Melvin Marcher

Outline of Methods for Estimating Ground-Water Supplies

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Outline of Methods for Estimating Ground-Water Supplies

By OSCAR EDWARD MEINZER

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By Oscar Edward Meinzer

INTRODUCTION

The most urgent problems in ground-water hydrology at present are those relating to the rate at which the rock formations will supply water to wells in specified areas—not during a day, a month, or a year but perennially. The permeable rock formations may be regarded as underground reservoirs or as underground conduits. Some water-bearing formations function chiefly as reservoirs, others chiefly as conduits, but all of them have some of the properties of both. In attempting to solve the quantitative problems above mentioned these two functions of rock formations must be recognized and differentiated. Quantitative methods based on the reservoir conception are applicable chiefly, though not exclusively, to formations or parts of formations that have water tables; methods based on the conduit conception are applicable most largely to artesian formations, in which the water moves laterally considerable distances from the intake to the discharge area.

ROCK FORMATIONS AS RESERVOIRS

GENERAL CONDITIONS

The natural reservoirs formed by rock formations generally have very great capacity compared with that of ordinary artificial surface reservoirs, just as most other natural features are of huge size in contrast to the works of man. However, if the water derived from them is taken chiefly out of storage without being replaced the supply will eventually fail.

The " safe yield " of an underground reservoir, or the practicable rate of withdrawing water from it perennially for human use, may be estimated by methods that are comparable to those used in estimating the safe yield of a surface reservoir, although different in

[1] The substance of this paper was presented Dec. 29, 1928, at a meeting of the Society of Economic Geologists in New York City, and mimeographed copies of the paper in abbreviated form had been sent to the members prior to the meeting. The paper has been revised and enlarged for the present publication.

technique. It is necessary to measure the rate of inflow (intake methods) or the rate of discharge (discharge methods), or else these rates must be estimated by determining changes in storage (water-table or storage methods).[2] In many areas the most effective method of studying the ground-water supply is the inventory method, in which all available methods are used to make periodic inventories of the entire water supply of the area, from its entrance into the area as precipitation, stream flow, or underground percolation to its exit as evaporation, transpiration, run-off, or underground leakage.

The work of Charles H. Lee[3] in Owens Valley, Calif., in 1908 to 1911, is of outstanding importance in the history of quantitative methods in ground-water work, because he devised an entire system of quantitative methods and applied them in orderly fashion to obtain an average value for each element of intake into and discharge out of the underground reservoir. The term "inventory method" is more accurately applied, however, to investigations in which a detailed accounting of the several elements of intake, storage, and discharge is attempted for each of successive months, years, or other units of time during the entire period of observations. Thus, in the investigation in the Pomperaug Basin, Conn.,[4] an attempt was made to organize the available data in a monthly inventory of the water supply during a period of three years.

METHODS FOR ESTIMATING INTAKE FROM SURFACE STREAMS

The principal intake method consists of establishing gaging stations on influent streams and determining the quantities of water lost between successive stations. The quantity of water that reaches the water table consists of this loss minus the loss by evaporation and transpiration either directly from the stream or from soil moisture supplied by the stream. The method has been used in many ground-water investigations in the present century, notably in California, Arizona, New Mexico, Nevada, Utah, Colorado, and Idaho. The earliest measurements of seepage losses from streams and canals were made chiefly to ascertain the losses of surface water—for example, the investigation by Grunsky,[5] in 1882, of losses from

[2] Meinzer, O. E., Quantitative methods of estimating ground-water supplies: Geol. Soc. America Bull., vol. 31, pp. 329–338, 1920.

[3] Lee, C. H., An intensive study of the water resources of a part of Owens Valley, Calif.: U. S. Geol. Survey Water-Supply Paper 294, 1912; The determination of safe yield of underground reservoirs of the closed-basin type: Am. Soc. Civil Eng. Trans., vol. 78, pp. 148–251, 1915.

[4] Meinzer, O. E., and Stearns, N. D., A study of ground water in the Pomperaug Basin, Conn., with special reference to intake and discharge: U. S. Geol. Survey Water-Supply Paper 597, pp. 73–146, 1928.

[5] Grunsky, C. E., Irrigation near Fresno, Calif.: U. S. Geol. Survey Water-Supply Paper 18, pp. 71–79, 1898.

irrigation canals in the vicinity of Fresno, Calif. Later, however, in several investigations, seepage losses have been determined chiefly for the purpose of estimating the amount of ground-water recharge. Notable among the early ground-water investigations in which the seepage method was used are those of Clapp[6] and Mendenhall[7] in southern California, Smith in Arizona,[8] and Lee in Owens Valley. One of the most intensive studies of this method was made by Bailey[9] in his work on ground-water recharge in the Niles cone, California, by seepage from Alameda Creek. Bailey undertook not only to determine the recharge by years or other long periods, but also to determine the daily recharge, which required accurate data on changes in channel storage with changes in gage heights. He also investigated the relation of rate of recharge to the temperature of the water and to the duration of high stages of the river. Ultimately he developed curves to show the relation of rate of seepage to rate of stream flow and produced an empirical formula by which the recharge in any day can be computed from the records of stream flow and temperature. Intensive use of the seepage method has recently been made by Conkling[10] in the drainage basin of the San Gabriel River, and by Post[11] in the drainage basin of the Santa Ana River, both in California.

The seepage method gives the most reliable results for streams with relatively constant flow and heavy losses in proportion to the total flow. It is generally not applicable in times of flood, when the measurements of flow are relatively inaccurate and the percentage of loss is small. It is not applicable or only difficultly applicable to perennial streams in which the loss is small in comparison to the total flow, because with such streams small percentage errors in the measurements of the flow may produce large percentage errors in the computed seepage losses.

The seepage method is not very applicable for determining recharge from intermittent and ephemeral streams and is wholly inapplicable for determining recharge directly from rain or melting

[6] Clapp, W. B., in Hoyt, J. C., Report of progress of stream measurements for the calendar year 1903, pt. 4, Interior Basin, Pacific, and Hudson Bay drainage: U. S. Geol. Survey Water-Supply Paper 100, pp. 339–356, 1904.

[7] Mendenhall, W. C., The hydrology of San Bernardino Valley, Calif.: U. S. Geol. Survey Water-Supply Paper 142, pp. 50, 51, 1905; Ground waters and irrigation enterprises in the foothill belt, southern California: U. S. Geol. Survey Water-Supply Paper 219, pp. 28, 29, 1908.

[8] Smith, G. E. P., Ground-water supply and irrigation in the Rillito Valley: Arizona Univ. Agr. Exper. Sta. Bull. 64, pp. 118, 119, Tucson, 1910.

[9] Bailey, Paul, Engineering investigation of percolation from Alameda Creek and ground-water studies on Niles cone: California State Water Comm. Third Bienn. Rept., pp. 95–131, 1921.

[10] Conkling, Harold, San Gabriel investigation: California Dept. Pub. Works Div. Water Rights Bull. 5, pp. 52–72, 1927.

[11] Post, W. S., Santa Ana investigation, flood control and conservation: California Dept. Pub. Works Div. Eng. and Irr. Bull. 19, pp. 165–179, 1929.

snow. In humid regions almost no streams except the ephemeral ones are influent, and therefore this method obviously has little application. The method as applied to ephemeral streams presents two difficulties—the storm run-off is too flashy to be measured accurately, and a considerable part of the surface water lost by seepage may be held by molecular attraction in the soil and subsoil and therefore may not reach the water table but may ultimately be returned to the atmosphere.

Studies have recently been made by White[12] of recharge from the flow of the Mimbres River, an intermittent stream in New Mexico, which generally flows about 20 to 36 hours after a torrential summer rain. The flow from some of the rainstorms was all lost within the stretch of the river that crosses the intake area, and the loss could be determined by measuring the flow at the upper end of this stretch. In the heaviest storms, however, the flow persisted beyond the intake area, and measurements were made at convenient intervals within this stretch to determine the approximate loss. Borings were made in the bed of the river before and after the storms, and the moisture content of samples obtained from these borings was determined in order to ascertain how much of the seepage water was held by the sand of the river bed.

METHODS FOR ESTIMATING INTAKE DIRECTLY FROM RAIN AND MELTING SNOW

Many attempts have been made to measure directly the percolation from the surface to the water table. Mariotte, in the seventeenth century, observed the percolation of rain water into the cellar of the Paris Observatory. Measurements of percolation have been made with lysimeters, or vessels to catch the downward percolating water, by several European[13] and American[14] investigators, and some very long records have been obtained in England. The first tests mentioned by Veatch were made by Dalton in Manchester, England, from 1796 to 1798, and by Maurice in Geneva, Switzerland, in 1796 and 1797. The longest record, that of Dickinson and Evans, in Hempel Hempstead, in England, extended over at least 49 years, from 1835 to 1884, and the records of Lawes and Gilbert and those of Greaves extended over 32 and 22 years, respectively. Many of

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[12] White, W. N., Preliminary report on the ground-water supply of Mimbres Valley, N. Mex.: U. S. Geol. Survey Water-Supply Paper 637, pp. 77–80, 1931. (Released in manuscript form in 1929.)

[13] Veatch, A. C., Fluctuations of the water level in wells: U. S. Geol. Survey Water-Supply Paper 155, pp. 32–34, 44–49, 1906. This paper summarizes the work that had been done with lysimeters in England and on the continent of Europe.

[14] Bark, D. H., Duty of water investigations, experiments, and results: Idaho State Engineer Ninth Bienn. Rept. (1911–12), pp. 302–309; Experiments on the economical use of irrigation water in Idaho: U. S. Dept. Agr. Bull. 339, pp. 37–39, 1916. Other American experiments could be cited.

the lysimeter tests have been made with disturbed materials and under other artificial conditions, which may have produced results very different from those that would be found under natural conditions. Moreover, even if tests are made under natural conditions, the fundamental difficulty will still be encountered that downward percolation is irregularly distributed, and it is difficult or impossible to find a place for the test that is even approximately representative of average conditions. An extensive area may have little or no recharge, but at a few places water may be pouring into the zone of saturation in large volumes.

Tests recently made by H. F. Blaney and C. A. Taylor[15] near Ontario, Calif., indicate that water percolating downward through unsaturated material tends to follow the surface of the constituent grains and hence to move around any vessel that is placed underground to catch it. If the material extends downward into the vessel, the percolating water will accumulate in the capillary openings of the material in the vessel and can be detected by laboratory determination of the increase in moisture content of that material. However, unless the material lies within the vessel to a height exceeding the capillary range, no free gravity water may appear in the vessel even though recharge is occurring.

The earliest attempts to determine the rate of ground-water recharge were doubtless based on measurements of the quantities of water that fall as rain or snow. This was the method of Mariotte,[16] who attempted a quantitative study of the water supply of the drainage basin of the Seine River by making crude measurements of the water that fell as rain on the basin and of the water discharged by the river. By means of these measurements he disproved the ancient assumption that the rainfall is inadequate in amount to supply the water discharged by the springs. In the years that have elapsed since Mariotte's work the most common method of estimating ground-water recharge has probably been to determine approximately the quantity of water that annually falls as rain or snow on a given area and apply to this quantity the percentage that is assumed to reach the zone of saturation. This method is of little value, however, except to give an idea of the maximum possibilities, unless there is a reliable basis for the percentage that is assumed. Commonly the assumed percentage and therefore also the computed recharge are much too large.

[15] Unpublished data. Since this paper was sent to press the following paper describing these methods has been published: Taylor, G. H., Investigations relating to the absorption of precipitation and its penetration to the zone of saturation: Am. Geophysical Union Trans. 12th meeting, pp. 206–211, 1931.

[16] Mariotte, Edmé, Traités du mouvements des eaux et des autres corps fluides, 1686.

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Attempts have been made to compute the recharge in an intake area by deducting the run-off and the evaporation (including transpiration) from the precipitation, but the results have generally been unreliable, especially because of uncertainty as to the amount of evaporation and transpiration. The method is inapplicable where the recharge is relatively so small that a moderate percentage error in precipitation, run-off, or evaporation may produce a large percentage error in the computed recharge. It apparently gave useful results as applied by McCombs [17] and Kunesh [18] in estimating the ground-water supply of the island of Oahu, where fairly accurate data are available as to precipitation and run-off and where the intake facilities are so good that a large proportion of the precipitation seeps into the rocks and percolates to the water table. The weakest element in the Oahu computations was the estimate of evaporation, for which reliable data were not available. To overcome this weakness, tests of soil evaporation and transpiration are now being made.

Recently a detailed study has been made with this method by Blaney [19] in certain intake areas in southern California. From the quantity of water produced by each rain during the winter rainy season were deducted the carefully estimated quantities of run-off produced by the rain and of soil evaporation and transpiration during the subsequent period of fair weather. Borings and tests of soil moisture proved that the remaining water was stored in the soil to satisfy the deficiency of soil moisture in the root zone. After this deficiency had been completely supplied and the soil moisture had been brought up to the specific retention, or quantity of water that the soil can hold by molecular attraction against the pull of gravity, any remaining surplus was assumed to percolate downward beyond the reach of plant roots and ultimately to reach the water table. If an impermeable bed intervened between the root zone and the water table, the downward percolating water would necessarily form a body of perched ground water above the impermeable bed.

The method of following the movements of water in the soil by making periodic borings with soil augers and determining the moisture content of soil samples obtained from different depths has been extensively used in connection with agricultural investigations and less frequently to determine the ground-water recharge. Re-

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[17] McCombs, John, Methods of estimating safe yield of Honolulu artesian area: Honolulu Sewer and Water Comm. Rept., 1927, pp. 55–61.

[18] Kunesh, J. F., Surface, spring, and tunnel water investigations: Honolulu Sewer and Water Comm. Rept., 1929, pp. 85–92.

[19] Blaney, H. F., Disposal of rainfall, in Post, W. S., op. cit. (Santa Ana investigation), pp. 152–157.

cently, however, earnest attempts have been made to use this method in ground-water investigations in California by Sonderegger,[20] by Blaney and C. A. Taylor, and by Stearns, Robinson, and G. H. Taylor.[21] In the latest work by G. H. Taylor the moisture equivalent, or amount of moisture retained by the soil in a centrifuge, was taken as a rough measure of the specific retention in computing deficiencies in the belt of soil moisture, or "root zone."

METHODS FOR ESTIMATING DISCHARGE BY OVERFLOW

An underground reservoir, like a surface reservoir, may lose water by overflow, evaporation, diversion, and underground leakage. The overflow of an underground reservoir may appear in definite springs or in general seepage that contributes to the flow of the surface streams. If it appears in a few large and definite springs the ground-water discharge can be determined by measuring the flow of these springs. If the springs are relatively constant it may be sufficient to make occasional measurements with current meter, weir, or other device; if they are relatively variable it may be necessary to establish gaging stations in order to obtain records of daily or continuous rate of flow. An excellent example of an extensive survey made largely by the first method is that by Crandall[22] and others of the very large and relatively constant springs that issue from the lava rocks into the Snake River, in Idaho. Examples of the second method are afforded by gaging stations, some of them with automatic water-stage recorders, maintained by the United States Geological Survey on several of the springs of first magnitude in Missouri and Texas that issue from limestone and have great fluctuations.[23]

If the overflow of the underground reservoir appears in many small, widely scattered springs or in general seepage its quantity may be estimated by establishing gaging stations on the effluent streams to measure the total run-off and then differentiating between the two components of the total—the direct run-off and the ground-water run-off. This differentiation can best be made by studying

[20] Sonderegger, A. L., Water supply from rainfall on valley floors: Am. Soc. Civil Eng. Proc., vol. 55, pp. 1144–1150, 1929. Since this paper was sent to press the following paper relating to this subject has been published: Blaney, H. F., Taylor, C. A., and Young, A. A., Rainfall penetration and consumptive use of water in Santa Ana River valley and coastal plain: California Div. Water Resources Bull. 33, 1931.

[21] Stearns, H. T., Robinson, T. W., and Taylor, G. H., Geology and water resources of the Mokelumne area, California: U. S. Geol. Survey Water-Supply Paper 619, pp. 174, 175, 1930; also more recent report released in manuscript form (on file in public library, Lodi, Calif.).

[22] Crandall, Lynn, The springs of Snake River Canyon: Idaho Irr. Eng. and Agr. Societies Joint Conf., 1918 and 1919, Proc., pp. 146–150. See also Meinzer, O. E., Large springs in the United States: U. S. Geol. Survey Water-Supply Paper 557, pp. 42–47, 1927.

[23] Meinzer, O. E., op. cit., pp. 17–41.

small representative drainage basins rather than by gaging the trunk stream of a large drainage basin, because the large basin is likely to have more frequent rainfall in some parts of its area and a longer and less definite period for complete delivery of the direct run-off.

In a study by the United States Geological Survey of the drainage basin of the Pomperaug River, in Connecticut, which has an area of 89 square miles, a gaging station was established near the mouth of the river. It was determined, from considerations of channel storage, velocity of the water, and rate of movement of flood crests, that the direct run-off is nearly all delivered from the basin within a week of the time the water falls as rain, and that therefore the stream flow a week or more after the latest rain is virtually all derived from ground water. A hydrograph was constructed of the total run-off for the years covered by the investigation, 1913 to 1916, and a hydrograph of the ground-water run-off was then made which coincided with that for total run-off in the periods of prolonged fair weather and was interpolated between these periods. The ground-water run-off was computed from this hydrograph.[24] This general method was earlier used in the flood-control investigation of the drainage basin of the Miami River, in Ohio, by Houk,[25] who prepared a hydrograph of the ground-water run-off of the Mad River for the period 1915 to 1919 by connecting the low points of the hydrograph for total run-off.

The method can be improved by obtaining gage-height records at enough points in the drainage system to permit an estimate of the channel storage of the system on each day. The daily ground-water run-off can then be computed as soon after a rain as the storm water has reached the streams. For any day with a falling stage after the storm water has reached the streams, the ground-water run-off will be the total run-off minus the decrease in channel storage. The decrease in channel storage can 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. Calculations show that in a drainage basin which is not larger than the Pomperaug the total quantity of water stored in the stream system at any time is rather small compared with the rate of discharge, and hence that errors in the measurement of decrease in storage will introduce relatively small errors in the estimates of ground-water run-off. The proposed method would have the advantage that the

[24] Meinzer, O. E., and Stearns, N. D., A study of ground water in the Pomperaug Basin, Conn., with special reference to intake and discharge: U. S. Geol. Survey Water-Supply Paper 597, pp. 107–116, pl. 19, 1929.

[25] Houk, I. E., Rainfall and run-off in Miami Valley, State of Ohio: Miami Conservancy Dist. Tech. Repts., pt. 8, 1921.

record would cover a much larger part of each period between rains and that the entire process would be one of observation and measurement with less of the arbitrary feature of the present method. It would not be necessary to make current-meter measurements to develop rating curves at the subsidiary stations, as only change in storage, indicated by change in gage height, would be involved.

METHODS FOR ESTIMATING DISCHARGE BY EVAPORATION FROM SOIL AND TRANSPIRATION FROM PLANTS

Where the water table is near the surface the underground reservoir loses water by evaporation from the soil and by transpiration from plants. In some arid regions all or nearly all the ground water is disposed of in this way, and little or no water remains to be discharged through springs and seeps. In humid regions, however, the ground-water recharge is greater, and more soil moisture is supplied directly by the rains and melting snow, with the result that evaporation from the soil and transpiration from plants make less heavy demands on the ground water. For these reasons in humid regions there is generally an excess of ground-water discharge over the consumption by soil evaporation and transpiration, and this excess feeds the streams and maintains their flow during dry seasons.

Tests of the rate of discharge of ground water by capillary rise from the zone of saturation and subsequent evaporation from the soil were made by Slichter[26] in 1905, Lee[27] in 1910 and 1911, Sleight[28] in 1916, and other investigators more recently. In the Escalante Valley investigation, by White,[29] records were obtained of evaporation from columns of undisturbed soil for comparison with evaporation from similar columns of soil that was dug up and artificially packed into tanks. The rate depends on the evaporativity of the atmosphere, the depth to the water table, and the texture of the soil, which controls both the height and the rate of capillary rise. The standard method of estimating the quantity of ground water lost by soil evaporation is to map the discharge area with reference to specified depths to the water table for different seasons, and to determine by means of tank experiments the rate of discharge for each range of depth. The problem is likely to be complicated by diversity in the texture of the soil, which may require a soil map of the discharge area and tank experiments for soils of different types.

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[26] Slichter, C. S., The underflow in Arkansas Valley in western Kansas: U. S. Geol. Survey Water-Supply Paper 153, pp. 43, 44, 1906.

[27] Lee, C. H., An intensive study of the water resources of a part of Owens Valley, Calif.: U. S. Geol. Survey Water-Supply Paper 294, pp. 57, 119, pl. 15, 1912.

[28] Sleight, R. B., Evaporation from the surfaces of water and river-bed materials: Jour. Agr. Research, vol. 10, pp. 209–261, 1917.

[29] White, W. N., A method of estimating ground-water supplies based on discharge of plants and evaporation from soil; results of investigations in Escalante Valley, Utah (released in manuscript form in 1930).

Not much work has yet been done to establish the relations of the rate of soil evaporation to the factors that determine the rate. In most tank experiments data are obtained as to the atmospheric conditions and the depth to the water table, which give some basis for generalizations as to the variations of soil evaporation with variations in these factors. Generally records are also obtained of the rate of evaporation from a free water surface, so that the data as to soil evaporation can be expressed in terms of evaporation opportunity, or ratio of soil evaporation to evaporation from a free water surface. In the experiments by Sleight mechanical analyses were made of five soils used in the evaporation tests, and these analyses gave a basis for determining the evaporation opportunity for different sizes of grain with each of several depths to the water table. It is desirable that in future tank experiments tests be made of the mechanical composition, porosity, permeability, and capillary range of the soils used.

The rate at which soil discharge occurs may be controlled either by the rate at which the water can be lifted by capillarity to the level at which it is evaporated or by the rate at which it can be evaporated. The rate at which water is lifted through the soil from the water table to the point at which it is evaporated depends chiefly on the permeability of the soil, the height the water is lifted, and the capillary head or difference between the potential and actual capillary lift. In other words, the law of capillary rise is essentially Darcy's law of flow, in which the energy is furnished by capillary attraction instead of by hydrostatic head. Therefore, if the water table is very near the surface and the soil is very permeable the rate of discharge is determined by the possible rate of evaporation; but if the water table is at greater depth and the soil is less permeable the limiting factor may be the possible rate of capillary rise. Further investigation of the processes involved in soil discharge and of the controlling laws should help to make the field methods more applicable and exact.

As a rule, vegetal discharge, which is effected by transpiration, is greater than soil discharge and occurs over wider areas, because the roots of certain plants lift water much higher than it is lifted by the capillary interstices of the soil. In arid regions nearly all the vegetal discharge of ground water is accomplished by plants of a few dominant and well-recognized species.[30] To determine the quantity of ground water discharged by the ground-water plants in any particular area it is necessary to make a map of the area showing, so far as may be practicable, the distribution, density, and growth of

[30] Meinzer, O. E., Plants as indicators of ground water: U. S. Geol. Survey Water-Supply Paper 577, p. 1, 1927.

the principal species, and to determine the rate at which they discharge water. Holes should also be dug to investigate the root habits of different species of plants and to determine to what depths their roots extend to reach the capillary fringe and feed on water from the zone of saturation.

Maps showing areas of ground-water discharge and the distribution of dominant species of ground-water plants have been made for several western valleys investigated by the Geological Survey—for example, the Independence district of Owens Valley, Calif.,[31] Sulphur Springs Valley, Ariz.,[32] Big Smoky Valley, Nev.,[33] Steptoe Valley, Nev.,[34] and parts of Escalante Valley, Utah.[35] Information that will be helpful in the construction of such maps is given in several of the United States Geological Survey water-supply papers.[36]

The rate of vegetal discharge has generally been determined by means of tanks in which the different species are grown with measured quantities of water and in some investigations with different depths to the water table. In Lee's investigations in Owens Valley the discharge of salt-grass areas was investigated by means of tank experiments in which natural conditions were reproduced as nearly as practicable. Since the work by Lee was done tank experiments have been made in connection with several ground-water investigations. In the recent investigation by White in the Escalante Valley records of discharge were obtained from soil tanks in which water tables were maintained and salt grass, alfalfa, and greasewood were grown. Comparative data were obtained on tanks with undisturbed salt-grass sod and on tanks which were filled and planted in the ordinary way. The method was also introduced of obtaining the dry weight of vegetable matter produced in the tank experiments and determining the ratio of weight of water consumed to weight of dry matter produced. This ratio was then applied to data on dry matter naturally produced in order to compute the natural discharge of ground water. The method of determining vegetal discharge from daily fluctuations of the water table is described on pages 116–119.

In humid regions the problems of both vegetal and soil discharge of ground water are complicated by the soil moisture that is derived

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[31] Lee, C. H., An intensive study of the water resources of a part of Owens Valley, Calif.: U. S. Geol. Survey Water-Supply Paper 294, pl. 25, 1912.

[32] Meinzer, O. E., and Kelton, F. C., Geology and water resources of Sulphur Spring Valley, Ariz.: U. S. Geol. Survey Water-Supply Paper 320, pls. 1, 2, 1913.

[33] Meinzer, O. E., Geology and water resources of Big Smoky, Clayton, and Alkali Spring Valleys, Nev.: U. S. Geol. Survey Water-Supply Paper 423, pl. 2, 1917.

[34] Clark, W. O., and Riddell, C. W., Exploratory drilling for water and use of ground water for irrigation in Steptoe Valley, Nev.: U. S. Geol. Survey Water-Supply Paper 467, pl. 2, 1920.

[35] White, W. N., op. cit., pl. 1 (map by W. N. White and Depue Falck).

[36] See especially Water-Supply Paper 423, pp. 92–102, and Water-Supply Paper 577, already cited.

directly from rain and snow. The problem of vegetal discharge is further complicated by the breaking down of the distinction between ground-water plants and plants that do not depend on ground water. No definite method has been developed for humid regions, but there are several possible lines of attack. More or less exact information on the extent of vegetal discharge of ground water in humid regions can be obtained from observations on the depth to the water table, the distribution of vegetation of different types, the relation of the root systems to the water table and overlying capillary fringe, and the amount of soil moisture and changes in soil moisture at different depths, especially in periods of drought. E. D. Burchard, district engineer in the Geological Survey, found from the record of an automatic water-stage recorder that the flow of a small stream in North Carolina had considerable daily fluctuations in the very dry month of July, 1926. These fluctuations were apparently caused by differences in the rate of evaporation in different parts of the 24-hour period, chiefly in vegetal and soil discharge of ground water, and they suggest a possible method for humid regions. It is likely, however, that such a method would give less than actual quantities. In the Pomperaug investigation some consideration was given to a possible method based on a ground-water rating curve (see p. 119), but the method actually used was merely to compute the ground-water evaporation as the difference between ground-water recharge and ground-water run-off.[37]

STORAGE METHODS IN GENERAL

If a capacity table or curve has been developed for a surface reservoir, a gage showing the changes in its water level will indicate for any period the net increase or decrease in storage. When the water level is rising it will register the excess in rate of recharge over rate of discharge, and vice versa. If it is known that there is no discharge in a certain period the record of rise will show the total rate of recharge during that period.

Because of the frictional resistance of water-bearing materials to the flow of the water from the intake to the discharge area of an underground reservoir the water table is almost nowhere a level surface, such as the water surface of an ordinary reservoir. Therefore, when this storage method is applied to underground reservoirs the depths to the water level in numerous wells are measured every week or month, or at other more or less frequent intervals, and continuous records are generally obtained on a few representative wells by means of automatic water-stage recorders. Levels are usually run

[37] Meinzer, O. E., and Stearns, N. D., A study of ground water in the Pomperaug Basin, Conn.: U. S. Geol. Survey Water-Supply Paper 597, pp. 138–139, 143, 1929.

to the observation wells, and a series of contour maps of the water table at successive dates are constructed. From these maps or from the base data it is possible to compute the volume of water-bearing material that became saturated or drained in a given period.

In order to evaluate these results in terms of water instead of water-bearing material it is necessary to apply a factor generally known as the specific yield. This factor represents the interstitial space that is emptied when the water table declines, expressed as a percentage of the total volume of material that is unwatered. The same factor generally represents the space that is occupied by water going into storage when the water table rises. If, however, the material into which the water rises has been further desiccated by soil evaporation or vegetal discharge the amount of recharge is greater for each unit of rise.

The water level in a well is sensitive to every force that acts upon the body of water with which the well communicates. Hence, in most wells the water level fluctuates almost constantly, often in a complicated manner. The curve produced by an automatic water-stage recorder over a well gives an accurate record of the resultant of the forces that act upon the ground water. A well that extends only slightly below the water table generally records faithfully both recharge and discharge, but great care must be taken in applying this method to use only wells in which the water level is a continuation of the water table and not the upper surface of a column of water that is supported by artesian pressure in the water-bearing formation and that therefore fluctuates as a result of atmospheric or other changes in pressure. Even in a true water-table well there may be fluctuations produced by changes in temperature which affect capillarity, by confinement of air in the interstices of the overlying zone of aeration by rain or frozen ground, or by other causes that have nothing to do with recharge or discharge.

In summer, when the vegetation makes heavy demands on the soil moisture, most rainstorms do not contribute at all to the supply of ground water and hence do not affect the water levels in wells. But in unusually heavy and prolonged storms or series of storms water may penetrate to the water table and the water levels may rise, or at least their rate of decline may be diminished.[38] Likewise recharge from stream flow or irrigation produces a rise in the water levels.[39]

In most regions of the United States recharge of the underground reservoirs occurs chiefly in the winter or in late autumn or early spring, when there is not much evaporation from the soil and little

[38] Meinzer, O. E., and Stearns, N. D., op. cit., pl. 19.

[39] Smith, G. E. P., Ground-water supply and irrigation in the Rillito Valley: Arizona Univ. Agr. Exper. Sta. Bull. 64, pp. 176–186, 1910.

or no transpiration from plants to consume the water that falls as rain or is produced by the melting of snow. Hence in most regions the water table is usually at its lowest stage in the autumn and at its highest stage in the spring. In very cold regions, however, the ground is likely to be frozen continuously during a long period in the winter, and the lowest stage may occur just before the ground thaws out in the spring. In arid regions such as New Mexico, in which the principal rainy season occurs in the summer, the annual recharge may be relatively small, but there may be a rise in the water table either in the summer or in the winter. In regions such as California, in which the rainy season is in the winter, nearly all the recharge may occur during this season of relatively small discharge. In such regions the increase in storage from the lowest stage in autumn to the highest stage in spring may represent approximately the annual recharge. In irrigated districts recharge may, of course, occur during the irrigation season.

The report by Smith [40] on the Rillito Valley, Ariz., includes a map on which are shown two sets of 5-foot contours of the water table—one set for the position of the water table on November 1, 1906, at the end of a dry season, and the other for its position on February 1, 1907, after the beginning of the spring flood season. It also contains a cross section of the valley on which are shown the profiles of the water table on successive dates in 1906 and 1907. From the rise in the water table shown in these illustrations and an assumed effective porosity, or specific yield, rough calculations were made of the recharge from the Rillito River in the flood season and of the probable supply of ground water annually available for irrigation. This method was applied more definitely and in more detail by Clark [41] in the Santa Clara Valley, Calif., from 1912 to 1920, and by Lee [42] in San Diego County, Calif., in 1914 and 1915. In both of these investigations numerous measurements were made of water levels in wells, data were obtained for estimating the specific yield, maps were prepared showing the contours of the water table at different dates, and computations were made of the seasonal increase in storage represented by the rise in the water table. More recently the same method has been used in making computations of net semiannual additions of water to or removals of water from the under-

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[40] Smith, G. E. P., Ground-water supply and irrigation in the Rillito Valley: Arizona Univ. Agr. Exper. Sta. Bull. 64, pp. 194, 195, figs. 55, 56, 1910.

[41] Clark, W. O., Ground-water resources of the Niles cone and adjacent areas, Calif.: U. S. Geol. Survey Water-Supply Paper 345, pp. 149–162, pl. 10, 1915; Ground water for irrigation in the Morgan Hill area, Calif.: U. S. Geol. Survey Water-Supply Paper 400, pp. 66–68, 76–105, pls. 6, 7, 1917; Ground water in Santa Clara Valley, Calif.: U. S. Geol. Survey Water-Supply Paper 519, pp. 58–75, 129–183, pls. 10–19, 1924.

[42] Ellis, A. J., and Lee, C. H., Geology and ground waters of the western part of San Diego County, Calif.: U. S. Geol. Survey Water-Supply Paper 446, pp. 121–155, pls. 20–25, 1919.

ground reservoir of the Mokelumne River area in California.[43] For this purpose monthly or semiannual measurements were made of the water levels in about 500 observation wells. On account of uncertainties as to the specific yield and as to pressure effects in observation wells there may be large percentage errors in the computed recharge in all these investigations, and in the present stage of development of this method it is impossible even to state what may be the limits of error.

SPECIFIC YIELD IN RELATION TO STORAGE METHODS

Perhaps the greatest difficulty in the application of quantitative methods lies in the variability in the texture and hence in the hydrologic properties of the water-bearing materials. The hydrologic properties vary greatly, even with apparently slight differences in texture. Hence the ordinary geologic descriptions are quite inadequate for hydrologic purposes, and quantitative descriptions based on laboratory determinations have become essential. However, the hard rocks that yield their water largely from joints or crevices are not amenable to laboratory tests, and the incoherent materials are difficult to handle without disturbing and repacking, which may considerably change their texture and hydrologic properties.

The two hydrologic properties of greatest significance are specific yield and permeability. Mechanical analyses and determinations of porosity and moisture equivalent are useful chiefly as indirect means of determining these two essential hydrologic properties. Any quantitative method that does not involve either of these properties has a great advantage in avoiding the complications that result from the heterogeneity of the water-bearing materials. Most storage methods involve the perplexing problem of specific yield.

Seven more or less distinct methods have been used to determine specific yield, but none of them have been thoroughly developed. These methods[44] are as follows: (1) Saturating samples in the laboratory and allowing them to drain; (2) saturating in the field a considerable body of material situated above the water table and above the capillary fringe and allowing it to drain downward naturally; (3) collecting samples immediately above the capillary fringe after the water table has gone down an appreciable distance, as it commonly does in summer and autumn; (4) ascertaining the

[43] Stearns, H. T., Robinson, T. W., and Taylor, G. H., Geology and water resources of the Mokelumne area, California: U. S. Geol. Survey Water-Supply Paper 619, pp. 112–172, 185–188, 292–398, pls. 6–8, 13, 1930; also later report released in manuscript form (on file in public library, Lodi, Calif.).

[44] Meinzer, O. E., The occurrence of ground water in the United States, with a discussion of principles: U. S. Geol. Survey Water-Supply Paper 489, pp. 67–76, 1923.

volume of sediments drained by heavy pumping, a record being kept of the quantity of water that is pumped; (5) ascertaining the volume of sediments saturated by a measured amount of seepage from one or more streams; (6) making indirect determinations in the laboratory with small samples by the application of centrifugal force; and (7) making mechanical analyses and determinations of porosity and estimating therefrom the specific retention and the specific yield.

The laboratory saturation and drainage method consists of draining high columns of saturated materials by gravity and determining both the volume of material drained and the volume of water yielded. The volume of water yielded can be measured directly or can be computed from the porosity and the moisture content after draining. The columns must be high enough to avoid the vitiating effects of the capillary fringe, and care must be taken to prevent loss by evaporation. As molecular forces vary with the temperature of the water, the tests must be made at a uniform temperature, or corrections for temperature must be applied. As drainage continues for a long time at a diminishing rate, the specific yield should be determined for specified periods of draining. If the material is fine, appreciable drainage may occur during a period of several weeks.

This method was used by Hazen [45] and by King.[46] In both sets of experiments the yield of successive segments of the different columns of water-bearing materials was determined by deducting the moisture content after drainage from the porosity, but in the King experiments the volume of water drained from each column was also measured at appropriate intervals of time.[47]

Much careful work has been done on this method by White [48] in 1925–1927, and by Taylor [49] since 1927. In White's work the following procedure was followed: A pit was dug and in the bottom a heavy metal cylinder, 36 or 42 inches high and 18 inches in diameter, was driven vertically into the water-bearing material within a few inches of the water table. The earth around the cylinder was then removed, and a metal plate was shoved under the cylinder and soldered to it, thus confining the column of undisturbed material in a water-tight vessel. Small wells were then sunk in the material in the cylinder,

[45] Hazen, Allen, Experiments upon the purification of sewage and water at the Lawrence experiment station: Massachusetts Board of Health Twenty-third Ann. Rept., pp. 428–434, 1892.

[46] King, F. H., Principles and conditions of the movements of ground water: U. S. Geol. Survey Nineteenth Ann. Rept., pt. 2, pp. 86–91, 1899.

[47] Meinzer, O. E., The occurrence of ground water in the United States, with a discussion of principles: U. S. Geol. Survey Water-Supply Paper 489, pp. 53–57, 1923. This paper gives a concise presentation and discussion of the Hazen and King experiments.

[48] White, W. N., Note in regard to investigation in the Escalante Desert, Utah: U. S. Geol. Survey Water-Resources Bulletin, March 10, 1927 (mimeographed reprint).

[49] Stearns, H. T., Robinson, T. W., and Taylor, G. H., Geology and water resources of the Mokelumne area, California: U. S. Geol. Survey Water-Supply Paper 619, pp. 151–172, 1930.

and the effects on the water levels in these wells produced by adding or withdrawing measured quantities of water were observed. From these data the specific yield was computed. This work was done in the autumn, when the water table was at its lowest stage; hence the tests were made on the material through which the water table fluctuates and which is therefore subjected to alternate draining and resaturation.

In the work by Taylor the general method developed by White was employed, but improvements were made in the technique, and extensive tests were made to determine the effects of changes in temperature and the periods required in different materials to reach approximate capillary equilibrium.

The method of field saturation and drainage is similar in principle to the laboratory method. A plot of land is selected where the water table and capillary fringe are at sufficient depth below the surface. The material underlying the plot is thoroughly wetted by applying water at the surface and is then allowed to drain, care being taken to avoid much evaporation. After a sufficient period of draining samples are taken for determination of moisture content and porosity, and the specific yield is computed as the difference between these two. This method has been used by Israelsen [50] and other students of soil water. If the specific yield of material below the soil that is more representative of the water-bearing material is desired, it is necessary to apply enough water to wet this deeper material and to obtain samples from it. In some situations it might be advantageous to dig pits to the material to be tested.

In the direct sampling method samples are obtained by boring into the material immediately above the capillary fringe after the water table has declined appreciably, and the moisture content and porosity of the samples are determined. This is essentially the method used by Lee [51] in San Diego County, Calif. The essential feature in this method is to take a sample where the water table has gone down, as it generally does during the dry summer season in California. For the most conclusive results the sample should be taken from a point which in the preceding wet season was below the water table and which at the time of taking the sample is far enough above the water table not to be seriously affected by the capillary fringe. Where the fluctuation of the water table is less than the thickness of the capillary fringe the most significant samples are those taken just above the fringe. As in practice this position can not be very definitely determined it is advisable to take samples at several

[50] Israelsen, O. W., Studies in capacities of soils for irrigation water: Jour. Agr. Research, vol. 13, pp. 1–28, April 1, 1918.

[51] Ellis, A. J., and Lee, C. H., Geology and ground waters of the western part of San Diego County, Calif.: U. S. Geol. Survey Water-Supply Paper 446, pp. 121–123, 1919.

levels, as was done by Lee. In making tests of this kind it is essential to ascertain that the part of the deposit from which the sample is taken has not received any recent contribution of water from rain or irrigation and has not been exposed to evaporation or to absorption by plants, both of which consume water that is retained against gravity by molecular attraction.

The pumping method consists in observing the lowering of the water table and hence the volume of sediments drained by pumping a measured volume of water. The specific yield is, of course, the ratio of the volume of water pumped to the volume of material drained. Serious errors may be introduced by pumped water returning to the water table and by other percolation into or out of that part of the underground reservoir which is supplying the pumped water. This method was used by Clark [52] with records obtained in the Morgan Hill area, California, and is more or less applicable to other areas having water-table conditions and heavy seasonal pumpage.

The recharge method, which is the converse of the pumping method, consists of observations on the seepage losses from streams or canals and on the resultant rise of the water table, from which is computed the volume of material saturated. It is, of course, merely the storage method of determining recharge, with the specific yield instead of the quantity of water as the unknown factor.

The moisture-equivalent and mechanical-analysis methods are based on investigations by Briggs, Veihmeyer, and others [53] on the relations between the moisture equivalent or mechanical composition of the water-bearing materials, on one hand, and the specific retention, on the other hand. The specific yield is computed as the difference between porosity and specific retention as determined by one of these indirect methods. Both of these methods are valuable, at least in giving a general check on the results from other methods, and they deserve further investigation.

METHODS FOR ESTIMATING DISCHARGE FROM DAILY FLUCTUATIONS OF THE WATER TABLE

In most localities in which plants draw water from the zone of saturation there are daily fluctuations of the water table ranging from a small fraction of an inch to several inches. To obtain curves of these daily fluctuations the most sensitive water-stage recorders are used.

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[52] Clark, W. O., Ground water for irrigation in the Morgan Hill area, Calif.: U. S. Geol. Survey Water-Supply Paper 400, pp. 84–86, 1917.

[53] Briggs, L. J., and McLane, J. W., The moisture equivalents of soils: U. S. Dept. Agr. Bur. Soils Bull. 45, 1907. Briggs, L. J., and Shantz, H. L., The wilting coefficient for different plants and its indirect determination: U. S. Dept. Agr. Bur. Plant Industry Bull. 230, 1912. Briggs, L. J., Martin, O. F., and Pearce, J. R., The centrifugal method of mechanical soil analysis: U. S. Dept. Agr. Bur. Soils Bull. 24, 1904. Veihmeyer, F. J., and Hendrickson, A. H., The moisture equivalent as a measure of the field capacity of soils (unpublished manuscript).

During the daytime the plants withdraw water and consequently lower the water table. This drawdown of the water table produces a head in the underlying strata, which causes continual upward percolation. Therefore at night, when there is little or no withdrawal by plants, the water table rises much as the water level in a well rises when pumping is stopped. The daily transpiration is equal to the upward percolation throughout the 24-hour period plus or minus the comparatively small amount of water represented by the net decline or rise of the water table during the same period. The average rate of upward percolation occurs when the water table stands about midway between its lowest and highest daily stages. This occurs twice a day—during the declining phase, some time in the afternoon, and during the rising phase, usually late in the night. The nocturnal mid-stage occurs at a time when there is little or no withdrawal from the zone of saturation, and hence at a time when the replenishment of storage that causes the rise of the water table is approximately the average rate of upward percolation for the 24-hour period.[54] In order to ascertain the quantity of water withdrawn by the plants from the zone of saturation it is therefore necessary to determine the quantity of water that is required to raise the water table an inch, a foot, or other unit of height—that is, to determine the specific yield of the material through which the water table fluctuates.

Daily fluctuations of the water table were observed by King[55] in his experiments at Madison, Wis., in 1888 and subsequent years, and he recognized their possible significance in recording the discharge of ground water through vegetation. However, he dismissed the subject as too complicated for solution without more elaborate investigation. The method was successfully applied by G. E. P. Smith, in 1917 and subsequent years, by the use of automatic water-stage recorders on wells in tracts of cottonwood and mesquite and later also on wells in tracts of salt grass and alkali sacaton. To Smith, therefore, belongs the credit for the introduction of this method. He also developed the theory of upward percolation and showed that the daily vegetal discharge could be computed from the rate of rise of the water table at the nocturnal mid-stage if the specific yield were known.[56]

In 1925 the United States Geological Survey undertook an intensive investigation to develop this promising method and especially to devise means for interpreting the fluctuations in terms of quantity

[54] Smith, G. E. P., The effect of transpiration of trees on the ground-water supply (unpublished paper given before the Geological Society of Washington, Nov. 22, 1922). See Washington Acad. Sci. Jour., vol. 14, p. 160, 1924.

[55] King, F. H., Observations and experiments on the fluctuations in the level and rate of movement of ground water on the Wisconsin Agricultural Experiment Station farm at Whitewater, Wis.: U. S. Weather Bur. Bull. 5, 1892.

[56] Smith, G. E. P., Washington Acad. Sci. Jour., vol. 14, p. 160, 1924.

of ground water discharged by the plants. For this purpose the Escalante Valley, Utah, was chosen, and W. N. White, who had already done considerable hydrologic work in that valley, was assigned to make the investigation. A preliminary statement of the results of the first two years of his work on this method was published in 1927,[57] and a comprehensive report on the results of all three years of his work is nearly ready for publication.[58]

In the course of the Escalante Valley investigation three methods of determining the specific yield were devised and put into practice. One of these methods consisted of driving cylinders and making direct tests as above described (p. 114). In another method several soil tanks were used in which the three principal ground-water plants of the region—alfalfa, salt grass, and greasewood—were grown. The natural hydraulic conditions were reproduced as closely as possible, daily water-table curves were obtained comparable to those obtained in the field, and observations of the quantity of water fed into the tanks to replace the upward percolation were made at frequent intervals. Thus data were obtained for the evaluation of every process involved, including the interpretation of the rise and fall of the water table in terms of quantity of water. In the third method the ratio of dry alfalfa hay produced in the tank experiment to the quantity of water consumed was determined, and this ratio was used to compute the water consumed by a known weight of alfalfa growing as a ground-water plant in a field in which a well was dug and a water-stage recorder was operated. Thus it became possible to interpret the curve of water-table fluctuations in terms of quantity of water and hence to compute the specific yield. This might be regarded as an eighth method of determining specific yield. Similar ratios were also determined for salt grass and greasewood, and these were applied directly to the seasonal growth of these species in the field to compute the quantities of water that they discharged. As these computations did not involve the fluctuations of the water table they served as a check on the daily fluctuation method.

The studies of the daily fluctuations of the water table in areas of vegetal discharge have given valuable results. However, the application of the method is complicated in several ways, chiefly by the variability in the texture of the soil and water-bearing material as it affects the specific yield, permeability, capillarity, and root

[57] White, W. N., Notes in regard to the work in the Escalante Valley, Utah, on the discharge method of estimating ground-water supplies: U. S. Geol. Survey Water-Resources Bulletin, Mar. 10, 1927, pp. 22–29. See also Washington Acad. Sci. Jour., vol. 17, pp. 238–240, 1927.

[58] White, W. N., A method of estimating ground-water supplies based on discharge by plants and evaporation from soil; results of investigations in Escalante Valley, Utah (released in manuscript form in 1930).

distribution. The method is not applicable where the water-bearing formation consists of very permeable material without interstratified confining beds, because the fluctuations are too slight to be accurately recorded by available instruments, and, moreover, the mid-stage of the rising phase, as well as that of the declining phase, occurs while discharge is occurring. The method will probably have to be used to determine the approximate consumption of water by the principal species of ground-water plants, and these results will have to be applied to maps that will show in as much detail as practicable the area, density, and growth of each of these species.

METHOD BASED ON GROUND-WATER RATING CURVE

In the Pomperaug investigation weekly measurements were made of the depths to the water levels in a number of observation wells, and the average depth for each week was plotted against the corresponding ground-water run-off. It was, of course, found that the run-off for a given stage of the water table was much less in summer, when vegetal discharge and soil evaporation were active, than in the colder part of the year, when there was very little loss by these processes. A curve was drawn through or near the coordinate points for winter weeks, according to the empiric method used in making rating curves for surface streams. This gave a rating curve for ground-water run-off during the winter. It was hoped that the curve would show approximately the total ground-water discharge in summer and that the ground-water evaporation could be obtained by subtracting the ground-water run-off from the total discharge as shown by the curve. In fact, however, this difference seems to give minimum rather than actual figures for ground-water evaporation, because the roots of plants and the soil capillaries do not wait for the water to appear at the surface but pump it up from some depth.[59] Further investigation will be required to determine whether it will be practicable under any circumstances to use the stage of the water table as an index either of ground-water run-off or of total ground-water discharge.

SAFE YIELD AS COMPARED WITH NATURAL INTAKE AND DISCHARGE

The safe yield of a water-bearing formation, or the practicable rate of withdrawing water from it perennially for human use, may be either greater or less than the rate of natural recharge or discharge as determined before heavy withdrawals are begun. It may

[59] Meinzer, O. E., and Stearns, N. D., A study of ground water in the Pomperaug Basin, Conn., with special reference to intake and discharge: U. S. Geol. Survey Water-Supply Paper 597, fig. 13, 1928.

be less because natural discharge can not be wholly prevented, even by heavy withdrawals through wells. On the other hand, it may be greater because of artificial increase of the rate of recharge.

Before any large ground-water developments are made the average rate of discharge for any long period is obviously about equal to the average rate of recharge. When heavy pumping from wells is begun the water table declines, and this results in curtailing the loss from the underground reservoir by spring flow, seepage, soil evaporation, vegetal discharge, and perhaps underground leakage. It is, however, not practicable to prevent all the natural discharge even if the water table is greatly lowered. Obviously the water that is at first pumped is largely water taken out of storage. As pumping progresses the water table continues to decline and the salvage from natural discharge increases, until, if the safe yield is not exceeded, the pumpage is eventually all salvage, and there is no further net withdrawal from storage and no further permanent lowering of the water table. If, however, the pumpage exceeds the safe yield, withdrawal from storage will continue and the water table will persistently decline. In all investigations of undeveloped underground reservoirs the factor of unsalvageable natural discharge should be recognized.

Surface reservoirs that are full have no capacity for additional storage. Underground reservoirs are also limited in their capacity, though not so definitely as surface reservoirs. In a long period of wet weather the water table may become so high and effluent seepage may become so general that the underground storage will not be increased even by heavy precipitation, and after the rains cease stored water will be rapidly discharged until the water table reaches a more normal stage. This temporary storage capacity depends largely on the relief of the land, as has been pointed out by George Otis Smith.[59a] If the land surface is nearly level there is little opportunity for the underlying water-bearing beds to drain, and hence the springs will decline or dry up completely while the underground reservoir remains nearly full. On the other hand, if the region is deeply dissected the lowland springs may continue to flow with little decrease in volume throughout long periods of drought, during which they are gradually draining the underground reservoir. If prior to a wet season the water table has been lowered either by the flow of springs or by pumping from wells, the available capacity of the underground reservoir is increased, and more of the surface water will go into underground storage than would otherwise be possible. In this respect the recharge may be increased by use.

In order to utilize fully the flow of a stream, storage is required to hold the excess water of high stages for use in times of small

[59a] Personal communication.

flow or whenever there may be need for the water. Many of the underground reservoirs are so large that they have capacity to carry over great quantities of water not only from a wet season to the following dry season but also from a period of wet years to a period of dry years. However, to utilize these reservoirs fully it is necessary to pump enough water out of them to make room for all the inflow during the wettest seasons and during the periods of successive years of heavy precipitation. This was well illustrated by some of the underground reservoirs of southern California, whose water tables, under heavy pumping for irrigation, went down a little lower each summer than they had risen in the previous winter, until it appeared that excessive depletion must inevitably compel reduction in irrigation. Then came a period of wet winters when recharge occurred to a remarkable extent and the water levels rose beyond all expectations.[60]

The principle of increasing recharge by use was recognized by Smith [61] and Wolff [62] in early quantitative work in this country and has generally received due consideration in more recent investigations. In all investigations of undeveloped ground-water supplies the possible effect of use in increasing the yield should be recognized and so far as practicable should be evaluated. In making investigations where heavy pumping is already in progress, estimates based on a few years of normal or subnormal precipitation may be too low because they do not sufficiently take into account the great recharge that may occur in a depleted underground reservoir during a season of exceptionally high precipitation.

Artificial recharge can be accomplished in some places by draining surface water into wells, spreading it over tracts underlain by permeable material, temporarily storing it in leaky reservoirs from which it may percolate to the water table, or storing it in relatively tight reservoirs from which it is released as fast as it can seep into the stream bed below the reservoir. Artificial recharge by some of these methods has been practiced in the United States and other countries. It was suggested by Hilgard [63] in 1902 for southern California, where it has since received considerable investigation and has been adopted as a conservation measure.[64] Drainage

[60] Ebert, F. C., Records of water levels in wells in southern California: U. S. Geol. Survey Water-Supply Paper 468, pls. 2, 3, 4, 1921.

[61] Smith, G. E. P., op. cit., p. 189.

[62] Wolff, H. C., The utilization of the underflow near St. Francis, Kans.: U. S. Geol. Survey Water-Supply Paper 258, pp. 104–107, 1911.

[63] Hilgard, E W., Subterranean water supply of the San Bernardino Valley: U. S. Dept. Agr. Bull. 119, pp. 133–134, 1902.

[64] Lee, C. H., Subterranean storage of flood water by artificial methods in San Bernardino Valley, Calif.: California Conservation Comm. Rept. for 1912, pp. 335–400, 1913. Tate, C. E., Spreading water for flood control: Southern California Assoc. Members Am. Soc. Civil Eng. Bull., vol. 1, No. 4, pp. 76–86, Los Angeles, 1919. Post, W. S., Santa Ana investigation, flood control and conservation: California Dept. Public Works Div. Eng. and Irr. Bull. 19, pp. 165–179, 1929.

into wells has been practiced in many parts of the United States, chiefly to reclaim swampy land or to dispose of sewage and other wastes.[65] The drainage of sewage or other wastes into wells can not be approved, because it may produce dangerous pollution of water supplies. Drainage of surface water into wells to increase the ground-water supply for rice irrigation in Arkansas is now under consideration.[66] Water spreading has been practiced to a considerable extent in southern California, partly to decrease the effects of flood but largely to increase the supply of ground water. Storage in ordinary reservoirs and subsequent release has frequently been considered, and the unavoidable leakage of some reservoirs has been used to increase the ground-water supply.[67] Artificial recharge by damming stream channels in the permeable lava rocks of the Hawaiian Islands has been considered.[68] In ground-water investigations that involve the question of safe yield attention should as a matter of course be given to the possibilities of artificial recharge.

STORAGE METHODS APPLIED WHERE USE HAS BECOME LARGE

If an underground reservoir is heavily pumped a study of the relations of the water levels to the amount of pumpage is likely to give more reliable information as to the safe yield than can be obtained by any method of studying an undeveloped reservoir. If the water levels in the wells remain virtually stationary during a considerable period of pumping it may be concluded that during this period the rate of recharge has been about equal to the rate of discharge, including both natural discharge and withdrawals from wells. This principle was utilized by Mendenhall [69] in his ground-water investigations in San Bernardino Valley, Calif., early in the present century, where already there was heavy pumping for irrigation. He found that the records on two wells under observation for several years showed a net decline in water level in all years, though in two years of somewhat more than average precipitation the decline was only slight. He therefore concluded that in those two years the pumpage was virtually supplied by recharge but that in all other years there was a draft on storage.

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[65] Horton, R. E., Drainage of ponds into drilled wells: U. S. Geol. Survey Water-Supply Paper 145, pp. 30–39, 1905. Fuller, M. L., Drainage by wells: U. S. Geol. Survey Water-Supply Paper 258, pp. 6–22, 1911.

[66] Thompson, D. G., Ground-water supplies for rice irrigation in the Grand Prairie region, Arkansas: U. S. Dept. Interior Press Mem. 49844, January 26, 1931.

[67] Meinzer, O. E., Renick, B. C., and Bryan, Kirk, Geology of No. 3 reservoir site of the Carlsbad irrigation project, New Mexico, with respect to water-tightness: U. S. Geol. Survey Water-Supply Paper 580, pp. 21–25, 1927.

[68] Stearns, H. T., and Clark, W. O., Geology and water resources of the Kau District, Hawaii: U. S. Geol. Survey Water-Supply Paper 616, p. 21, 1930. The method had been suggested by Lincoln McCandless in Honolulu Advertiser, June 21. 1926.

[69] Mendenhall, W. C., The hydrology of San Bernardino Valley, Calif.: U. S. Geol. Survey Water-Supply Paper 142, pp. 56–67, 1905.

Regardless of the manner in which the water levels fluctuate, if at the end of any period they return approximately to the position they had at the beginning of the period the record of pumpage furnishes a measure of the recharge during the same period minus the natural loss. In the Niles cone, in California, the water levels in the autumn of 1918 had returned to almost the identical levels of the autumn of 1913, although in the intervening period there were notable fluctuations in level as a result of wet and dry years. Bailey's inventory of the ground-water supply[70] covered this 5-year period in which there was practically no net change in storage.

Under such conditions two uncertainties remain which affect the practical conclusions as to safe yield—first, as to how much natural discharge there was during the period and especially how much salvageable natural discharge; and, second, as to how nearly the recharge during the period represents the average rate of recharge or how much it was above or below the average. If it is known that during a period of such stationary or recurring conditions of the water table the rate of recharge has not been above the average, it may be concluded that pumpage has not exceeded the safe yield. On the other hand, the mere fact of essential equilibrium between recharge and discharge gives no indication of the maximum salvageable discharge—that is, of the maximum pumpage that would be practicable. If the rate of pumping is increased, the water table will decline and consequently the rate of natural discharge will be diminished. If the safe yield has not been exceeded, this process of lowering the water table and decreasing the natural discharge will proceed until eventually the diminution in the rate of natural discharge will equal the increase in the rate of pumping, the aggregate of the two kinds of discharge will again be balanced by recharge, there will no longer be any net withdrawal from storage, and the water table will be approximately stabilized in its new position. If the safe yield has been exceeded and the new rate of pumping is greater than the maximum possible salvage, then, obviously, the pumps will continue to draw upon the storage and the water table will continue to decline indefinitely. If with a given rate of pumping the water table is approximately stabilized and the natural discharge is known to be small or to have been reduced nearly to the practicable limit, it may be concluded that the pumpage furnishes an approximate measure of the safe yield and that the ground-water supply has been fully developed but not overdeveloped.

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[70] Bailey, Paul, Engineering investigation of percolation from Alameda Creek and ground-water studies on Niles cone: California State Water Comm. Third Bienn. Rept., pp. 105, 106, 1921.

The great advantage in basing computations on a period of stationary water levels or on a period in which the water levels are about the same at the end as at the beginning of the period is that there is no net withdrawal from or addition to the storage, and consequently the specific yield of the water-bearing material is not involved. If, however, the water table declines persistently or if it is desired to make monthly or annual estimates of the recharge, it becomes necessary to determine the specific yield or at least the capacity of the underground reservoir for each foot of saturation within the belt through which the water table migrates. In regions that have distinct wet and dry seasons, such as California and parts of the Hawaiian Islands, the recharge during the dry season may be negligible in quantity. If natural discharge has become relatively small, the pumpage in the dry season may therefore represent approximately the reduction in storage. If both the pumpage and the average decline of the water table are known, the approximate capacity of the underground reservoir for each foot of depth can be computed, and this figure can be used in computations of recharge during the wet season. In applying this method adequate precautions must, of course, be taken to get the average position of the water table for the entire area underlain by the underground reservoir and to distinguish between water levels that represent the water table and those that represent artesian pressure. This method was used by McCombs [71] in the artesian basin in the vicinity of Honolulu, where there is continuous heavy withdrawal from wells but only a negligible amount of natural discharge and practically no recharge during the dry seasons, and where the water-bearing lava rock is so permeable that there is almost no loss of head due to friction and the water table in the intake area is almost exactly recorded by the water levels in the artesian wells.

By this method the average specific yield of the water-bearing material that has been drained can be determined, provided, of course, that the area of the water table in the underground reservoir is known with reasonable accuracy. This is the fourth or pumping method described above. However, it should be remembered that in the procedure followed by McCombs the specific yield is not required but only the storage for each foot of depth.

LEAKAGE METHODS

In some places there is good evidence of leakage from one underground reservoir into another. If the leakage occurs as underflow

[71] McCombs, John, Methods of estimating safe yield of Honolulu artesian area: Honolulu Sewer and Water Comm. Rept., 1927, pp. 61–65.

through a relatively definite underground channel it can be estimated by one of the conduit methods described below, as, for example, in the investigation made by Slichter[72] in 1904 of the underflow at the narrows of the Rio Grande near El Paso, Tex. If, however, the leakage occurs by percolation through the rocks at considerable depths it may be quite impracticable to estimate the quantity of water that is lost. An example of such a condition is afforded by the probable underground escape of water from Alkali Spring Valley into Clayton Valley, in Nevada.[73]

After wells have been drilled into an artesian formation much artesian water may escape into permeable beds above the confining bed through defective or corroded well casings. Recently a successful method has been developed for measuring this underground loss of artesian water. The principle involved is very simple. A specially constructed current meter is lowered into the well, where it records the velocity of the water at any depth desired. With the velocity and the diameter of the well known, the rate of flow at any point can be determined. Hence a series of measurements taken at successive levels from the bottom to the top of the well will show just where the water enters and in what amounts, where it leaks out and in what amounts, and how much is delivered at the surface. For certain investigations it may be necessary to use apparatus for measuring the diameter of a well at any depth, but thus far no serious difficulties have arisen in the leakage work on account of uncertainty as to the diameter.

In connection with drainage investigations on the Twin Falls South Side project, in Idaho, from 1910 to 1914, a Price current meter, such as is used by the United States Geological Survey in stream gaging, was used by W. G. Sloan, of the Department of Agriculture, and W. B. Heroy, of the Geological Survey, to measure the discharge of wells and also to determine the levels at which the water entered the wells. Beginning in 1918, a Price meter was successfully used by the Geological Survey for making a systematic survey of leakage of artesian wells in Honolulu, first by R. D. Klise and later by John McCombs and others. In 1925 a current meter that is much better adapted for the well work was designed by Carl H. Au and was used by A. G. Fiedler, who further developed the method in leakage investigations in the Roswell artesian basin, New Mexico, and in other localities.[74] A so-called "kink finder," or

[72] Slichter, C. S., Observations on the ground waters of Rio Grande Valley: U. S. Geol. Survey Water-Supply Paper 141, pp. 9–13, 1905.

[73] Meinzer, O. E., Geology and water resources of Big Smoky, Clayton, and Alkali Spring Valleys, Nev.: U. S. Geol. Survey Water-Supply Paper 423, pp. 147–150, 1917.

[74] McCombs, John, and Fiedler, A. G., Methods of exploring and repairing leaky artesian wells, with a preface by O. E. Meinzer: U. S. Geol. Survey Water-Supply Paper 596, pp. 1–32, 1927. The work of Sloan and Heroy was not mentioned in this paper because it was not at that time known to any of the authors.

automatic gage for determining the diameter of a well from bottom to top, invented by P. B. Whitney, was successfully used, in 1927, in a 4,700-foot well of the Repollo Oil Co., near Boulder, Colo.[75]

In an investigation in the Winter Garden area, in Texas, it was necessary to determine whether salty water from one formation is entering a fresh-water formation through wells with defective casings. For this purpose an apparatus was devised which can be lowered into the wells and will indicate the electric conductivity and therefore the degree of salinity of the water at different depths.[76]

ROCK FORMATIONS AS CONDUITS

GENERAL CONDITIONS

Ground water travels laterally from the intake areas to the discharge areas. On account of the viscosity of the water and the small size of the interstices of the water-bearing material through which it usually percolates, there is a very appreciable resistance to its flow. Hence the rate of recharge is controlled not only by the intake facilities and the quantity of surface water that reaches the intake area but also by the capacity of the formation to carry water from the intake area to the area of natural discharge or to the wells where the water supply is to be recovered. If the distance from the intake area to the discharge area is great, as it is in some of the large artesian basins, the problem of safe yield relates to the capacity of the formation as a conduit rather than to its capacity as a reservoir. Quantitative methods are therefore required to estimate the transmission capacities of water-bearing formations.

METHODS BASED ON FIELD TESTS OF VELOCITY

The rate of flow of a surface stream at a selected cross section is determined by measuring the velocity of the water in each unit of the cross section. The rate of flow through each unit is the product of the velocity by the area of the unit, provided the flow is at right angles to the cross section. The total flow of the stream is, then, the sum of the flows through all the units of the cross section.[77]

The same principle is applied in estimating the flow of ground water, but different means must be used for measuring the velocity and cross-section area, and a third factor, which may be called the

[75] Described by Mr. Whitney in letter of May 14, 1927, to the Director, U. S. Geological Survey.

[76] Survey of the underground waters of Texas: U. S. Dept. Interior Press Mem. 50678, Feb. 16, 1931.

[77] Hoyt, J. C., and Grover, N. C., River discharge, 4th ed., pp. 3, 4, 44–81, New York, John Wiley & Sons, 1927.

effective porosity, is involved. The velocity of a surface stream is generally measured with a current meter, but this instrument is obviously not available for measuring the velocity of ground water. For this purpose use has been made of salts and dyes, which are introduced into the ground water through an upstream well and later detected in one or more downstream wells, the velocity being computed from the observed interval of time and the measured distance between the wells. Common salt was first used by European hydrologists and was detected in the downstream wells by chemical tests for chloride in successive samples taken from these wells. Later the more convenient electrolytic method was devised in this country by Slichter.

One of the pioneers in developing methods for measuring the velocity of ground water was the German hydrologist A. Thiem,[78] who made investigations leading to the development of ground-water supplies in many cities of Germany and published the results of his work in several papers that appeared between 1879 and 1892. He used common salt in his experiments because it is inexpensive, readily soluble, easily detected by the test for chloride, and not objectionable in any way. His method was to dig two test wells approximately in line with the direction of the movement of the ground water as determined from the slope of the water table. He then dosed the upper well with salt and at suitable intervals took samples from the lower well which he tested for their chloride content.

A notable advance in the methods of measuring the velocity of ground water was made in this country when the electrolytic method was developed by Slichter.[79] His first work was done in 1901 in the Arkansas River Valley, near Garden City, Kans.[80] In succeeding years he and his associates made measurements of the velocity of ground water by the electrolytic method on Long Island [81] and in the valleys of the Los Angeles,[82] Hondo,[83] San Gabriel,[83] and Mohave [83]

[78] Thiem, A., Verfahren für Messung natürlicher Grundwasser-geschwindigkeiten: Poly. Notizblatt, vol. 42, p. 229, 1887. Slichter, C. S., Theoretical investigation of the motion of ground waters: U. S. Geol. Survey Nineteenth Ann. Rept., pt. 2, pp. 297–384 (see especially the bibliography, pp. 380–384), 1898; The motions of underground waters: U. S. Geol. Survey Water-Supply Paper 67, pp. 46–48, 1902.

[79] Slichter, C. S., The motions of underground waters: U. S. Geol. Survey Water-Supply Paper 67, 1902; Field measurements of the rate of movement of underground waters: U. S. Geol. Survey Water-Supply Paper 140, 1905.

[80] Slichter, C. S., The underflow in Arkansas Valley in western Kansas: U. S. Geol. Survey Water-Supply Paper 153, 1906.

[81] Slichter, C. S., Field measurements of the rate of movement of underground waters: U. S. Geol. Survey Water-Supply Paper 140, pp. 65–85, 1905. Veatch, A. C., Slichter, C. S., Bowman, Isaiah, Crosby, W. O., and Horton, R. E., Underground water resources of Long Island, N. Y.: U. S. Geol. Survey Prof. Paper 44, pp. 86–115, 1906.

[82] Hamlin, Homer, Underflow tests in the drainage basin of Los Angeles River: U. S. Geol. Survey Water-Supply Paper 112, 1905.

[83] Slichter, C. S., Field measurements of the rate of movement of underground waters: U. S. Geol. Survey Water-Supply Paper 140, pp. 50–64, 1905.

Rivers, in California, the Rio Grande near El Paso, Tex.,[84] the South Platte River, in Nebraska,[85] and the Republican River, in Kansas.[86] The method has been standardized and has been extensively used in ground-water work, both in the United States and in other countries.

The electrolytic method depends on the fact that the electric conductivity of water increases with its concentration of dissolved salt, which acts as an electrolyte. Simple electric apparatus is used with a battery to generate the current and an ordinary or recording ammeter to measure the current. The progress of the salt is recorded when the current is passed through the water from the upstream well to a downstream well, and its arrival in one of the downstream wells is indicated when the current passing from the casing of that well to an electrode inside of it suddenly increases. Many mechanical difficulties are involved in the installation of the wells and the application of the electrolyte, which may introduce large errors in the results or may completely vitiate the experiments. The difficulties are especially serious where the velocities are small.

Dye has been used chiefly in sanitary investigations, largely to trace the course of rather definite underground streams, such as occur in limestone, and to determine whether certain water supplies receive contributions from polluted sources. The dye most commonly used is the chemical called uranin, which is the sodium salt of fluorescein and which has often been incorrectly called fluorescein. This is a harmless but very strong and stable dye that colors the water green. The dye method was described by Dole[87] in 1906, in a paper which also contains a historical sketch of the development of the method and a bibliography. Dye tests were made by Dr. Dionis des Carrières in 1882, during a severe epidemic of typhoid fever in the city of Auxerre, in France, and since that time frequent use has been made in France of various dyes to study the movements of ground water. In 1899 elaborate investigations were made by Trillat of the value of certain dyes as flow indicators and of the effects on them of passage through common soils. The fluoroscope, for detecting dyes when present in great dilution, was invented by Trillat and was perfected by Marboutin. Dye tests have frequently been made in the United States by sanitary engineers and

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[84] Slichter, C. S., Observations on the ground waters of Rio Grande Valley: U. S. Geol. Survey Water-Supply Paper 141, pp. 9–13, 1905.

[85] Slichter, C. S., and Wolff, H. C., The underflow of the South Platte Valley: U. S. Geol. Survey Water-Supply Paper 184, pp. 9–12, 1906.

[86] Wolff, H. C., The utilization of the underflow near St. Francis, Kans.: U. S. Geol. Survey Water-Supply Paper 258, pp. 98–119, 1911.

[87] Dole, R. B., Use of fluorescein in the study of underground water: U. S. Geol. Survey Water-Supply Paper 160, pp. 73–85, 1906.

others, and a good example is afforded by the series of tests made by Stabler.[88]

In 1921 and subsequent years uranin dye was successfully used by Stiles and his assistants [89] in an investigation to determine the extent to which bacteria are carried through formations of sand. This investigation, which was made at an experiment station near Fort Caswell, N. C., by the United States Public Health Service, with the cooperation of the United States Geological Survey, involved a minute 3-dimensional survey of the direction and rate of movement of the ground water and made a notable contribution in demonstrating the use of the dye method in fine-grained materials and in providing means for detailed study of the movement of ground water.

The use of a dye probably affords the most accurate method of studying in detail the movements of ground water. It may be rather easily applied in some creviced rocks that have relatively definite underground streams, but in porous rocks with small interstices the dye may be very elusive, and the method may be found to be much more laborious and difficult than would appear on casual consideration. In the Fort Caswell experiment about 550 test wells were sunk in addition to some trenches, and the exact distribution of the uranin was determined.

To estimate the flow of ground water, whether by electrolytic, chemical, or dye methods, it is necessary to ascertain with some degree of accuracy the cross section through which the water flows and the velocity of flow through each unit of the cross section. This is accomplished by sinking test wells and to some extent by studying the records of existing wells. It is possible that the recently devised method of underground exploration based on the electric conductivity of the different rock formations can be used in finding the limits of shallow underground conduits, but that method seems not to be available for determining the permeability of different units of the cross section or the velocity of the ground water flowing through them.

The flow of a surface stream, in cubic feet per second, can be found by multiplying the average downstream component of the velocity of the water in feet per second by the area of the cross section of the stream in square feet. To determine the flow of ground water, however, a third factor, which has been called the effective

[88] Stabler, Herman, Fluorescein an aid to tracing waters underground: Reclamation Record, vol. 12, pp. 122, 123, U. S. Recl. Service, 1921.

[89] Stiles, C. W., Crohurst, H. R., Thomson, G. E., and Stearns, N. D., Experimental bacterial and chemical pollution of wells via ground water, with a report on the geology and ground-water hydrology of the experimental area at Fort Caswell, N. C.: U. S. Pub. Health Service Hygienic Lab. Bull. 147, 1927.

porosity, must be applied. Much of the cross section is occupied by rock and by water that is securely attached to the rock surfaces by molecular attraction. The area through which the water is flowing is therefore less than the area of the cross section of the water-bearing material and may be only a small fraction of that area. In a coarse, clean gravel, which has only large interstices, the effective porosity may be virtually the same as the actual porosity, or percentage of pore space; but in a fine-grained or poorly assorted material the effect of attached water may become very great, and the effective porosity may be much less than the actual porosity. Clay may have a high porosity but may be entirely impermeable and hence have an effective porosity of zero. The effective porosity of very fine grained materials is generally not of great consequence in determinations of total flow, because in these materials the velocity is so slow that the computed flow, with any assumed effective porosity, is likely to be relatively slight or entirely negligible. The problem of determining effective porosity, as distinguished from actual porosity, is, however, important in studying the general run of water-bearing materials, which are neither extremely fine nor extremely coarse and clean.

Hitherto not much work has been done on this phase of the velocity methods of determining rate of flow. No distinction has generally been made between actual and effective porosity, and frequently a factor of 33⅓ per cent has been used, apparently without even making a test of the porosity. It is certain that the effective porosity of different water-bearing materials ranges between wide limits and that it must be at least roughly determined if reliable results as to rate of flow are to be obtained. It would seem that each field test of velocity should be supplemented by a laboratory test of effective porosity, for which the laboratory apparatus devised by Slichter could be used.[90]

Velocity methods have been used chiefly on water-bearing formations lying at or near the surface. The electrolytic method has been used chiefly to determine the underflow of streams through alluvial deposits confined in rock troughs. These tests have produced data on the velocity of ground water through materials of certain kinds and under known hydraulic gradients. The data are perhaps of the greatest value in furnishing a general comparative basis for making rough estimates of velocity of ground water in other places and under other conditions. Their value in the specific investigations for which they were made has been largely in giving negative information, for they have generally shown that the rate of under-

[90] Slichter, C. S., Field measurements of the rate of movement of underground waters: U. S. Geol. Survey Water-Supply Paper 140, pp. 29–49, 1905.

flow is small in comparison with the recharge that occurs all along the valley by downward percolation of rain or stream water. At the present time methods to determine the rate of flow are most needed in connection with water-bearing formations which lie beneath confining beds and in which the water must travel long distances from the intake areas to the areas of recovery. One of the principal difficulties in using velocity methods on such formations is the cost of sinking the deep test wells that would generally be required.

METHODS BASED ON LABORATORY DETERMINATIONS OF PERMEABILITY

About 1843 Poiseuille,[91] in connection with his studies of the circulation of the blood, discovered the law of flow through capillary tubes—namely, that the rate of flow is proportional to the hydraulic gradient. In 1856 Darcy,[92] inspector general of the Paris waterworks, whose extensive and valuable experiments on the flow of water through pipes are well known, verified this law and demonstrated its application to water percolating through the capillary interstices of sand or other porous medium. He expressed this law by means of the formula $v=\frac{kp}{h}$, in which v is velocity of the water through a column of permeable material, p the difference in head at the two ends of the column, h the length of the column, and k a constant that depends on the character of the material, especially on the size of the grains, which was to be experimentally determined.

In the 75 years since the results of Darcy's work were published laboratory investigations of various phases of the problem of the flow of liquids and gases through permeable materials have been made, many of them by French and German physicists and engineers.[93] A critical review of the early investigations was made by King,[94] an American physicist and hydrologist. He reviewed the laboratory studies on the flow of water through permeable materials that were made by the European investigators Hagen, Seelheim, Welitschkowsky, and Wollny and by the American engineer J. C. Trautwine; the work of Thomas Graham and O. E. Meyer, who demonstrated that the law of Poiseuille holds for gases as well as for liquids; and the later investigations by Fleck, Welitschkowsky, Renk,

[91] Poiseuille, J., Recherches expérimentales sur le mouvement des liquides dans les tubes de très petit diamètre: Mém. savants étrang., vol. 9, p. 433, 1846.

[92] Darcy, H. P. G., Les fontaines publiques de la ville de Dijon, Paris, 1856. See Slichter, C. S., The motions of underground waters: U. S. Geol. Survey Water-Supply Paper 67, pp. 18, 19, 1902.

[93] See bibliography in Slichter, C. S., Theoretical investigation of the motion of ground waters: U. S. Geol. Survey Nineteenth Ann. Rept., pt. 2, pp. 380–384, 1899.

[94] King, F. H., Principles and conditions of the movements of ground water: U. S. Geol. Survey Nineteenth Ann. Rept., pt. 2, pp. 178–189, 1899.

Ammon, and Wollny concerning the flow of gases. The whole subject of the law of flow of liquids and gases through capillary tubes and permeable materials was reinvestigated by King, and his results are given in the paper cited. During the years 1902 to 1904 laboratory experiments on the flow of water through sand and gravel were made under the direction of Slichter [95] with lower hydraulic gradients than had previously been used, in order to approximate natural conditions more closely. Since that time permeability experiments with both liquids and gases have been made by other American physicists and engineers in various fields of investigation. In 1923 a hydrologic laboratory was established in the Geological Survey, and tests of permeability to water were made under hydraulic gradients as low as 5 feet to the mile.[96] In this laboratory standard permeability tests under low gradients are now made of many water-bearing materials found in the areas that are investigated by the Geological Survey.

The foregoing historical review shows that Darcy's law, that the flow of water through a permeable material varies directly as the hydraulic gradient, has been thoroughly tested by experiment and has been demonstrated to be correct for practical purposes if it is not applied with excessive extrapolation. Hence, the flow through a formation can be computed if its permeability, the hydraulic gradient of the water it contains, and the area of its cross section are determined.

The hydraulic gradient can be rather satisfactorily determined by measuring the depth to the water level in wells properly distributed over the area underlain by the formation. A well should not be yielding water at the time it is measured or for some time prior to its measurement; otherwise there will be a local drawdown that will introduce an error. There may also be a drawdown caused by heavy flow or pumping from other wells in the same vicinity, and precautions must be taken with respect to errors that might be produced by changes in atmospheric pressure or other causes. The water levels are measured from definite bench marks, or reference points, to which levels are run. A contour map of the pressure-indicating surface is then constructed. Such a map shows approximately the direction of movement of the ground water and the hydraulic gradient at every point, the direction in any locality being at right angles to the contour and the gradient being the ratio of the contour interval to the distance between successive contours.

The area of cross section can be estimated from well records and outcrops that show the thickness and extent of the formation. A

[95] Slichter, C. S., Field measurements of the rate of movement of underground waters: U. S. Geol. Survey Water-Supply Paper 140, pp. 29–49, 1905.

[96] Stearns, N. D., Laboratory tests on physical properties of water-bearing materials: U. S. Geol. Survey Water-Supply Paper 596, pp. 144–163, 1927.

cross section that is pertinent for this purpose must obviously have the same general trend as the contours of the pressure-indicating surface, in order to be approximately at right angles to the direction of flow.

The most serious difficulty with this method and one that up to the present time has not been effectively overcome is that of determining the true average permeability of the material that constitutes the water-bearing formation. Laboratory methods are available to determine accurately the permeability of the samples that are tested, but the difficulty lies in obtaining representative samples. Even apparently slight differences in texture may make great differences in permeability. A rather inconspicuous admixture of colloidal clay to an otherwise permeable sand may cut down greatly its capacity to conduct water. In a sand formation a few thin strata of coarse, clean sand may conduct more water than all the rest of the formation. These permeable strata may be overlooked in the sampling, or if samples from them are taken it may be impossible to give them the proper weight in comparison with samples from other parts of the formation. Consolidated rocks are likely to contain joints and crevices which conduct much of the water and which therefore render laboratory methods inapplicable. On the other hand, unconsolidated samples can not easily be recovered and tested without disturbing the texture of the material and thus introducing errors of unknown but conceivably great amount. Moreover, samples taken at the outcrop of a formation may not be representative because of changes produced by weathering, and samples obtained from wells are generally nonvolumetric and greatly disturbed and may be either washed or mixed with clay of foreign origin. If the conditions of drilling can be controlled it may be possible to obtain an undisturbed or only moderately disturbed sample, especially if a core barrel is used, but such favorable conditions are rarely obtainable. In spite of the difficulties, the laboratory tests are of value in giving general limits of permeability, and it will be highly desirable to complete the project, now in progress, of making a set of permeability tests of all the leading formations in this country that conduct water through their pore spaces.

Efforts have been made by Hazen, King, Slichter, and others[97] to compute the permeability of water-bearing material from its mechanical composition and porosity. For this purpose Hazen,[98] in his

[97] For a review of the work of Allen Hazen, F. H. King, and C. S. Slichter on this subject, with references, see Stearns, N. D., op. cit., pp. 170–176.

[98] Hazen, Allen, Experiments upon the purification of sewage and water at the Lawrence Experiment Station, Nov. 1, 1889, to Dec. 31, 1891: Massachusetts Board of Health Twenty-third Ann. Rept., p. 431, 1892; Some physical properties of sands and gravels with special reference to their use in filtration: Massachusetts Board of Health Twenty-fourth Ann. Rept., p. 541, 1893.

work on filter sands in 1889 to 1893, used the term "effective size of grain," or size of grain that would give the actual permeability of a more or less heterogenous material. He found that in the materials with which he was dealing the effective size was best shown by the "10 per cent size"—that is, the size that is not exceeded by the grains in 10 per cent of the material by weight. These indirect methods of computing permeability are useful for some purposes but have not always given consistent results and are not in general to be recommended. Direct tests of permeability require no more work and are generally more satisfactory.

METHOD BASED ON PERMEABILITY DETERMINED FROM DISCHARGE AND DRAWDOWN OF WELLS

One of the most promising methods of determining permeability is a field method proposed by G. Thiem,[99] son of the German hydrologist A. Thiem, based on the performance of wells that enter the water-bearing formation. In order to make a test by this method it is necessary to have one well from which water can be withdrawn, either by pumping or by artesian flow, at a rate that will produce considerable drawdown, and two or more other wells that are located at different distances from the discharge well and near enough to it to have appreciable drawdown when it is discharging. This method has the advantage over the laboratory methods in that it deals with all the water-bearing materials in the vicinity of a well, undisturbed and in place.

The method is very simple in principle. The water level is measured in each of the observation wells at a time when the discharge well has been idle long enough for the water levels to have reached the static level. The well is then pumped or allowed to discharge by artesian pressure at a uniform rate until equilibrium has again been virtually established and the new water levels have become essentially stationary. If the test is made in an area of artesian flow, the pressure may be measured instead of the depth to the water level. From the ultimate drawdown in the several observation wells produced by the pumping or artesian discharge a profile of the cone of depression can be constructed, and from this profile the hydraulic gradient at any distance from the discharge well can be determined. If the water-bearing formation is of uniform character and thickness in the vicinity of the wells and the thickness is known, all three factors are available for computing the permeability. If the water in the formation is confined under pressure, the cross-section area is obviously $2\pi rt$, in which r is the distance from the discharge well to the point where the hydraulic

[99] Thiem, G., Hydrologische Methoden, Leipzig, 1906.

gradient is determined and t is the thickness of the formation. If the water is not confined but forms a water table, the effective cross-section area is $2\pi rt'$, in which t' is the thickness of the saturated part of the formation or the height of the water table above the base of the formation during pumping at the point where the hydraulic gradient is determined. In either case the coefficient of permeability as defined in Water-Supply Paper 596 is equal to $\frac{q}{ag}$, in which q represents the rate of pumping or artesian discharge in gallons a day, a the cross-section area in square feet, and g the hydraulic gradient as defined on page 132. Further investigation is needed as to inaccuracies that may be introduced by irregularities in the texture, thickness, and stratigraphy of the formation, by the natural slope or irregularity of the pressure-indicating surface, and by decrease in permeability that may result from compression of the formation when the artesian pressure of the water is relieved. The permeability determined by this method is used in exactly the same manner as the permeability determined by any other method in computing the rate of flow through a formation when its cross-section area and the natural hydraulic gradient of its water are known.

In 1906 G. Thiem published as a dissertation for the degree of doctor-engineer from the Konigliche Technische Hochschule at Stuttgart the results of his experiments and mathematical study relating to his field method for determining permeability. The experiments were made as a part of an investigation to find an additional water supply for the city of Prague. Ten sets of wells were sunk for the purpose, each set including one well that was pumped and two observation wells. The observation wells were placed in line with the pumped well but in any direction from it regardless of the direction of the natural hydraulic gradient. A formula was developed for computing the permeability from these tests. This formula, slightly modified in form, is as follows:

$$P=\frac{(q \log_e a_1-\log_e a)}{2\pi m (S - S_1)}$$

in which P is the coefficient of permeability, q is the rate of discharge, a and a_1 are the respective distances of the two observation wells from the discharge well, S and S_1 are the respective drawdowns in the two observation wells, and m is the thickness of the water-bearing bed, if artesian conditions exist, or the average thickness of the saturated part of the bed at the two observation wells, if water-table conditions exist. To obtain the coefficient of permeability as defined in Water-Supply Paper 596 all these factors should be expressed in feet except q, which should be expressed in gallons a day.

Theoretically this method can also be used for computing specific yield, but no test has been made of its practicability for this purpose. For example, during a specified period in the initial stage of a test under water-table conditions an accurate record can be obtained of total pumpage and of the progressive lowering of the water levels in the observation wells. From the record of water levels and the permeability, as determined by the test, the flow through the selected cylindrical cross-section area during the period can be computed. The quantity taken from storage within the cylinder can then be computed as the total pumpage minus the inflow, and from the records of the water levels the volume of material in the cylinder that was unwatered during the period can also be computed. Better results can doubtless be obtained, however, if the specific-yield test is made during a period immediately after the well has been shut down, when the quantity of water taken into storage in the cylinder will be equal to the total inflow during the period and the volume of material saturated can be computed from the rise of the water levels.

Closely akin to the specific yield, but generally of much smaller magnitude, is the change in storage capacity of an artesian formation with changes in the artesian pressure on account of the volume elasticity of the formation. Obviously the increase in storage in an artesian formation can be computed from progressive changes in artesian pressure immediately after the discharge well has been shut down, by the procedure that has just been outlined for computing specific yield under water-table conditions. Rough computations of this kind have been made.[1]

METHOD BASED ON AREA OF INFLUENCE OF WELLS

Ground water obeys the law of all fluids in that it always flows away from a point of high pressure toward one of lower pressure. In other words, it flows in the direction of the hydraulic gradient. In a formation that has a water table the ground water flows from a high area of the water table toward a low area, much as surface water flows down the slope of a land surface. In an artesian formation the water flows in the direction of the hydraulic gradient; it may flow downward or upward, as in the mains and service pipes of a system of waterworks. In some places the artesian water moves in the direction of the dip of the formation, and in other places it moves in a direction opposite to the dip or at some angle to the dip. An accurate contour map of a water table or of the pressure-indicating surface of a formation filled with water under pressure shows the direction of flow at every point.

[1] Meinzer, O. E., Compressibility and elasticity of artesian aquifers: Econ. Geology, vol. 23, pp. 263–291, 1928.

So long as no wells have been sunk into a formation its water flows from the intake area toward the area of natural discharge. If, however, wells are sunk and water is withdrawn through them, either by pumping or by artesian flow, the water table or other pressure-indicating surface will be depressed in the vicinity of the wells, and water that would normally percolate to an area of natural discharge will be diverted toward the wells. An accurate contour map showing the depression in the pressure-indicating surface caused by the operation of the wells will show the area within which the water is diverted toward the wells.

So long as the withdrawals are relatively light the depression will remain small, and water which in its natural course does not flow near the wells will not be diverted. As the withdrawals become heavier, however, the depression is enlarged and a greater proportion of the ground water is diverted toward the wells, until eventually the depression may extend across the entire width of the formation, and practically all of the flow may be recovered through the wells.

The flow of ground water through a formation can be estimated from a record of the pumpage or artesian flow and the extent of the depression as determined from water levels in wells, due allowance being made for the quantity of water that is taken out of storage in developing the depression. This method has the advantage that it does not require determination of the permeability and involves specific yield only in so far as there is a change in storage. Except for changes in storage it involves only the rate of pumping or artesian flow and the altitude of the water levels in the observation wells. The method should prove applicable where a considerable quantity of water is withdrawn and sufficient wells are available for water-level observations.

In applying this method attention must be given to the modified condition of the pressure-indicating surface on the downstream side of the depression produced by the operation of the wells. If no water is permitted to pass the wells this surface will eventually be lowered about to the level of the natural outlet of the ground water or to the level of the crest of an effective underground barrier. In many places this drawdown would make the cost of pumping prohibitive. In order to prevent the lowering of the water levels in the wells below the economic limit it may be necessary to allow a part—perhaps a large part—of the water to flow past the depression, so that it will maintain a certain minimum gradient to the natural outlet.

METHOD BASED ON MOVEMENTS OF WATER LEVELS IN RELATION TO RATES OF WITHDRAWAL

When a well is pumped some water is inevitably taken out of storage from the well and from the material surrounding it. This reduces the pressure, creates a hydraulic gradient toward the well, and causes the ground water to flow into the well. If the water-bearing formation has a water table, considerable ground water may have to be removed from storage before a gradient will be developed that is steep enough to make the water flow toward the well at the same rate that it is pumped and to establish approximate equilibrium. If the formation is filled with water under pressure only a comparatively small amount of water has to be removed from storage in order to give the required gradient, and hence drawdown will be rapid and approximate equilibrium will be quickly established. The quantity to be removed from storage in order to reach equilibrium varies in proportion to the compressibility of the water-bearing formation.[2] When, with a constant rate of pumping, equilibrium is established, water is no longer removed from storage around the well but flows to the well as rapidly as it is withdrawn. Even after there is approximate equilibrium and the water level in the well remains virtually stationary the water may be taken from storage in distant parts of the formation rather than from the increments at its intake area. However, unless the withdrawal exceeds the recharge minus the unavoidable loss, the water level in the well will approach a limit at a constantly decreasing rate of drawdown. In most ground-water developments the supply available is conditioned by the permissible lift. A vital question, therefore, relates to the rate at which water will flow to the well under the gradient that will be established by the drawdown involved in the permissible lift.

This elementary principle of hydraulics controls not only the practicable capacity of an individual well but also the quantity of water that can be recovered by a group of wells or by a large number of wells distributed throughout a wide area. The details are more complicated, but the principle is the same. If the ground-water supply in an area is extensively developed by drilling many wells and drawing heavily on them the water levels or artesian pressures in the wells will inevitably decline. This decline will occur whether the conditions are primarily of the water-table or the artesian type, and whether the artesian wells discharge by artesian flow or by pumping. Other things being equal, however, the decline will occur more rapidly under artesian conditions than under water-table conditions. The mere fact that there is a decline in the water levels or artesian

[2] Meinzer, O. E., Compressibility and elasticity of artesian aquifers: Econ. Geology, vol. 23, pp. 263–291, 1928.

pressures during the period of development, when the rate of withdrawal is constantly increasing, is not an indication of overdevelopment as long as the decline is not so great as to indicate the approach of lifts beyond the economic limit.

If the rate of pumping becomes constant the water levels in the wells will not immediately become stationary, but they will decline at a diminishing rate. If the rate of pumping exceeds the salvageable recharge the water levels will eventually decline to the bottoms of the wells, and the rate of recovery will unavoidably be decreased. If, however, the rate of pumping does not exceed the salvageable recharge, the water levels will approach as a limit some level above the bottom of the wells, though perhaps too low for practicable operation. To determine the limit even approximately requires the lapse of sufficient time to permit the development of a curve in which the water levels in a key well or group of wells are plotted against time and which can with some confidence be projected into the future. The problem will invariably be complicated by irregularities in the rate and location of the pumping, adequate records of which can not easily be obtained, and also by fluctuations in water levels produced by causes other than pumping.

In most areas there are fluctuations in the rate of pumping or of artesian flow, as the case may be. In many areas the most pronounced fluctuations are seasonal. On the whole these fluctuations increase the applicability of this method. It has been shown that some decline in water levels must be expected even after the rate of pumping has become constant. Furthermore, decline may even continue for some time after the rate of pumping has been reduced without indicating a dangerous amount of ultimate drawdown. However, if with either constant or reduced rate of pumping the water levels continue to decline persistently and without marked slowing up, excessive ultimate drawdown is indicated, although it may not be possible to determine the ultimate drawdown or to estimate the amount of overdevelopment. In the development of many of the areas of artesian flow there is an early period of original high pressure and active drilling during which the artesian discharge increases greatly and the pressure drops rapidly. This early period is generally followed by a long period in which there is progressive but gradual decline both in discharge and in pressure. It appears obvious that under these conditions water is being withdrawn from storage and that equilibrium will not be established until the pressure or the discharge or both are still further reduced.[3]

[3] Meinzer, O. E., and Hard, H. A., The artesian water supply of the Dakota sandstone in North Dakota, with special reference to the Edgeley quadrangle: U. S. Geol. Survey Water-Supply Paper 520, pp. 73–95, 1925. Hard, H. A., Geology and water resources of the Edgeley and LaMoure quadrangles, N. Dak.: U. S. Geol. Survey Bull. 801, pp. 57–87, 1929.

If the rate of pumping or of artesian discharge is radically reduced, as it is in the autumn in irrigation districts and in some public waterworks, the water levels or artesian pressures that have been declining are likely to reverse their direction and begin to rise. This means that the point of equilibrium has been reached and passed—that, with the gradient which was established by the heavier withdrawal, water is flowing into the area more rapidly than it is being removed by the reduced rate of withdrawal. In the next spring or summer the rate of withdrawal may be again increased and another reversal of the water levels may occur. If the rate of withdrawal is greater than in the previous summer a new low stage may be established. By critical study of this game of tag between the water levels and the rates of withdrawal a number of points of equilibrium can be approximately determined, but exact determination is complicated by the lag in the movements of the water levels due to storage, and further study needs to be given to this subject to develop a systematic method of interpretation that is mathematically sound. These points of equilibrium are points on a rating curve of the inflowing stream of ground water. To the extent that this curve can be projected it will indicate the drawdown that will be caused by increased rates of withdrawal and will thus establish the safe yield. The rating-curve method for conduit conditions was developed by Thompson in his work in New Jersey from 1923 to 1928 and was successfully applied by him in making estimates of safe yield for Atlantic City and other municipalities.[4]

EVALUATION OF EXTRANEOUS INFLUENCES ON THE WATER LEVELS

The movements of the water levels in wells are not the effects of simple forces acting singly but rather the resultants of a complex of interacting forces. This complexity, of course, increases the difficulties of interpretation. For example, a study of the effects produced by different rates of pumping may be complicated through changes produced by variations in atmospheric pressure or by the ebbing and flowing of oceanic tides. The barometric or tidal effects must be evaluated, and corrections must then be made for them. These extraneous agencies are, however, not wholly a detriment, for if they are studied in a time and place where other agencies are absent or can be eliminated, they are likely to afford valuable information on the ground-water conditions.

If a shallow well ends in a formation that lies at the surface and is freely permeable throughout, only slight barometric effects or

[4] Thompson, D. G., Ground-water supplies in the Atlantic City region: New Jersey Dept. Conservation and Development Bull. 30, pp. 35–88, 1928.

none at all are to be expected, because any change in the atmospheric pressure is transmitted almost as freely to the water table through the permeable material as to the water level in the well. In the Escalante Valley investigation, by White, water-stage recorders were used on about 70 shallow water-table wells. With few exceptions, the records of these wells showed no evidence of fluctuations due to changes in atmospheric pressure. The exceptions occurred chiefly in the early spring, when aparently the soil was wet enough to be impervious to air. Similarly, in a shallow observation well near Washington, D. C., barometric fluctuations, amounting to about 20 per cent of the full fluctuations of a water barometer, are apparently confined to periods when the water in the soil tends to reduce notably its permeability to the air. Barometric fluctuations have been found to be virtually absent in observation wells equipped with automatic recorders in different parts of the country where water-table conditions exist.

If a well ends in an artesian formation and this formation or the overlying confining beds have sufficient strength to resist deformation by slight changes in pressure at the surface, the well will act as a barometer. The fluctuations of its water level will have virtually the same range of fluctuations as would be shown by a water barometer, or 13.5 times the range in a mercury barometer. However, for obvious reasons, the movements of the water level in the well will always be in the opposite direction from those in an ordinary mercury barometer. In a deep well at The Dalles, Oreg., that ends in an artesian stratum in volcanic rocks, the water level was found by Piper[5] to respond practically 100 per cent to changes in atmospheric pressure as indicated by the barographic records of the United States Weather Bureau. The record of a deep well in the Roswell artesian basin, New Mexico, obtained by means of a water-stage recorder, showed that on certain days when no great disturbances were caused by pumping the fluctuations of the water level in the well closely followed the fluctuations of the barometer. In general an inch of change in the mercury barometer was represented by about a foot of fluctuation in the water level.[6] In another artesian well situated near the railroad in the same basin no rise in the water level was recorded when trains passed. Both of these results indicated artesian conditions in competent formations. If a similar well were situated near the seashore its water level would not be expected to show any tidal influence.

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[5] Piper, A. M., Geology and ground-water resources of the Dalles region, Oregon: U. S. Dept. Interior Press Mem. 52343, Apr. 7, 1931.

[6] Fiedler, A. G., Report on investigations of the Roswell artesian basin, New Mexico, during the year ending June 30, 1926: New Mexico State Engineer Seventh Bienn. Rept., p. 37, pl. 7, 1926.

If a well ends in an artesian formation that has volume elasticity, such as incoherent sand, and is confined beneath beds of soft shale that is impermeable but yields to even slight pressure, its water level will have smaller fluctuations resulting from atmospheric changes than that of a water barometer, and the ratio of the movements in the well to the corresponding movements in the barometer should give a measure of the resistance of the water-bearing and confining beds. If such a well is near the seashore its water level will fluctuate more or less with the tide. In the Atlantic City investigation, by Thompson, it was found that artesian wells near the seashore ending in Coastal Plain deposits of sand overlain by clay fluctuated notably with the tide but showed no marked barometric fluctuations. These results seem to indicate artesian conditions in incompetent strata. In such wells the water level might be expected to rise as a result of the compression produced by a passing railroad train. In the Mokelumne area it was found that the water level in a well drilled through valley fill to a stratum at some depth below the water table rose whenever a train passed over the railroad 117 feet away, but in a well drilled only to the water table at the same place the water level was not affected by passing trains.[7]

If a well ends in a formation that is effectively covered by an impermeable bed but is unsaturated in its upper part, thus having an air chamber between the water table and the overlying impermeable bed, and if the well is tightly cased to a level below the water table, the water level in the well will behave like that in an artesian well in competent beds. It will fluctuate through about the same range as the water level in a water barometer, because, as in the other case, the counter pressure will remain nearly constant. These conditions were noted by Thompson[8] in the Grand Prairie region of Arkansas, where some of the wells were found to perform as perfect barometers, apparently with 100 per cent of barometric fluctuation. These are examples of actual conditions which have been encountered in quantitative ground-water investigations and which must be understood if the data on water levels are to be correctly interpreted.

PRESENT STATUS AND OUTLOOK

The foregoing discussion shows that there are many ways in which the problem of available supplies of ground water can be attacked, but it also shows that the subject is complicated and that

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[7] Stearns, H. T., Robinson, T. W., and Taylor, G. H., Geology and water resources of the Mokelumne area, California: U. S. Geol. Survey Water-Supply Paper 619, pp. 148, 149, 1930.

[8] Thompson, D. G., Ground-water supplies for rice irrigation in the Grand Prairie region, Arkansas: U. S. Dept. Interior Press Mem. 49844, pp. 8, 9, Jan. 26, 1931.

there are difficulties in applying any of the methods that have been outlined. Only those who have worked on ground-water problems can appreciate how numerous and baffling are these difficulties. No one method is applicable to all conditions. In some areas several more or less independent methods can be used which serve to give checks on the accuracy of the work. In other areas only a single method may be applicable, or it may be almost impossible to make effective use of any of the known methods.

Other methods must be used in humid than in arid regions, in areas of large ground-water developments than in undeveloped areas, and in artesian areas than in areas with water-table conditions. Arid regions are, as a rule, better adapted for quantitative studies than humid regions, and many of the available methods have been developed in arid regions. More methods are available in developed than in undeveloped areas, and the demand for quantitative work commonly comes after there has been enough development to cause considerable drawdown. A number of the methods, however, do not depend at all on development, and considerable successful quantitative work has been done in areas that were virtually undeveloped. In the past most of the quantitative work has been done in areas in which the conditions are dominantly of the water-table or reservoir type, but recently quantitative attacks have been made on a number of artesian basins in which the capacity of the formation as a conduit is important, and methods are being developed for these conditions.

The most formidable difficulties result from the complexity of the texture of the water-bearing formations, which make it hard to get reliable figures for their two properties of chief significance—specific yield and permeability. Some of the most promising methods are those which do not involve either of these properties or else determine them from water levels in wells or from water levels and pumpage.

It is becoming evident that the chief instrument of precision in ground-water hydrology is the water-stage recorder installed over a well. The water levels in wells are sensitive to every change that takes place in the ground water, and they can be almost perfectly recorded by a high-grade automatic recorder. In the interpretation of such records lie possibilities of a new and fruitful phase of ground-water hydrology that are not now fully appreciated and that can be realized only as a result of much critical investigation. It will be necessary to understand the various agencies that influence the delicate equilibrium that exists in every body of ground water and also the nature of the container that holds the ground water and how it can be expected to respond to these agencies. It will be necessary to recognize clearly the basic distinctions between a water table and the pressure-indicating surface of an artesian formation, and between changes that represent increments to or withdrawals from storage and those that represent only variations in pressure.

It is also becoming evident that time is required to obtain reliable results in most quantitative investigations of ground water. In this respect ground-water work differs from most other geologic work. It does not deal primarily with features which are the fossilized product of events that occurred in past ages and which can be observed and studied at any convenient time, but it deals rather with forces that are now operating and producing changes which can be kept under observation. Past changes that were not recorded are gone forever; future changes can be observed only with the lapse of time. For this reason systematic observations should be made and records kept in every area in which large ground-water developments have been made or are contemplated, so that reliable data will accumulate and will be available for interpretation in the future.

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