of the wells at least, the water table is considerably higher than the nearby stream where ground-water discharge occurs, even in the fall when the water table is lowest. If the winter relation between base-flow and ground-water stage for the three index wells and the creek (fig. 7) held true beyond the lower limits observed for 1952–53, zero flow would occur at a ground-water stage 10.7 inches above the 28-foot depth datum. Yet the water table corresponding to this stage in each of the three wells would still be substantially above the nearby stream, particularly in well De—3 (fig. 3). This condition cannot, of course, really exist so long as any drainable water remains in the rocks. Therefore, the slope of the relation line in figure 7 would have to steepen sharply below the limits of the 1952–53 data (stage of about 27 inches). However, a more gradual steepening appears more reasonable, and would be consistent with the curved stage-flow relation line suggested above. It is concluded that the drainable void space in the rocks decreases rapidly, perhaps abruptly, below the lower limit of water-table fluctuation.

Meinzer and Stearns (1929, p. 129) discuss the significance of the ground-water stage corresponding to zero flow. They state:

> As the water table comes near this point the ground-water runoff approaches zero, and therefore the actual arrival of the water table at the point of zero flow would be indefinitely delayed were it not for the fact that evaporation of ground water would continue and might easily carry the water table below this point—a condition common in arid regions.

Owing to the lack of suitable data for a drought period the stage-flow relations could not be defined for flows approaching zero.

Some of the difference between the winter recession curve and the hypothetical curve in figure 8 may be caused by an error in defining the winter curve. The upper and lower parts are extrapolated from Riggs' data (written communication, 1957) and therefore are less certain than the central part. Also, the upper part might be too steep and the lower part too flat because most of the higher base flows occur in the spring when evapotranspiration is increasing rapidly, and most of the lower base flows occur in the fall when evapotranspiration is decreasing rapidly.

During periods of little or no rain in October and November the discharge of Brandywine Creek, and many other streams in the region, may decrease at much lower rates than normal, remain practically constant, or even rise slightly because of the decreases in evapotranspiration from ground water.

The slope of the recession curve is steeper and the effects of changing rates of evapotranspiration from ground water are less apparent at the higher base flows that prevail in the spring months than in the fall but these factors probably are of equal magnitude in spring and fall and cause the curves defined by conventional methods to be steeper than a recession curve for a constant rate of loss.

In summary, it is concluded that if index wells had been selected to indicate the average water-table gradients adjacent to streams rather than average fluctuation of water table in the basin, the relation line between ground-water stage and base flow of the creek probably would be curved rather than straight as shown in figure 7. However, the curvature probably would not be as great as that in a relation line derived from the winter base-flow recession and figure 7 because of the probable effects of evapotranspiration from ground water on the winter recession curve and the uncertainties in definition of the upper and lower parts of that curve. The evapotranspiration from ground water calculated from a curved relation line would be less than that calculated from the straight line shown in figure 7, and the average ground-water discharge for 1952–53 would be correspondingly less than 489 cfs (316 mgd) given in table 4.

COMPARISON WITH RESULTS OF OTHER STUDIES

Comparison of relations between ground-water stage and base flow in the Brandywine Creek basin with those reported in other areas reveals not only similarities but also significant differences.

In their detailed study of the Beaverdam Creek basin, Maryland, Rasmussen and Andreasen (1959, fig. 6) found that the slope of the stage-flow relation increases markedly with decrease in stage, and that the points representing monthly averages do not define a loop. Part of the curvature may be due to the changes in evapotranspiration in spring and fall, as described above. The absence of a loop is probably due to the fact that evapotranspiration from ground water occurs over the entire Beaverdam Creek basin instead of being restricted largely to the stream valleys as in the Brandywine Creek basin. Thus, in the Beaverdam Creek basin, increased evapotranspiration in the growing season would tend to lower ground-water levels almost uniformly throughout the basin instead of only in the zones adjacent to the streams, and the relation of average ground-water stage to base flow would be nearly the same at all times. The average ground-water stage recession rate at a given stage would vary with the season instead of being nearly constant as in the Brandywine Creek basin.

In the Dilldown Creek basin, a 1,530-acre area in the Delaware-Lehigh Experimental Forest (Pennsylvania Dept. of Forests and Waters, 1951) the relation of ground-water stage in two index wells to base flow is similar to that determined by Rasmussen and Andreasen (1959), except that the stage-flow relation is even more strongly curved, and significant seasonal differences in the stage-flow relation were observed. Two curves are shown by the Pennsylvania Department of Forests and Waters (1951, fig. 16): one for the growing season, May to September; the other for the nongrowing season, October to April.

Albertson[2] (1942), who studied the Rapid Creek watershed, a small basin in Iowa, found a logarithmic relation between base flow and ground-water stage which is similar in a qualitative way to the relations observed in the Beaverdam Creek and Dilldown Creek basins. Albertson found a different stage-flow relation from year to year, however.

In the Pomperaug River basin, Conn., Meinzer and Stearns (1929) found a curved relation similar to but having decidedly less curvature than the relations for the Beaverdam Creek, Dilldown Creek, and Rapid Creek basins. Meinzer and Stearns also found seasonal changes in the stage-flow relation which they attributed to ground-water evaporation (evapotranspiration as presently defined).

Whelan (1950) in a study of the Baker River basin, an area of 143 square miles in the White Mountains of New Hampshire, noted markedly different ground-water depletion (base-flow recession) curves for different months; the differences apparently were due to seasonal changes in evapotranspiration from ground water.

Not all investigators have observed the curved stage-flow relation described above. Clark (1956, fig. 4c)

[2] Albertson, M. L., 1942, Ground-water flow in Rapid Creek watershed: Iowa Univ. M.S. thesis.

determined a nearly linear relation of ground-water stage in one index well and base flow in Pond Creek, Okla., after he had removed the effects of evapotranspiration from ground water (which he called index of streamflow loss) by using an empirical formula. This relation is very similar to that determined in the Brandywine Creek basin (fig. 7).

COMPUTATION OF GROUND-WATER PARAMETERS WITH THE AID OF STREAMFLOW DATA

Two hydrologic parameters [3] that commonly are most useful in appraising the water-yielding potential of rocks or earth materials are: (a) coefficient of storage and (b) coefficient of transmissibility. For the water-table conditions that probably prevail in most of Brandywine Creek basin, the coefficient of storage is about equal to the specific yield or long-term gravity yield.

GRAVITY YIELD OF THE ZONE OF WATER-TABLE FLUCTUATION

As defined by Rasmussen and Andreasen (1959, p. 83): "The gravity yield of a rock or soil after saturation or partial saturation is the ratio of (a) the volume of water it will yield by gravity to (b) its own volume during the period of ground-water recession." Conversely, gravity yield of the material after drainage or partial drainage may be considered as the ratio of the volume of water it will take into storage to its own volume during the period of ground-water rise. In a strict sense this is a short-term coefficient of storage, which for convenience in this paper is referred to as gravity yield. Gravity yield, thus defined, ordinarily is less than specific yield, which is attained only after long periods of recession or rise, when gravity drainage or refilling is complete. For periods of several months the gravity yield probably is not significantly less than the specific yield.

In the following analyses, it is assumed that the observation wells and index wells indicate the average fluctuations in the water table throughout the basin and that almost all the ground-water discharge to streams is from the water-table (unconfined) zone.

COMPUTATION BY METHOD A

Gravity yield may be calculated from the simple equation

$$Y_g = \frac{\Delta S_g}{\Delta H_g}$$

in which Y_g is gravity yield (a dimensionless ratio), ΔS_g is the increase in ground-water storage in a specified period (expressed in inches of water over the area), and ΔH_g is the corresponding increase in ground-water stage (expressed in inches). The change in stage may be measured in observation wells, but the change in storage must be determined indirectly from base-flow data.

During periods of no evapotranspiration and no recharge to ground water the change in storage is equal to the runoff from ground water, which may be substituted for increase in storage in the equation above. (Withdrawal of water from wells is considered negligible.) These ideal conditions seldom, if ever, occur. However, the base-flow recession curve for winter (fig. 8) represents conditions when the evapotranspiration loss from ground water is relatively small. The runoff from natural storage derived by integration of the recession curve is considered a fair estimate of the runoff from ground water and hence of change in storage. The actual change in storage is somewhat greater than the runoff from ground water because of a small amount of evapotranspiration from ground water.

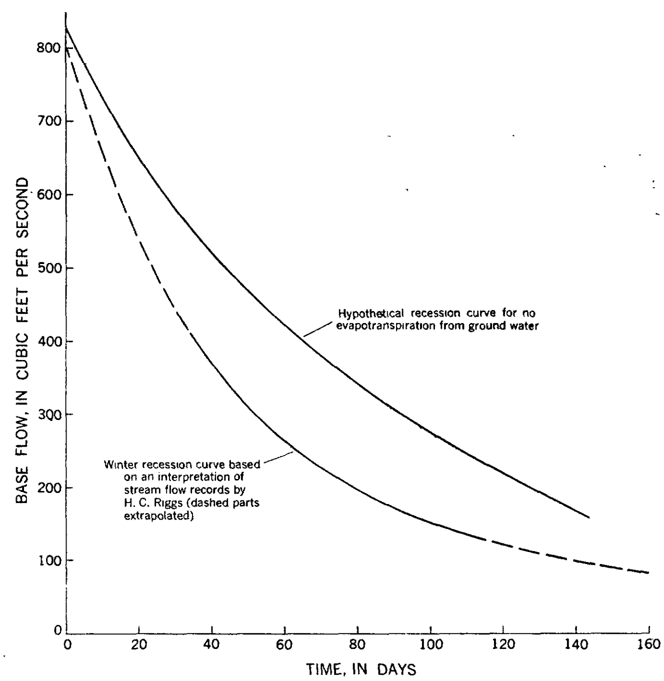

FIGURE 8.—Base-flow recession curves for Brandywine Creek at Chadds Ford, Pa. Hypothetical curve computed from figures 6 and 7; see text.

The part of the base-flow recession curve between about 410 cfs and 120 cfs is plotted from data furnished by H. C. Riggs (written communication, 1957) The extrapolated portions above and below these limits, shown by dashed lines in figure 8, are needed to cover the range in data for the study period.

Three seasonal periods were analyzed: two recession periods, one from late April or early May to mid-

[3] Parameter, in this report, refers to a characteristic that can be measured, in contrast to a characteristic that can only be described.

November 1952, and the other from middle or late March to late October 1953; and one period of rise, from mid-November 1952 to middle or late March 1953. The ground-water stages and seasonal changes in stage at each of the 16 observation wells and the averages for these periods are listed in table 6. The corresponding base flows are obtained from the base flow hydrograph in figure 5.

TABLE 6.—*Seasonal changes in ground-water stage in 16 wells, 1952–53*

Well	Spring 1952		Change in stage, spring to fall 1952 (feet)	Fall 1952		Change in stage, fall 1952 to spring 1953 (feet)	Spring 1953		Change in stage, spring to fall 1953 (feet)	Fall 1953	
	Date	Depth to water (feet)		Date	Depth to water (feet)		Date	Depth to water (feet)		Date	Depth to water (feet)
Bd-10	May 2	10.30	−3.79	Nov. 19	[1] 14.09	2.74	Mar. 31	11.35	−2.72	Oct. 30	14.07
Ch-2	Apr. 29	4.20	−6.50	Nov. 18	10.70	4.97	Mar. 24	5.73	−5.74	Oct. 25	11.47
3	May 2	18.48	−7.35	Nov. 19	[1] 25.83	6.21	Mar. 20	19.62	−9.51	Oct. 28	[1] 29.13
4	do	19.45	−6.75	Nov. 16	26.20	5.70	do	20.50	−6.90	Oct. 23	27.40
5	Apr. 29	1.55	−4.66	Nov. 20	6.21	4.46	Mar. 28	1.75	−5.22	Oct. 28	6.97
6	do	12.19	−4.67	Nov. 19	[1] 16.86	3.82	Mar. 20	13.04	−5.11	Oct. 23	18.15
7	May 2	28.05	−7.81	do	[1] 35.86	6.48	do	29.38	−7.20	do	36.58
8	Apr. 27	6.86	−9.24	Nov. 16	16.10	7.10	do	9.00	−9.30	do	18.30
9	Apr. 29	12.53	−5.89	Nov. 19	[1] 18.42	6.83	do	11.59	−9.07	do	20.66
10	do	8.46	−5.17	do	[1] 13.63	4.91	do	8.72	−5.34	do	14.06
11	May 2	10.36	−2.40	Nov. 17	12.76	2.54	do	10.24	−3.23	do	13.47
12	do	30.90	−5.40	Nov. 16	36.30	5.70	Mar. 22	30.60	−6.15	do	36.75
13	do	12.62	−3.43	Nov. 19	[1] 16.05	3.65	Mar. 20	12.40	−4.38	do	16.78
14	do	17.42	−4.98	do	[1] 22.40	5.05	Mar. 30	17.35	−5.43	do	22.78
15	do	22.38	−7.54	Nov. 17	[2] 29.92	8.67	Mar. 20	21.25	−10.73	do	31.98
De-3	May 4	12.09	−4.60	Nov. 21	16.69	4.79	Mar. 30	11.90	−5.87	Oct. 29	17.77
Average		14.24	−5.64		19.88	5.23		14.65	−6.37		21.02

[1] Estimated.
[2] Depth reported, 24.92 ft, probably 5 ft in error.

The calculation of the gravity yield for each of the three periods is shown below; the base flow at the ends of the periods are taken from figure 2 and the change in stage is taken from table 6.

Spring to fall 1952

S_g=integral of winter base-flow recession curve in figure 8 from 750 cfs to 162 cfs=33,100 cfs–days=−4.29 in.

H_g=−5.64 ft.=−67.7 in.

$$Y_g=\frac{-4.29\text{ in.}}{-67.7\text{ in.}}=0.063,\text{ or }6.3\text{ percent}$$

Fall 1952 to spring 1953

S_g=integral from 162 cfs to 806 cfs=35,900 cfs–days=4.65 in.

H_g=5.23 ft.=62.8 in.

$$Y_g=\frac{4.65\text{ in.}}{62.8\text{ in.}}=0.074,\text{ or }7.4\text{ percent}$$

Spring to fall 1953

S_g=integral from 806 cfs to 95 cfs=42,000 cfs–days=−5.44 in.

H_g=−6.37 ft.=−76.4 in.

$$Y_g=\frac{-5.44\text{ in.}}{-76.4\text{ in.}}=0.071,\text{ or }7.1\text{ percent}$$

Arithmetic sum of seasonal fluctuations

S_g=14.38 in.

H_g=206.9 in.

$$Y_g=\frac{14.38\text{ in.}}{206.9\text{ in.}}=0.070,\text{ or }7.0\text{ percent}$$

If the calculation is based on the 3 index wells instead of 16 wells, the change in storage is practically the same as above, but the total fluctuation in ground-water stage is reduced to 191.9 inches, and

$$Y_g=\frac{14.38\text{ in.}}{191.9\text{ in.}}=0.074,\text{ or }7.4\text{ percent.}$$

This result is more nearly comparable with the gravity yield obtained by method B from the index wells.

COMPUTATION BY METHOD B

The effects of evapotranspiration from ground water (which are not entirely absent in the winter base-flow recession curve) can be eliminated by using the hypothetical recession curve (fig. 8). This advantage may be offset, at least in part, by the greater chance for error in the more indirect method described below.

For any period the ground-water discharge to streams corresponding to no loss by evapotranspiration from ground water (which is equal to change in storage) may be obtained by integrating the hypothetical recession curve in figure 8 between the limits corresponding to the beginning and end of the period. Using these values for increase in storage and the same procedure as used in method A the gravity yield is computed as 10.7 percent.

Because of the linear relation between ground-water stage and base flow, values of gravity yield computed by this method are the same at all stages. However, as previously noted, the available evidence indicates that gravity yield probably decreases sharply at depths below the lower limit of water-table fluctuation.

EVALUATION OF RESULTS

The difference in values of gravity yield computed by the two methods is due to the differences in the base-flow recession curves, both of which are subject to error.

Earlier discussion indicated that the winter recession curve may be too flat at the lower end, and that the hypothetical curve representing no loss probably should resemble the winter recession curve more closely than it does. Although part of the difference between the curves is due to inclusion in the winter curve of effects of evapotranspiration from ground water, probably a large part of the difference is due to errors in definition of the curves.

The accuracy of the estimates of gravity yield by both methods depends in part on whether the index wells and observation wells provide a representative sample of average water-table fluctuation in the basin, and in part on whether essentially all the ground-water discharge to streams is from the water-table (unconfined) zone. For example, if the average fluctuation of the water table in the basin were double that in the index wells or observation wells, the gravity yield would be only half as much as the values estimated in the study. If a significant amount of ground-water discharge occurs from perched or semiperched water bodies above the true water table, or from confined or semiconfined water bodies below the water-table zone, the estimates of gravity yield are too high. Ward [4] made the following pertinent observation:

> In some places two or more water-bearing zones may be encountered in deep excavations. The first occurs in the weathered zone and the water there can probably be drained off with tiles. Below the zone of weathering massive rock may be encountered that will yield no water. Upon continued excavation, zones of shearing or closely spaced joints may be encountered. These are usually saturated, and flows up to several gallons per minute may be encountered.

This suggests that, at some places at least, more than one water body exists in the crystalline rocks, and that appreciable ground-water discharge to streams may take place from fractures below, and not freely connected with, the water-table zone.

In summary, the average gravity yield of the zone of water-table fluctuation probably is between about 7½ and 10 percent. However, this estimate may require modification if information becomes available to show that the magnitude of average water-table fluctuation in the basin is appreciably different from that indicated by the wells used in the study, or that a significant amount of natural ground-water discharge occurs from zones above or below the water-table zone tapped by the wells.

By comparison, Herpers and Barksdale (1951, p. 27) estimated that the average specific yield (long-term gravity yield) of the zone within 300 feet of the land surface in the Brunswick formation (red shale and sandstone of Triassic age) in the Newark, N.J., area is on the order of 1 or 2 percent. The average gravity yield of the unconsolidated sandy sediments in the zone of water-table fluctuation in Beaverdam Creek basin, in the Coastal Plain province in Maryland, was estimated to be about 11 percent by Rasmussen and Andreasen (1959).

[4] Ward, R. F., 1956, The geology of the Wissahickon formation of Delaware: Unpublished report prepared in cooperation between Delaware State Highway Dept., Delaware Geol. Survey, and U.S. Geol. Survey.

COEFFICIENT OF TRANSMISSIBILITY OF ROCKS IN THE BASIN

Coefficient of transmissibility is ordinarily measured by means of a pumping test made with a pumped well and one or more observation wells. The methods have become standardized and are described by Wenzel (1942) and in more recent papers. The standard test procedures, although yielding reasonably accurate results, may be interpreted only in terms of the relatively small sample of material in the vicinity of the pumped well. Moreover, in many areas, including Brandywine Creek basin, pumping-test data are not available.

Methods of approximating the coefficient of transmissibility from numerical analysis of water-level data are described by Stallman (1956). For hypothetical cases where a very thick uniform aquifer discharging to a stream or other body of surface water is assumed, the coefficient of transmissibility may be calculated from formulas devised by Jacob (1943).

In Brandywine Creek basin, both the water-level data required for Stallman's method and the thick, uniform aquifer assumed by Jacob are lacking. However, a very rough estimate of the average coefficient of transmissibility may be made, using the total ground-water discharge (assumed equal to base flow plus evapotranspiration from ground water) estimated in the preceding section, the average hydraulic gradient (water-table slope) adjacent to the discharge areas, and the length of the discharge areas. The equation is

$$T=\frac{R_g+ET_g}{I(2L)}$$

in which T is the coefficient of transmissibility in gallons per day per foot, R_g is the base flow in gallons per day, ET_g is evapotranspiration from ground water in gallons per day, I is the average water-table slope adjacent to discharge areas in feet per mile, and L is the length of discharge areas in miles (as indicated by length of blue lines representing streams on standard topographic quadrangle maps).

The average ground-water discharge (including evapotranspiration from ground water) for the 2-year period 1952–53 (col. 3, table 4) was 489 cfs or 316×10^6 gpd.

The average water-table slope was computed as the average slope of the water table between the observation wells and the discharge outlets. The elevations of the natural discharge outlets (usually small stream valleys) downgradient from the wells in a direction perpendicular to the landslope were obtained from topographic maps. The average depths to water were calculated as the arithmetic average of the seasonal extremes for the calendar years 1952–53 (last col. table 1). This depth was subtracted from the elevation of the land surface at each of the wells to obtain the elevation of the water table. The difference in elevation was divided by the horizontal distance from each well to the discharge outlet to obtain the average water-table slope in feet per mile. The slopes at the three index wells are illustrated in figure 3; the values for 15 observation wells are listed below (well Bd–10 was omitted because it was too near the discharge outlet to permit an accurate estimate of the slope).

Well	Distance to discharge outlet (miles)	Average difference in elevation of water table (feet)	Average slope of water table (feet per mile)
Ch–2	0.40	17	42
3	.25	19	76
4	.18	59	330
5	.23	46	200
6	.18	5	28
7	.20	13	65
8	.16	12	75
9	.21	44	210
10	.18	5	28
11	.20	38	190
12	.08	11	140
13	.16	70	140
14	.07	20	290
15	.13	39	300
De–3	.17	70	410
Average			190

The discharge areas in the basin are arbitrarily assumed to coincide with the streams shown by blue lines on the 1:62,500 topographic maps of the basin, although these are not accurate measures of lengths of discharge areas, which vary with season and water-table conditions. The total length of streams, calculated from Langbein's figure for stream density of 2.26 miles per square mile (Langbein and others, 1947), is 649 miles. Ground water moves toward the discharge outlet from both sides of each stream, so that the total length of transmitting section is 2×649 miles=1,298 miles.

The coefficient of transmissibility is then calculated as follows:

$$T=\frac{316\times10^6\ (\mathrm{gal})(\mathrm{day})^{-1}}{190\ (\mathrm{ft})(\mathrm{mi})^{-1}\times2\times649\ \mathrm{mi}}=1{,}300(\mathrm{gal})(\mathrm{day})^{-1}(\mathrm{ft})^{-1}$$

Another rough estimate of average water-table gradient was made by using stream density, average landslope, and an assumed average depth to the water table under drainage divides. In this estimate it is assumed for the sake of simplicity that the average slope of the water table adjacent to the discharge areas is the same as that from drainage divides to streams. Stream density is 2.26 miles per square mile; average interstream distance is the reciprocal of stream density; and the average distance from streams to divides is half the average interstream distance

$$\frac{1}{2\times2.26(\mathrm{mi})(\mathrm{mi})^{-2}}=0.22\ \mathrm{mi}$$

Average landslope is 450 feet per mile (Hely and Olmsted, written communication, 1961); the average elevation of divides above the streams (the average relief) is

$$450\ \text{feet per mile}\times0.22\ \text{mile}=100\ \text{feet}$$

The average depth to the water table beneath divides is assumed to be 35 feet—slightly more than the deepest levels in the 16 observation wells. The average slope of the water table from divides to streams is therefore $(100-35)$ feet$\div$0.22 mile=300 feet per mile. Substituting this value for the value of 190 feet per mile used in the previous estimate, the coefficient of transmissionibility is 800 gpd per foot, instead of 1,300.

The estimates of the coefficient of transmissibility probably are less accurate than those of gravity yield. The calculated water-table gradients and length of discharge areas are the most uncertain quantities used in the estimates; the value for total ground-water discharge is considerably more accurate. The assumption of a uniform, continuous water-table gradient adjacent to the discharge areas, as obtains in homogeneous granular material, is not applicable to parts of Brandywine Creek basin. In fractured crystalline rock, the zone of saturation is discontinuous, and the water table may stand at different levels in different sets of fractures. The gradients adjacent to the discharge areas almost certainly are not the same as the average gradients from ground-water divides to streams, as assumed in the second estimate.

In permeable materials extending to great depth the amplitude of water-table fluctuation is small in proportion to the total thickness of water-bearing material, so that the thickness of that portion of the material transmitting water varies only slightly as the water table fluctuates. In the weathered and fractured crystalline rocks of Brandywine Creek basin however, the drainable void space decreases rapidly with depth below the water table so that, as the water table declines, the effective thickness, and consequently the transmissibility, of the water-bearing zone likewise decreases rapidly. Therefore, a coefficient of transmissibility determined for a wet period, like 1952–53, would be significantly higher than that determined for a drier period when the water table was lower.

It is concluded that the average coefficient of transmissibility of the rocks in the basin for the calendar years 1952–53 is on the order of 1,000 gpd per foot. In comparison, the coefficient of transmissibility estimated from average specific capacity of wells in the Piedmont Upland of the Delaware River region is about 1,500 gpd per foot (Hely and Olmsted, written communication, 1961). Callahan and Stewart (1959) report that pumping-test data from the weathered crystalline rocks in the Piedmont province in Georgia indicate a coefficient of transmissibility ranging from 500 to 4,000 gpd per foot. The rocks there are similar to those in Brandywine Creek basin, but the weathered zone is thicker and the degree of weathering possibly is more intense.

CONCLUSIONS

As an average for periods of several years, about two-thirds of the total runoff of Brandywine Creek is base flow (chiefly ground-water discharge to the streams). This proportion is significantly higher than in streams draining areas underlain by most other types of consolidated rocks in the region and is nearly as high as the base flows of streams draining areas underlain by permeable sandy sediments in the Coastal Plain.

A direct, linear relation exists between the monthly average ground-water stage in three index wells (which are assumed to have indicated average water-table conditions in Brandywine Creek basin) and base flow of the creek in winter months, when evapotranspiration from ground water is very small. However, if index wells had been selected so as to indicate average gradient of the water table adjacent to areas of ground-water discharge rather than average fluctuation of the water table in the basin, the relation line between ground-water stage and base flow would have been curved rather than straight, and the slope of such a relation line would steepen with decreasing stage.

The average ground-water discharge in the basin for the calendar years 1952–53 was 489 cfs (316 mgd), of which slightly less than one-fourth was estimated to be loss by evapotranspiration (based on the linear relation between ground-water stage and base flow). However, loss estimated from the curvilinear relation described above would be less, and the total discharge correspondingly less.

The average gravity yield of the zone of water-table fluctuation in the basin probably is between about 7½ and 10 percent—surprisingly high for such aquifers. This estimate may require modification if future information indicates that the wells used in the study do not accurately reflect the average magnitude of water-table fluctuation in the basin, or that a significant amount of ground water is discharged from zones above or below the water-table zone tapped by the wells.

Even if allowance is made for possible errors in the estimate, the gravity yield is several times as great as that of some shale and sandstone of the Brunswick formation of Triassic age, and may not be much less than that of some of the unconsolidated sandy sediments of the Coastal Plain. However, the gravity yield in Brandywine Creek basin probably decreases rapidly, perhaps abruptly, below the lower limit of water-table fluctuation.

The average coefficient of transmissibility of the rocks in the basin is relatively low and decreases with decreasing ground-water stage. The average value for the period 1952–53 is on the order of 1,000 gpd per foot. It agrees generally with an estimate made from average specific capacity of wells in the region and is within the limits obtained from pumping tests of wells in similar rocks in Georgia.

Average gradients of the water table in the basin are rather steep—one estimate is 190 feet per mile, and another 300 feet per mile. Most of the drainable water in the rocks is above the stream channels, which are the major discharge outlets.

The seeming inconsistency of low to moderate ground-water yield to wells and high yield to streams in the crystalline rocks of the Piedmont Upland was cited in the introduction. Several of the conclusions stated above help to explain this paradox.

The most plausible explanation concerns the nature of the bedrock, the overlying weathered zone and soil, and the topography. Most of the water-bearing openings in crystalline rocks are in the weathered zone and in the immediately underlying fractured rock. Normally the water table lies within the lower part of the weathered zone, or locally in the upper part of the fractured zone. The gravity yield and consequent ground-water storage capacity of the zone of water-table fluctuation are high, and the deep, permeable soils permit much of the precipitation to infiltrate to the water table to be stored for later ground-water discharge to streams as base flow. The rocks have comparatively low transmissibility and their gravity yield or drainable void space decreases rapidly with depth. Because of the low transmissibility, the rapid decrease of gravity yield with depth, and the fairly rugged topography, water-table gradients are steep. Consequently, relatively high ground-water discharge to streams is maintained by drainage from the substantial ground-water reservoir, despite the fairly low transmissibility of the rocks.

On the other hand, well yields in the basin are restricted by the small thickness and fairly low transmissibility of the water-bearing material. A well has a small length and area of contact with water-bearing material in comparison with the stream system, and the yield of most wells increases slightly, if at all, when they are deepened below the zone of weathering and fracturing or when the pumping drawdown is increased and lowers the water level much below that zone.

The above conclusions apply to problems of estimating the potential productivity of aquifers. In general, no simple relation exists between the amount of ground-water discharge in an area and the proportion of this discharge that can be diverted to wells. Natural ground-water recharge and discharge are almost as large in the Brandywine Creek basin as in the sandy parts of the Coastal Plain, but the productivity of wells in crystalline rocks in Brandywine Creek basin is generally much less than that of wells in the sandy aquifers of the Coastal Plain.

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