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CHARLES D. WALCOTT, DIRECTOR

THE MOTIONS OF UNDERGROUND WATERS

BY

CHARLES S. SLICHTER

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DEPARTMENT OF THE INTERIOR,
UNITED STATES GEOLOGICAL SURVEY,
DIVISION OF HYDROGRAPHY,
Washington, D. C., March 26, 1902.

SIR: I have the honor to transmit herewith a manuscript by Prof. Charles S. Slichter on the motions of underground waters, and request that it be printed in the series of Water-Supply and Irrigation Papers. Professor Slichter has been giving considerable time and attention to experimental and theoretical considerations of the movements of underground waters, the preliminary results of which were published in Part II of the Nineteenth Annual Report. The present paper treats of the simpler and more general topics connected with the movements of water underground, being intended to answer the more elementary questions which arise in a consideration of the subject. Examples are given of the various areas in which water occurs underground, the origin and extent of the waters are discussed, and methods of bringing them to the surface and making them available are touched upon.

Very respectfully,

F. H. NEWELL,
Hydrographer in Charge.

Hon. CHARLES D. WALCOTT,
Director United States Geological Survey.

THE MOTIONS OF UNDERGROUND WATERS.

By Charles S. Slichter.

CHAPTER I.

ORIGIN AND EXTENT OF UNDERGROUND WATERS.

All underground waters have their origin in rainfall. A part of the rainfall immediately runs off the surface of the ground into the streams and rivers, just as the rain is diverted by the roof of a house into the gutters and spouts. Another portion of it is absorbed temporarily by the surface soil, but is again returned to the atmosphere by evaporation, either directly or through the agency of vegetation. A third portion penetrates the lower levels of the soil, there to become a part of the great mass of underground water and help to furnish the perennial supply of streams and lakes, of springs and wells, and to take part in geologic work of the most profound importance.

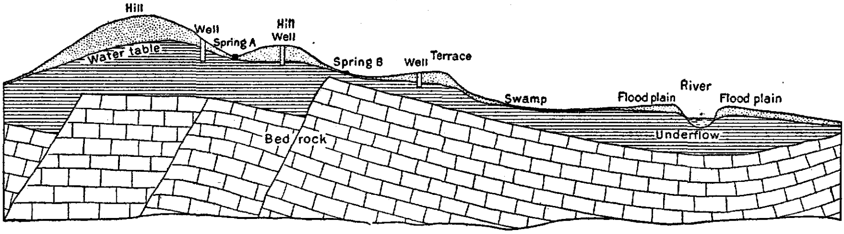


Fig. 1.—Ideal section across a river valley, showing the position of the ground water and the undulations of the water table with reference to the surface of the ground and bed rock.

In its downward course through the soil the rain water soon reaches a level at which the soil is completely saturated. The surface of this saturated zone is known as the water table or water plane. It is the water existing in the soil or rocks below this water table which is usually included in the term underground water. Above the zone of saturation the soil usually contains a large percentage of moisture, which plays a most important part in the growth of plants and in the physical, chemical, and biological phenomena of the soil. The consideration of the phenomena in this nonsaturated portion of the soil does not, however, come within the scope of this paper.

The depth of the water table below the surface of the ground varies much in different localities. In regions of copious rainfall it is usually but a few feet below the surface. In arid regions its depth may be measured in hundreds of feet. In general the water table follows contours very similar to those of the land surface, but its undulations and slopes are of much less magnitude than those of the surface of the ground. This is shown in the ideal section forming fig. 1. A region in which the water table lies at great depth below the surface of the ground is, of course, a region of deep wells. A locality where the water table coincides with the surface of the ground is a swamp or marsh.

DEPTH OF GROUND WATER.

The lowest theoretical limit at which ground waters can exist is reached when the pressure in the rocks, due to the weight of the superincumbent material, is so enormous that all cavities and pores in the rock are completely closed. This limit has been shown by Professor Van Hise to be at a depth of approximately 6 miles.[a] The region above this depth Van Hise has distinguished as the zone of fracture, for in it pressures and stresses result in the actual breaking and fracturing of the rock. The region below this depth has been called by Van Hise the zone of flowage, for in it the enormous pressures will not permit the formation of cracks and cavities, and the rocks, when transformed by stresses, must actually flow like clay under the pressure of the hand. Adopting, then, this fundamental classification of Van Hise, we may say that the lower limit of the existence of ground water is found when we reach the lower boundary of the zone of fracture. This limit, however, refers primarily to the geologic work of ground waters and not to the practical limit of its occurrence in quantities sufficient for economic uses, for, as will be seen later, the principal zones from which underground water is actually recoverable in useful quantities are almost completely confined to the domain of sedimentary rocks and surface deposits. Even the mineral and thermal springs from beds of crystalline or metamorphic rocks are from a level much above the geologic limit of depth.

TOTAL AMOUNT OF UNDERGROUND WATER.

The amount of ground water within the crust of the earth is enormous. The writer estimates the entire amount to be about 565,000 million million cubic yards, or about 430,000 million million cubic meters. He has arrived at this result by considering that the geologic limit of the existence of ground water is at an average depth of 6 miles below the surface of the land and 5 miles below the floor of the ocean. The land surface and water surface he has assumed to be 52,000,000 square miles and 144,700,000 square miles, respectively. The average pore space of the surface rocks which is occupied by water or moisture he has taken as 10 per cent of their total volume. He believes that this estimate of 10 per cent is too large rather than too small. It forms, however, a convenient basis for the estimates.

[a] Principles of North American pre-Cambrian geology, by C. R. Van Hise: Sixteenth Ann. Rept U. S. Geol. Survey, Pt. I, 1896, p. 593.

According to these estimates, the total amount of underground water is sufficient to cover the entire surface of the earth to a uniform depth of from 3,000 to 3,500 feet. Assuming a mean depth of the ocean of 12,000 feet leads to the conclusion that the total amount of oceanic water is about 1,800,000 million million cubic yards, so that the total quantity of ground water is nearly one-third the amount of the oceanic water.

Former estimates of the quantity of underground water the writer believes to be entirely too large. The most frequently quoted estimate is that of Achille Delesse.[a] He estimates that the average amount of water in the surface rocks is 5 per cent by weight or 12½ per cent by volume. He supposes that the increase of temperature in the interior of the earth (1° C. for 100 feet) would limit the existence of water in the liquid form to a depth of 3,300 meters if it were not for the increase of pressure due to this great depth. Allowing for this, Delesse estimates that liquid water may exist at a depth of 18,500 meters and at a temperature of 600° C. These considerations lead him to the enormous estimate of 1,530,000 million million cubic yards (1,175,089 million million cubic meters). This, he says, is one nine hundred and twenty-first of the volume of the globe, which is nearly equal to Beaumont's estimate of the volume of the sea, one eight hundred and twenty-seventh of the the earth's volume.

The discovery of Van Hise that pores can not exist in the rocks at a depth much greater than 10,000 meters requires us to throw out these older estimates based upon the temperature gradient of the earth's crust.

PERMEABILITY OF ROCKS.

It has been tentatively assumed that all rocks are pervious to water. In a very general sense this may be said to be the case. Practically all rocks, even those of greatest strength, are not entirely solid masses of matter, but instead are collections of discrete particles and crystals more or less perfectly cemented and compacted together. This fact is much better understood when it is remembered that all except igneous rocks were at one time deposits laid down on the bottom of the sea by rivers and currents. These sediments may in time be changed to sedimentary rocks, and the sedimentary rocks in turn be changed to harder and more compact metamorphic and crystalline rocks, yet the original granular character of the mass is rarely completely lost, even in its changed condition, and the open texture persists, in some degree, forever. The small spaces or pores between the rock particles are usually occupied by water.

In building the St. Gothard tunnel the rocks encountered from the southern end were "principally mica-schist, hornblende rock and gneiss containing more hornblende and mica and quartzite," yet the

[a] Bull. Soc. géol. France, second series, Vol. XIX, 1861-62, p. 64.

contained water was so great that the grade of the tunnel had to be changed from 1:1,000 to 1:500 in order to obtain proper drainage.[a]

A portion of the underground water may be found in the larger cracks and crevices of the rocks, but by far the larger portion exists in the minute pores and openings between the rock particles themselves or in the small interstices between the separate grains of the soil.

POROSITY OF ROCKS AND SOILS.

The fractional part of a rock or of soil which is occupied by open spaces or voids determines its porosity. Thus, if a gallon of sand will hold, when saturated, three-tenths of a gallon of water, the porosity of the sand is said to be 30 per cent, three-tenths of its volume being made up of pores between the grains of sand. Likewise, if a cubic foot of sandstone will hold, when saturated, one-quarter of a cubic foot of water, the sandstone is said to have a porosity of 25 per cent, for one-fourth of its volume is pore space or voids.

The following table (I) contains several determinations of the porosities of Wisconsin building stones, made by Mr. E. R. Buckley, State geologist. They are probably the most carefully made determinations yet published. Special attention is called to the fact that all of the rocks listed in the table are building stones and that the important water-bearing rocks naturally show higher porosities than the more compact of these. From the table it is seen that the amount of open or unoccupied space in these building stones varies from about 1 part in 400 for the Montello granite to more than 1 part in 4 for the Dunnville sandstone. It is worthy of remark in this connection that the Montello granite was selected for the sarcophagus of the tomb of Gen. U. S. Grant, because comparative tests showed it to be the strongest granite in this country. Its porosity here given (about one-fourth of 1 per cent) shows that even the strongest rocks are measurably porous.

TABLE I.

Porosities of Wisconsin building stones, as determined by E. R. Buckley, State geologist.

Kind of stone.	Name of quarry.	Average porosity of two specimens.
		Per cent.
Granite	Berlin (Wis.) Granite Co	0.384
Do	Montello (Wis.) Granite Co	0.237
Niagara limestone	Marblehead (Wis.) Lime and Stone Co	0.77
Do	Story Bros., Wauwatosa, Wis	6.4
Lower Magnesian limestone.	Bridgeport, Wis	13.19
Sandstone	Chicago and Northwestern Railway., Ableman, Wis	5.6
Do	Bass Island (Wis.) and Lake Superior Sandstone Co., Ashland, Wis.	20.7
Do	Dunnville, Wis	28.26

[a] Tunneling, Explosive Compounds, and Rock Drills, by Henry S. Drinker, New York, 1878, p. 276.

The porosity of quartz sand will usually vary between 30 and 40 per cent, and that of clay loams between 40 and 50 per cent, depending upon the variety of sizes in the mixture and the manner of packing the particles.

MOTION OF THE WATER.

In general the water contained in the porous soils and rocks is not stationary, but possesses an exceedingly slow, although perfectly definite, motion. The best evidence of this is supplied by geology, which shows that nearly all of the rocks have been vastly changed by the work of underground waters. Enormous amounts of material have been slowly deposited in the pores of loose sandstones and limestones, until they have been converted into strong and nearly impervious rocks. Professor Van Hise estimates that thousands of cubic miles of dissolved quartz have been deposited by the slowly moving ground waters in sands and sandstones, until in some instances the latter have been converted into solid quartzite.[a] Almost any specimen of rock, especially if examined in thin sections under the microscope, will show some evidence of the work of moving ground waters. In many cases broken and worn fragments of crystals will be observed to have been patched and mended and added to until this so-called "secondary growth" has become a most important part of the rock. In other cases cementing material and minerals not originally present are found distributed throughout the rock in great abundance. Indeed the solution, transportation, and redepositing of material through the agency of ground waters is a geologic fact of profoundest importance, and is in evidence everywhere within the surface rocks of the earth.

CAUSE AND RATE OF MOVEMENT.

The cause of the motion of water through a porous medium is the same as the cause of the water movement through the pipes and conduits of a water-supply plant of a city—the difference of pressure from point to point. The difference in pressure in the case of ground waters is nearly always due to gravity alone. The water moves in the underground current for the same reason that water moves in the surface streams—it flows from a higher to a lower level.

The rate of the movement of water through a porous soil or rock depends upon several important elements, which may be enumerated as follows: (1) The size of the pores in the water-bearing medium, the capacity to transmit water being enormously greater for large pores than for small pores; (2) the porosity of the material, the flow being much greater for high porosity than for low porosity, other things being equal; (3) the pressure gradient, or the change in the pressure, or head, per unit of length measured in the direction of the motion,

[a] Earth movements; presidential address of C. R. Van Hise: Trans. Wisconsin Acad. Sci., Vol. XI, 1896, p. 511.

the flow being greater, of course, for high gradients than for low gradients; and (4) the temperature of the water, the flow being noticeably greater for high temperatures than for low temperatures.

THE LAWS OF FLOW.

Many experimenters have attempted to discover the exact law followed by water when traversing a porous medium, and to express that law by means of a mathematical formula. The earliest attempt to deduce such a law was made in 1856 by a Frenchman named Darcy,[a] who announced that the flow of water in a certain direction through a column of soil is proportional to the difference in pressure at the ends of the column and inversely proportional to the length of the column. According to this law, doubling the length of the soil

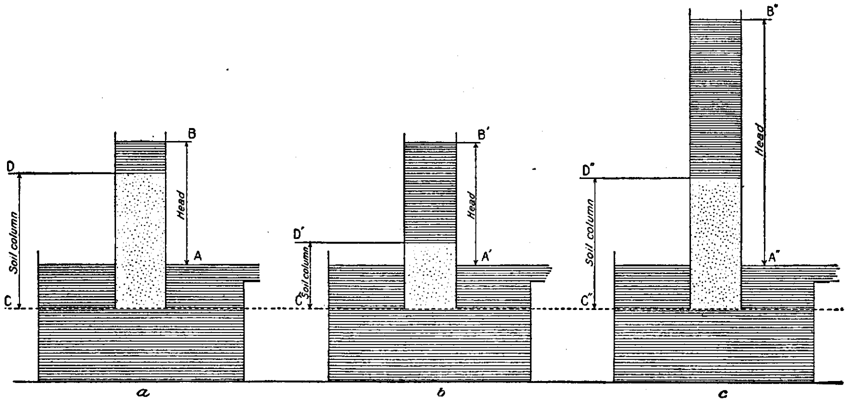


FIG. 2.—Diagram illustrating Darcy's law of the flow of water through a column of soil. The length of the soil column in case *a* is C D, and the difference in pressure at the ends of the column is A B. In case *b* the length of the soil column (C′ D′) is one-half of C D of case *a*, while the pressure difference (A′ B′) is the same as A B of case *a*. Therefore, according to Darcy's law, the flow in case *b* is twice that in case *a*. Likewise the flow in case *c* is twice that in case *a*, since the soil column is of the same length as in case *a*, while the pressure difference (A″ B″) is twice A B of case *a*. The amounts of the flows in cases *b* and *c* are equal to each other, for the length of the column of soil and the pressure difference are each double those of case *b*, which, according to Darcy's law, makes the flows equal.

column and keeping the water pressure the same at the ends will result in halving the flow, while doubling the pressure differences at the ends of the column and keeping the length of the column the same will result in doubling the flow. These facts are illustrated in fig. 2. Darcy expressed his conclusions by the following formula:

$$v = k\frac{p}{h}, \tag{1}$$

in which v stands for the velocity of the moving ground water, p the difference in pressure at the ends of the column of soil (measured usually by the height of the water column, as in fig. 2), h the length of the column, and k a constant depending upon the character of

[a] Les fontaines publiques de la ville de Dijon, by H. Darcy, Paris, 1856.

the soil, especially upon the size of the soil grains, which must be experimentally determined in each case by the use of apparatus similar to that shown in fig. 2.

Numerous other experimenters have investigated the same problem, with results substantially the same. Among the most important of these is the result of an American engineer, Mr. Allen Hazen,[a] whose conclusions are expressed by the following formula:

$$v = cd^2 \frac{h}{l}(0.70 + 0.03t), \qquad (2)$$

in which v is velocity of the water, in meters daily, in a solid column of the same area as that of the sand, c is a constant factor which

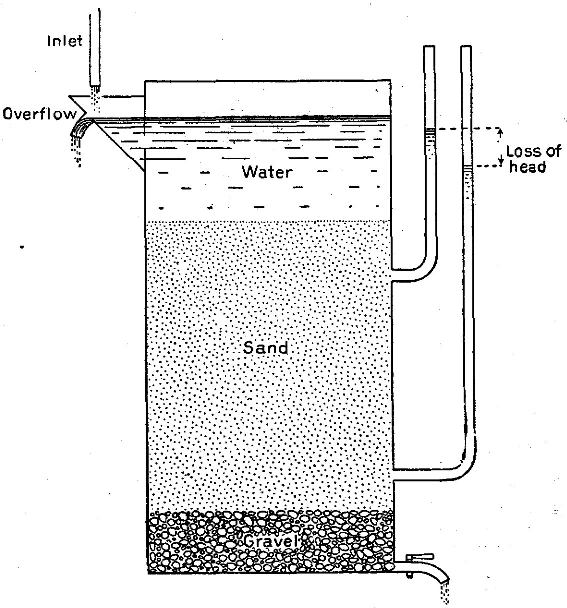


FIG. 3.—Hazen's experimental filter for the determination of the laws of flow of water through sand and the evaluation of the transmission constant of the same. The length of the soil column must be measured between the points at which the pressure gages enter the large cylinder filled with sand or other material subject to experimentation.

Hazen's experiments indicate to be approximately 1,000, d is the "effective size" of sand grain (measured in millimeters), which is such that 10 per cent of the material is of smaller grains and 90 per cent of larger grains than the size given, h is the loss of head, l is the thickness of sand through which water passes, and t is the temperature on the centigrade scale. Hazen states that the loss of head should be measured from points just inside the ends of the soil column, as is shown in the accompanying illustration of his apparatus, fig. 3.

[a] Some physical properties of sands and gravels, by Allen Hazen: Rept. Massachusetts State Board of Health, 1892, p. 541.

Hazen's work has been the basis on which the estimates of American engineers have been founded and has become standard.

Prof. F. H. King, in a long series of carefully conducted experiments, showed that the flow of water may increase somewhat faster than the pressure.[a] The formulas given herewith, however, are quite accurate enough for all engineering purposes to which they properly apply, although the refinement of a second approximation may result in a slight modification.

In another paper[b] the writer has given an expression for the flow ot water or other fluid through a column of soil made up of uncemented grains of nearly uniform size and of a well-rounded or approximately spherical form. It was there shown that a mass of spheres thrown together in a haphazard way will present great variety in the amount

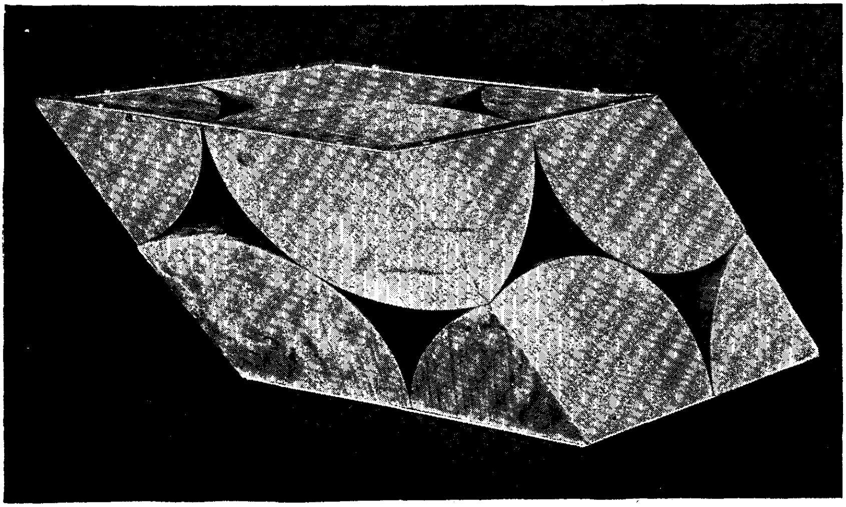


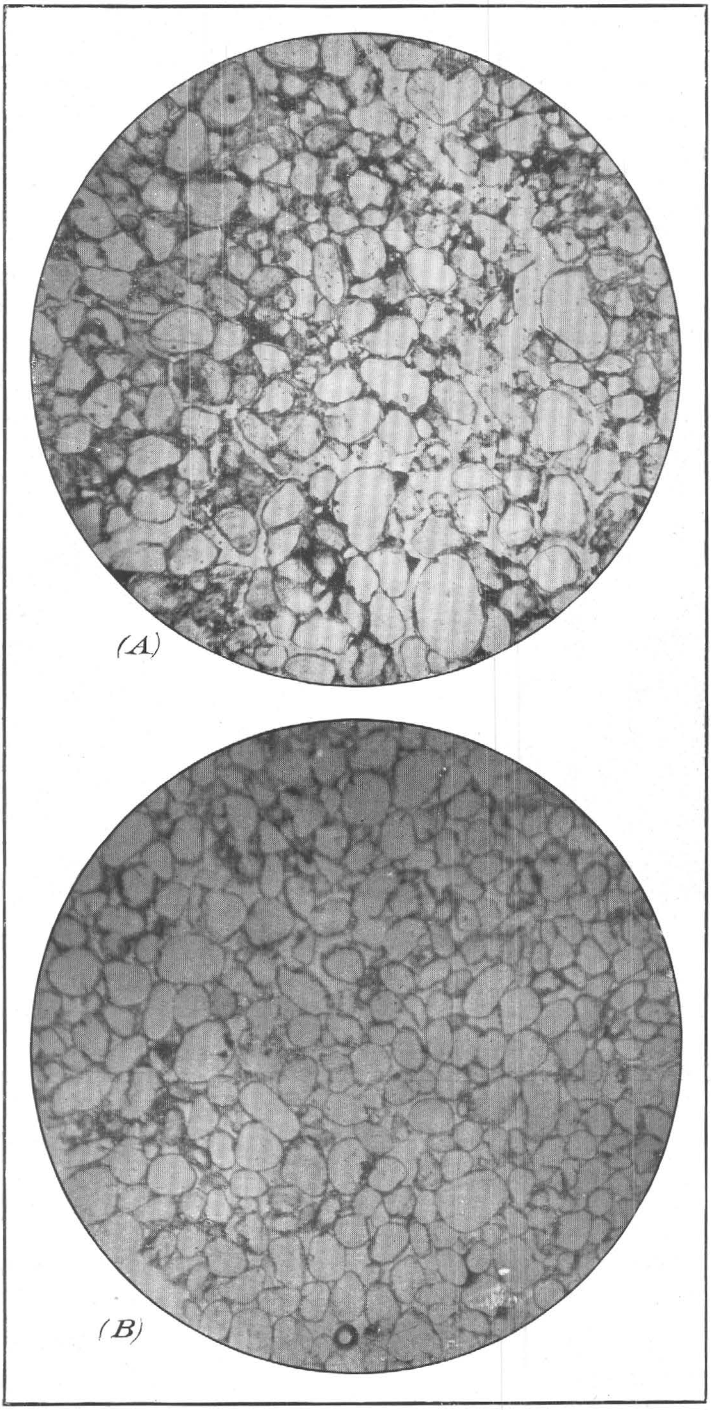
FIG. 4.—Unit rhombohedron formed by passing planes through the centers of eight contiguous spheres in the most compact packing of a mass of spheres, and showing the shape of the pores in such a mass.

of open space or porosity within the mass. The value of the porosity is shown to be independent of the size of the grains but dependent merely upon the manner of packing. The minimum porosity of a mass of spheres, or the porosity when the spheres are packed in the most compact manner possible, was shown to be 25.95 per cent of the whole space occupied by the spheres. The maximum porosity that is probable in a mass of spheres was shown to be 47.64 per cent of the whole space occupied.

If a quantity of common shot be poured into a glass and the amount of open space between the shot be determined by measuring the

[a] Principles and conditions of the movements of ground water, by F. H. King: Nineteenth Ann. Rept. U. S. Geol. Survey, Pt. II, 1899, p. 59.

[b] Theoretical investigation of the motion of ground waters, by C. S. Slichter: Nineteenth Ann. Rept. U. S. Geol. Survey, Pt. II, 1899, p. 295.



SECTIONS OF WISCONSIN SANDSTONE.

A, Red sandstone from Laville; *B*, brown sandstone from Argyle quantity of water required to fill the pores between the shot, it will be found that the resulting porosity will be well within the limits just given. It will also be found that the porosity will vary greatly if different methods be used in filling the glass, but that it is possible to obtain the same porosity with small shot as with large shot. The pores between the shot will be seen to be somewhat triangular in form, as shown in figs. 4 and 5, and they are both larger in diameter and shorter in length for a packing of spheres having a large porosity than they are for a packing of low porosity. This very important element in the transmission capacity of a soil the writer has attempted to take proper account of in the derivation of the formula given on page 24. The individual grains in a mass of sand are not, of course,

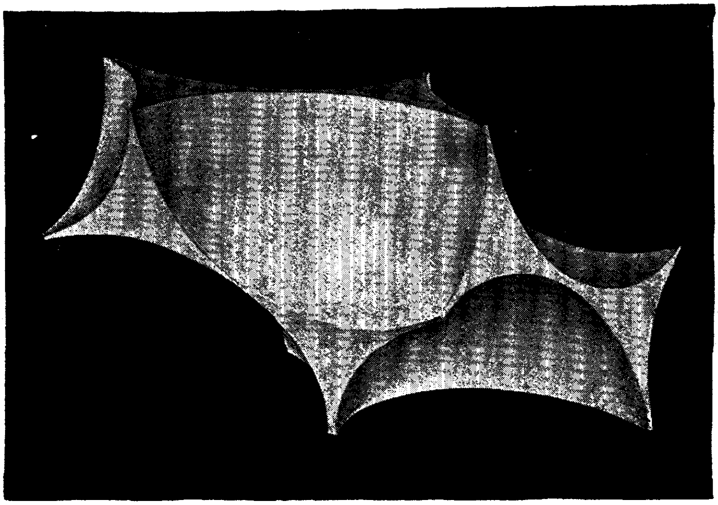


FIG. 5.—Unit element of the pore space in a mass of spheres packed in the most compact manner possible, being a plaster cast of the interior of the solid of fig. 4. By piling up a number of solids like this, with their similar faces in contact, the continuous pores in a mass of spheres would be represented.

exactly spherical in form, but it is true that the length, shape, and number of the pores, and other factors affecting the transmission capacity, are quite similar in a waterworn sand to those found in a mass of shot of proper size. Small variations in the shape of the particles have little or no effect on the result.

MECHANICAL ANALYSIS OF SOILS.

A water-washed quartz sand when viewed under the microscope looks very like a mass of common gravel or a heap of cobblestones. Pl. I, taken from the report of E. R. Buckley, State geologist of Wisconsin, on Wisconsin building stones, illustrates the general appearance very well. Soils and sands which have been deposited by running water are often well sorted as to size, but in general a naturally occurring deposit is to be regarded as a mixture of grains of a great variety of sizes. It therefore becomes necessary to determine the extent to which grains of the various sizes are present in any sample and to be able to express the results in a convenient way.

The determination of the structure of the soil is known as soil analysis, or, more properly, as the mechanical analysis of soils. In order to compare one soil with another as to its capacity to transmit water, it is necessary to have some way of arriving at a mean or average-sized grain which it is appropriate to associate with each sample. This mean diameter is known as the *effective size*, and is such that if all grains were of that diameter the soil would have the same transmission capacity that it actually has. Hazen's method of determining the effective size consists in first separating or analyzing the sand or soil into several grades by use of sieves of known mesh. The effective size is determined from the dimensions of the mesh of a sieve which will permit 10 per cent of the sample to pass through it, but will retain the other 90 per cent. That is, in any soil, 10 per cent of the grains are smaller than the effective size and 90 per cent are larger.

In order to give expression to the variety of sizes present in a sample, Hazen introduces a number known as the *uniformity coefficient*. To determine this magnitude, first find the size of sand grain which is such that 60 per cent of the material is of smaller grains and 40 per cent of larger grains. This result, when divided by the effective size of soil grain of the entire sample, gives the uniformity coefficient. Thus, if 60 per cent of a sample be finer than 0.62 mm. and 10 per cent be finer than 0.25 mm., the uniformity coefficient is $\frac{0.62}{0.25}$, or $2\frac{1}{2}$. Hazen concludes from his experimental work that the 10 per cent of small grains in a sample of a natural sand or soil has the same influence on the rate of flow of water as the 90 per cent of large grains, provided the uniformity coefficient does not exceed 5.

To illustrate Hazen's method, the analysis and diagram (fig. 6) on the next page are reprinted from his paper in the report of the Massachusetts State board of health for 1892, page 547. A curve similar to the one here reproduced (fig. 6) should be drawn for each analysis, so that the points of crossing of the 10 per cent and the 60 per cent lines can be determined; from the ordinates of these the effective size and the uniformity coefficient are derived.

Weight [of sample], dry, 110.9 grams. It was put into a series of sieves in a mechanical shaker and given 100 turns (equal to about 700 single shakes). The sieves were then taken apart, and the portion passing the finest sieve weighed. After noting the weight, the sand remaining on the finest sieve but passing all the coarser sieves was added to the first and again weighed, this process being repeated until all the sample was upon the scale, weighing 110.7 grams, showing a loss by handling of only 0.2 gram. The figures were as follows:

Sieve marked.	Size of separation of this sieve.	Quantity of sand passing.	Per cent of total weight.
	Millimeters.	*Grams.*	
190	0.105	0.5	0.5
140	0.135	1.3	1.2
100	0.182	4.1	3.7
60	0.320	23.2	21.0
40	0.46	56.7	51.2
20	0.93	89.1	80.5
10	2.04	104.6	94.3
6	3.90	110.7	100.0

Plotting the figures [in columns 2 and 4], we find from the curve that 10 and 60 per cent respectively are finer than .25 and .62 millimeter, and we have for the effective size, as described above, .25 and for the uniformity coefficient 2.5.

The most promising method of soil analysis for the purpose of determining its transmission capacity is that devised by Professor King.

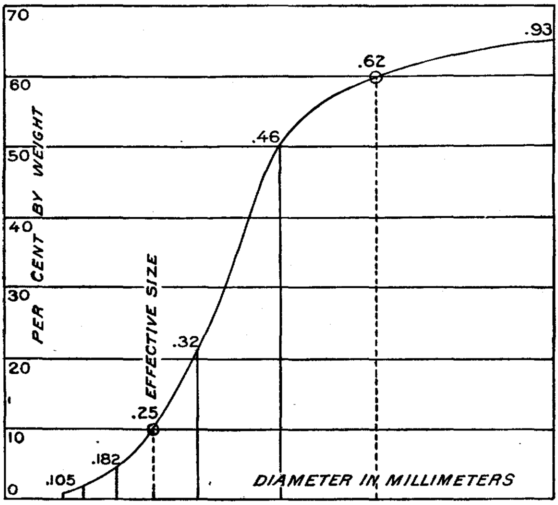


FIG. 6.—Diagram illustrating Hazen's method of making mechanical analyses of soils.

The analysis is accomplished without the use of sieves, by means of an apparatus known as King's aspirator. In this method the effective size is determined by measuring the time required for the flow of a known amount of air through the sample, the measurements being made under a known pressure. It seems that the results yielded by this method are much more concordant than those given by other methods, and the apparatus deserves a thorough test by engineers interested in soil analysis.[a]

[a] A new method for the mechanical analysis of soils, by F. H. King: Fifteenth Ann. Rept. Agr. Exp. Station Univ. Wisconsin, Madison, 1898, p. 123.

THE AUTHOR'S FORMULA FOR DETERMINING FLOW.

The formula which the writer has devised for determining the flow of water through a column of sand is as follows:

$$q=0.2012\,\frac{pd^2s}{\mu hK}\text{ cubic feet per minute.} \tag{3}$$

In this formula q stands for the quantity of water transmitted by the column of sand in one minute; p is the difference in pressure at the ends of the columns, or the head under which the flow takes place, measured in feet of water; s is the area of the cross section of the sand column, measured in square feet; h is the length of the column, in feet; d is the mean diameter of the soil grains, measured in millimeters, or the so-called "effective size;" μ is the number which takes account of the friction between the particles of water, and is known as the coefficient of viscosity (it is defined as the amount of force necessary to maintain unit difference in velocity between two layers of water unit distance apart; its value, which decreases rapidly with an increase in the temperature of the water, for temperatures from 32° to 100° is given in Table II, below); K is a constant which depends upon the porosity of the sand, and its value for porosities, varying from 26 to 47 per cent, has been computed and is given in Table III, on the opposite page.

TABLE II.

Variation of the viscosity of water, with temperature, and the relative flow of water of various temperatures through a soil, 50° F. being taken as the standard temperature.

Temperature.	Coefficient of viscosity μ.	Relative flow.[a]
Degrees F.		
32	0.0178	0.74
35	0.0168	0.78
40	0.0154	0.85
45	0.0142	0.92
50	0.0131	1.00
55	0.0121	1.08
60	0.0113	1.16
65	0.0105	1.25
70	0.098	1.34
75	0.092	1.42
80	0.087	1.51
85	0.081	1.62
90	0.077	1.70
95	0.073	1.80
100	0.069	1.90

[a] "Relative flow" means flow at given temperature compared with flow at 50° F. It is expressed as a percentage.

TABLE III.

Constants for various porosities of an ideal soil.

Porosity m.	$\frac{1}{K}$.	Log. K.	Diff.	Colog. K.
Per cent.				
0.26	0.001187	1.9258	563	8.0742
0.27	0.001350	1.8695	504	8.1305
0.28	0.001517	1.8191	490	8.1809
0.29	0.001694	1.7701	502	8.2299
0.30	0.001905	1.7199	467	8.2801
0.31	0.002122	1.6732	455	8.3268
0.32	0.002356	1.6277	430	8.3723
0.33	0.002601	1.5847	438	8.4152
0.34	0.002878	1.5409	410	8.4591
0.35	0.003163	1.4999	407	8.5001
0.36	0.003473	1.4592	400	8.5408
0.37	0.003808	1.4193	377	8.5807
0.38	0.004154	1.3816	371	8.6184
0.39	0.004524	1.3445	367	8.6555
0.40	0.004922	1.3078	353	8.6922
0.41	0.005339	1.2725	351	8.7275
0.42	0.005789	1.2374	345	8.7626
0.43	0.006267	1.2029	339	8.7971
0.44	0.006776	1.1690	320	8.8310
0.45	0.007295	1.1370	312	8.8630
0.46	0.007838	1.1058	329	8.8942
0.47	0.008455	1.0729	--------	8.9271

If t stands for temperature of the water Fahrenheit, the author's formula, in which the coefficient of viscosity has been replaced by an expression varying with the temperature similar to that given in the formula of Hazen, may be written as follows:

$$q=11.3\frac{pd^2s}{hK}[1+0.0187(t-32)] \text{ cubic feet per minute.} \qquad (4)$$

It is seen from the above formula that the quantity of water transmitted by a column of sand not only depends upon the length of the column and the head of water as expressed by Darcy's law, but varies in a most remarkable way with the effective size of the soil grain, with the temperature of the water, and with the porosity. Since the flow varies as the square of the size of the soil grain this element in the formula has a most important effect, as doubling the size of the soil grain will quadruple the flow of water. Thus the flow through a sand whose effective size of grain is 1 mm. is 10,000 times the flow through a soil whose effective size of grain is 0.01 mm. The variation of flow with temperature is also important, as the flow at 70° F. is about double that at 32° F. The variation in porosity is quite as important as the variation in temperature.

From Table III it appears that if two samples of the same sand are packed, one sample so that its porosity is 26 per cent and the other sample so that its porosity is 47 per cent, the flow through the latter sample will be more than seven times the flow through the former sample. If the two samples of the same sand are packed so that their porosities are 30 per cent and 40 per cent, the flow through the latter sample will be about 2.6 times the flow through the former sample. These facts should make clear the enormous influence of porosity on flow, and the inadequacy of a formula of flow which does not take it into account.

Part of the expression on the right side of formula (3) or (4) depends only upon the character of the soil through which the water is passing. Representing this by k we have:

$$k=0.2012\frac{d^2}{\mu K}=Md^2, \tag{5}$$

and the formula for the flow becomes

$$q=k\frac{ps}{h}, \tag{6}$$

which is essentially Darcy's formula. The constant k is the quantity of water that is transmitted in unit time through a cylinder of the soil of unit length and unit cross section under unit difference in head at the ends. We shall frequently refer to k as the transmission constant, or merely as the constant of a soil.

It should be especially noted that the velocity of flow through a soil for the pressure gradients and size of grain that commonly occur is exceedingly slow, and much less than might at first be supposed. Darton states that the rate of flow in the sands of the Dakota formation, from which the remarkable artesian wells of South Dakota draw their supply, does not exceed a mile or two a year.[a] Mr. E. L. Rogers reported to the Denver society of civil engineers[b] that American estimates agree with careful and exhaustive studies of French engineers, which show the average velocity in sands to be about a mile a year, or about an eighth of an inch a minute. In Arizona the rate has been figured out as between one-fourth and one-third of an inch per minute, while on Arkansas River above Dodge, Kans., a ditch a mile long and 5 feet below the water table in the sand developed a flow of about three-eighths inch per minute.

[a] New developments in well boring and irrigation in eastern South Dakota, by N. H. Darton: Eighteenth Ann. Rept. U. S. Geol. Survey, Part IV, 1897, p. 609.

[b] Engineering Record, Vol. XXV, p. 4351.

TABLE IV.

Velocity of water in sands of various effective sizes of soil grain and the maximum flow or transmission constant for each soil.

[Porosity, 32 per cent; temperature, 50° F. Results for other porosities can be found by the use of Table V, and for other temperatures by the use of Table II.]

1.	2.	3.	4.	5.	6.	7.
Diameter of soil grain.	Velocity, pressure gradient 1:1.	Velocity, pressure gradient 1:1.	Velocity, pressure gradient 100 feet to 1 mile.	Maximum flow, or transmission constant, k.	Logarithm of numbers in column 5.	Kind of soil.
Mm.	*Ins. per min.*	*Miles per year.*	*Miles per year.*	*Cu. ft. per min.*		
0.01	0.0014	0.0113	0.00026	0.000036	5.5569	
0.02	0.0054	0.0452	0.00102	0.000144	6.1590	Silt.
0.03	0.0122	0.1016	0.00230	0.000324	6.5111	
0.04	0.0218	0.1807	0.00408	0.000577	6.7610	
0.05	0.0340	0.2823	0.00638	0.000901	6.9548	
0.06	0.0490	0.4065	0.00918	0.001298	7.1132	
0.07	0.0667	0.5534	0.01250	0.001766	7.2471	Very fine sand.
0.08	0.0871	0.7228	0.01633	0.002308	7.3631	
0.09	0.1103	0.9147	0.02066	0.002920	7.4654	
0.10	0.1361	1.129	0.02551	0.003605	7.5569	
0.12	0.1961	1.627	0.03674	0.005192	7.7153	
0.14	0.2668	2.213	0.05011	0.007065	7.8491	
0.15	0.3063	2.541	0.05753	0.008112	7.9091	Fine sand.
0.16	0.3485	2.892	0.06382	0.009228	7.9651	
0.18	0.4412	3.659	0.08266	0.01168	8.0675	
0.20	0.5446	4.518	0.1021	0.01442	8.1590	
0.25	0.8509	7.058	0.1594	0.02253	8.3528	
0.30	1.225	10.16	0.2296	0.03244	8.5111	
0.35	1.668	13.84	0.3125	0.04417	8.6451	Medium sand.
0.40	2.178	18.07	0.4081	0.05768	8.7610	
0.45	2.757	22.87	0.5165	0.07300	8.8633	
0.50	3.403	28.23	0.6377	0.09012	8.9548	
0.55	4.119	34.17	0.7718	0.1090	9.0377	
0.60	4.901	40.65	0.9183	0.1298	9.1132	
0.65	5.751	47.81	1.077	0.1523	9.1827	
0.70	6.671	55.34	1.250	0.1766	9.2471	Coarse sand.
0.75	7.660	63.53	1.435	0.2028	9.3071	
0.80	8.714	72.28	1.633	0.2308	9.3631	
0.85	9.835	81.57	1.843	0.2604	9.4157	
0.90	11.03	91.47	2.066	0.2920	9.4654	
0.95	12.28	101.9	2.302	0.3253	9.5123	
1.00	13.61	112.9	2.551	0.3605	9.5569	
2.00	54.46	451.8	10.21	1.442	0.1590	
3.00	122.5	1,016	22.96	3.244	0.5111	Fine gravel.
4.00	217.8	1,807	40.81	5.768	0.7610	
5.00	340.3	2,823	63.77	9.012	0.9548	

Table IV, above, gives the velocity of movement of water in sands of various grades for different pressure gradients. Column 1 gives the effective size of the soil grains in millimeters. As already stated, this size is such that if all grains were of that diameter the soil would have the same transmission capacity that it actually has. Column 2 gives the velocity of flow, or the rate at which the water moves through the ground in inches per minute under a pressure gradient of 1 foot difference in head to each foot of distance. Column 3 gives the velocity of flow reduced to miles per year, the pressure gradient being the same as in column 2. Column 4 gives the velocity of flow in miles per year under a pressure gradient of 100 feet to the mile. The velocity for a pressure gradient of 10 feet to the mile would be one-tenth of the numbers in this (fourth) column, and so on for other gradients. Column 5 gives the actual discharge in cubic feet per minute for each square foot of cross section if the pressure gradient be 1 foot difference in head for each foot of distance. For a pressure gradient of 1 foot difference in head for each 100 feet of distance, the flow per square foot will be 0.01 of the tabulated numbers, and so on for other gradients. The numbers in this (fifth) column have also been called the "transmission constants," and have been represented in the formulas by k.

TABLE V.

Relative flow of water through sands of same effective size grain, but packed so as to possess different porosities.

Porosity, or per cent of voids.	Relative flow.[a]
30	0.81
32	1.00
34	1.22
36	1.47
38	1.76
40	2.09

[a] "Relative flow" means flow for the given porosity compared with flow for porosity 32 per cent as standard. It is expressed in the table as a percentage of the flow for a sample having a porosity of 32 per cent.

MAXIMUM FLOW.

Inasmuch as the flow of ground water is nearly always caused by a difference in head due to gravity only, the maximum flow that is pos-

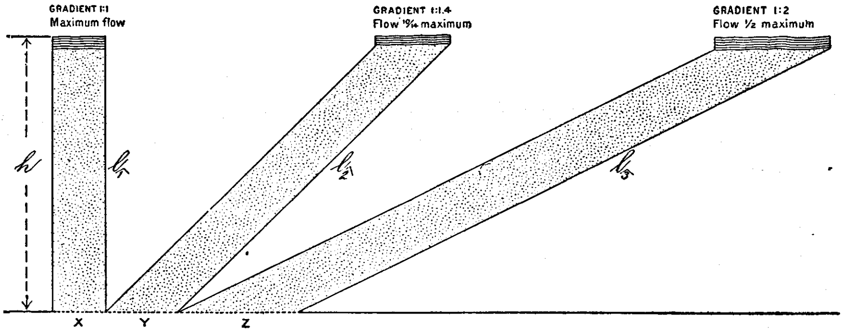


FIG. 7.—Diagram illustrating various pressure gradients and the maximum flow. In these three cases the upper portions of the soil columns are supposed to be supplied with water as fast as it can flow through the columns. The escape at X, Y, Z is supposed to be perfectly free. The head under which the flow takes place is h, in each case, as shown at the left of the figure. The various lengths of the soil columns, l_1, l_2, and l_3 produce the pressure gradients $h/l_1 = 1$; $h/l_2 = \frac{1}{1.4}$; and $h/l_3 = \frac{1}{2}$, respectively, with the resulting flows in proportion if the material in the various columns be the same.

sible is found in the case in which the ground water is free to move in a vertical direction, as in a perfectly underdrained sand-filter bed. The motion in this case is due to the weight of the water of saturation, and the flow is of course greater than would be the case if the water were obliged to flow in a direction inclined to the vertical, instead of in the vertical direction as supposed. These facts are illustrated in fig. 7. The flow in the case of pressure gradient 1:1 forms a most convenient basis for calculation, and it is frequently called, as suggested by Hazen, the maximum flow. The flow for any other gradient is immediately calculable from the maximum flow—for a gradient 1:100 the flow being, of course, one one-hundredth of the maximum flow.

The appropriateness of the term maximum flow is illustrated by fig. 8, which shows the original water table and the depressed water table due to the construction of a drainage ditch. It is plain that the pressure gradient for all of the streams of flow marked by arrow heads is less than the gradient 1:1. If the wetted area of the ditch be multiplied by the maximum flow for the kind of material in which the ditch has been excavated the flow thus computed will in every case exceed the flow actually determined by measurements of the yield of the ditch.

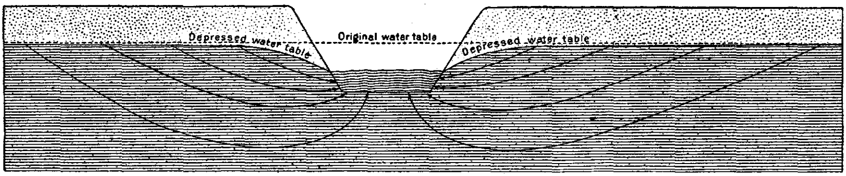


FIG. 8.—Diagram showing lines of flow into a drainage ditch and the shape of the water table in its neighborhood. The head under which the flow takes place is the difference in height of the original water table and the level of the surface of the water in the ditch. This is much less than the lengths of the curved lines of the flow into the ditch, hence the rate of flow must be much less than the so-called maximum flow.

There is not uniformity in the use of the term velocity as applied to the motion of ground waters. We use the term to express the rate (measured as so many feet a day, etc.) at which the water advances through the porous medium, irrespective of the amount of water thus advancing. The amount of ground water (measured in cubic feet per minute, etc.) passing through a given cross section the writer has called the flow or the discharge. It is equal to the velocity multiplied by the porosity. Some measure velocity as a rate of motion in a solid column of same area as the cross section of the porous medium. This is the same magnitude which we have called flow.

In using Table IV one should use the numbers in columns 2, 3, or 4 if the velocity of ground water is wanted, but should pass to column 5 if the flow or yield is required. Thus, suppose it is desired to find the rate of motion of ground water through a bed of sand which slopes 10 feet to the mile. The results can be found for various materials and grades of material by dividing the numbers in column 4 by 10, since a slope of 10 feet to a mile will cause but one-tenth of the velocity existing for a slope of 100 feet to a mile. For materials of various grades we obtain the following results:

Velocity of ground water in materials of different grades, pressure gradient 10 feet per mile.

Material.	Miles.	Feet per year.
Fine sand, 0.2 mm. diameter	0.010	52.8
Medium sand, 0.4 mm. diameter	0.041	216.0
Coarse sand, 0.8 mm. diameter	0.16	845.0
Fine gravel, 2 mm. diameter	1.02	5,386.0

Suppose that it is desired to ascertain the amount of water that will pass through a bed 200 feet deep and 1,000 feet wide, having the same slope as that just mentioned. This problem requires us to find the flow, and the numbers used in the computation should therefore be taken from column 5 of Table IV. The flow for 1 square foot of cross section of the bed will be $\frac{10}{5280}$ of the maximum flow given in that column for material of various grades, and the total flow is found by multiplying the maximum flow by $\frac{10}{5280} \times 200 \times 1{,}000$, which gives the following results for the same materials described in the preceding table:

Flow of ground water in materials of different grades through a bed of vertical cross section 200 by 1,000 feet, sloping 10 feet per mile.

	Cu. ft. per min.
Fine sand	5.5
Medium sand	22.0
Coarse sand	87.0
Fine gravel	546.0

The estimates in Table IV were based upon a porosity of 32 per cent. For other porosities the results must be changed by the percentages shown in Table V. Thus all of the results just found must be increased by about 37 per cent if the porosity of the material be 35 instead of 32 per cent.

CHAPTER II.

SURFACE ZONE OF FLOW OF GROUND WATERS.

Most people obtain their ideas of underground streams of water from the descriptions of the underground torrents in the Mammoth and other caves. Such notions, however, are erroneous. The rivers found in caverns are almost exclusively peculiar to limestone or calcareous formations and are not typical of subterranean streams. The underground drainage of calcareous rocks is, nevertheless, of great interest. The well-known solvent action of rain water percolating through limestone has no more beautiful demonstration than the existence of caverns like the Mammoth Cave, with its scores of miles of ramifying passages and its enormous vaulted chambers. As is well known, these galleries and passages represent the channels of former subterranean streams which for the most part have now found escape at lower levels. Maps of such caves show that the passages are arranged somewhat like the branches of a tree—very similar, indeed, to the divisions and subdivisions of surface drainage as represented by the main river, the lesser rivers, and the tributaries.

As a rule, the running streams of limestone caverns, as well as those found in the seams, joints, fissures, etc., of crystalline and other rocks, join the surface-drainage system before they have attained any considerable size. Therefore large cavern streams are very rare, and when discovered they have attracted wide attention. The Echo River of the Mammoth Cave, one of the largest and best known of this type, is from 20 to 200 feet wide and from 10 to 40 feet deep. Sometimes little can be directly observed of a subterranean stream except the enormous spring which marks its mouth. The famous Wyandot Spring, near Columbus, Ohio, representing the drainage of a considerable area collected into ramifying passages in the Corniferous limestone, is a good example of this type, as are also the enormous fresh-water springs in the ocean off the coast of Florida.

Pl. II shows a sink hole connecting with an underground drainage system, and an underground channel, as observed by Mr. Willard D. Johnson on the southwestern plains. The writer is indebted to Mr. Johnson for the photographs.

Putting the subterranean streams of the limestone type into a class by themselves, there remain the great systems of underground drainage represented by the slowly percolating waters of sand and gravel deposits, sandstones, and other porous materials. It is this class of subsurface drainage that we must regard as the important one, both because of its actual areal extent and because of its economic importance as a possible source of water supply. It is the water of this type of underground drainage that is properly designated ground water or underground water, the running water of crevices and caverns rarely being uppermost in mind when this term is used.

Underground waters may be divided into three principal zones: (1) The unsaturated zone; (2) the surface zone of flow; and (3) the deeper zones of flow.

The motion of water in the unsaturated zone is essentially in a vertical direction, downward, supplying the saturated sheet below it in times of rainfall, and upward supplying the surface evaporation and the requirements of vegetation by means of the capillary action of the soil during rainless periods. A special discussion of this zone of ground waters does not come within the scope of this paper, and is therefore omitted.

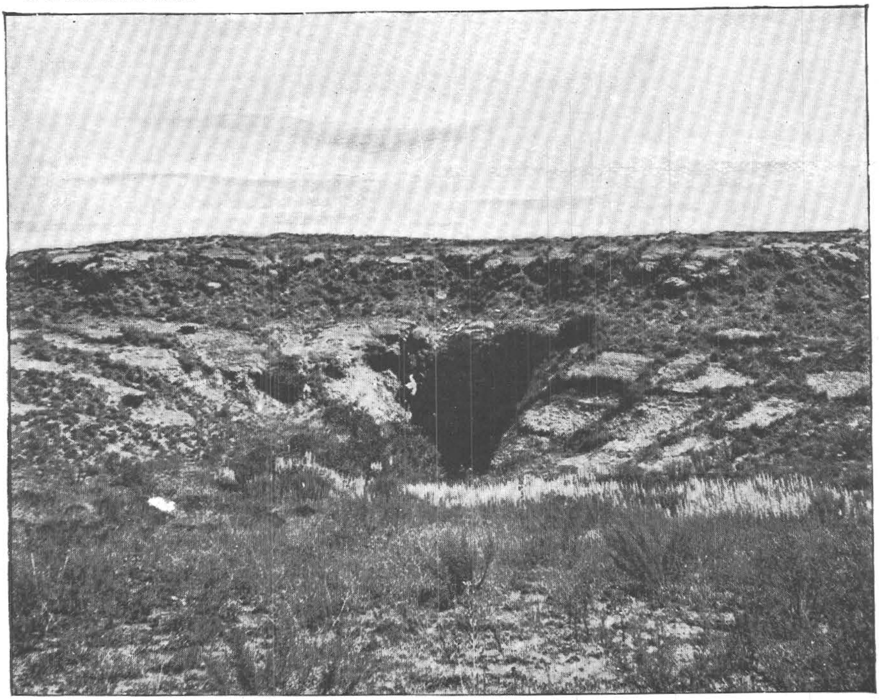
The surface or upper zone of flow extends from the level of the water table to the first impervious material of general extent reached by the underground water in its downward percolation.

The deeper zones of flow are those that lie below the first impervious stratum. There may be several zones falling within this class. In these zones the direction and character of the flow are usually quite independent of surface topography and are almost entirely controlled by large regional and geologic conditions. Special consideration of the deeper zones of flow will be postponed to a later chapter of this paper.

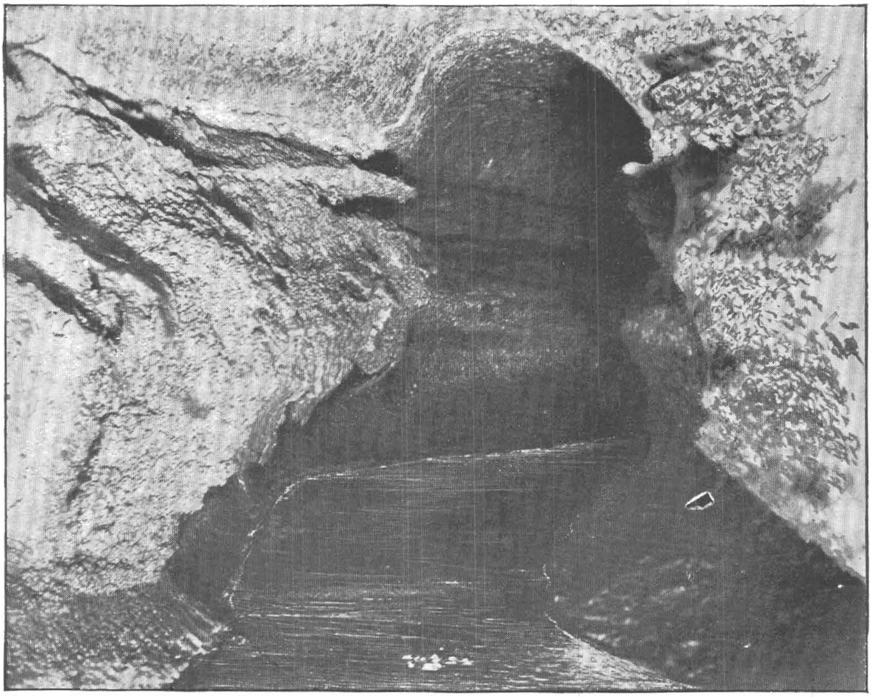
UNDERGROUND DRAINAGE BASINS.

The unit of the surface zone of flow of ground waters is the river valley. In the surface zone the rate and direction of motion of the underground water conforms primarily to the slopes and grades of the land surface. The effective principle in the surface zone is that underground flow follows the trend and direction of the surface drainage. The direction taken by the surface waters in their course into streams and drainage channels is in general the same as that taken by the seepage waters of the upper zone of flow. As previously pointed out, actual determinations usually show that the water table has a slope which is essentially similar to the slope of the surface of the ground, differing from the latter principally in being less steep. The surface divide or watershed usually coincides with the line of the underground water divide or watershed, and the motion of the underground seepage into the streams and rivers is similar to the lines followed by the surface drainage into the same streams.

The lowest line of drainage of the valley is known technically as the thalweg. Topographically it is a line upon a contour map which is a natural water course. Beneath the thalweg there is usually a similar drainage line for the underground current, in general coincident



A. SINK-HOLE CONNECTION WITH UNDERGROUND DRAINAGE SYSTEM.



B. UNDERGROUND CHANNEL.

with the thalweg. For other parts of the valley the actual lines of motion of the underground water are represented by a set of curves which cut the contour lines of the water table at right angles, as shown in fig. 9. The similarity of the contours of the water table to those of the land surface enables one to sketch approximately the lines of underground seepage from a contour map of the surface.

For the most part the lines of flow run into the surface streams or thalwegs, but between A and B and X and Y there is indication of an underflow or general movement in the direction of the surface streams and independent of the same.

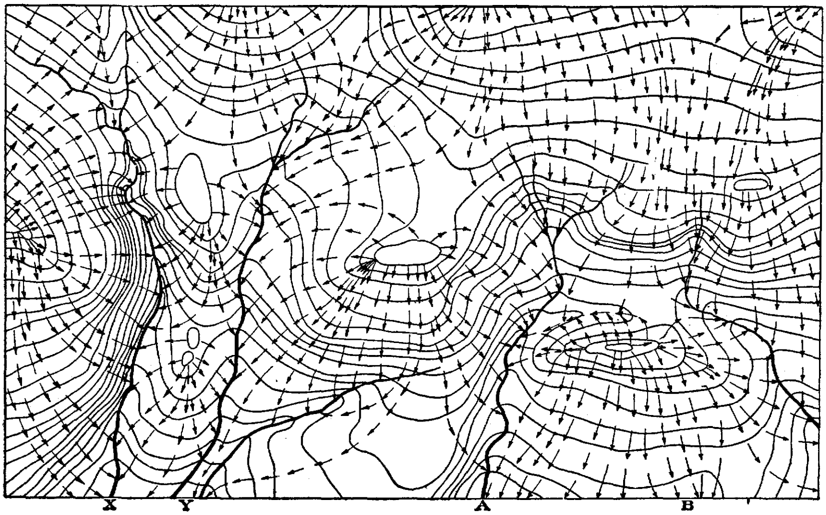


FIG. 9.—Contour map showing position of water table (continuous lines), supposed lines of motion of ground water (arrowed lines), and the thalwegs or drainage lines (heavy lines).

It is claimed by Chalon,[a] a French engineer, that the subterranean thalweg on the main line of underground drainage is usually nearer the steeper side of the valley than is the surface stream. (See fig. 10.) Such principles are emphasized in foreign treatises, but American experience has not found them of great value.

SHAPE OF THE WATER TABLE.

The similarity between the contours of the land surface and the contours of the water table just explained must not be taken too literally. The coincidence of the surface and subterranean thalwegs and of the surface and subterranean watersheds is a common occurrence, but is not a geologic necessity. In itself the surface topog-

[a] Sur la recherche des eaux souterraines: Mém. Soc. ingénieurs civils de France, 1897, Vol. II, p. 38. See also L'Art de découvrir les sources, by L'Abbé Paramelle, fourth ed., Paris, 1896, p. 138.

raphy is only one, and often not the most important, element in the control of the underground current.

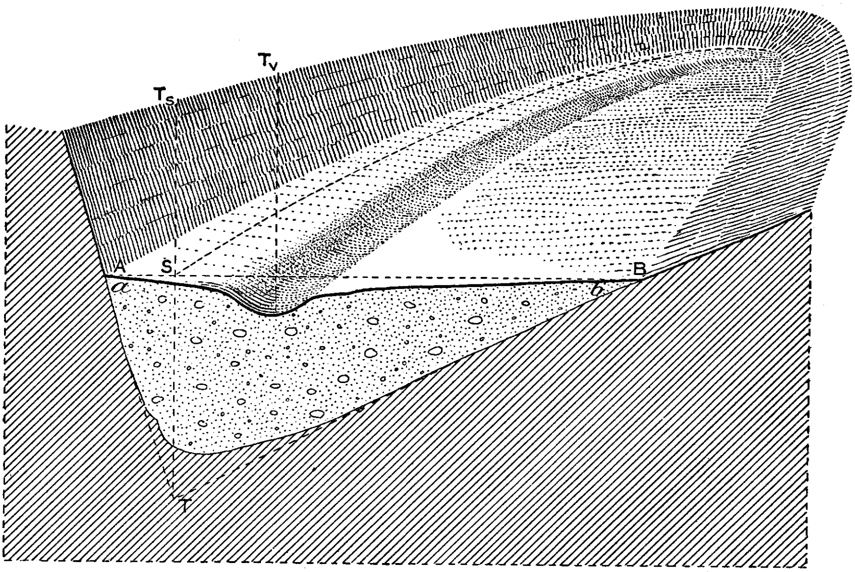


FIG. 10.—Diagram illustrating Chalon's principle that the main core of the underground drainage, or subterranean thalweg, naturally lies nearer the steeper side of the valley than does the surface stream. The surface stream is at T_v and the subterranean thalweg at T_s.

The horizontal distribution and motion of the ground water is influenced first of all by the form of the surface of the impervious layer.

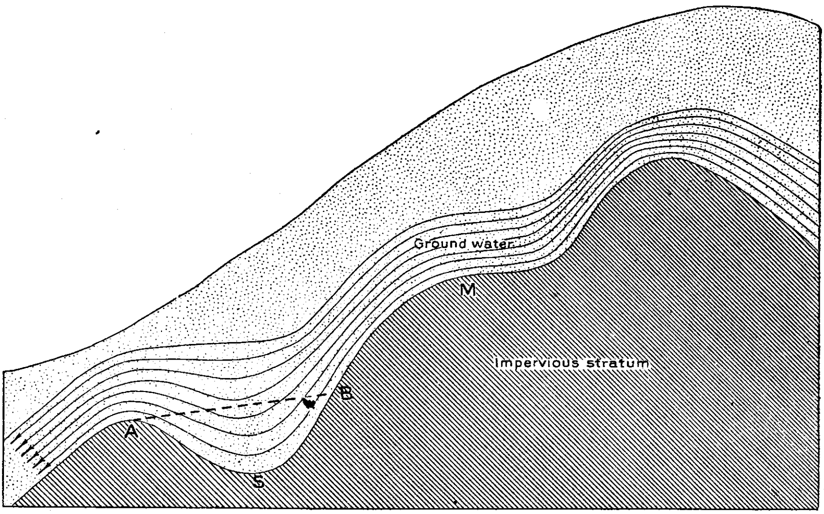


FIG. 11.—Diagram illustrating lines of flow of ground water over a series of monoclines, synclines, and anticlines, as at M, S, and A, respectively.

It is also influenced in a marked degree by the varying altitude of the surface or receiving area, by the character of the pervious layer, by the altitude and distance of the nearest thalweg or drainage channel, and finally by the amount of the rainfall. These elements acting together in a region determine the depth of the water table below the surface and the direction and rate of motion of the underground current. They form a complicated system, and it is not easy to describe the precise part which each plays in a given case. Fine material and large rainfall tend to make the ground water stand high within the hills and elevated places and to give steep gradients to the water table. Likewise coarse material and light rainfall tend to a low water table and to light gradients. From these considerations it is obvious that the form of the impervious stratum affects the water table much less in humid climates than in semiarid or arid climates.

It is also important to remember that synclines in the impervious material (as at S, fig. 11) crossed by the lines of motion of the ground

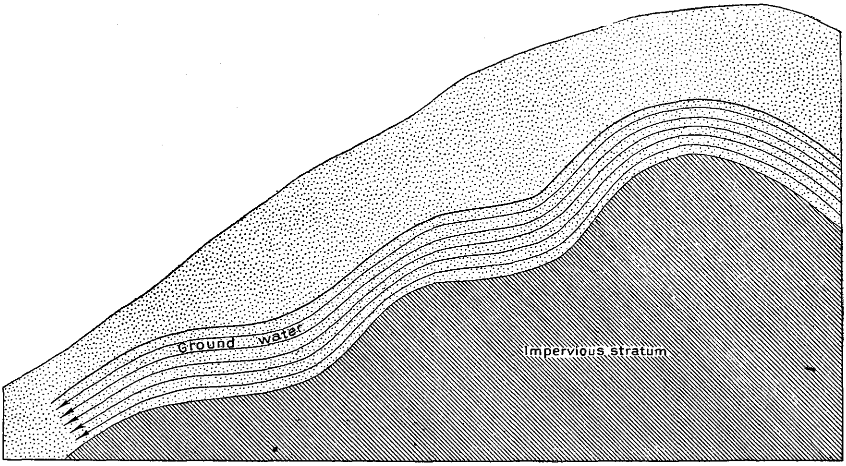


FIG. 12.—Diagram illustrating lines of flow of ground water over a series of monoclines in an impervious bed.

water have no effect on the form of the water table. However, if the lines of motion follow a syncline instead of crossing it, the water table also will probably have a synclinal form, although less pronounced than that of the impervious layer. Monoclines in an impervious floor crossed by moving ground water will give a similar form to the water table, as shown in fig. 12. We may normally expect, then, that the water table in the direction of the lines of motion will be a series of monoclines, as shown in figs. 1, 11, and 12, with only occasional deviations during periods of rainfall.

The motion of the ground water as a whole is somewhat like the slow motion of a very viscous sirup or the slowly creeping ice of a glacier. These comparisons, however, are likely to be misleading. In the slow motion of a viscous liquid the bounding layer sticks fast to the walls of the vessel containing the liquid, so that motion nearly

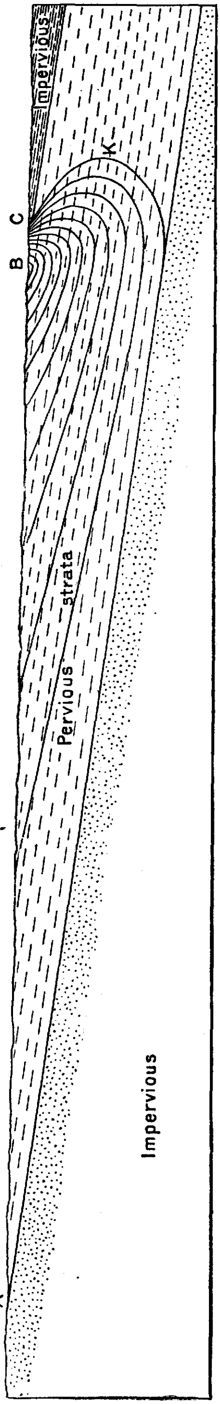


FIG. 13.—Diagram illustrating lines of motion of ground water in an outcrop of a pervious bed underlain and overlain with impervious strata. The diagram represents a vertical section taken along the dip of the strata. The area from A to B is a receiving area, which absorbs the rainfall. From B to C is an area of outflow or of returning ground waters. The pervious bed is supposed to have no drainage to the right of the figure, so that all return seepage must take place around the edge, C, of the impervious cover. The diagram was drawn by the author to illustrate the paths of the ground waters which have deposited the ores in the Mesabi iron range.

ceases in the neighborhood of the boundary. This is not the case, however, in the movements of ground waters, as the frictional resistance which the water meets is present in each individual pore between the soil particles, and is not transmitted from layer to layer through the water itself, as in the case of a viscous liquid. If the material is no finer near the impervious layer or boundary the resistance to motion per unit length is no greater at the boundary of the region than in its interior. This uniform distribution of the resistance to motion throughout the mass of the ground water, and the separation of each individual stream in a capillary pore from every other similar stream in the neighboring pores makes the character of the motion difficult to understand. The motion of water in pipes, canals, and surface streams is an exceedingly poor and a very misleading analogy. If it were not for the ever-present controlling influence of gravity the motion would be entirely analogous to the flow of heat or electricity in a conducting medium, as the writer has shown in another paper.[a]

A remarkable conclusion from the paper just cited is that ground-water motions resemble in general character the motions of a "perfect" or frictionless liquid under similar conditions but with the porous medium entirely absent. The chief difference lies in the fact that in the case of a perfect liquid gravity produces a rapidly accelerated motion, the energy represented by the increase in the motion being the equivalent of the work done by the external forces (gravity), while in the case of the motion of the water in the pores of the medium

[a]See Chapter II of Theoretical investigation of the motion of ground waters: Nineteenth Ann. Rept. U. S. Geol. Survey, Pt. II, 1899, p. 295.

most of the energy contributed by the work of the external forces is transformed into heat by the enormous friction in the capillary spaces. Thus in fig. 12 the velocity along the various lines of flow, represented by the continuous curves with arrowheads on them, is substantially the same for each line, since they are of practically the same length. The water flows along them just as heat would flow along a conducting material of the same shape. Likewise in fig. 11 the velocity along the lower lines of flow is somewhat less than along the upper lines, for the lower lines have the greater length. Again we may imagine that the ground water flows along them just as heat would flow along a conductor of similar shape, unless the ground water is about to escape to the surface drainage, as at S in figs. 25 and 26; in this case the effect of gravity is to give relatively high velocities to the water lying next to the impervious floor. The writer has shown that the velocities and lines of motion can be calculated for such cases, some results being shown in figs. 14 and 15.

The contention of some German hydrographers that there can be no motion[a] in a region like ASB in fig. 11 must be entirely abandoned. The water must circulate in all parts of the enlargements in the porous medium, for the same reasons that heat would be conducted

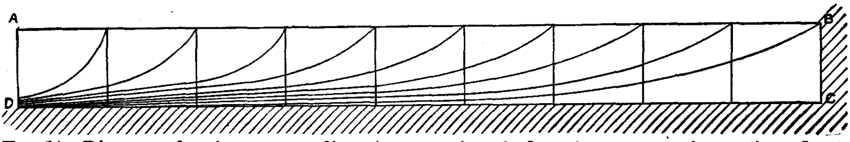


FIG. 14.—Diagram showing seepage lines in a pervious bed resting upon an impervious floor. A, B, C, D is a vertical section. Rain falls upon the surface AB and finds free escape at the surface AD. The curved lines show the paths along which the seepage takes place. The effect of gravity is to draw these lines close together near the impervious floor.

over similar enlargements in a conductor. All lines of motion must begin and end in the boundaries of the water-bearing medium, and must entirely traverse and completely occupy all enlargements in the porous strata.

As already indicated, the general trend of the moving underground water, under the influence of gravity, is into the neighboring streams and lakes. This motion must be materially modified by many causes, which frequently present most complex combinations. While the return flow of ground water to the water courses by means of diffused and almost imperceptible seepage is the rule, yet we must remember that geologic conditions may be such—for example, the outcropping of an impervious stratum—as to force the water table above the surface of the ground and converge and concentrate the lines of flow into a strong current. In the latter case we have the phenomenon of the flowing spring. Ground water returning to the surface in the form of springs so quickly attracts observation

[a] Otto Lueger, Wasserversorgung der Städte, p. 127.

that the more important but less obvious return in the form of diffused seepage almost entirely escapes attention.

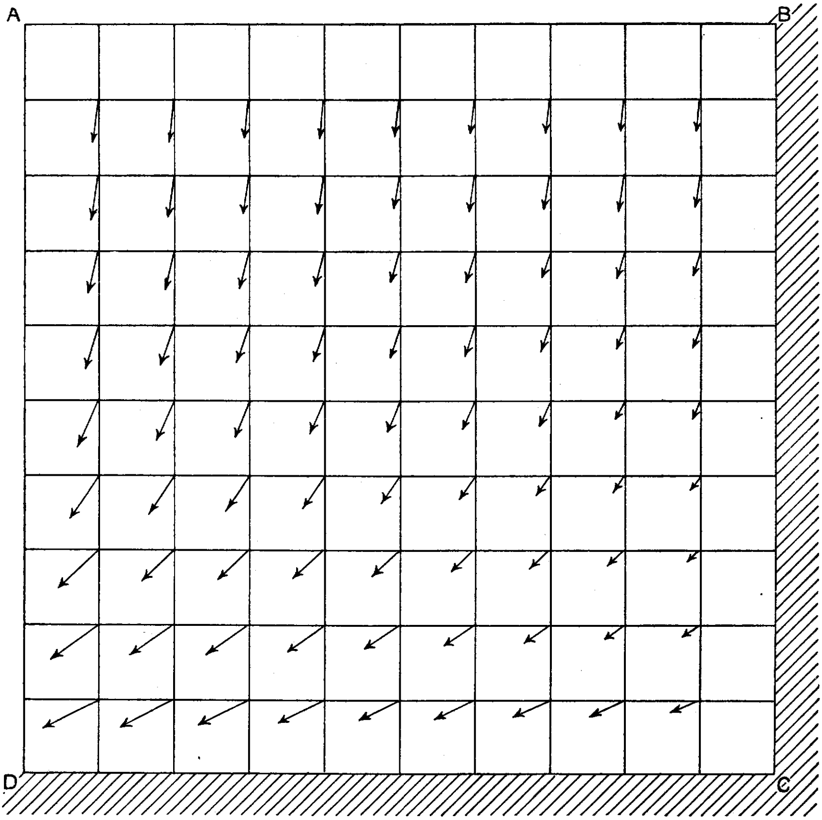
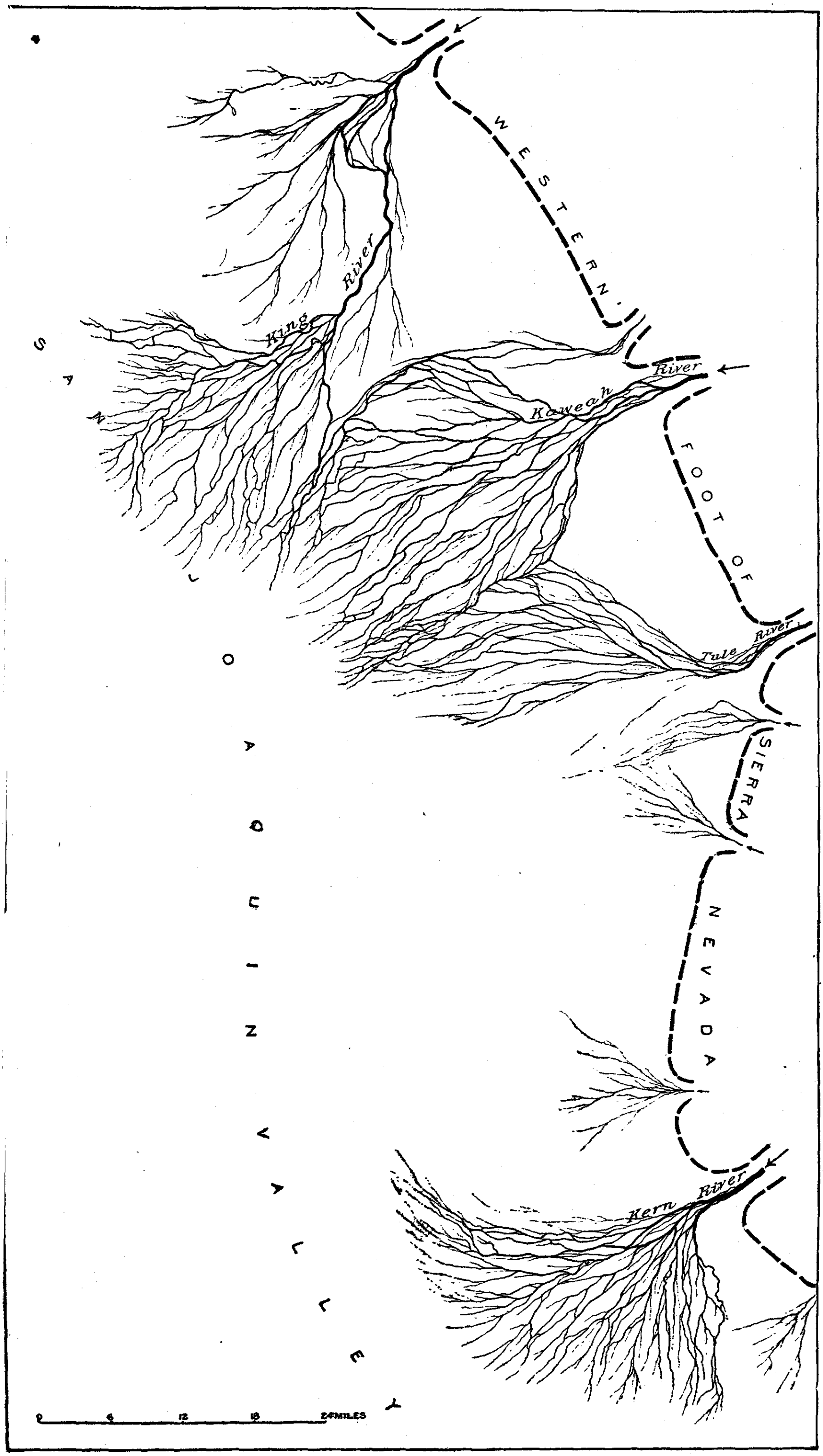


FIG. 15.—Diagram showing direction and rate of seepage in a pervious bed resting upon an impervious floor. The directions and lengths of the arrows show, respectively, the direction and the velocity of motion at various points in the pervious bed. Rain falls upon AB and escapes at AD, as in fig. 14.

THE UNDERFLOW.

The ground water after starting on its journey toward the river valley may not after all find its way immediately into the channel. Sometimes it takes a general course down the thalweg and toward the sea within the porous medium itself. This movement may be so great as to constitute a large underground stream, scores of feet in depth and miles in breadth.

The moving sheet of water beneath the bed and banks of a stream is the underflow, properly so called. This term is also extensively used in the West to designate the nearly stationary ground water of the Great Plains. It originated with those who formerly believed that there was a great sheet of water beneath the surface of the Plains, originating in the melting snows of the Rocky Mountains and flowing rapidly through the subsoil toward Missouri River and the sea.



MAP SHOWING THE DELTAS OR FANS OF DISAPPEARING STREAMS AS THEY LEAVE THEIR MOUNTAIN CANYONS.

It is evident that a considerable underflow is impossible with material as fine as is sometimes found filling the valleys of rivers. Such material may play a very important part in the storage of water, but it can not be an important element in its transportation over great distances. An entirely different condition may exist where the sands and gravel beneath the stream are sufficiently coarse, as is likely to be the case near the source of streams in the mountains. Here the material deposited by the stream is the coarser sands, gravel, and bowlders brought down by the mountain torrents. The running water will not deposit the finer material until the stream bed reaches a gentle slope and the current has lost its high velocity. The periodic floods sweeping down the mountain sides during the heavy rains soon fill the river canyons with coarse débris, on the surface of which the perennial stream has its bed. Below the bed of the stream there is a vast body of water slowly percolating through the coarse material. Occasionally as we pass down the stream we may find the rock walls of the canyon nearer together, or perhaps the bed rock beneath the débris is closer to the surface, the result in either case being a marked augmentation of the amount of water in the river, because the percolating waters of the underflow are forced to the surface and mingle with the waters of the surface stream.

At the mouth of the canyon the valley broadens out and the bed rock sinks more deeply beneath the surface of the land, while the stream deposits more and more of its suspended matter. The underflow may also broaden and deepen to fill the greatly enlarged channel, the finer material and lesser slope decreasing the speed with which it flows down the valley. The valley may be now so broad that if the rainfall be copious the constant increments to the underflow from seepage of the rainfall from the higher land must give rise to regular flow of the subsurface waters into the perennial stream itself, causing its constant growth during its course toward the sea. On the other hand, if the stream issues from its mountain canyon in an arid region the surface stream may gradually disappear, until a dry wash through the valley, only occasionally swept with floods from the mountains, marks the general course pursued by the silent underflow.

Numerous instances of the disappearance of mountain streams occur in the arid regions of the West. There are many interesting cases along the Coast Range in California, two of which are shown on the Cucamonga topographic atlas sheet of the Geological Survey. From that map will be seen the way in which the disappearing streams fan out into numerous branches, which are very appropriately called, from analogy, the deltas of the rivers. In Pl. III are shown the distributing deltas of King, Kaweah, and Kern rivers, California, as they enter San Joaquin Valley from their canyons in the western edge of the Sierra Nevada. There can be no question that the underflow in this case follows the general trend indicated by the surface branches.

It has been pointed out how the broadening of the valley, as a stream flows out of its mountain canyon, will have effects entirely opposite, depending upon whether the climate of the valley be arid or humid. If the climate be arid the broadening of the valley will expose the underflow to increasing evaporation during its slow movement in a broad belt, while if the climate be humid the broadening of the valley will greatly increase the collecting area of the rainfall and constantly contribute a part of the underflow to the water of the river itself. In one case the broadening of the valley is the occasion of the constant diminution of the volume of the river; in the other case it is the occasion of the constant growth of the river.

The relation of the underflow to the waters of the river channel presents many interesting phenomena and variations. In some cases silt may have rendered the channel of the river so impervious that for considerable distances little interchange can take place between the river waters and the underflow waters. This sealing effect of river silt is illustrated in an attempt by Nettleton to measure the velocity of the underflow of the Rio Grande.[a] The plan was to sink holes in a sandbar a little below the water line, with the expectation that the water would rise in them to the level of the river surface. It was then proposed to note the time it took a colored liquid to travel from one hole to the next one downstream. But even on this island, surrounded by water and still under water a few days prior to the experiment, and to all appearances composed of the same material as the river bed and its banks, they did not succeed in finding water 3 feet below the surface. The holes were within a foot of the river and yet no water came into them within twenty-four hours.

That underflow waters are sometimes quite independent of the water flowing in the surface stream is abundantly shown by the experience of German water-supply engineers. B. Salbach cited some interesting examples before the engineering congress at the Chicago World's Fair in 1893.[b] In 1867 Salbach was called to report upon the preliminary works then under way for testing the sources of supply of the city of Halle. These works consisted of borings in the Aue, near Beesen, above the junction of the Elster and the Saale, in a thick and widely extended bed of gravel, and in pumping large quantities of water from a well which a chemical analysis had shown to be suitable for the proposed supply.

It had been assumed, before undertaking the examinations, that the water from the neighboring rivers had penetrated this bed of gravel. That this assumption was erroneous was speedily proved. In the first place, the water taken from the gravel was shown to have

[a] See Final Report of Chief Engineer E. S. Nettleton: Ex. Doc. 41, Pt. II, Fifty-second Congress, first session, p. 35.

[b] Experiences had during the last twenty-five years with waterworks having an underground source of supply, by B. Salbach, Baurath at Dresden, Saxony: Trans. Am. Soc. Civil Eng., Vol. XXX, 1893, pp. 293-329.

an entirely different chemical composition from that of the river water, the water from the gravel being harder than that taken from the river, while the water of the Saale gravel was still harder than the water of the Elster gravel. In the next place, it was shown by a series of experiments that a natural filtration of the river water was wholly impossible. To prove this several wells were dug along the shore of the river, and the ground-water level in them was reduced, by pumping, about 4 feet lower than the river level. When this was done the earth over the whole water level on the river side was entirely dry. Extending the well laterally until within about 14 inches of the bank of the river caused no flow of water from the bed of the river into the well, this last 14 inches of soil being dense and impervious. These examinations proved that in this case at least there was no flow of water from the river into the bed of gravel of the valley, and that the water drawn from the wells was an independent underground source of supply which flowed parallel to the river through the gravel and followed the general slope of the valley.

Salbach also gave an account of the water-bearing gravels of the valley of the Elbe. Several bore holes put down on the right bank of the Elbe showed a steep slope of the water table at right angles to the river. The ground water taken from these holes was found to be exceedingly pure and of an appreciably less degree of hardness than the water of the river. Borings on the left bank showed a less slope of the water table at right angles to the river, while the samples of water taken from the borings were pure, but somewhat harder than the water of the Elbe. Borings put down in the river itself to a depth of from 23 to 26 feet showed that the ground water rose in the bore holes 6 inches above the level of the river, while the water was softer than the river water. It is a remarkable fact that the ground water in almost the entire valley of the Elbe is softer than that of the river.

RATE OF MOVEMENT OF THE UNDERFLOW.

The magnitude of the underflow depends, obviously, upon many important factors. One of these is the average gradient of the river valleys. Others are the depth, width, and composition of the beds which constitute the alluvial deposits of the stream. The fineness of this alluvium is of special importance. An inspection of Table IV (p. 27) will show that a coarse sand or gravel will permit an underflow of considerable magnitude. On the other hand, a fine material will render the underflow quite insignificant.

Direct observations of the rate of motion of the underflow are almost completely lacking in this country. A number of indirect observations of various kinds enable us, however, to form an idea of the possible rates under special circumstances. One of the highest determinations of the rate of underflow is that made by Hicks for Loup

River in Nebraska.[a] In the region considered September is a relatively dry month, much drier than August. The maximum rainfall usually occurs about the middle of July. Notwithstanding the marked decline in precipitation, and quite independent of the state of the weather at the time, the Loup rivers generally rise in September. The increase in volume is slight, yet perceptible, and, occurring in a dry month, can not be explained by contemporary causes. Hicks believes the rise to be due to the July rainfall upon the neighboring undrained areas, which amount to about one-third of the entire basin. The percolating waters reach the rivers about two months after they have fallen as rain, and estimating the average distance traversed by the subterranean water as 20 miles, Hicks computes the rate of flow to be about one-third of a mile per day. This rate of flow is so enormous, as can be seen by reference to the fourth column of Table IV, that the explanation must be defective. Possibly the phenomena may be explained by the lesser evaporation of September.

More satisfactory conclusions as to the rate of the underflow may be deduced from the measurements of seepage reported by Professor Carpenter.[b] Estimates based upon the seepage from the Fort Morgan canal and a new section of the same canal gave for the first a velocity of about 3 feet a day and for the second a velocity of about 15 feet a day. Water Commissioner J. T. Hurley reports to Carpenter that the seepage from the Weldon Valley canal has progressed 1½ miles in five years. N. C. Alford reports that in one case, under the Larimer County canal, it was five years before seepage showed at a distance of 40 rods from the canal, though the slope was considerable. In one case, near Greeley, according to Mr. S. A. Bradfield, it seems to have taken about ten years for the water to move 2½ miles.[c]

C. E. Grunsky has measured the seepage loss in King River and the Fresno canal, California,[d] and found that the losses for various sections of the canal were 3.77, 3.48, and 1.25 second-feet per mile. The average velocities of seepage through the canal bed, calculated on a basis of average width of 40 feet and a porosity of 33 per cent, are, respectively, 4.8, 4.3, and 1.6 feet per day. According to the same authority the losses on various portions of the Fresno canal were 8.49, 0.74, and 0.95 second-feet per mile. The average velocities of seepage, calculated on a basis of width of canal bed of 50 feet, are, respectively, 8.4, 0.7, and 0.1 feet per day. The average losses on a 6-mile and a 1-mile section of the Centerville and Kingsbury canal were, respectively, 15.63 and 52.35 second-feet per mile. These figures lead, on a basis of canal width of 50 feet, to the enormous seepage velocities of 16 and 52 feet per day, respectively. For a number of miles this

[a] On the underflow and sheet water, etc., by L E. Hicks: Ex. Doc. 41, Pt. III, Fifty-second Congress, first session, Washington, 1892, p. 187.

[b] Seepage or return waters from irrigation: Bull. Colorado Agr. Exp. Sta. No. 33, p. 45 et seq.

[c] Op. cit.

[d] Irrigation near Fresno, Cal., by C. E. Grunsky: Water-Supply and Irrigation Paper U. S. Geol. Survey No. 18, 1898, pp. 76, 77.

canal is within several hundred yards of the edge of the descent to the bottoms, which lie from 20 to 30 feet below the surface of the upland. This location in porous soils and subsoils causes great loss of water by percolation.

The writer has made determinations by the chlorine and other methods (see pp. 48–50) of the rate of movement of the underflow beneath the channel of Arkansas River in western Kansas. Six miles below Garden, at a level of 10 feet below the river bed, the velocity was found to be $2\frac{1}{2}$ feet per day. The fall or gradient of the river bed is about 7 feet to the mile. The material below the 10-foot level is coarser than that above, and the velocity in it is undoubtedly higher, although the determined rate must be considered rather high for a gradient of 7 feet to the mile. South of the island, near Garden, the rate was found to be very high, a motion as great as 12 feet per day being determined. The motion at these points seemed to be entirely in the direction of the thalweg, it being impossible to detect any side or cross motion at the particular points where the experiments were made.

Determinations of the rate of underflow in the narrows of the Hondo and San Gabriel rivers, in southern California, by the author's electrical method, gave rates of $3\frac{1}{2}$, 4, $5\frac{1}{2}$, and 7 feet per day. This work is in progress as this paper goes to press.

Gilbert reports that the underflow sands of the Arkansas River Valley in eastern Colorado have a breadth approximately as great as the bottom lands of the river, and range in depth from 5 to 20 and 30 feet.[a] The slope of the water table is from 7 to 15 feet to the mile. The sands are exceedingly variable in texture. One of the coarser of two samples gave a porosity of 29 per cent. The effective size of the grains is not given by Gilbert, but he states that a saturated sample, if freely drained, would part with about one-third of its water in several days. These data, however, are not sufficient to estimate the effective size of grain very closely, but the writer believes that they indicate that the effective size was between 0.05 and 0.1 mm. If the former, the magnitude of the underflow in the river valley, if the sand be assumed to be 20 feet thick and 50 miles wide, with a slope of 10 feet to the mile, is barely 10 cubic feet per minute. If the effective size be assumed to be 0.1 mm., the underflow would be about 36 cubic feet per minute. A sand as coarse as 1 mm. would, under the same conditions, furnish an underflow of 3,600 cubic feet per minute.

RETURN WATERS FROM IRRIGATION.

Some remarkable instances of changes from the condition of diminishing rivers of arid valleys to the condition of growing rivers of humid valleys have been noted in various parts of the West, these being the results of the irrigation of considerable portions of river lands. Data relating to these changes have been given in various

[a] The underground waters of the Arkansas Valley, by G. K. Gilbert: Seventeenth Ann. Rept. U. S. Geol. Survey, Pt. II, 1896, p. 557.

annual reports of the United States Geological Survey, in connection with the results of river measurements, notably on the Gila, Boise,

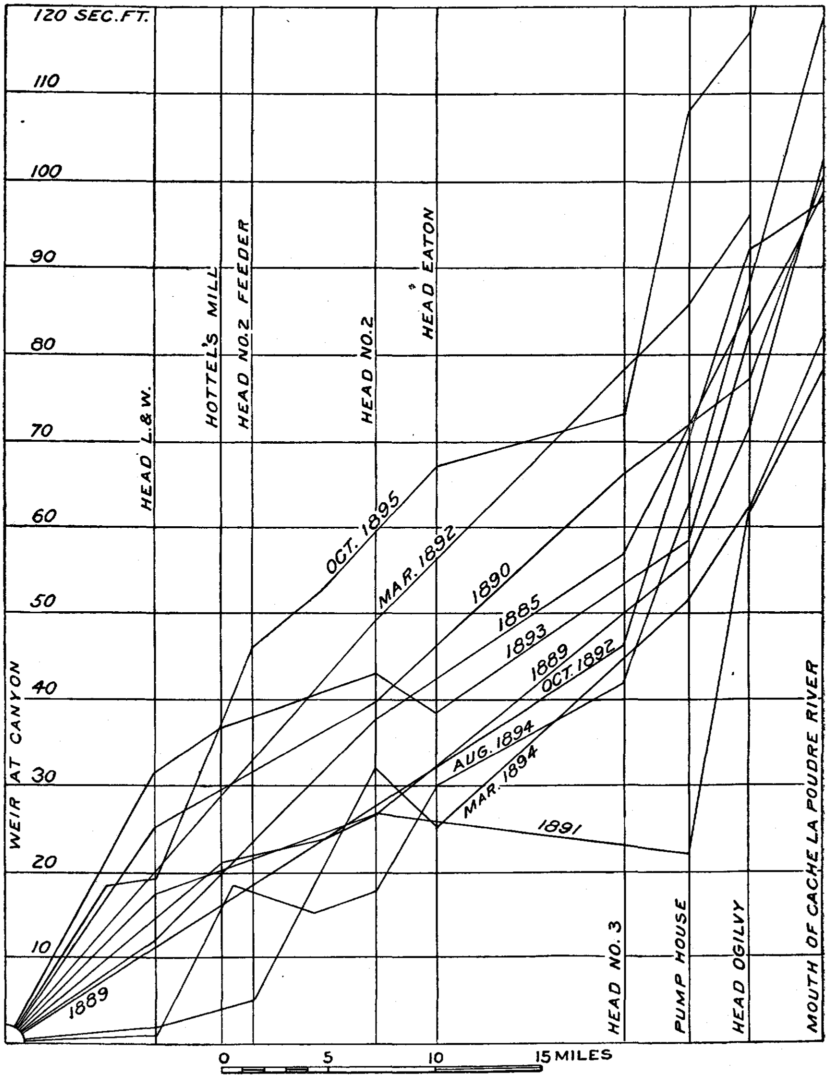


FIG. 16.—Diagram showing growth of Cache la Poudre River. The horizontal distances or abscissæ give the distances in miles from the first weir at the canyon to the point of measurement. The vertical distances or ordinates represent the amount of return waters in cubic feet per second. The distances were measured not along the curves of the river but on the map, taking generally a straight course across the bottoms, because it was thought that the amount of inflow will not be increased by the curves of the river, but, instead, will depend upon the straight course of the stream, other things being equal. It is evident that there is a general agreement between the various lines, notwithstanding certain irregularities, such as the small inflow in the first section in March and August, 1894, and actual losses recorded in the fifth or adjacent sections in 1891, 1893, and 1894.

Platte, and other rivers. One of the most complete series of measurements is that made by Prof. L. G. Carpenter in the valley of Cache la Poudre River, a tributary of South Platte River, Colorado. As the lands of the valley were gradually brought under cultivation one of the effects first noticed was the gradual saturation of the subsoil and the rise of the water table, a phenomenon very common in all irrigated countries, and usually evidenced by a decided rising of the waters in the wells. In many places in the Cache la Poudre Valley the rise in the water table was from 20 to 40 feet. This addition to the water table and increase in the pressure gradient brought about a gradual seepage toward the river and even the creation of flowing springs and marshy areas. The river was changed from a diminishing stream to an increasing one, so that waters used for irrigation upstream returned as seepage waters to the river and were again withdrawn for irrigation farther downstream, until the same water was used several times. A summary of the measurements of this river down to 1896 is presented graphically in fig. 16.

TABLE VI.

Return waters of Cache la Poudre River, Colorado, from measurements made in 1895.[a]

[There are no tributaries between stations, and all measurements were made in October, when water in stream is as low as it usually gets.]

Station.	Water in river.	Drawn out by canals.	Water in river, plus amount diverted.	Ground-water inflow between stations.	Total ground-water inflow.
	Cu.ft.per sec.	*Cu.ft.per sec.*	*Cu.ft.per sec.*	*Cu.ft.per sec.*	*Cu.ft.per sec.*
No. 1	127.609				
		5.498			
No. 2	133.973		139.471	11.862	11.862
		9.485			
No. 3	149.985		159.470	25.497	37.359
		7.453			
No. 4	161.863		169.316	19.331	56.690
		38.955			
No. 5	153.117		192.072	30.209	86.899

The following are Professor Carpenter's conclusions:[b]

(1) There is a real increase in the volume of the streams as they pass through the irrigated sections.

(2) There is no such increase in the streams as they pass through the unirrigated sections. On the contrary, there is an actual loss, even when the drainage of a large area enters.

(3) The increase is more as the irrigated area is greater.

(4) The increase is approximately proportional to the irrigated area, and it seems probable that with more intimate knowledge of the amount of water applied and the features of the drainage the proportions would be found to be close.

(5) The amount of the increase depends very slightly, if at all, upon the rainfall, and so far as it does it is influenced principally by the rainfall on the irrigated lands. Only where the lands are already saturated is the rainfall sufficient to cause seepage.

(6) There is no perceptible underflow from the side channels even where they drain several thousand square miles.

[a] From Engineering Record, Vol. XIV, p. 48.

[b] Tenth Biennial Report State Engineer of Colorado, 1901, p. 210.

(7) The inflow is practically the same throughout the year. It is more in summer, less in winter, principally because of the effect of the temperature of the soil.

(8) The passage of the seepage water through the soil is very slow, so that it may take years for the seepage from the outlying lands to reach the river.

(9) The amount of seepage is slowly but constantly increasing.

(10) It may be expected to increase for some years to come.

(11) An increased amount of land may be brought under cultivation with time, more especially on the lower portions of the streams.

(12) The seepage being nearly constant throughout the year, while the needs are greatest in summer, the use of storage will best utilize the water from inflow.

* * * * * * *

(15) On the Poudre River about 30 per cent of the water applied in irrigation returned to the river.

(16) The use of water on the upper portions of a stream, when water is not immediately needed by prior appropriations, will increase the flow of the stream late in summer and prevent such low stages as it would have without this resulting action.

(17) The seepage water is already an important factor in the water supply for the agriculture of the State. The capital value of the water thus received in the valley of the Cache la Poudre alone is not less than $300,000, and perhaps $500,000, and for the Platte is from $2,000,000 to $3,000,000. It is large for the other streams, but of unknown amount.

(18) An actual loss is incurred in carrying a stream like the Platte through sandy beds.

(19) Ultimately the returns from seepage will make the lower portions of such valleys as the Platte more certain of water and probably enable a larger acreage to be grown.

EFFECT OF TILE DRAINAGE, ETC.

The effect of the return waters from irrigation in increasing the dry-months flow of a stream and making the available supply of water more constant finds its exact counterpart in the effect of the deforesting and cultivation of the land upon the perennial flow of rivers. The changed condition of the streams and the influence of the light flow in the dry months upon the smaller water powers which have usually followed the settlement of the country are too well known to need special comment. Another striking example of the same kind is observed in river valleys in which large portions of the marshes and lowlands have been reclaimed by tile drainage. The tiling of the land permits the percolating waters from the rains to find almost immediate passage into the streams, while under natural conditions the return seepage would require weeks or months to reach the rivers. Part of the falling off in the reliability of the water power at Aurora, Ill., and other points on Fox River must be attributed to the reclamation of thousands of acres of lowlands in the upper part of its valley.

CHLORINE METHOD OF DETERMINING THE RATE OF THE UNDERFLOW.

Continental engineers formerly attempted to determine the rate of movement of ground water by experimental pumping from a number of test wells, deriving the velocity of flow from computations based upon these data. A chemical method of tracing the moving ground water was occasionally used with success. The earliest instance of this nature was probably the successful attempt to trace the origin of an underground vein of water encountered in building a tunnel in Switzerland by charging the ground water with common salt.

A. Thiem introduced a similar method in his investigations of ground-water movements near Leipzig. His method is as follows: In line with the probable direction of motion he digs two test wells at a known distance apart. In the upper well he places a substance which is readily dissolved by water and can easily be tested by chemical analysis. He then takes samples of the water at suitable intervals from the lower well, from the analysis of which he is able to draw conclusions as to the rate of flow. The dissolved chemical penetrates

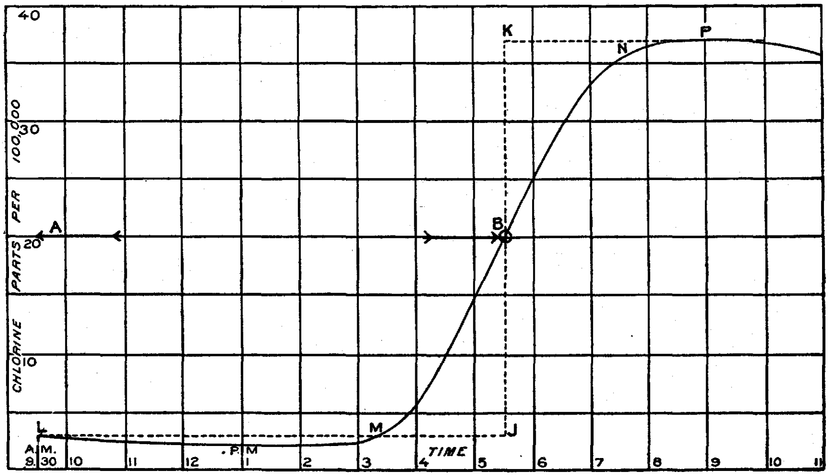


FIG. 17.—Curve obtained in determining the velocity of ground water by the chlorine method. The distance A B represents the time of passage of the ground water between two wells. B is the point of inflection of the curve. The dotted line L J K P is the form that the curve would take if there were no diffusion of the salt. The rounded parts M and N are due to diffusion. German hydrographers have incorrectly used the maximum point P for the determination of the velocity instead of the point of inflection B.

in all directions radially from the place of its introduction in the water-bearing strata, with greater or less velocity, according to the permeability of the ground, and in concentrically decreasing strength of solution. Since the entire body of water during the distribution of the solution moves at the same time in the direction of the second well, the movement of the chemical is greater in half of the circuit of the charged water than in the other half, according as the diffusion takes place with or against the natural current of the ground water. The shorter the distance between the test wells the greater the degree of concentration. The velocity of the ground-water stream is determined by plotting the results of simultaneous analyses from both wells. (See fig. 17.)

Thiem used common salt in his experiments, because it is rapidly soluble and is easily determined chemically; and it has the further advantages of neither poisoning nor discoloring the ground water. He used this method in the new water-supply systems at Greiswald and Stralsund with good results.[a]

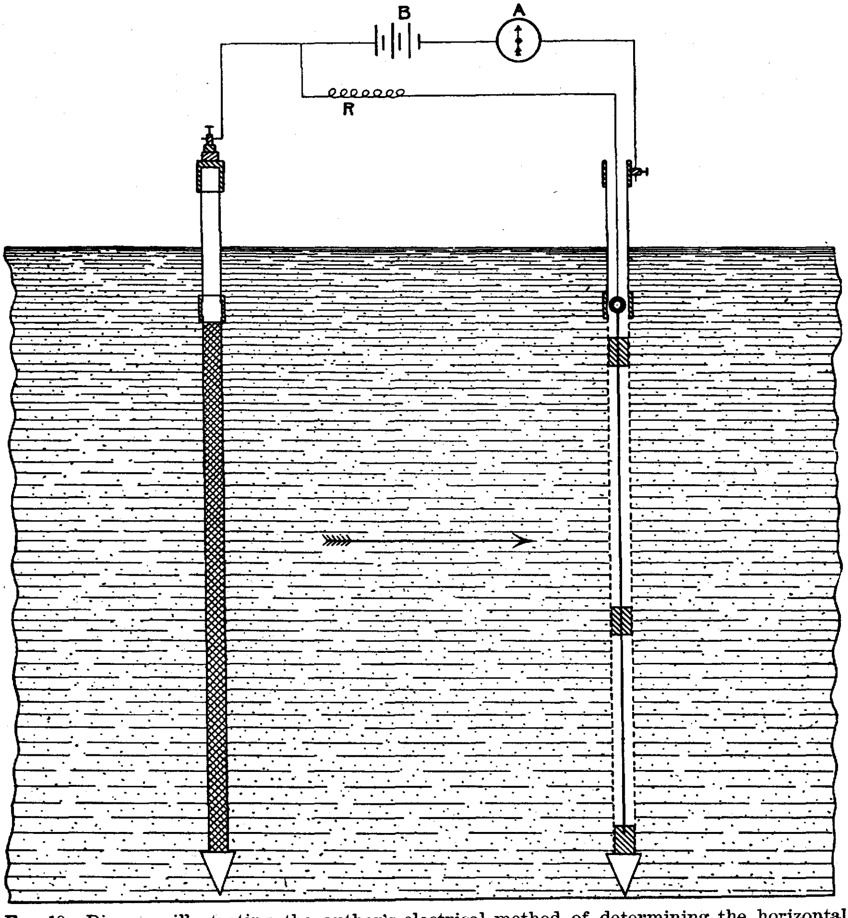


FIG. 18.—Diagram illustrating the author's electrical method of determining the horizontal velocity of ground water. The ground water is supposed to be moving in the direction of the arrow. The upstream drive well is charged with an electrolyte. The gradual motion of the ground water toward the lower well and its final arrival at that well are registered by the ammeter A. B is the battery and R a suitable resistance.

THE AUTHOR'S METHOD OF DETERMINING THE RATE OF THE UNDERFLOW.

A more rapid and less laborious method of surveying underground currents of ground waters than the chlorine method becomes almost imperative when extensive surveys are undertaken. Since July, 1901, the writer has given much attention to the development of a new

[a] Verfahren für Messung natürlicher Grundwassergeschwindigkeiten, by A. Thiem: Polyt. Notizblatt, Vol. XLII, 1887, p. 229.

method intended to be especially applicable to the survey of the underflow streams of the Plains. Acting under the authority of the chief hydrographer of the United States Geological Survey, he made preliminary field tests along the valley of Arkansas River in western Kansas during the month of August, 1901, and has continued the work in the laboratory and field since that time. The experiments seem to leave little to be desired either in the certainty of the results or in the rapidity with which work can be prosecuted. The method is an electrical one. A double row of 1¼-inch drive wells is sunk across the channel of the river whose underflow is to be tested. The upstream wells are then charged with a strong electrolyte, which dissolves and passes downstream with the moving water. (See fig 18.)

The passage of the electrolyte toward the lower well is shown by the gradual movement of the needle of an electrical instrument, and

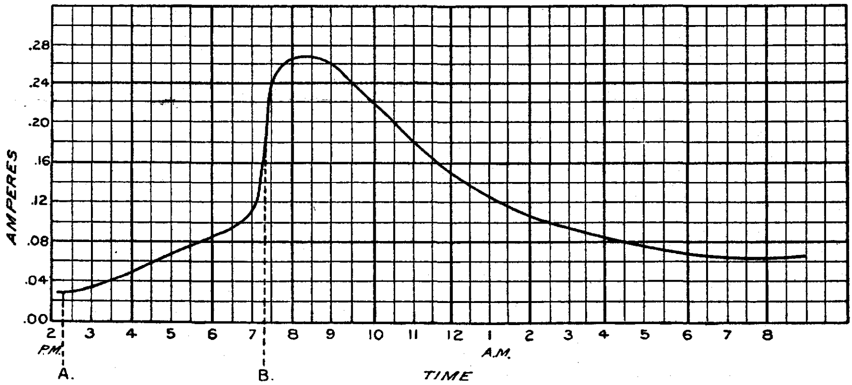


FIG. 19.—Curve obtained by the author's electrical method of determining the velocity of ground water. The distance A B represents the time of passage of the ground water from the upper to the lower well. The sudden deflection at B shows the final arrival of the electrolyte at the lower well. The point B should be taken at the point of inflection of the curve and not at the highest or maximum point. If the point of inflection be taken the effect of the diffusion of the electrolyte will be nullified.

the final arrival at the well is shown by a sudden and strong deflection of the needle. (See fig. 19.) It is exceedingly interesting to watch the gradual movement of the water in this indirect way. In an hour or two a very good indication of the rate can be obtained by noting the slope of the current curve obtained. If several of a row of wells are being used simultaneously, the main stem of the underground current will show itself by the rapidly rising electrical currents corresponding to certain of the wells. Thus not only are the tedious chemical analyses of the chlorine method obviated, but the actual movement of the water is traced from the beginning of the experiment, which is impossible in the older methods.

Numerous electrolytes have been tried, both in the laboratory and in the field. It is obvious that the electrolyte must possess the following properties: (1) It must be readily soluble in water; (2) it must be chemically inactive to the dissolved matter in natural waters or to the material of the pervious medium; (3) it must possess a low coefficient of diffusion; (4) it must possess a high conductivity; and (5) its cost must be low. Of the numerous salts tried up to the present time ammonium chloride has given the best results. With this salt a current of sufficient intensity to throw the needle of a recording ammeter can be obtained by the use of a few dry cells.

The electric circuit to the wells can be made in various ways. A brass rod insulated from the well tube and lowered into the downstream well by means of a rubber-covered wire may serve as one electrode, the casing of the same well constituting the other electrode. In this case the indication of the movement of the ground water will not be noted until the electrolyte has reached the lower well, where its presence will be shown by a sharp rise in the current curve. Instead of this the two wells themselves may be used as electrodes, in which case the gradual passage of the electrolyte downstream can be observed from the beginning, but the final indications of its arrival at the lower well are less pronounced. The best method is a combination of these two: Run the wire from the casing of the lower well to one pole of the battery, with the ammeter in circuit, and connect the other pole of the battery both to the internal electrode of the lower well and to the casing of the upper well. In this case the gradual movement of the ground water from one well to the other is shown on the ammeter chart, and an abrupt indication of the arrival of the electrolyte at the lower well is also recorded.

The drive wells may be made of perforated sections from 4 to 8 feet long, so that a well will offer free passage to the water throughout its entire length. The internal electrode may be made in several insulated links with separate external wires, so that varying velocities at different depths can readily be determined.

While making tests in the underflow of Hondo River, California, the author was able to determine the different velocities in three different strata of gravel penetrated by the wells by a single setting of the wells. The curve developed three well-marked points of inflection, corresponding to the velocities in the several strata, and the brass screens on the well point and well-point extension were blackened by the electrolyte at depths corresponding to the three strata.

Underflow measurements were made by the author on Mohave River at the site of the proposed dam near Victorville, Cal. At this point the river passes through a narrow gorge, barely 120 feet wide at the river surface. Here the greatest depth to bed rock is 47 feet, and the material is a very coarse granite sand or fine gravel. A double row of test wells across the narrows, with extra downstream wells for the determination of the direction of flow, gave velocities at various points of 6, 8, 20, 35, 48, and 64 feet per day. Notwithstanding these high rates the total volume of underflow is quite small on account of the limited cross section (4,160 square feet). For a cross section at the dam site see Eighteenth Ann. Rept. U. S. Geol. Survey, Part IV, page 709. Estimating a porosity of 33 per cent and an average velocity of 50 feet per day gives a total underflow of 0.8 second-foot, or 520,000 gallons per twenty-four hours.

The velocities at this point were so high that the electrolyte passed between wells only 18 inches apart without touching them, and a small quantity of caustic potash had to be added to spread the electrolyte into a broad stream.

CONCLUSIONS.

The important facts concerning the underflow of rivers are brought together in the following summary:

(1) In the mountainous portions of streams, where the river slope is great and the material deposited in the river channel is coarse, the underflow in the direction of the thalweg may be relatively very large.

(2) The seepage in the steeper slopes and terraces of a river valley is primarily in a direction toward the thalweg and not along it.

(3) The underflow in the alluvium beneath the surface of the bottom lands of the valley is greater in the direction toward the channel than in the direction of the thalweg, except (*a*) where the water table has a greater slope downstream than toward the channel, and (*b*) where the fine silt of the river channel covers deeper deposits of coarser material, rendering seepage into the channel difficult and underground passage downstream easy.

(4) The velocity of the underflow is very small and the total amount is commonly greatly exaggerated.

(5) The rate of the underflow is especially small in the beds of overloaded, silt-depositing streams, although the storage capacity in such cases may be correspondingly great.

CHAPTER III.

DEEP ZONES OF FLOW.

The sedimentary rocks usually occur in alternating layers of sandstone, shale, and limestone, or in alternating successions of sandstone and shale. These classes of rocks represent the various kinds of deposits which are laid down on the floor of the sea. Originally the sandstone was sand, deposited near shore in the more rapidly moving water. The shale was mud, deposited in the belt of sluggish currents in shallow water. The limestone was limy ooze, the remains of lime-secreting organisms upon the deeper and quiescent floor of the sea. Sand and mud are therefore called mechanical sediments and the calcareous ooze is called an organic sediment.

Water is, then, not only the transporter of the material worn from the shore and the surface of the land, but it is also the sorter of that material. The gradual change in the shore line by erosion and by alternating elevation and subsidence of the land has caused rocks of these various classes to be deposited above one another in a more or less regular threefold succession, shale above sandstone and limestone above shale. The various strata are usually not arranged, therefore, like the leaves of a book, in perfectly horizontal and coextensive layers, but the arrangement is better likened to the overlapping of the shingles on the roof of a house, sediments of one class being imbricated with those of the other classes. The higher ends of the strata correspond to the deposits laid down near the shore, unless earth movements have inverted the direction of the original dip.

The common alternation of sandstone, shale, and limestone found in the sedimentary rocks is a fact of the greatest importance in the consideration of the motion of underground waters. The sandstones are in general the most pervious of all water-bearing rocks, and the shales are usually the most impervious. Generally limestone is to be classed as an impervious rock, although frequently it is a water-bearing rock of great importance.

Important classes of pervious limestones are the following: (1) Chalk and partially consolidated shelly limestone, such as coquina, etc.; (2) limestone that has at one time been exposed in the belt of weathering (The solvent action of surface waters upon limestones and the readiness with which they are rendered cavernous and prolific bearers of water are too well known to require special comment.); and (3) deeply buried limestone in which a portion of the carbonate of lime has been replaced by dolomite or carbonate of magnesium. The dolomite being less bulky than the chemical equivalent of calcium carbonate, the substitution may render the limestone sufficiently porous to become an important water-bearing rock. A well-known example is the Trenton limestone in Ohio, which bears enormous quantities of water, oil, and natural gas in its upper dolomite layers.[a] The substitution of basal magnesium for calcium may in itself account for a porosity of from 10 to 12½ per cent.

The sandstones are not only confined by impervious material above them and often below them, but their exposures above the surface of the land are normally covered with a very pervious, coarse, sandy soil which readily imbibes the rainfall. Thus all of the elements are present for the storage and circulation of underground waters in the zones of the sedimentary rocks far below the surface zone of flow.

DISTINGUISHING FEATURES.

The pervious and water-bearing sandstones and limestones beneath the surface zone of flow constitute what we have called the deeper zones of flow. There may be several of these deeper zones, or they may be absent altogether. When present, they are distinguished from the surface zone of flow by the following characteristics:

(1) The surface zone of flow has a free, unconfined upper boundary (the water table) and an impervious lower boundary. The deep zone of flow has an impervious upper boundary as well as an impervious lower boundary.

(2) The unit of the upper zone of flow is the drainage area or river valley. The unit of the deep zone of flow is regional and geologic and not dependent upon surface contours. However, it must not be forgotten that the deeper geologic structure is frequently the principal determining factor controlling the surface drainage, so that the deep zones of flow do not commonly run counter to the direction of the surface flow.

(3) The surface zone of flow is dependent upon the local rainfall of the immediate region. The deeper zones of flow receive their waters from distant areas.

(4) The surface zone of flow is in part above the level of surface-drainage channels, while the deeper zones are entirely below the local drainage level.

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[a] For more complete discussion see Rock waters of Ohio, by Edward Orton: Nineteenth Ann. Rept. U. S. Geol. Survey, Pt. IV, 1899, p. 640.

(5) There is commonly a difference in the chemical composition of the waters from the two zones. It is difficult in our present state of knowledge to make valuable generalizations. The waters of the surface zones are usually less mineralized than those of deep strata, but in arid regions this general rule is frequently reversed. The carbonates are the predominant salts of the surface waters. The deeper waters are usually characterized by rather high amounts of dissolved chlorides. Waters of the surface zone contain dissolved oxygen gas, which is almost entirely absent from the deep waters.

Some of these distinctions (1, 2, and 4 above) are illustrated by the accompanying ideal cross section through South Dakota (fig. 20). This section is along a line from Sioux Falls to Chamberlain and thence along the thalweg of White River to the Black Hills. The diagram is based upon similar drawings published by Darton, Todd, and others, but without attempt to show with absolute accuracy the geologic details. The deep zone of flow is approximately from west to east, through the porous strata marked Dakota sandstone, the water entering at an elevation of more than 3,000 feet, at B. The impervious upper layer is the Benton shale or clay. This cross section presents considerable variety in the surface zones of flow. Between C and D the upper zone of ground water drains from east and west into Cheyenne River, which here flows north. From D to E the surface ground waters drain from north and south into White River. From E to F is the valley of the Missouri. East of this river the surface of the land is covered with glacial drift, represented by the light surface shading in the diagram. At G is James River; at H and I are the forks of Vermilion River, and at J is Big Sioux River. These streams flow to the south, so that the surface ground-water seepage is substantially east and west.

PERCOLATION INTO DEEP ZONES.

The rainfall finds its way into the deeper zones of flow in various ways, either directly by percolation into the exposed ends of the pervious strata, or indirectly by the seepage from streams and rivers which have cut their valleys through the outcrops. The water must leave the deep zone at a lower level than that at which it enters, and hence the outlets for the water must be sought in the lower outcrops of the rock, where, perhaps, a river has eroded a channel through it, or where the impervious cover is locally absent. There is also always the possibility of the escape of water through joints and faults, and even by seepage through the covering strata, for imperviousness as applied to rocks is merely a relative term. The covering layer of shale or hard rock has an enormously greater area than the vertical transmitting section of the pervious rock, so what the covering rock lacks in permeability it can make up in area.

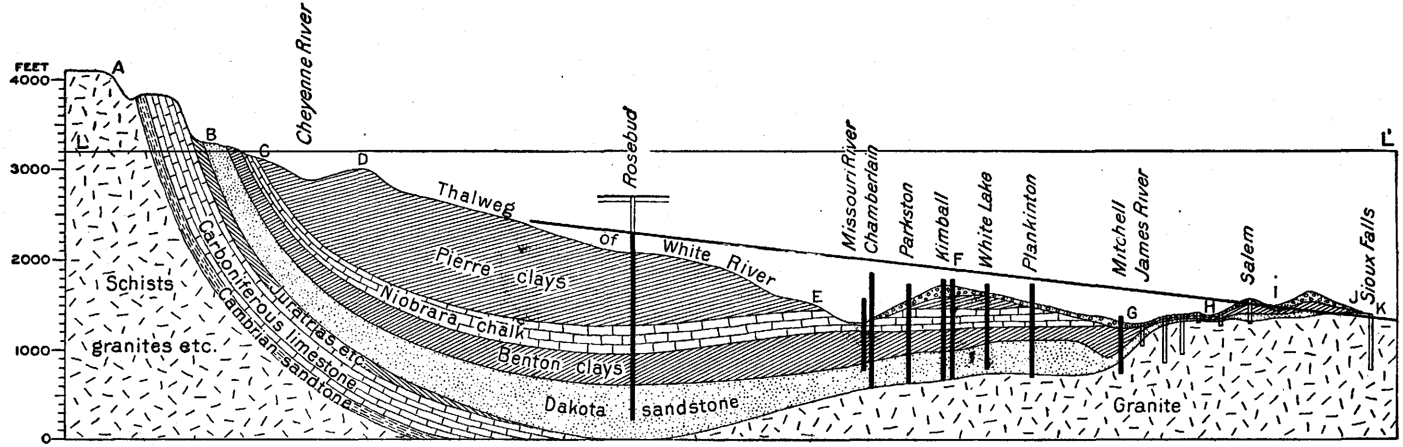

FIG. 20.—Cross section through South Dakota artesian basin, based upon published data of Darton, Todd, and others. The section is along a line through Sioux Falls and Chamberlain and thence along the thalweg of White River. All of the wells shown are within a few miles of this line, excepting the Rosebud well, which is 25 or 30 miles from the line and on much higher land. The base of altitudes is the sea level. The height to which the water would rise in a closed tube at the various wells is shown by the heavy black lines. In studying the drawing one should remember that it has been necessary to use a scale which greatly exaggerates vertical, in comparison with horizontal, distances.

The ratio of the cross section of the Potsdam sandstone to the area of the covering strata is certainly no less than 1: 3,000, for the average thickness of the Potsdam is about 700 feet, while the formation is known to exist as a water-bearing rock for at least 400 miles south of the catchment area. The average thickness of the Dakota sandstone is approximately 200 feet, while the distance from the catchment area in the Black Hills to the eastern outcrop at Yankton is about 350 miles, so that the ratio of the area of the transmitting cross section of the porous strata to the area of the covering strata is about 1: 9,000. Thus if, on the average, the covering material had in the one case only $\frac{1}{3000}$ and in the other only $\frac{1}{9000}$ of the transmission capacity of the water-bearing stone, there would still be as great opportunity for the upward escape of the water as for its onward movement. This upward seepage would cause motion of the water in the porous strata, so that motion is possible even if there be no actual outcrop of the transmitting rock.

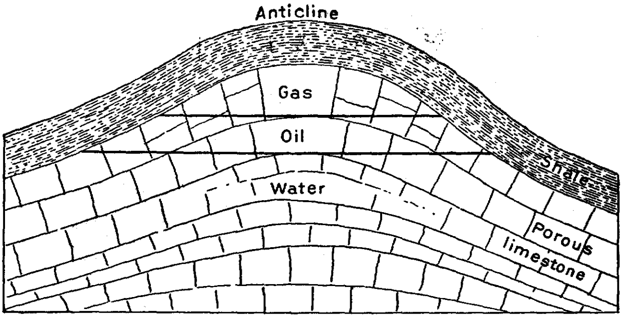

FIG. 21.—Diagram illustrating how gas and oil have been retained beneath anticlines of shale during protracted geologic intervals, thus proving the substantial imperviousness of the covering.

Local upward seepage may explain, perhaps, some of the anomalous variations in pressure of underground water in artesian basins. It is certain, however, that upward leakage is entirely absent over large areas. In the Ohio oil regions, for example, all of the oil and natural gas, not being continually supplied to the rock as water is, would have escaped ages ago if we suppose the most minute permeability to the overlying strata of shale. The gas and oil are held under the anticline of shale as shown in fig. 21, just as air may be held under a basin inverted in a tub of water. It would seem equally reasonable to suppose that the briny water which underlies the petroleum has occupied its present position during the same geologic interval as that in which the petroleum has been impounded. While this may be true in special instances, it is far from being the necessary conclusion, for the water on which the oil floats is a continuous body, and is free to flow toward an outlet in a distant portion of the formation, while the oil and gas, confined by their low density to a particular anticline, can not escape so long as their own covering holds, even though a neighboring anticline be pervious and leaky.

The following water-bearing formations of wide extent and of high economic importance may be specially mentioned as good illustrations of deep zones of flow:

(1) The Dakota formation, commonly referred to as the Dakota sandstone, outcropping in the Black Hills and in the foothills of the Rocky Mountains, and underlying the Dakotas, Nebraska, and Kansas, and of unknown extent northward.

(2) The Potsdam sandstone, outcropping in a V-shaped area in central Wisconsin, dipping southward under Illinois, Iowa, Indiana, and Ohio at great depths, and outcropping again in the great uplift which has formed the Rocky Mountains.

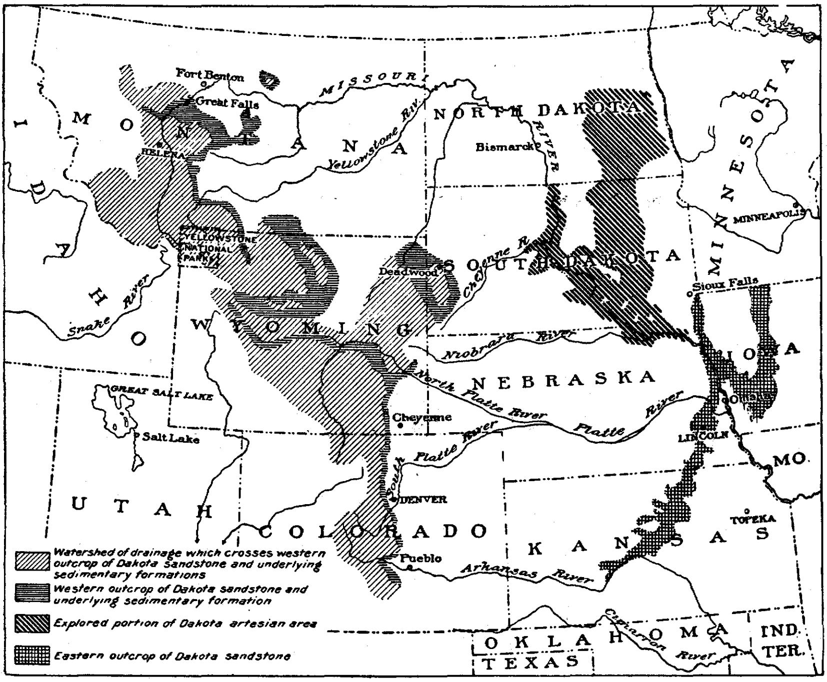

FIG. 22.—Darton's map of catchment area of the Dakota and the Dakota artesian basin.

(3) The Atlantic Coastal Plain deposits, extending along the eastern margin of the Alantic slope from Long Island to Texas, and about 100 miles in width. These consist of a succession of unconsolidated deposits consisting mainly of sands, clays, and marls, instead of sandstone, shale, and limestone, which lie on the east-sloping floor of the crystalline rocks. The western edge has an altitude of from 300 to 400 feet above the sea.

DAKOTA FORMATION.

Darton's map (fig. 22) of the outcrop of this formation in the Black Hills and in the foothills of the Rockies shows the source of the supply of water to the formation so far as known. The map shows not only the surface outcrop, but also the drainage area that is tributary

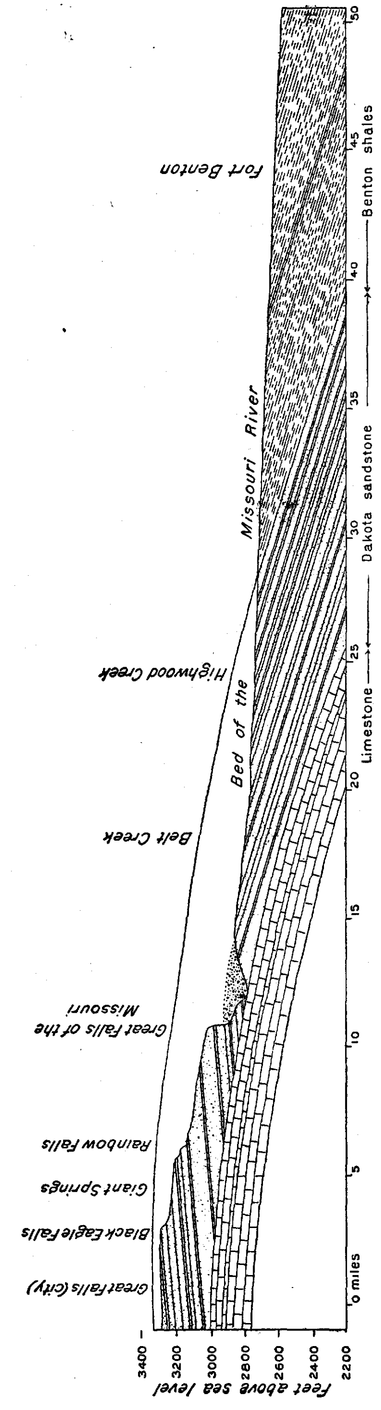

FIG. 23.—Section along Missouri River near Great Falls, Mont. It here appears that Missouri River flows across the Dakota formation, in which it loses a large amount of water.

to it. Darton gives the approximate area of these outcrops of permeable beds as follows: Black Hills, 2,400 square miles; Bighorn Mountains, 7,000 square miles; Rocky Mountain foothills in Montana, more than 3,000 square miles; Rocky Mountain foothills in Wyoming, about 1,500 square miles; total, 13,900 square miles.

The Dakota sandstone in the Black Hills receives water as follows:

(1) Of the rain which falls on the Dakota sandstone probably one-fourth is directly absorbed by this rock.

(2) All the drainage of the Hills crosses the Dakota sandstone, which therefore receives a second accession.

(3) A probable third accession is from the Carboniferous group which underlies the Dakota sandstone. Many streams of the Hills either sink or lose considerable water in the lower beds of the Carboniferous. Inasmuch as this group thins out and is apparently absent in the eastern part of the basin, it is probable that a portion of this water is passed along to the Dakota sandstone.

Missouri River flows over the outcrop of the Dakota sandstone in passing from Great Falls to Fort Benton, Mont., as can be seen from consulting Darton's map (fig. 22) and the accompanying section (fig. 23). In this part of its course the river is believed to lose a considerable volume of water.

The elevation of the outcrop of the Dakota sandstone in the Black Hills is from 3,100 to 3,800 feet above sea level. The outcrop farther west, in the foothills, reaches an elevation much higher, as can be seen by consulting fig. 22. The sandstone comes to the surface again in southeastern Dakota, near Yankton, and along Missouri and Big Sioux rivers, at an elevation of about 1,200 feet. In this area there is evidence of leakage from the sandstone, probably greatest in the neighborhood of Yankton. At Chamberlain, Running Water,

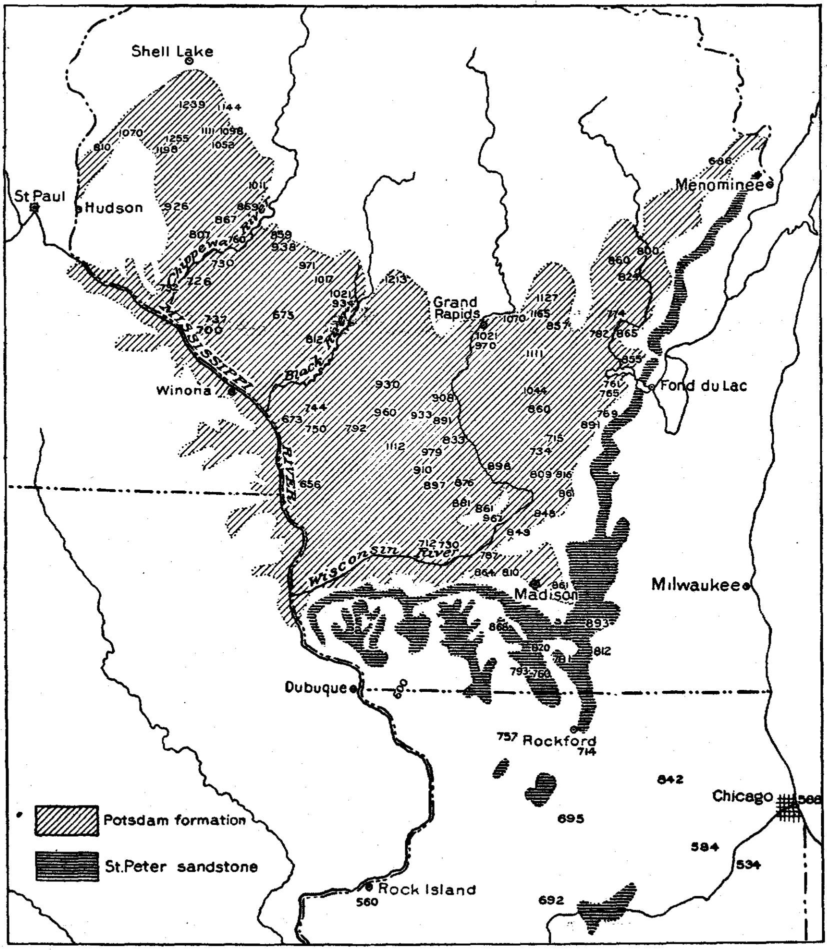

FIG. 24.—Wisconsin outcrop of Potsdam and St. Peter sandstones. Figures indicate height, in feet, above mean sea level.

and Randall, springs are seen rising from the bed of Missouri River. Through extensive areas in Dakota this subterranean water is under enormous pressure and constitutes one of the most remarkable artesian areas in the world. More will be said of this area later.

POTSDAM FORMATION.

The Wisconsin outcrop of this water-bearing series is shown in fig. 24. The altitude of the V-shaped outcrop is in the neighborhood of from 900 to 1,000 feet above sea level. The soil above the outcrop is sandy and porous over extensive areas, making the catchment area especially favorable for the imbibition of water. The area of the outcrop is nearly 12,000 square miles, and the thickness of the formation reaches a maximum of nearly 1,000 feet. The strata of the series are mostly porous sandstones, with occasional layers of shale and some beds of limestone. The Potsdam is overlain by the Lower Magnesian limestone, varying in thickness from 65 to 250 feet. Overlying the Lower Magnesian is another water-bearing sandstone of great importance, known as the St. Peter. It varies in thickness from 212 feet to a fraction of a foot, the average being probably between 80 and 100 feet. The rock is of an exceedingly friable character, being, as a rule, hardly compact enough to permit handling. Overlying the St. Peter is the Trenton limestone. The Potsdam and its overlying strata dip gently away from the V-shaped outcrop.

The water-bearing sandstones of these formations supply artesian water of exceptional purity to large areas in southern Wisconsin, northern Illinois, and northeastern Iowa. The head of the flowing wells is not high anywhere within the area, and at many places the wells have to be pumped. This artesian basin is remarkable on account of its large areal extent and the excellent quality of the water rather than the large flow of individual wells. The usual yield of carefully constructed wells varies from 50 to 300 gallons a minute.[a]

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[a] A tabulation of the data from wells in Illinois will be found on page 810 of the Nineteenth Annual Report of the United States Geological Survey, Part II; see also Vol VI, Iowa Geological Survey.

CHAPTER IV.

RECOVERY OF UNDERGROUND WATER FROM THE SURFACE FLOWS.

It has already been explained that underground water comes to the surface and joins the surface waters by diffused seepage into rivers, lakes, and marshes, and into the bed of the sea. It has also been pointed

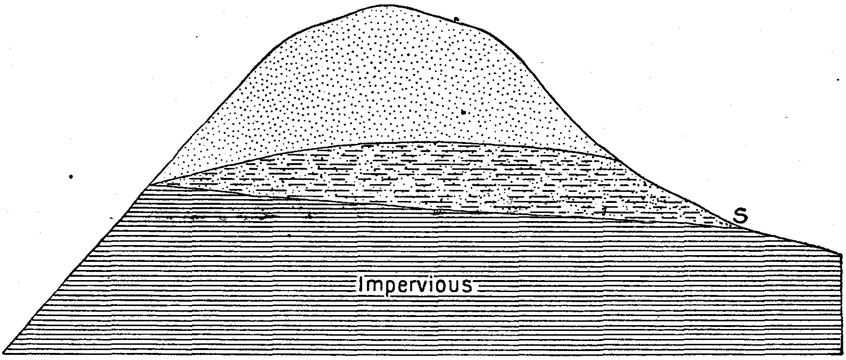


Fig. 25.—Diagram showing the formation of a stratum spring at an outcrop of an inclined impervious stratum, as at S.

out that if the return seepage, owing to unusual geologic conditions, is strongly concentrated in special areas, the returning water constitutes a spring or a natural fountain. A surface spring is usually

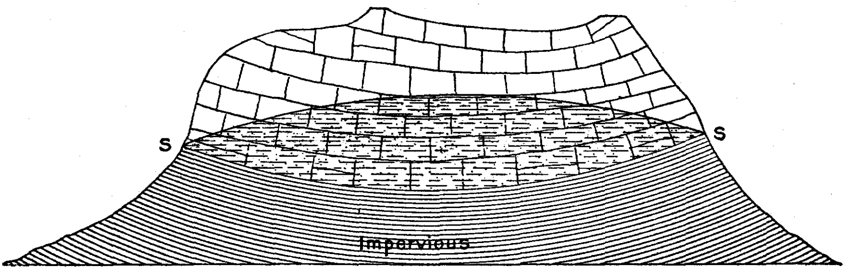


Fig. 26.—Diagram showing the formation (at SS) of overflow springs at the edges of a basin of impervious material.

associated with an outcrop of an impervious stratum. The simplest form is shown at S, fig. 25, which, if we follow the classification of Haas, may be called a stratum spring. A fault in the bed rock may force the ground water to the surface, forming a so-called fault spring, as at "Spring B," fig. 1, page 13. Other forms are overflow springs, as at SS, fig. 26; chasm springs, as in fig. 27, or valley springs, as at "Spring A," fig. 1, page 13. A stratum spring is shown in Pl. IV, *A*.

COMMON OPEN WELLS.

Besides these natural processes which are constantly returning underground waters to the surface waters, there are several well-known ways of artifically restoring ground waters to the surface. The most common of all methods is, of course, the construction of an open well and the raising of the water to the surface by appropriate mechanical means, as shown in fig. 28. This is so universal a practice that it requires but little discussion. The well is simply excavated to a suitable depth below the water table, so that the ground water of the surface zone of flow is free to flow into the excavation.

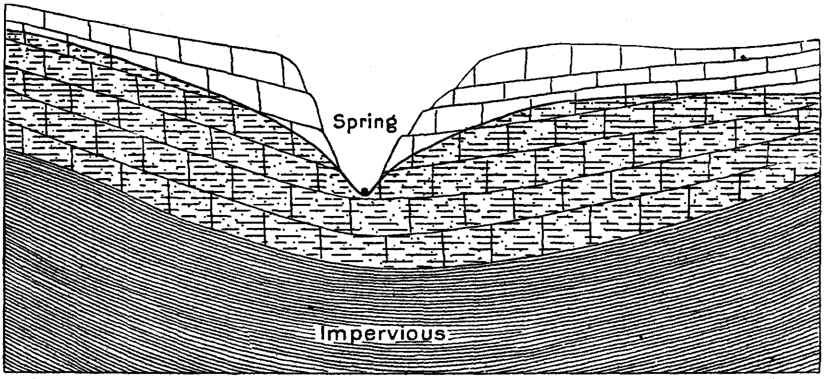
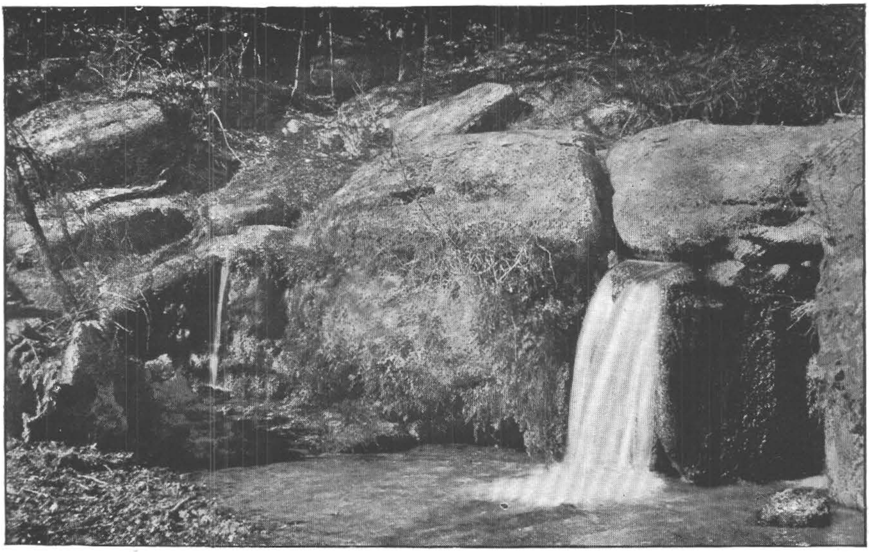


FIG. 27.—Diagram showing the formation of chasm springs in a cut which extends below the natural level of ground waters.

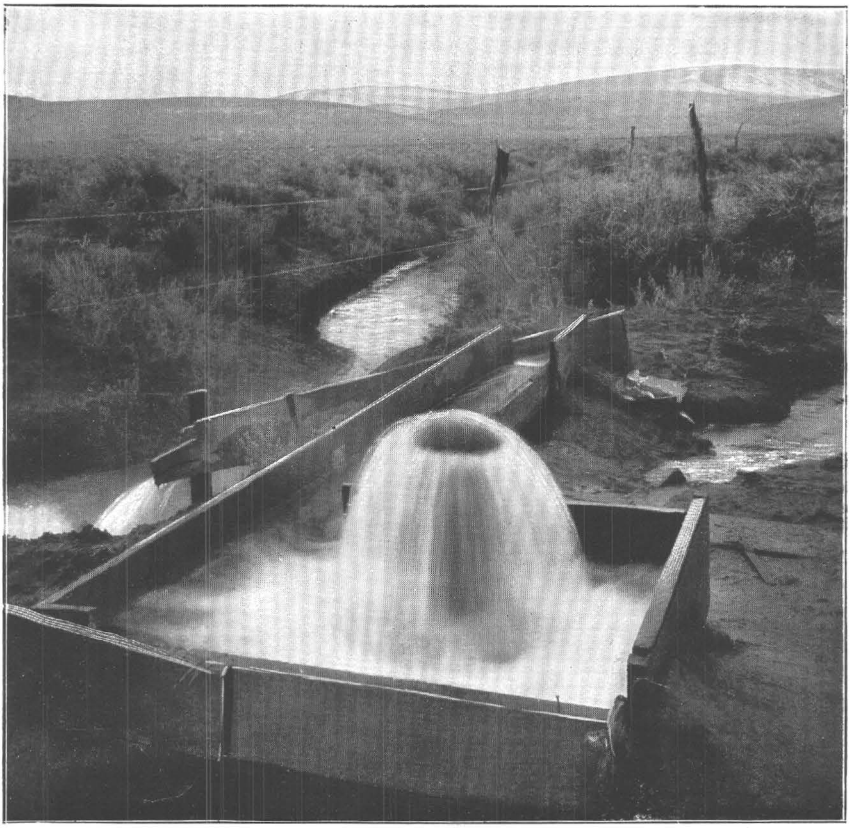
The result of the regular removal of water from the well is the lowering of the water table in its neighborhood, the water table assuming the very regular curve shown in figs. 28, 29, and 31. The form of this surface can be worked out from theoretical considerations, and the results seem to agree substantially with experiment. These figures (29, 30, and 31) are taken from the report of a very carefully conducted series of experiments by J. C. Hoadley, C. E., of Boston, Mass.[a] The investigations are known as the Melrose or Malden experiments, because the first series was carried out in the town of Melrose, near the boundary of Malden, and on the border of the uncultivated tract known as Middlesex Fells.

Fig. 29 shows the varying position of the water table on May 24, 1882, near a 3-inch pipe, during ten hours of continuous pumping at

[a] See Sanitary Engineer, Vol. XI, 1884, pp. 11, 35, 60, 84, and Proc. Soc. Arts, Boston, 1882-83, p. 115.



A. SPRING FORMED AT OUTCROP OF AN IMPERVIOUS FLOOR.



B. ARTESIAN WELL.

a mean and nearly uniform rate of 2,313 gallons an hour. The pumping was from a $1\frac{1}{2}$-inch suction pipe dropped loosely into the well. There had been no pumping from the well for seventy-five hours preceding the experiment. The head required to force the water into the $1\frac{1}{2}$-inch pipe with open end (no holes) and to overcome bends and friction in pipe was 29.5−20.75=8.75 feet. The barometer stood at

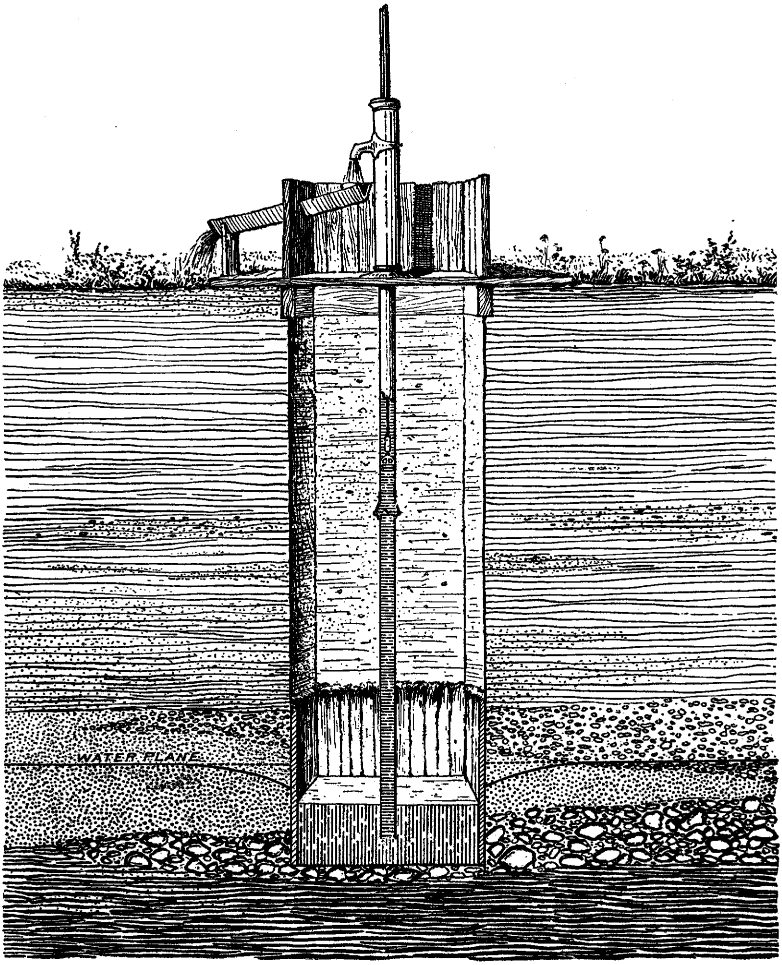


FIG. 28.—Common dug well; shallow in water.

30.015 inches, the thermometer at 67.6° F. The height of partial vacuum in suction pipe, as indicated by vacuum gage, was 20.75 feet above base; height of absolute vacuum, as indicated by vacuum gage above the base line, was 6.75 feet; height of valve seat of steam pump above base, 40.80 feet. Fig. 30 shows the effect of pumping, of stopping the pump, and of rain, upon the height of ground water in one of the wells. The pumping was at the rate of 2,010 gallons an hour, except for several stops indicated on the diagram.

Fig. 31 illustrates a very excellent apparatus designed by Hoadley for the experimental study of well phenomena in the laboratory. The

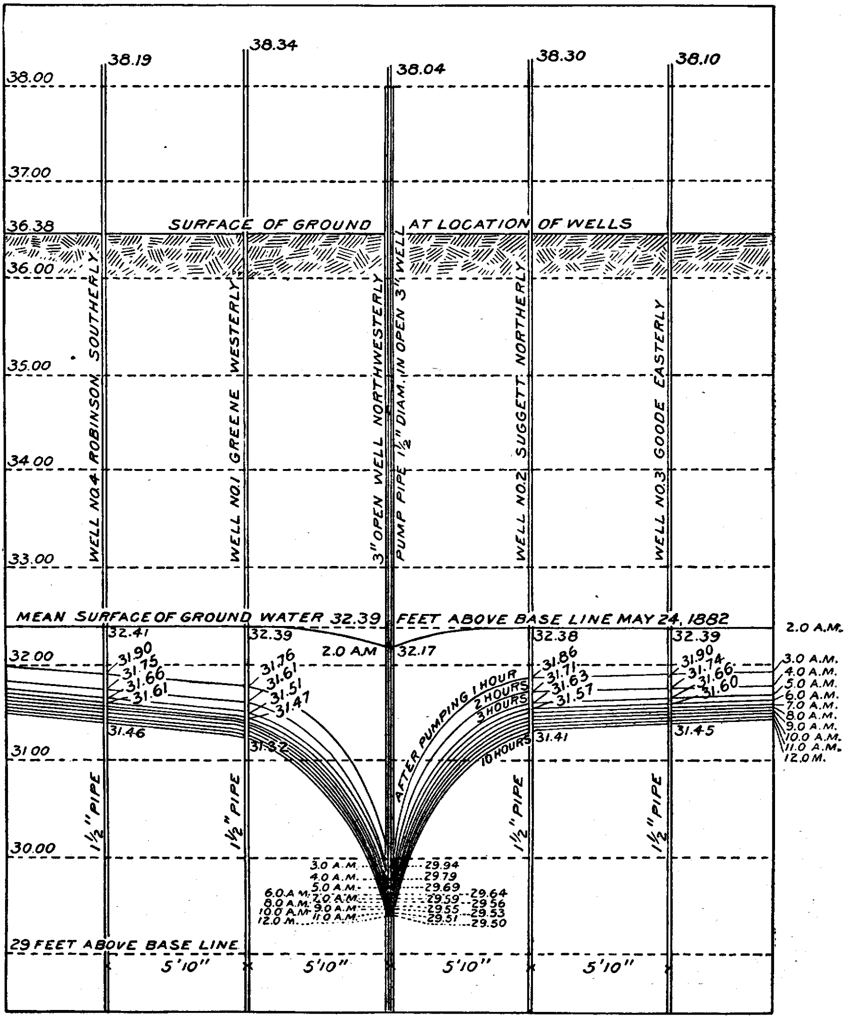


Fig. 29.—Diagram illustrating the changing position of the water table near a 3-inch well during ten consecutive hours of continuous pumping at an average rate of 2,313 gallons an hour. (After Hoadley.)

sand is held within a 48-inch cylindrical curb of wire cloth, so placed in a larger tank as to leave a 1-inch space all round, which can be kept filled with water to supply the sand during experimental pumping from a well at the center of the cylinder. Numerous brass tubes drilled with fine holes occupy positions in the sand at various distances from the central well and communicate with appropriate gages of glass tubes, from which the position of the ground-water surface in the sand can be determined at any time, as shown in the lower part of fig. 31.

The capacity of surface wells depends upon several factors. Perhaps the most important is the degree of fineness of the material of the water-bearing stratum. The size of the soil grains not only controls the rate at which water can be transmitted to the well, but also determines the proportion of the contained water which the soil will freely part with. The fine-grained soils retain a considerable proportion of the water of saturation as capillary water, even after free means of

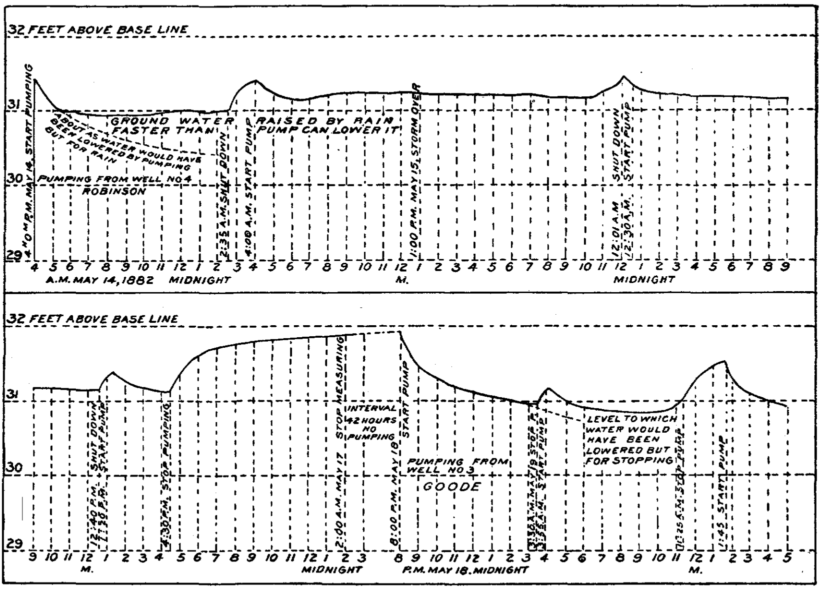


FIG. 30.—Diagram illustrating the varying position of the water table in a well due to pumping for various intervals and modified by rainfall and stops of various duration. (After Hoadley.)

drainage are established, so that fine-grained material will not only deliver water slowly but will furnish only a small total amount. The yield of the well also depends upon the amount that the water table is lowered by the withdrawal of water, or the head under which the flow takes place, and upon the size and shape of the excavation and the character of the walls and casing. Theoretical considerations seem to show that the yield of a well is directly proportional to the transmission constant (k of formula on page 26 of this paper) of the porous material, and nearly proportional to the head, or the distance the water table is lowered. If the well is deep and relatively small in diameter (say from 6 to 12 inches), increasing the diameter seems to have little effect on the volume, doubling the diameter adding barely 8 or 10 per cent to the theoretical yield; while if the well is shallow and porportionately large in diameter the yield seems to be proportional to the diameter, so that doubling the diameter would double the flow. This dependence of yield upon diameter is only true when the well is being used to nearly its full capacity. When the well is drawn upon for but a small percentage of its full capacity the indications are that doubling the diameter will have a greater effect no the capacity of the well than these figures indicate.[a]

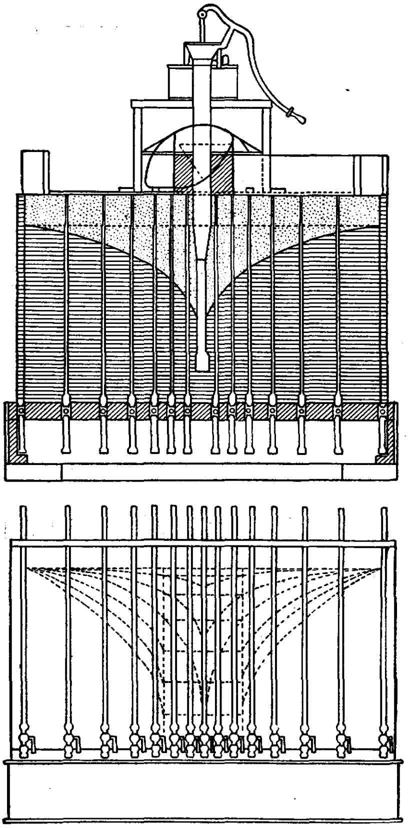


FIG. 31.—Hoadley's laboratory apparatus for the investigation of the phenomena of common open and driven wells.

As already stated, the flow of ground water into a well, if it be not too shallow, varies directly as the distance the surface of the water in the well is lowered by pumping. Thus, if the water in a well is lowered 2 feet below the natural level by pumping from it at the rate of 20 gallons a minute, the same well may be expected to yield approximately 40 gallons a minute if the water is lowered 4 feet below the natural level. For shallow wells the yield will not increase in this direct ratio, but will be considerably less, on account of the decrease in percolating surface, due to the lowering of the water table in the neighborhood of the well. Besides the advantages just mentioned, tubular wells, owing to their greater depth, are much more likely to strike a vein of coarse material, a small stratum of which may be expected to furnish much more water than a considerable depth of fine material. This accounts for the well-known superiority of deep tubular wells over common dug wells. When a large supply of water is required, as for irrigation or for village supply, it is a common practice to sink several tubular wells in the bottom of a large dug well (see fig. 32). In this way not only is a large supply obtained, but the large well acts as

[a] See the author's theoretical investigation of the motion of ground wateis: Nineteenth Ann. Rept. U. S. Geol. Survey, Pt. II, 1899, pp. 358, 362, 364.

a storage reservoir, equalizing the load upon the pumps and permitting temporary overdrafts of the capacity of the wells. Some careful meaurements of the rate of rise of the water surface in a well after pumping has ceased have been made by Mr. Willard D. Johnson. Data from the city well at Garden, Kans., as obtained by him, are shown in figs. 33 and 34. The curve in fig. 34 agrees very closely with

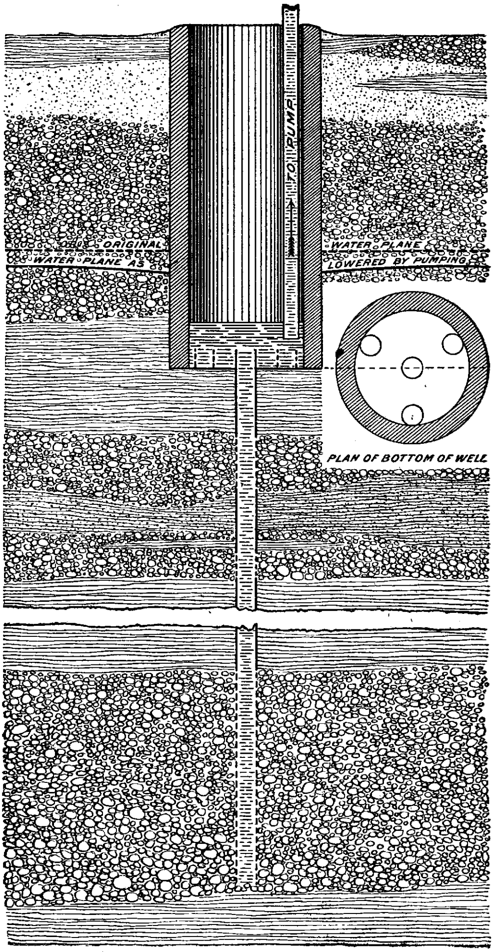


FIG. 32.—Combination dug and tubular well.

estimates based on theoretical considerations, which indicate that if at the end of a certain period of time (say fifteen minutes) the depression is half of the original amount, then at the end of twice that period of time (thirty minutes) the depression will be one-fourth of the original amount; at the end of thrice the period of time (forty-five minutes) it will be one-eighth of the original amount, etc.

CONTAMINATION OF SURFACE WELLS.

Since the upper surface of the ground water in the surface zone of flow is everywhere exposed to contamination by seepage of impurities

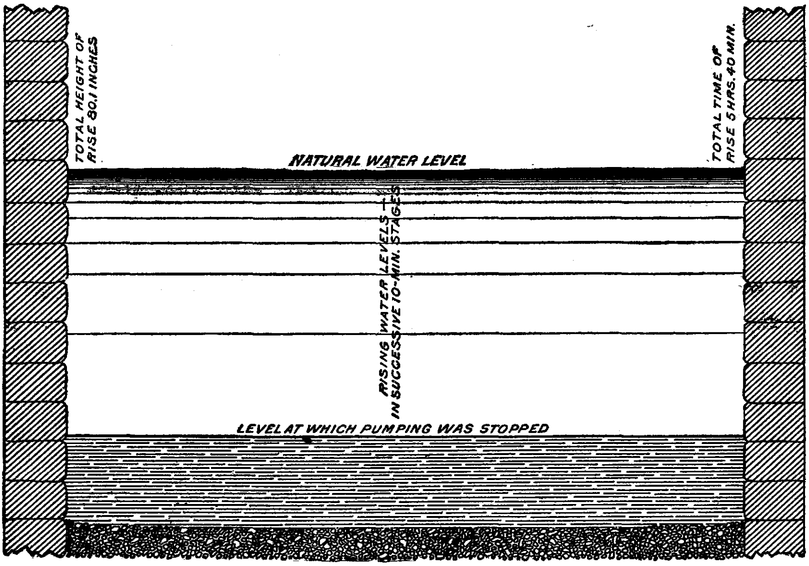


FIG. 33.—Diagram showing rise of water in city well at Garden, Kans., after heavy pumping.

from the surface of the ground, wells in this zone of ground waters are especially subject to pollution.

The organic impurities—such as decaying vegetable and animal matter, and the products of their decomposition, animal excreta,

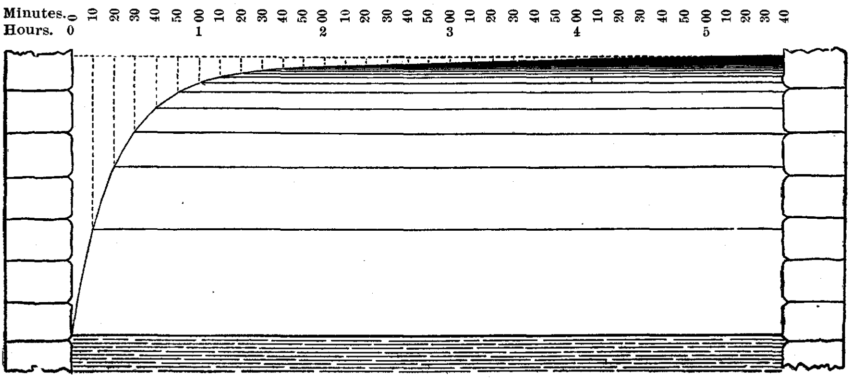


FIG. 34.—Curve showing rise of water surface in well at Garden, Kans.

especially disease germs common to such matter, etc.—constitute the most dangerous elements in drinking water. The possibility of such contamination can frequently be avoided by a proper location of the well. The slope of the ground will usually give an indication of the direction in which the ground water is moving, and sources of contamination above the well can therefore be foreseen. It is true, however, that the depression of the water table in the immediate vicinity of the well will permit it to draw impurities from all directions, especially in the dry months, when the depression extends considerable distances. In the thickly settled districts of cities and villages it is extremely difficult to locate surface wells so that they will not be polluted to a greater or less extent.

Contamination from cesspools is much more dangerous than contamination from surface filth, for the upper few feet of the soil contain living organisms which purify and destroy organic impurities slowly seeping downward, while the organic matter in cesspools is immediately contributed to the subsoil without the action of the nitrifying organisms. The author has noted 18 cases of typhoid fever in one district of a city, the families drawing their water from 18

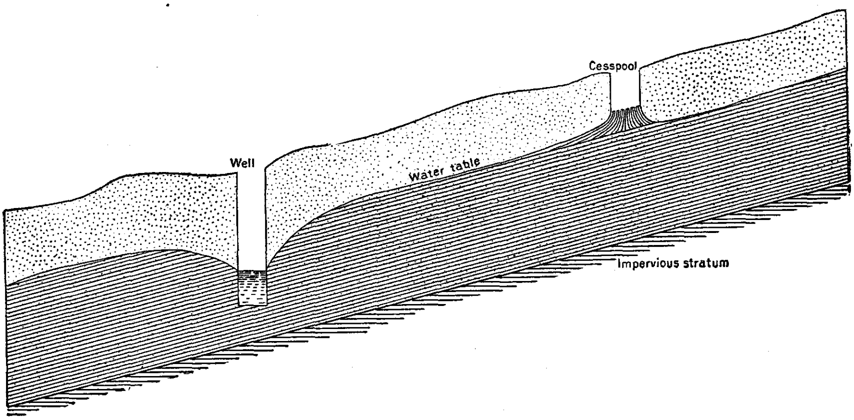


FIG. 35.—Diagram illustrating danger of contamination of surface wells from cesspools.

neighboring wells. The city was built upon the slope of a river terrace, and the alternation of wells and cesspools was essentially as illustrated in fig. 35. In this particular case pure artesian water from flowing wells could have been had by drilling about 1,000 feet.

NONUNIFORMITY OF NEIGHBORING WELLS.

It is a not uncommon experience to find that water can not be obtained a short distance from a good well. Such a discovery always causes considerable comment, while the large regions in which ground water is found at very uniform depths call forth no comment whatever. Irregularities in the strata, especially in the deposits of clay in the glacial drift, account for most of the observed diversity of supply. Fig. 36 illustrates an extreme form of irregularity which was observed in East Schleswig-Holstein. By the action of the ice and drift in Glacial times this formation was so thrown together that there were formed in the soil a number of trough-shaped and irregular cavities which have been filled with permeable glacial sand. These troughs or cavities have collected ground water at various depths into masses which are only in partial communication with one another. In fact, the ground water is for the most part separated into distinct streams of greater or less size, which are in communication with one another only during high water. A–G, fig. 36, represent wells whose water levels, on account of these conditions, show a great variety of height, and some of which, as at G, have no water in dry weather. Another cause of irregularity in the depth and supply of wells is found in localities where gravelly beds of ancient streams or the beaches of former lakes lie buried beneath a later formation of surface material. These lines of gravel are usually very narrow, but may extend for miles in a sinuous course, or in a gently curving course, as the case may be that of a buried stream or a buried lake.

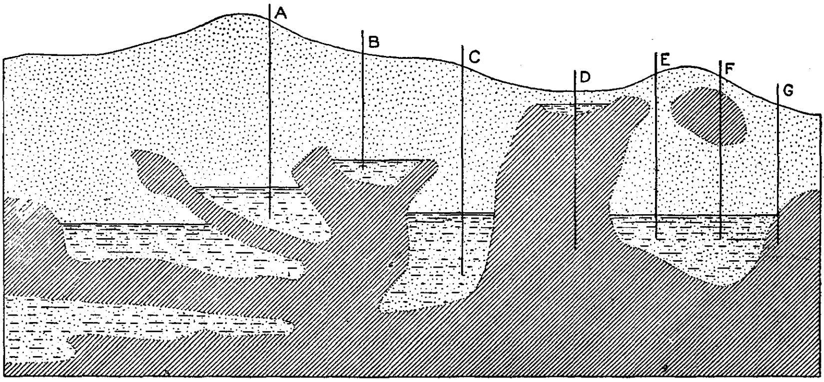


FIG. 36.—Diagram showing extreme irregularities of the water table and the depth of wells (A, B, C, etc.) in East Schleswig-Holstein. (After Haas.)

FLUCTUATIONS IN THE POSITION OF THE WATER TABLE AND THE EFFECT ON WELLS AND SPRINGS.

The height of the water table is constantly subject to variation. Its position at any time is dependent upon the varying relations between rainfall and evaporation. The influence of the former upon the position of the water table can readily be ascertained by actual measurement. There will be observed, of course, not only variations corresponding to single rainstorms, but variations from the rainy months to the dry months and from rainy years to dry years, etc. The average position of the water table at Munich, determined from observations made in three wells and extending over a period of thirty years, is shown in fig. 37.[a] A similar diagram for the city of Berlin, made from observations in thirty-seven wells and extending over sixteen years, is shown in fig. 38. Everyone is familiar with the common experience of the low stage of water in wells during dry months or

[a] The diagrams are taken from the article by I. P. Gerhardt in Der Wasserbau, Leipsic, Vol. I, Pt. I, p 47.

exceptionally dry years, which is caused by fluctuations in the water table like those represented graphically in these diagrams.

Changes in the barometer produce interesting changes in the position of the water table and still greater changes in the flow of springs and wells.

Observations show that the flow of springs and wells increases with a lowering of the barometer. This phenomenon has been carefully noted from early times, accurate descriptions dating from the eighteenth century. Baldwin Latham, in 1881, gave the results of some very interesting observations in England. He reports that some of

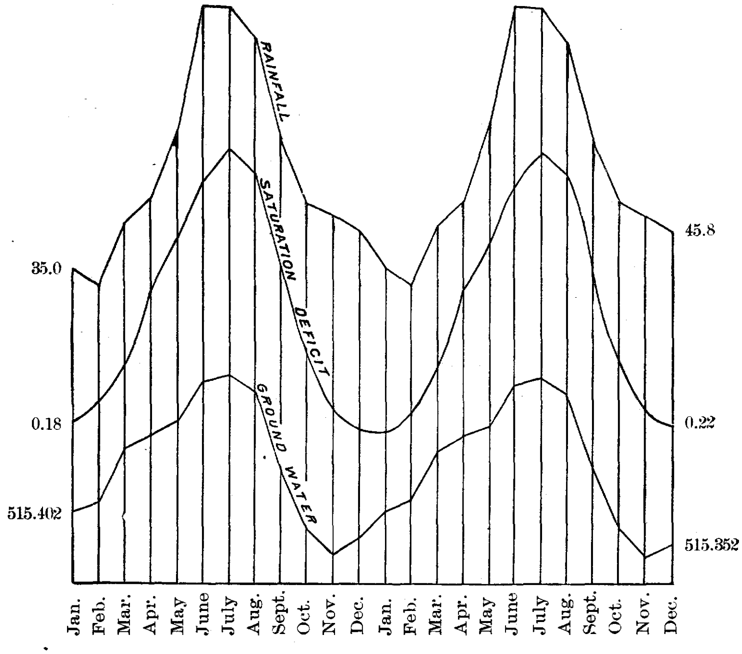


FIG. 37.—Diagram showing fluctuation of the height of the ground water and associated phenomena as observed at Munich. Averages are for thirty years and from observations in three wells. Ground water and rainfall are in centimeters, saturation deficit in millimeters. The saturation deficit indicates the amount of water lacking for the complete saturation of the air at the prevailing temperature.

the long-established millers on the chalk streams claimed that they were able to foretell the approach of a rainstorm by a sensible increase in the flow of the river before any rain had fallen. He undertook a series of observations to investigate the phenomenon, and found, by setting up gages in the spring of 1881, on the Bourne, in the Caterham Valley, near Croydon, and selecting periods when there was no rain to vitiate the results, that whenever there was a rapid fall in the barometer there was a corresponding increase in the volume of water flowing, and that with a rise in the barometer there was a diminution in the flow. The fluctuations in the flow of the Croydon Bourne, due to barometric pressure, at one period exceeded 500,000 gallons a day.

Observations upon the yield of wells and springs and upon percolation gages gave similar evidence. In 1883 Latham made observations on the effect of barometric pressure upon the flow of water from artesian wells, with results in accordance with the other observations.[a] Similar results were later observed in Germany and in America.[b] (See fig. 39.)

The phenomenon is usually explained by the expansibility of the air confined in the porous medium in the neighborhood of the ground water and of the air dissolved in the ground water itself. Even the air in the soil above the water table meets more or less resistance to its escape when the barometer falls, and consequently a differential

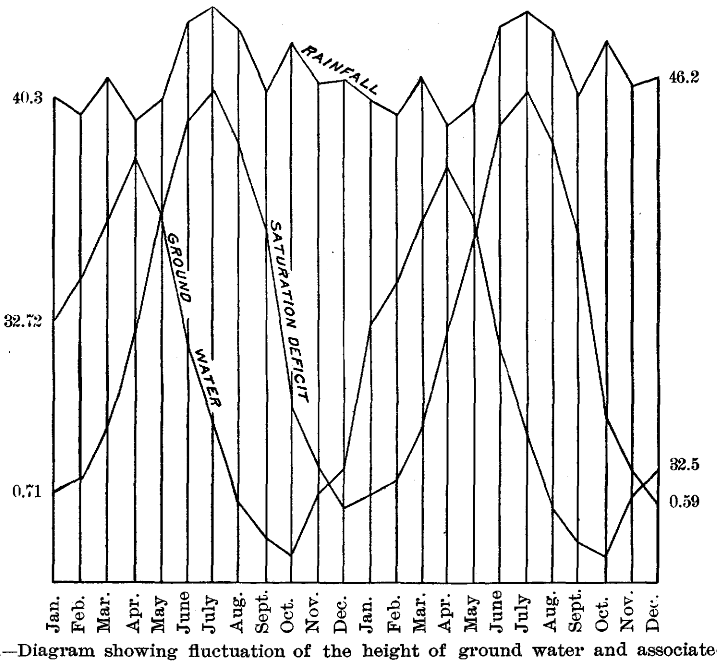


FIG. 38.—Diagram showing fluctuation of the height of ground water and associated phenomena as observed at Berlin. Averages are for sixteen years and from observations in thirty seven wells. Ground water and rainfall are in centimeters, saturation deficit in millimeters.

pressure must exist over the surface of the water table and the surface exposed at the well or spring.

A change of an inch in the height of the barometer corresponds to a change of pressure of about 1 foot of water. It is evident that such a differential pressure exerted over the surface of the water table is sufficient to materially affect the flow into wells and springs.

A very unique and not uncommon phenomenon, due to the same cause, is the "blowing" of wells just before a storm, as has been observed in the West. The low barometer before a storm causes

[a] See British Association Reports, 1881, p. 614; 1883, p. 495.

[b] Einfluss des Atmosphärendrucks auf die Ergiebigkeit von Brunnen und Quellen, by Otto Lueger: Centralblatt d. Bauverw., 1882, p. 8. F. H. King, U. S. Dept. Agr. Weather Bulletin No. 5, 1892, p. 50.

large quantities of air to be forced from the well with a loud, roaring noise.

AVAILABLE SUPPLY OF WELLS AND SPRINGS.

We may summarize the conditions affecting the amount of ground water available for the supply of wells and springs as follows: (1)

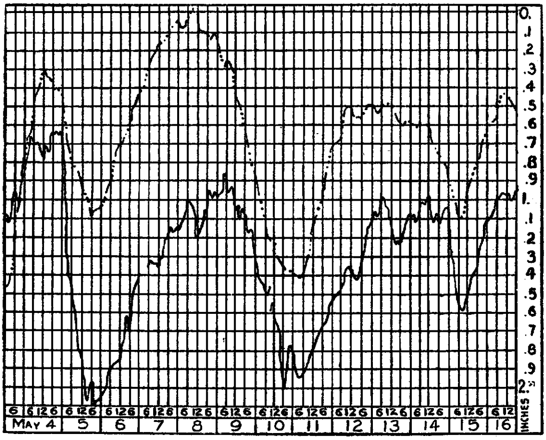


FIG. 39.—Autograph record of simultaneous changes in flow of spring and atmospheric pressure. Upper curve is air pressure, lower curve flow of spring. (After F. H. King.)

Magnitude of the area contributory to the wells; (2) amount of the rainfall upon this area; (3) geologic structure, such as (*a*) the arrangement or stratification of the material, (*b*) the breadth and depth of the water-bearing medium, and (*c*) the character or composition of this material, such as its fineness and porosity; and (4) physiographic features of the land surface, such as mountains, plateaus, hills, valleys, plains, forests, prairies, cultivated areas, etc.

COLLECTING GALLERIES AND SUBCANALS.

Another method of obtaining water from the surface underflow of ground water is by excavating a ditch or tunnel, with its bottom

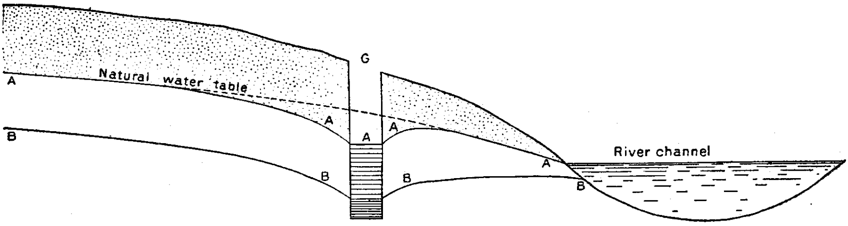


FIG. 40.—Diagram illustrating Salbach's theory of the contamination of wells or collecting galleries by reversed flow from rivers. Liability of contamination begins, according to Salbach when the water table falls so as to intersect the bed of the stream, as it does at B B B B.

below the water table, and running as nearly as possible at right angles to the direction of the underflow. A ditch so constructed is

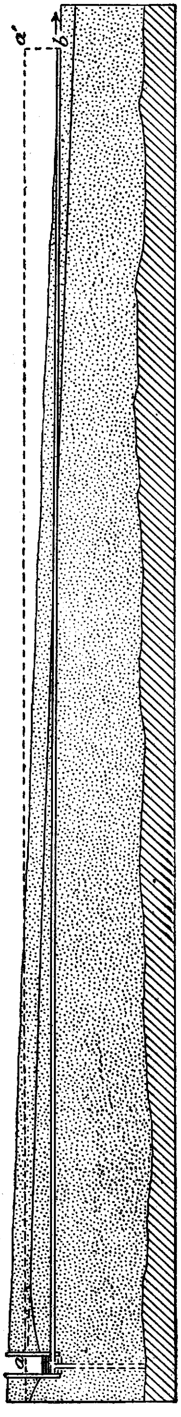
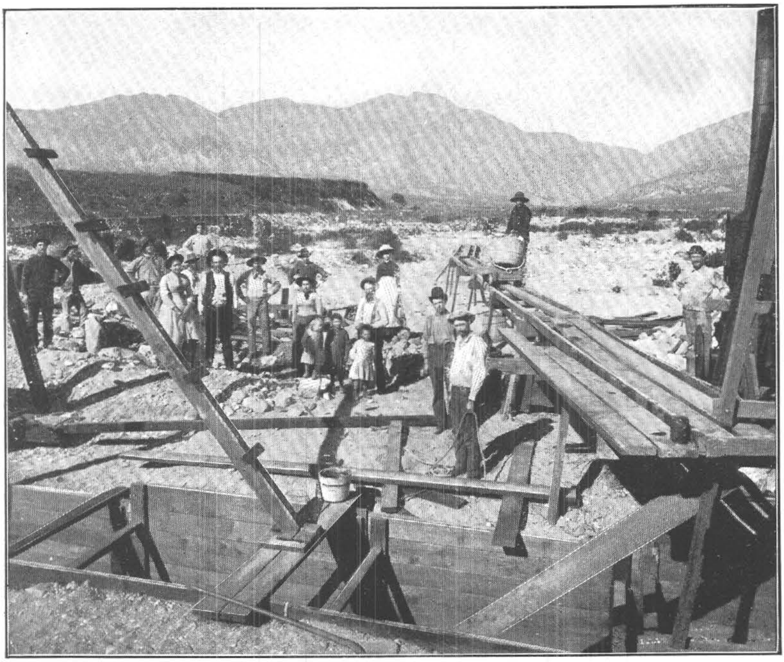


FIG. 41.—Diagram illustrating principle involved in construction of infiltration galleries for gravity flow, by Willard D. Johnson.

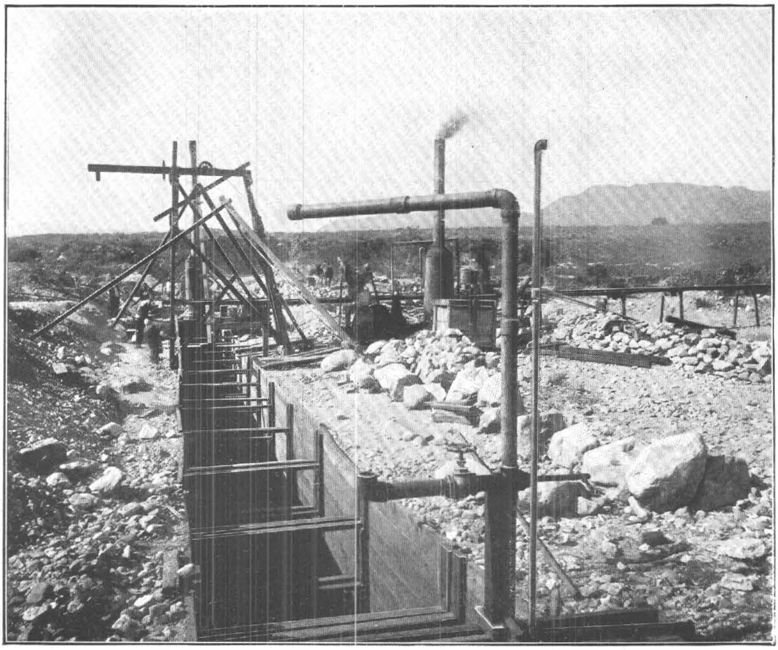
nothing but a very elongated surface well. Its great length often makes it possible to bring the water to the surface of the ground without pumping, by simply extending the ditch with a slope less than that of the surface of the ground until the flow comes to the surface. When such a channel is excavated for the purpose of furnishing a city water supply, it is usually called a collecting or infiltration gallery. Such excavations may be either open and unwalled or covered and walled. A variety of terms have been applied by various authors to similar devices constructed for obtaining water for irrigation. The open ditches are called gravity cuts, fountains, and underflow canals, the first term applying especially when the recovery is made without pumping. If the water is secured by tunneling or "mining" for water, the excavation is frequently called a subsurface canal or underflow tunnel.

Several gravity cuts and tunnels made in the porous gravels so common in California have yielded water in abundance. Fountains constructed in material having a relatively low transmission constant, while not furnishing water for irrigation works of great magnitude, have, nevertheless, been important factors in many localities. The yield of a fountain or infiltration gallery, like the capacity of a well, must depend very largely upon the character of the water-bearing material in which the excavation is made, the yield being proportionate to the transmission constant. Other things being equal, the yield must be nearly proportional to the length of the canal. The depth is likewise an important factor. Theoretical considerations indicate that the flow should be proportional to the square of the depth of the water surface in the canal below the natural water table. This conclusion seems to be borne out by experience.

The usual practice in recovering underflow by means of gravity works is to follow a grade much less than the slope of the ground surface until the floor of the canal is about 6 feet below the level of ground water, and then to follow, in the further excavation from that point, the same slope as the surface of the ground. It is obvious that



A. VIEW LOOKING UP THE PACOIMA WASH FROM THE SITE OF THE SUBSURFACE DAM.



B. VIEW OF 50-FOOT SECTION OF SUBSURFACE DAM ON THE PACOIMA WASH. SHOWING EXCAVATION AND LAGGING.

gravity works are impracticable where the ground water is at great depths or where there is an insignificant slope to the water table.

The effect of the removal of ground water from the soil by means of an infiltration gallery is, of course, the lowering of the water table in the immediate vicinity of the excavation. The curved form assumed by the water table is approximately that shown in figs. 8 (p. 29) and 40 (p. 73), and is similar to the form of the water table in the neighbor-

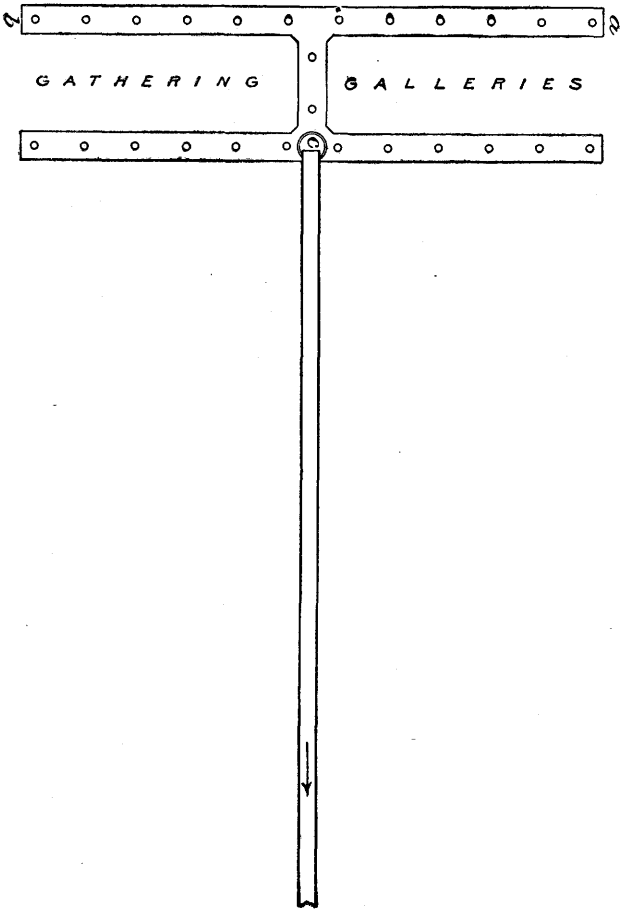


FIG. 42.—Plan of infiltration galleries proposed by Willard D. Johnson.

hood of a well. If the infiltration gallery is constructed in close proximity to the bank of the river, and if the water collected is used for a city water supply, it is important to know under what circumstances pollution may take place by seepage of river water into the gallery. According to Salbach such pollution will not take place unless the curved water table actually intersects the bed of the stream, as shown by B B B B in fig. 40.

Disappointment has sometimes followed the construction of infiltration works of the types described, especially in regions of low rainfall or where the underflow had not previously been subjected to careful survey. The underflow cut above Dodge, Kans., is now entirely abandoned. The supply of water was disappointing and the river floods have filled the excavation, which was dredged at great expense, with silt and sand.

Mr. Willard D. Johnson has made important suggestions for the construction of infiltration galleries for gravity works, as shown in the accompanying figures (41, 42, and 43). He would use one or more batteries of tubular wells to reach the best portions of the water-bearing medium, as shown by a careful survey made prior to the construction of the works. The best flow of water is usually

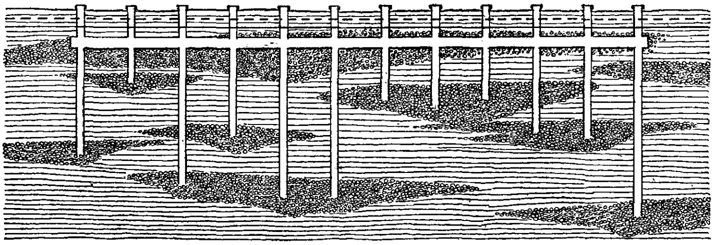


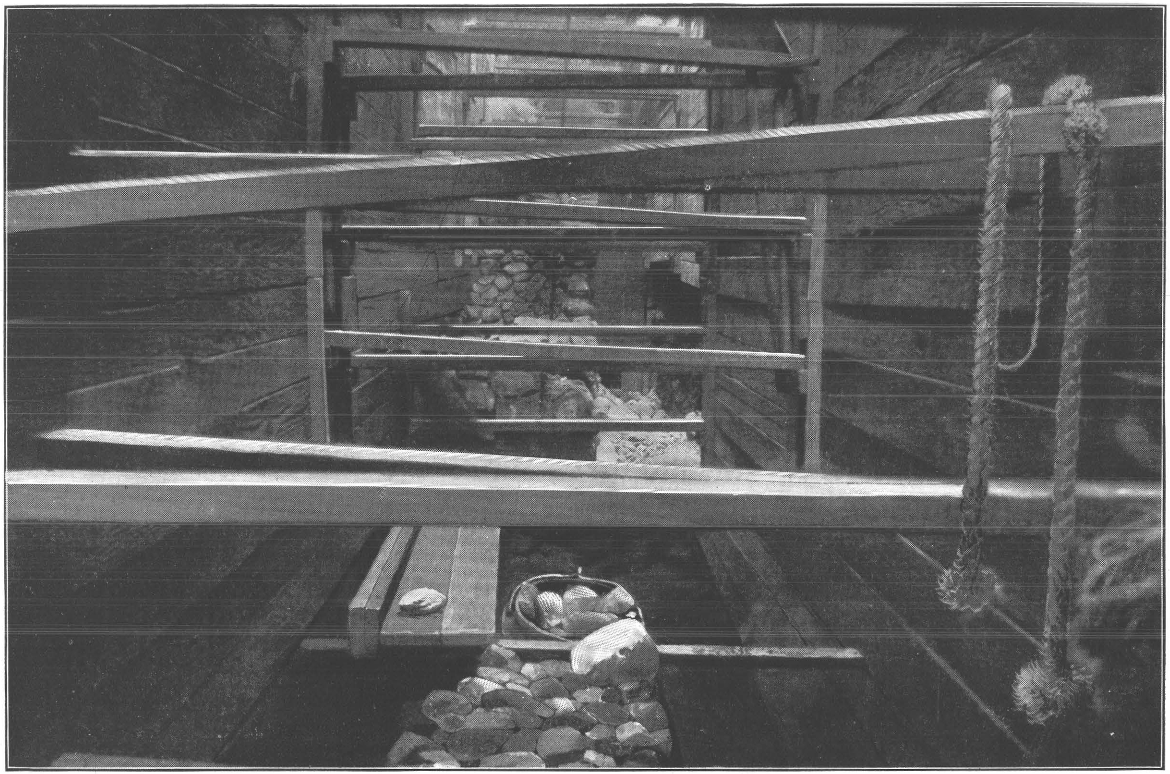
FIG. 43.—Longitudinal section of infiltration galleries designed by Willard D. Johnson.

found at a considerble depth below the water table, and Mr. Johnson's plan offers a simple way of reaching it without expensive excavation.

SUBSURFACE DAMS.

Another method of recovering the underflow of a stream is by means of a subsurface dam. Such a dam is constructed by excavating a trench at right angles to the direction of the underflow and extending in depth to the impervious stratum, and then filling the trench with impervious material. If the underflow is confined within an impervious trough or canyon, it is obvious that such a construction must result in bringing it to the surface. An example of this is found on Pacoima Creek, Los Angeles County, Cal., where a subsurface dam was constructed in 1887–1890. It is claimed that by means of this dam the owners have been enabled to use the bed-rock flow of water for the three dry years, 1898–1900, and thereby to successfully carry through the orange, lemon, and olive growing in Fernando Valley. This dam is described in the Eighteenth Annual Report of the United States Geological Survey, Part IV, pages 693 to 695; also in Reservoirs for Irrigation, Water Power, etc., by James D. Schuyler, 1901, page 205.

A view of the wash up the canyon from the site of the dam is shown in Pl. V, *A*. An underflow was known to exist, as the stream had a good perennial flow back in the hills, shown in the background of



VIEW FROM BOTTOM OF 50-FOOT SECTION OF PACOIMA SUBSURFACE DAM, SHOWING LAGGING IN PLACE AND CONSTRUCTION OF CONCRETE WALL.

Pl. V, *A*, which was observed to gradually sink into the deep gravel wash before reaching the plains. At the site of the dam the canyon walls are about 600 feet apart, and the rock floor is at irregular depths, generally from 25 to 50 feet below the surface of the gravel.

In constructing the dam, a trench was excavated in successive sections at right angles to the course of the wash. Each section was 50 feet long and 5 feet wide, as shown in Pl. V, *B*. The excavation extended to bed rock and was boarded up with 2-inch by 8-inch by 8-foot timbers and lagged with 3-inch by 4-inch by 5-foot timbers, as shown in Pl. V, *B*, and Pl. VI. When bed rock was reached, a small trench about 12 inches deep and 10 inches wide was excavated in it, and the construction of the cement and gravel dam was begun. The wall is about 2 feet thick, and was built up to the surface in the usual manner, loose surface material being replaced around the dam as built, completely inclosing and supporting it. Centrifugal pumps kept the trenches free from water while working at bed rock, except at the closing section. Two masonry collecting wells were built in the wall of the dam, as a part of the same structure and extending to bed rock. The wells are 4 feet in internal diameter, and are built heavy on the downstream side, but lighter and with a number of openings to admit water on the upstream side.

As constructed, the dam shuts off the underflow, with the exception of some leakage which developed in places, the dam not being completely water-tight. The underground reservoir formed in the upstream gravel can not be definitely outlined, but its surface area is about 300 acres. The water is collected by means of four lines of 10-inch to 12-inch concrete tile pipes laid above one another in horizontal rows from 8 to 10 feet apart on the upper side of the dam, beginning within a few feet of the bed rock. The tiles are in 3-foot lengths, laid with open joints, and discharge into the collecting wells described.

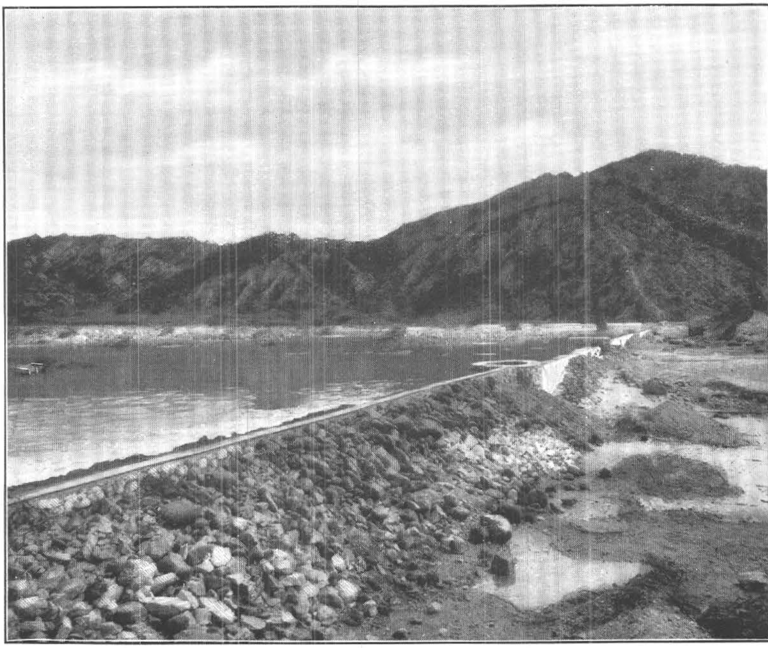
A distributing main is laid into each well about 10 feet below the surface, and is bent down inside of the well so as to take water about 30 feet below the surface. These mains are 14 and 20 inches in diameter, and extend underground to the towns of San Fernando and Pacoima, furnishing water for irrigation and domestic use.

The end sections of the dam were carried up about 2 feet higher than the central portion, so as to prevent an overflow of water cutting out the banks of the wash. The total cost of the structure was about $40,000, cement costing $5 a barrel and labor from $2 to $4 a day. The top of the completed dam is shown in Pl. VII, *A*, and a portion of the central section in Pl. VII, *B*.

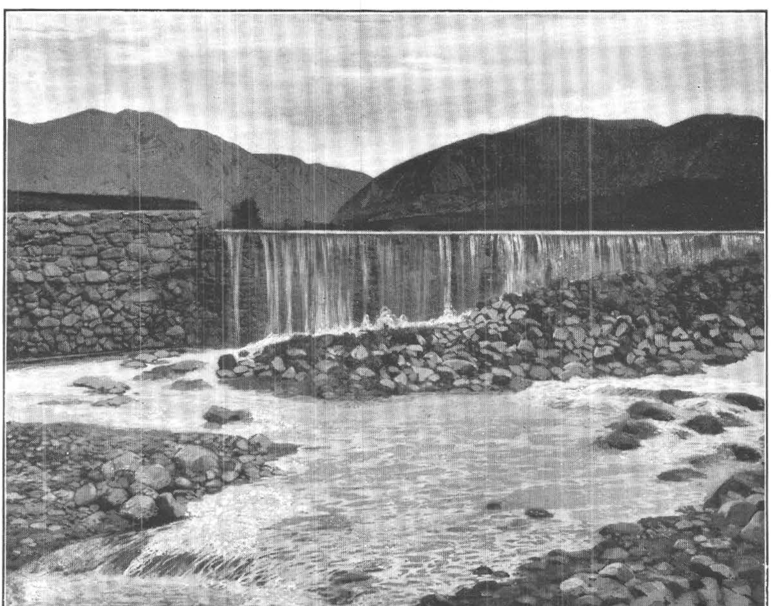
ADVANTAGES OF UNDERGROUND SUPPLY.

The recovery of ground water for irrigation by pumping from wells or by means of gravity cuts, infiltration galleries, or submerged dams is, considering the amount of water recovered, more complicated and expensive than the common practice of using the perennial flow of surface streams. But as the waters of the streams become more and more appropriated other sources of supply must be sought. The principal alternatives are the storage of storm waters and the utilization of the underflow. The appropriation of the perennial flow of the rivers by the mountain States leaves the settlers along the valleys of the Great Plains dependent upon the underground waters for irrigation. Fortunately portions of western Kansas and Nebraska are supplied with underground water within a reasonable distance from the surface. While the size of underflow works must be small in comparison with the usual works of stream diversion or reservoir storage, yet there are compensating advantages associated with them which make their development especially worthy of encouragement where conditions are favorable. In the first place, the utilization of the underflow is making use of water which in most cases would otherwise go to waste. Frequently the underflow drawn from the alluvium of a river is completely replaced by return seepage of river water during flood stages of the stream, so that the sands and gravels are really being used as inexpensive and indestructible reservoirs for the storage of storm waters. In the second place, there is usually greater regularity and uniformity in the quantity of ground water than there is in the perennial flow of rivers, although the former is by no means free from the influence of dry months and dry seasons, as has already been shown. Furthermore, irrigation by the use of windmills or other power along the river valleys of the Plains, even if practiced on a small percentage of the area, must have an appreciable effect in raising the level of the water table, in turn augmenting the acreage which can be cultivated and irrigated, and increasing the percentage of the rainfall which will enter the soil. The combination of intensive farming on small irrigated tracts with cattle raising on adjoining natural grass lands would seem to indicate a most attractive future for vast portions of the Great Plains and to contain possibilities of unique economic and social interest.

The fundamental disadvantage in the utilization of underground sources of water is the danger of overdrawing the natural supply. In regions in which the rainfall is light and catchment areas are small, as in parts of southern California, it is easy to extend development of underground sources so as to greatly exceed the natural rate of annual replenishment. In this way underground reservoirs are depleted which have been ages in filling; principal as well as interest is drawn upon, and much disappointment must inevitably follow.



A. VIEW OF TOP OF COMPLETED SUBSURFACE DAM ON THE PACOIMA WASH, SHOWING WELLS ON UPPER SIDE OF SAME.



B. VIEW OF CENTRAL PORTION OF SUBSURFACE DAM ON THE PACOIMA WASH DURING OVERFLOW.

CHAPTER V.

ARTESIAN AND DEEP WELLS.

In Chapter III several ways in which the water from the deep zones may again reach the surface have been pointed out. If outcrops of the porous stratum exist at lower levels than the exposures which constitute the receiving or catchment area, we may expect to find associated with them numerous springs and diffused return seepage. In places the covering rock may be defective or locally pervious, so that, if the pressure be great enough, water from the deep zone may reach the surface, or at least pass into a higher zone of flow. Occasionally joints and faults in the overlying rocks give opportunity for deep springs to show themselves. In some places the association of deep springs with the faulting system of the covering rocks is a phenomenon of great persistency. Dr. Peale, in his report on the mineral waters of the United States, remarks the almost universal association of mineral and thermal springs with the faulting of rocks.[a] Prof. W. H. Hobbs, who has made a study of the faulting system of the Pomperaug Valley, Connecticut, has kindly furnished a map (fig. 44, on next page) showing the association of springs and faults in that region. By referring to the map it will be noticed that the springs not only occur along the line of faults, but usually occur where two, three, or more faults intersect. "The Triangle" shown on the map is a swampy area caused by the dropping of a triangular block of surface rock left unsupported by three lines of faults. It is undoubtedly supplied by seepage around its margin. It is described in Professor Hobbs's paper in the Twenty-first Annual Report of the United States Geological Survey, Part II.

The springs arising from the deep zones of flow are usually distinguished from surface springs by a greater constancy of flow from season to season and by a greater uniformity in the summer and winter temperatures of the water. Many surface springs are found to dry up and permanently disappear with the settlement and deforesting of the country, but this phenomenon is much less common with deep springs. The deep springs are also much less liable to contamination than surface springs, for reasons which are very obvious. Even if the source of the deep zone of flow be polluted, the time required for the water to reach the outlet at the spring is usually so great that all organic impurities are effectually broken down and oxidized.

[a] Fourteenth Ann. Rept. U. S. Geol. Survey, Pt. II, 1894, p. 49.

In order to artificially recover water from the deep zone of flow, it is obvious that it must first be reached by drilling through the overlying material, the practicability of which depends upon the character and thickness of the covering rock and the expense which it is feasible to incur for the purpose. If the water in the water-bearing stratum is found under high pressure, it will rise in the drill hole until the pressure head due to the weight of the water column just balances the pressure in the water-bearing rock, or, if the pressure be

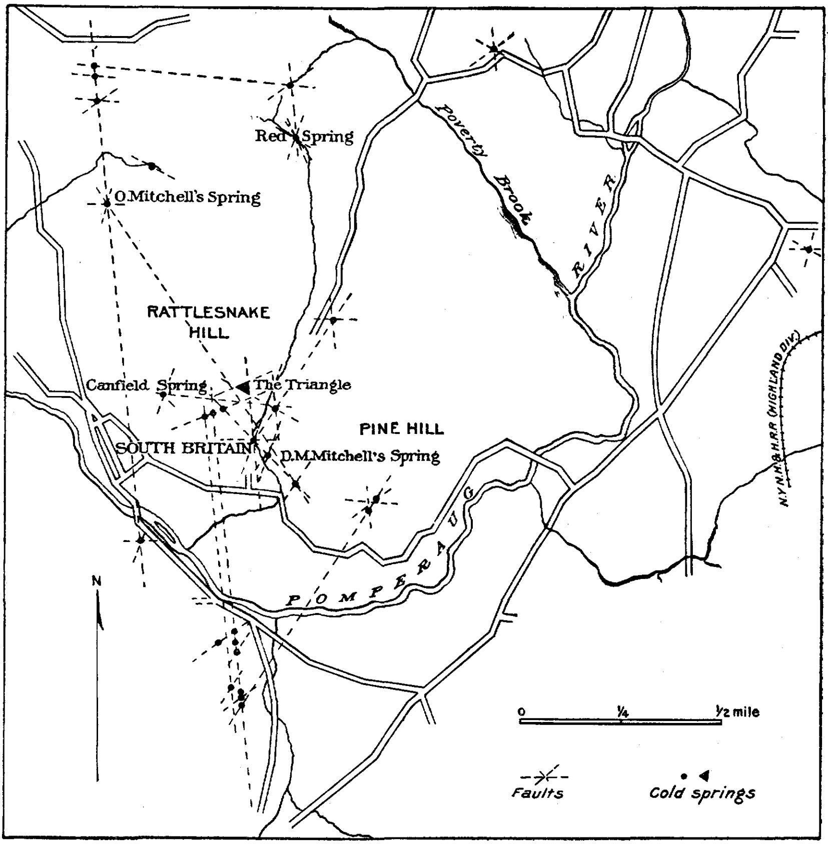


FIG. 44.—Map showing location of springs in the vicinity of South Britain, Conn.

great enough, until the water actually overflows above the surface. In order, however, that the water may rise as high as possible, it is necessary that the drill hole be properly cased with suitable well tubing, so that there will be no opportunity for the escape of water into pervious beds above the water-bearing rock; otherwise the drill hole may serve merely as a passage way from a lower to a higher stratum.

If the water actually overflows at the surface of the ground, we have an artesian well, properly so called. If the water does not rise to the surface, we have a nonflowing or deep well. This distinction is an important one to the owner of the wells, for in one case water is obtained without the constant expenditure of work and the wear and tear of pumping machinery, while in the other case a continual expenditure of work and money is required to bring the water to the surface. In the explanation of the phenomena and of the difficulties to be met with and overcome the distinction is one of little consequence and is entirely disregarded by many writers. The determination of the question whether or not a given well will overflow depends upon the altitude of the top of the well; so that of two wells exactly alike in all respects except that one is drilled on higher ground than the other, both tapping the same water-bearing rock, the one may overflow while the other may not. For this reason many writers call both wells artesian, and distinguish them as flowing and nonflowing wells. If the adjectives flowing and nonflowing are used, we may designate as artesian any well made into a deep zone of flow where the water is found under a static pressure so that it will rise in the well above the impervious covering stratum. The essential differences between such a well and a surface well are due to the existence of the upper impervious stratum, which is absent in the surface zone of flow. Owing to the presence of this impervious cover, we may emphasize the following distinguishing characteristics between surface wells and artesian wells: (1) The water in the water-bearing veins of an artesian well is under a static pressure and will rise when the impervious cover is penetrated, while the ground water supplying a surface well has a free upper surface which is lowered and changed in shape when water is removed from the well; and (2) an artesian well is supplied by the rainfall of districts not in the immediate vicinity of the well, but often distant hundreds of miles, while a surface well draws upon the rainfall of the immediate vicinity.

The conditions upon which the existence of artesian wells depends presupposes a knowledge of the principles involved in the motion and storage of underground waters in the deep zones of flow. The essential facts involved in such flow have been discussed in Chapter III. The diagram of the cross section of the Dakota artesian basin (fig. 20, p. 55) will serve to illustrate the conditions present in any artesian basin. In that diagram it will be noticed that the high elevation of the western exposure of the porous Dakota sandstone in the Black Hills tends to give a high pressure to the water in the sandstone in its eastern and nearly horizontal portion. If there were absolutely no escape for the water to the east of the catchment area, we should expect the water to rise in each well, if the casing were extended high enough, to the horizontal line L-L′ (fig. 20). The actual observed pressure, however, departs widely from that line, due undoubtedly to leakage from the eastern outcrop of the sandstone, especially along the Missouri and Great Sioux rivers in the southeastern part of South Dakota. At Running Water, Randall, and Chamberlain springs are seen rising from the bed of Missouri River, which probably accounts for the tendency to lower pressure in the neighborhood of the river. The high pressure at Kimball, Plankinton, and other places east of Chamberlain, probably reaches these points in part by a more northerly route, along which the loss by seepage is less than at Chamberlain and southerly points. This is verified by the high pressure of 205 pounds at Cheyenne Agency and by a pressure of more than 180 pounds at Crow Creek Agency, the latter place being halfway between Chamberlain and Pierre, and the pressure higher than is found at either of those places.

As already stated, the essential features of an artesian basin are well illustrated by this example of the Dakota area. The same principles are involved in the Potsdam and St. Peter artesian basin in southern Wisconsin and northern Illinois, and in the other large artesian basins in the country, except that the high pressure and large yield of the Dakota wells are in excess of those common in other basins. For a more extensive discussion of the general principles involved the reader is referred to Prof. T. C. Chamberlin's paper entitled, The Requisite and Qualifying Conditions of Artesian Wells.[a] The principal conditions on which artesian flows depend are summarized by Chamberlin as follows:

I. A pervious stratum to permit the entrance and the passage of the water.

II. A water-tight bed below to prevent the escape of the water downward.

III. A like impervious bed above to prevent escape upward, for the water, being under pressure from the fountain-head, would otherwise find relief in that direction.

IV. An inclination of these beds, so that the edge at which the waters enter will be higher than the surface of the well.

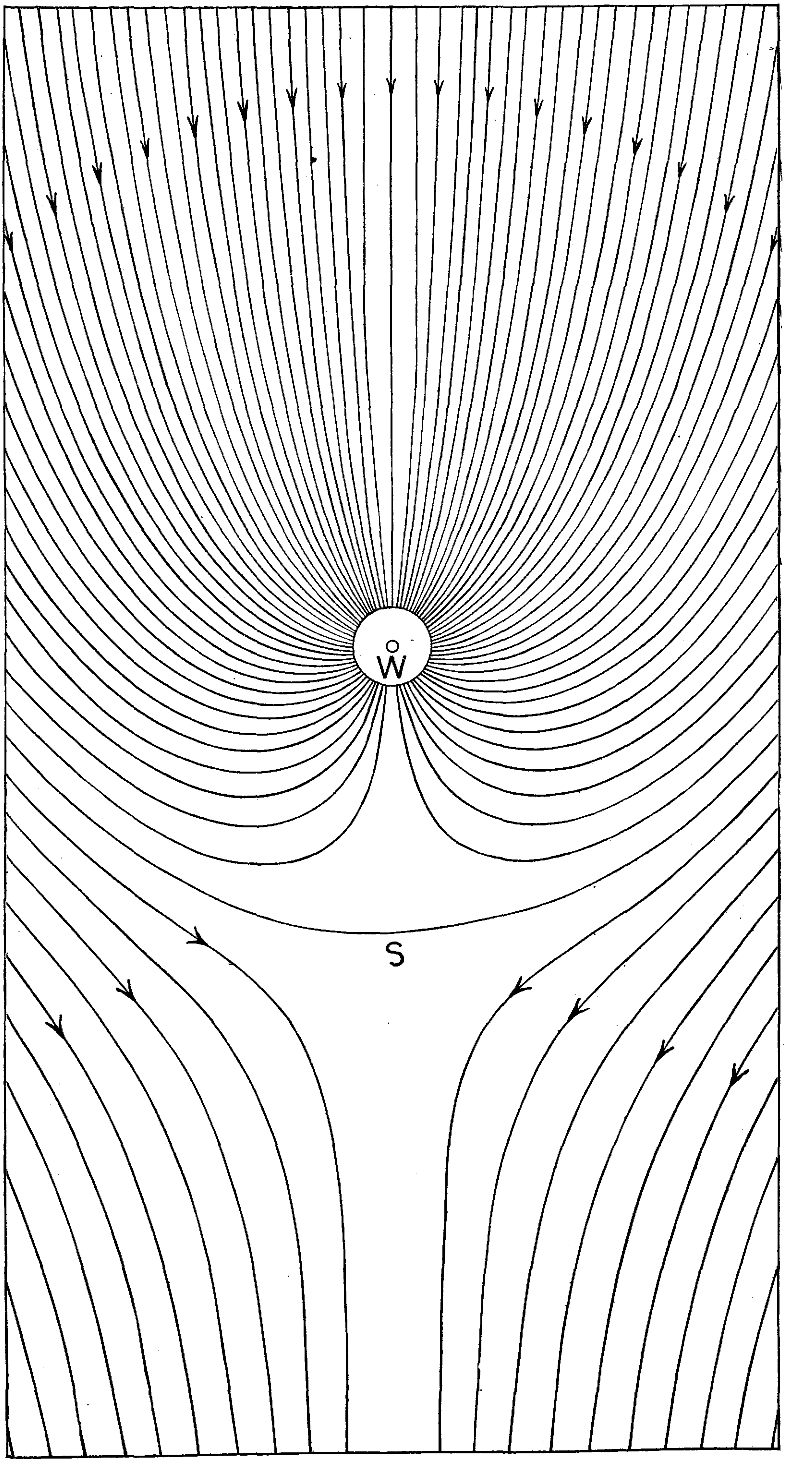
V. A suitable exposure of the edge of the porous stratum, so that it may ta in a sufficient supply of water.

VI. An adequate rain-fall to furnish this supply.

VII. An absence of any [*easy*] escape for the water at a lower level than the surface at the well.

These seven prerequisites as stated by Chamberlin have been widely quoted by various writers and have become classic. The writer has made no change in the statements except the substitution of the word "easy" for "any" in VII, which is undoubtedly in accord with the meaning intended by Professor Chamberlin. Of course the less opportunity there is for the water of the water-bearing rock to escape at a level below that of the well the higher will be the head of pressure at the well, but the water-bearing stone is usually so fine that the frictional resistance offered by it to the movement of the water is sufficient to establish considerable pressure a few miles back from an outcrop. As water flows through a horizontal porous medium under a head at

[a] Fifth Ann. Rept. U. S. Geol. Survey, 1885, pp. 131-173.



LINES OF FLOW INTO A WELL IN A REGION IN WHICH THE GROUND WATER HAS A CONSTANT MOTION IN A GENERAL DIRECTION.

one end, the pressure gradually drops from the maximum value at the end at which the water enters to nothing at the free end from which the water flows. This is illustrated by fig. 45. If we should prevent the escape of water at K_7 G_2, it would rise to corresponding heights (H', H'', H''') in all of the tubes. In spite of the free escape at the right end of the sand column K_7 G_2, the pressure gradient H_1 H_2 H_3 H_4 H_5 H_6 is sufficient to produce a flow at any point along the tube K_1 K_6 if the wall of the tube be punctured. In this case we may imagine that the water-bearing sand and its confining tube represent, respectively, the water-bearing stratum and the upper and lower impervious strata of an artesian basin, the water column at the left of the figure corresponding to the higher outcrop of the pervious

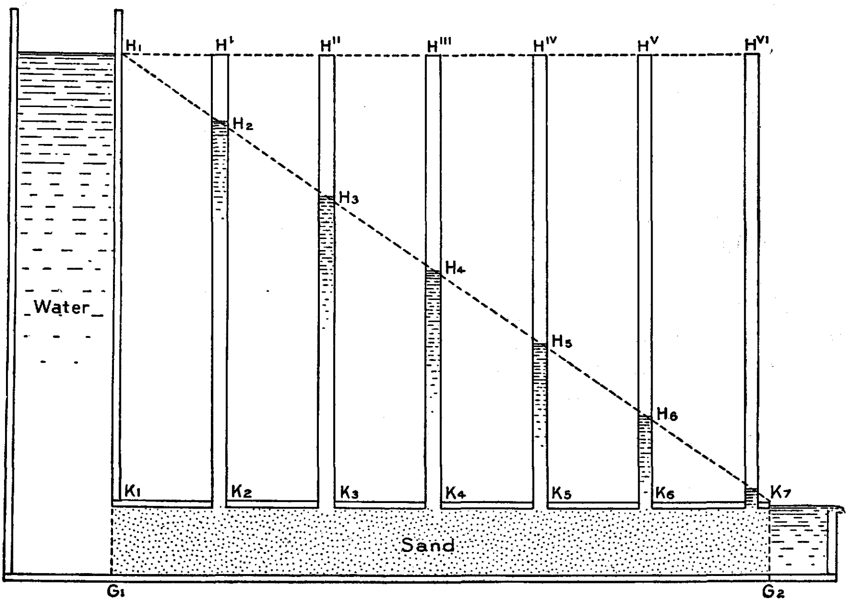


FIG. 45.—Diagram illustrating the fall of pressure as water moves through a pervious material.

rock in the catchment area, while a puncture made in the tube anywhere between K_1 and K_7 represents an artesian well.

The general character of the lines of flow in the neighborhood of an artesian well is shown in Pl. VIII.

YIELD OR CAPACITY OF ARTESIAN WELLS.

The yield of an artesian well, or the amount of water delivered in a given time, is usually measured in cubic feet per minute or per second, or in gallons per twenty-four hours. The yield depends upon the same factors as noted in Chapter IV, on surface wells, but in addition to the causes there enumerated as affecting the flow in common wells, we must now consider the frictional resistance which the water suffers in flowing through the casing and drill hole of the well. This resistance increases with the length of the casing, and also increases very rapidly with a decrease in the diameter of the casing. For long pipes the following formula will give the approximate discharge in cubic feet per minute:

$$q=378\sqrt{\frac{d^5h}{fl}},$$

in which d is the diameter of the pipe in feet; l is the length of the pipe in feet; h is the head under which the flow takes place, measured in feet of water; f is the friction factor, which depends for its value upon the character of the surface of the pipe, and varies with the size of the pipe and the velocity of flow (for rough approximations its mean value 0.02 can be used).

The foregoing formula states that the discharge from a long pipe varies directly as the $2\frac{1}{2}$ power of the diameter of the pipe, directly as the square root of the head, and inversely as the square root of the length of the pipe. The discharge of pipes, however, is ordinarily taken directly from hydraulic tables, thus saving the labor of computation.[a] If we attempt to estimate by the formula or by a hydraulic table the discharge from the well tubing of a flowing well, using as the head the observed static head when the mouth of the well is closed, and using the known size and depth of the well and tubing for the diameter and length of the pipe, a result will be obtained which will always be found to be in excess of the actual flow from the well. The reason for this is that no account is taken of the enormous resistance offered to the flow of water into the well through the pores of the water-bearing strata. The amount of water yielded by the porous strata to the well can also be expressed by a formula, as the writer has shown in another paper.[b] In order to estimate the amount of water delivered by the porous medium its transmission constant must be known. Suppose this constant to be given, the following is the formula which expresses the yield of the well, neglecting, as already stated, the resistance due to pipe friction:

$$Q_0=\frac{2\,\pi\,h\,k\,a}{\log_e\left(1+\dfrac{1{,}200}{D}\right)}\text{ cubic feet per minute,}$$

in which h is the observed static head, in feet, a is the thickness of the water-bearing stratum (of transmission capacity k), and D is the diameter of the well in feet. The logarithm used here is the natural

[a] See Tables Showing Loss of Head Due to Friction of Water in Pipes, by E. B. Weston, Van Nostrand, N. Y., 1896; and Graphical Solution of Hydraulic Problems, by F. C. Coffin, Wiley & Sons, New York, 1897.

[b] Theoretical investigation of the motion of ground waters: Nineteenth Ann. Rept. U. S. Geol. Survey, Pt. II, 1899, p. 360.

or napierian logarithm.[a] The denominator of the fraction in this formula is the only part that depends for its value upon D, the diameter of the well. This part of the formula, namely, $\frac{1}{\log_e\left(1+\frac{1,200}{D}\right)}$, changes value but slightly as D is changed. This fact is readily understood when we remember that 2 inches and 12 inches are the practical limits of size of artesian wells. In Table VII are written the various values which this expression takes for certain common sizes of wells. It is seen that the values differ but slightly for the various sizes, so that if friction in the pores of the sandstone be the only resistance that need be considered the capacity of such wells would be but slightly dependent upon the diameter of the well.

TABLE VII.

Values for various diameters of wells of the factor in artesian-well formula which depends upon the diameter of the well.

Diameter of well.	Value of $\frac{1}{\log_e\left(1+\frac{1,200}{D}\right)}$	Compared with value for 6-inch well.
Inches.		
2	0.1125	0.876
3	0.1179	0.918
4	0.1220	0.951
4½	0.1238	0.964
5	0.1254	0.977
6	0.1283	1.000
8	0.1333	1.039
9	0.1354	1.055
10	0.1374	1.070
12	0.1408	1.097

The formula for the actual flow from the well must take account of both the resistance offered by the water-bearing medium and the resistance due to pipe friction. The writer finds that the following formula is the expression of the actual yield of the well, if both of these factors are taken into account:

$$Q = q\left(\sqrt{1+\frac{q^2}{4\,Q_o^2}} - \frac{q}{2\,Q_o}\right) \text{ cubic feet per minute,}$$

in which q is the estimated free discharge of water through the well and casing under head h, including in the estimate the influence of all valves, bends, reduction in size of pipe, etc., and Q_o is the theoretical yield of the well under head h, if friction in the well and pipe be neglected. In this expression Q_o varies only slightly with a change in the diameter of the well, while q varies rapidly for such change.

[a] This formula was obtained by the writer in 1892. While the paper referred to was in press he discovered that substantially the same formula had been previously worked out by a German hydrographer.

Table VIII gives values of Q_o, or the theoretical yield of an artesian well, computed for various effective sizes of grains of the water-bearing medium. The table is calculated on the basis of a 6-inch well and 100 feet thickness of water-bearing strata. Other thicknesses of strata and other heads of pressure will give proportional results. Such a table may be found useful in estimating the yield of wells in unconsolidated sands of known character, but of course it must not be expected to give results of very great refinement. A nicety of numerical results is impracticable in such a subject, and the table, to be of value, must be used with discretion.

TABLE VIII.

Theoretical capacity (no allowance for pipe friction) of 6-inch artesian well extending 100 feet in materials of various kinds for various heads of pressure.

[Proportional yields for other heads and other thicknesses of material. Yields for wells of other diameters can be obtained by the use of the last column of Table VII. Porosity, 32 per cent; temperature, 50° F.]

Effective size of grain.	4-foot head.	8-foot head.	10-foot head.	12-foot head.	16-foot head.	25-foot head.	Conventional names of size of grains.
Mm.	*Cu. ft. per min.*	*Cu. ft. per min.*	*Cu. ft. per min.*	*Cu. ft. per min.*	*Cu. ft. per min.*	*Cu. ft. per min.*	
0.02	0.047	0.093	0.116	0.140	0.186	0.291	Silt.
0.04	0.186	0.372	0.465	0.558	0.744	1.163	
0.06	0.419	0.837	1.047	1.256	1.674	2.616	Very fine sand.
0.08	0.744	1.488	1.860	2.233	2.977	4.651	
0.10	1.163	2.326	2.907	3.492	4.652	7.270	
0.12	1.675	3.350	4.186	5.025	6.700	10.46	Fine sand.
0.14	2.279	4.558	5.697	6.837	9.116	14.24	
0.16	2.977	5.954	7.440	8.931	11.91	18.60	
0.18	3.768	7.536	9.419	11.30	15.07	23.55	
0.20	4.651	9.302	11.63	13.95	18.60	29.07	
0.25	7.266	14.53	18.17	21.80	29.06	45.47	
0.30	10.46	20.92	26.16	31.38	41.84	65.41	Medium sand.
0.40	18.60	37.20	46.51	55.80	74.40	116.3	
0.50	29.07	58.14	72.67	87.21	116.3	181.8	
0.60	41.86	83.72	104.7	125.6	167.4	261.6	
0.70	56.96	113.9	142.4	170.9	227.8	356.1	Coarse sand.
0.80	74.42	148.8	186.0	223.3	297.7	465.1	
0.90	94.19	188.4	235.5	282.6	376.8	588.7	
1.00	116.3	232.6	290.7	349.2	465.2	726.7	
2.00	465.1	930.2	1163.0	1395.0	1860.0	2907.0	Fine gravel.
3.00	1046.0	2092.0	2616.0	3138.0	4184.0	6541.0	

The writer has devised the following graphical method of solving problems connected with the capacity of artesian wells, and has found it to be very useful. We will first apply the method to a problem in which it is supposed that we know the theoretical yield of the well when pipe friction is neglected (that is, the value of Q_o from Table VIII), and the value of the theoretical discharge (q) of the well tube under the given static head, which last may be taken from a hydraulic table, as already suggested. In fig. 46, let AB and BC be each laid off equal to Q_o, so that $AC=2Q_o$. Then lay off CD at right angles to AC and equal to q. Join AD and lay off $DE=CD$ and draw CE. Also lay off $AF=CD$ and draw FG parallel to CE and make

$AH = AG$. Then $AH = Q$, the total actual yield of the well. The line HB represents the loss of flow due to friction in the well and casing. If the well and casing were larger, CD would be longer, and likewise AH and AG would be greater, so that the flow Q would be increased.

As a numerical example of this graphical construction, suppose that the water-bearing stratum is 100 feet thick, of effective size grain 0.2 mm., and that the static head is equal to a 50-foot column of water. Let us determine the yield of a 4-inch well 1,000 feet deep. From Table VIII we find that $Q_0 = 55.3$ cubic feet per minute. On page 98 of Weston's hydraulic tables the discharge of 1,000 feet of 4-inch pipe under a head of 50 feet is found to be about 257 gallons per minute, or 34.3 cubic feet per minute, so that $q = 34.3$. Draw $AB = BC = 55.3$ and also $CD = 34.3$. Making the construction indicated, we get AH or $Q = 24.6$ cubic feet per minute, which is some-

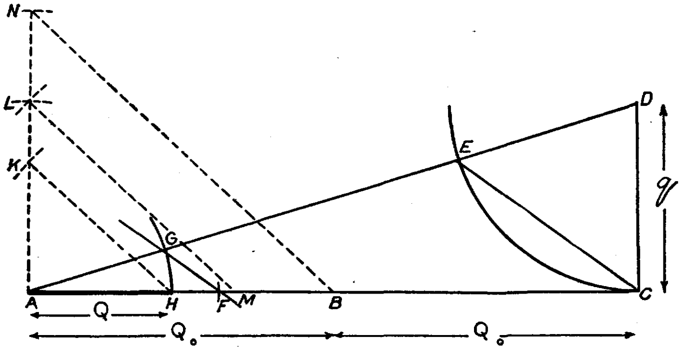


FIG. 46.—Diagram illustrating the author's graphical solution of problems in the yield of artesian wells.

what less than half of the theoretical yield, with pipe friction neglected.

Uncertainty as to the value of k, or the transmission constant of the rock, makes the application of this graphical solution more or less uncertain so far as precision of results is concerned. Nevertheless, it is capable of giving much valuable information concerning the relations of the two fundamental causes affecting the capacity of wells, pipe friction, and capillary friction in the pores of the rock.

Perhaps the most useful application of the method is to the inverse problem of determining the value of Q_0, or the theoretical flow, when the static head and the actual discharge of the well have been measured. The value of Q_0 can be found graphically in the following manner, as is shown by the dotted lines in fig. 46: Lay off on any straight line, AC, the known yield of the well, or $Q = AH$. Draw the line $AKLN$ perpendicular to AH, and from H as center describe an arc with radius equal to q, cutting AN in K. Then lay off

$AL=HK=q$, and draw LM parallel to KH. Next lay off $AN=LM$ and draw NB parallel to HK. Then AB is Q_0, or the theoretical flow of the well if pipe friction were entirely absent.

The usefulness of the inverse solution lies chiefly in the fact that it enables us to determine the increased yield which results from enlarging the diameter of a flowing well. Thus let us suppose that a 4-inch well 1,000 feet deep yields 26.6 cubic feet per minute under a static head of 50 feet; what yield should we expect from a 6-inch well, all other things being similar? We first lay off $AH=24.6$ in fig. 46, and make the construction shown by the dotted lines. This gives us $AB=55.3$, so that the yield of the 4-inch well, if pipe friction could be avoided, would be 55.3 cubic feet per minute.

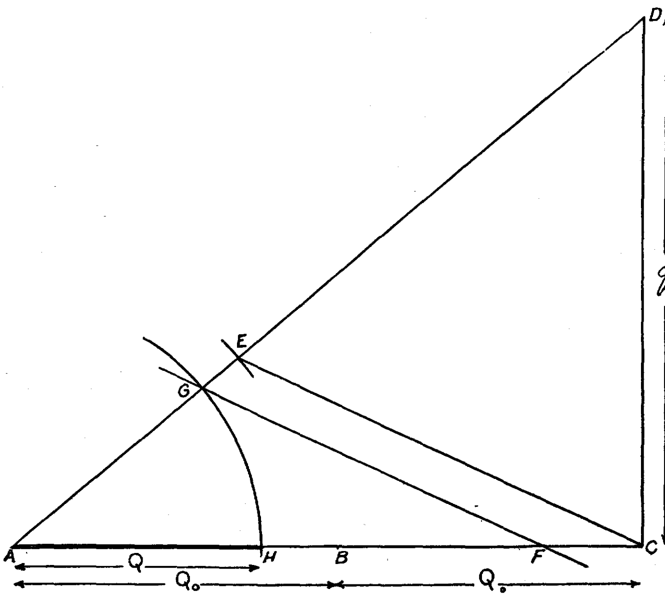


FIG. 47.—Diagram illustrating graphical solution of an artesian-well problem.

From Table VII, column 3, we observe that this value should be increased about 5 per cent if the diameter of the well be changed to 6 inches. This gives us 58.1 cubic feet per minute for Q_0 for a 6-inch well. The discharge of 1,000 feet of 6-inch pipe under a head of 50 feet is found on page 102 of Weston's tables to be 730 gallons, or 97.3 cubic feet, per minute. Then in fig. 47 lay off $AB=BC=58.1$ and $CD=97.3$. Following the construction explained, we arrive at the value of Q or AH, which is found to be 45.6. Therefore the yield of a 6-inch well will be 45.6 cubic feet per minute. The yield of the 4-inch well was 24.6 cubic feet per minute, so that a 6-inch well under these conditions should be expected to have a capacity about 85 per cent greater than that of a 4-inch well.

The foregoing problems may also be solved by means of the curves of fig. 48. These curves show the relation of Q, the discharge of a well, to Q_0, the theoretical yield, but on a scale in which the unit of discharge is equal to q, the capacity of the well tube under the given static head. In order to illustrate the use of the diagram, we shall use the same examples that we have just solved. The first problem calls for the yield of a 4-inch well 1,000 feet deep whose static head is 50 feet, the transmission constant and thickness of water-bearing stratum being given as before. Table VIII gives the theoretical yield Q_0= 55.3 cubic feet per minute. From a hydraulic table the capacity of

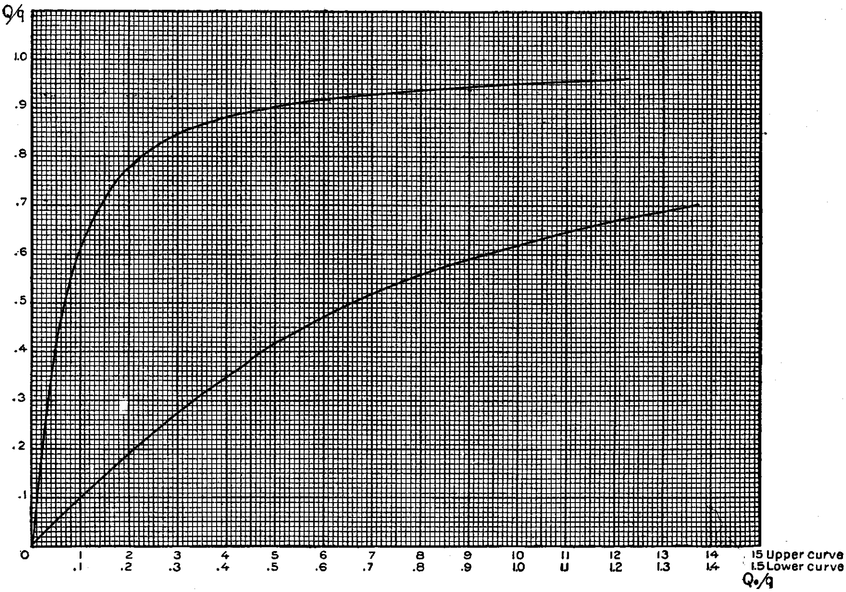


FIG. 48.—Graphical representation of the relations between Q and Q_0 in the equation on page 85, q being taken as equal to unity. The lower curve is a replat on larger scale of the portion of the upper curve between Q_0=0 and Q_0=1.35. From this diagram the value of Q_0 can be obtained when Q and q are given. The unit of measure is q=1, so that Q must always be less than unity on this scale. In other words, the ordinates in the diagram are equal to Q/q, the abscissas to Q_0/q.

1,000 feet of 4-inch pipe under a head of 50 feet is found to be 34.3 cubic feet per minute. Using this value of q as the unit of measure gives Q_0/q=55.3÷34.3=1.6. Finding this value on the horizonal scale of the diagram (fig. 48), we determine Q/q to be 0.725. To reduce to cubic feet per minute, multiply 0.725 by 34.3, the value of q, which gives 24.9 cubic feet per minute as the yield of the well. The result is slightly different from that obtained by the graphical construction.

The same diagram can also be used to solve the inverse problem previously discussed: If a 4-inch well 1,000 feet deep, having a static head of 50 feet, yields 24.6 cubic feet per minute, what will be the yield of a 6-inch well under the same conditions? From a hydraulic table we obtain, as before, the yield of 1,000 feet of 4-inch pipe, viz, 34.3 cubic feet per minute. This is the value of q. Then Q/q is $24.6 \div 34.3$, or 0.72. Finding 0.72 in the vertical scale of the diagram (fig. 48) gives $Q_0/q=1.55$. According to Table VII this should be increased about 5 per cent in order to give the value of Q_0/q for a 6-inch well, making its value 1.63, so that $Q_0=1.63\times 34.3$, or 56 cubic feet per minute. A hydraulic table gives q for 1,000 feet of 6-inch pipe, head 50 feet, equal to 97.3 cubic feet per minute, so that for a 6-inch well $Q_0/q=56\div 97.3=0.60$. Using the lower curve of fig. 48, we find corresponding to 0.60 that $Q/q=0.47$, and since $q=97.3$, $Q=97.3\times 0.47=45.7$, which is the yield of the 6-inch well in cubic feet per minute.

SIMPLE METHODS OF MEASURING THE YIELD OF FLOWING WELLS.

Prof. J. E. Todd, State geologist of South Dakota, has recently issued a bulletin describing very simple methods of determining the yield of an artesian well, which give fairly accurate results with little trouble and in a short time. The following tables and explanations are from his bulletin. All that is necessary for the purpose is that the water be discharged through a pipe of uniform diameter, a foot rule, still air, and care in taking measurements. Two methods are proposed, one for pipes discharging vertically, which is particularly applicable before the well is permanently finished, and one for horizontal discharge, which is the most usual way of finishing a well.

The table below is adapted to wells of moderate size as well as to large wells. In case the well is of other diameter than that given in the table its discharge can without much difficulty be obtained from the table by remembering that, other things being equal, the discharge varies as the square of the diameter of the pipe. If, for example, the pipe is one-half inch in diameter its discharge will be one-fourth of that of a pipe 1 inch in diameter for a stream of the same height. In a similar manner the discharge of a pipe 8 inches in diameter can be obtained by multiplying the discharge of the 4-inch pipe by 4.

In the first method the inside diameter of the pipe should first be measured, then the distance from the end of the pipe to the highest point of the dome of the water above, in a strictly vertical direction—*a* to *b* in the diagram, fig. 49. Find these distances in the table (IX, A), and the corresponding figure will give the number of gallons discharged each minute. Wind would not interfere in this case, so long as the measurements are taken vertically.

The method for determining the discharge of horizontal pipes requires a little more care. First, measure the diameter of the pipe, as before, then the vertical distance from the center of the opening of the pipe, or some convenient point corresponding to it on the side of the pipe, vertically downward 6 inches, *a* to *b* of the diagram, then from this point strictly horizontally to the center of the stream, *b* to *e*. With these data the flow in gallons per minute can be obtained from the table (IX, B). It will readily be seen that a slight error may make much difference in the discharge. Care must be taken to measure horizontally and also to the center of the stream. Because of

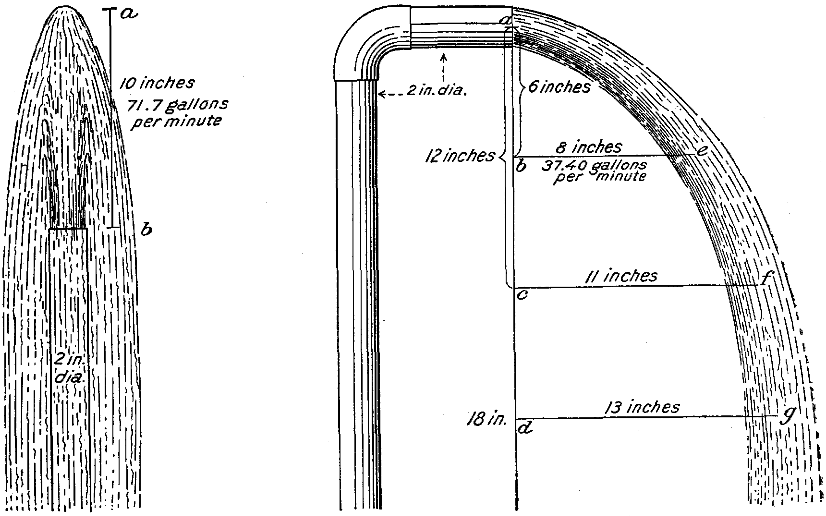


FIG. 49.—Diagram illustrating flow from vertical and horizontal pipes.

this difficulty, it is desirable to check the first determination by a second. For this purpose columns are given in the tables for corresponding measurements 12 inches below the center of the pipe. Of course the discharge from the same pipe should be the same in the two measurements of the same stream. Wind blowing either with or against the water may vitiate results to an indefinite amount; therefore measurements should be taken while the air is still.

TABLE IX.

Table for determining yield of artesian wells.

A. Flow from vertical pipes.

Height of jet.	Discharge in gallons per minute. Diameter of pipe in inches. 1	1¼	1½	2	3
Ins.					
½	3.96	6.2	8.91	15.8	35.6
1	5.60	8.7	12.6	22.4	51.4
2	7.99	12.5	18.0	32.0	71.9
3	9.81	15.3	22.1	39.2	88.3
4	11.33	17.7	25.5	45.3	102.0
5	12.68	19.8	28.5	50.7	113.8
6	13.88	21.7	31.2	55.5	124.9
7	14.96	23.6	33.7	59.8	134.9
8	16.00	25.1	36.0	64.0	144.1
9	17.01	26.6	38.3	68.0	153.1
10	17.93	28.1	40.3	71.6	161.3
11	18.80	29.5	42.3	75.2	169.3
12	19.65	30.7	44.2	78.6	176.9
13	20.46	31.8	45.9	81.8	184.1
14	21.22	33.0	47.6	84.9	190.9
15	21.95	34.2	49.3	87.8	197.5
16	22.67	35.2	50.9	90.7	203.9
17	23.37	36.3	52.5	93.5	210.3
18	24.06	37.5	54.1	96.2	216.5
19	24.72	38.6	55.6	98.9	222.5
20	25.37	39.6	57.0	101.6	228.5
21	26.02	40.6	58.4	104.2	234.3
22	26.66	41.6	59.9	106.7	240.0
23	27.28	42.6	61.4	109.2	245.6
24	27.90	43.5	62.8	111.6	251.1
25	28.49	44.4	64.1	114.0	256.4
26	29.05	45.3	65.3	116.2	261.4
27	29.59	46.1	66.4	118.2	266.1
28	30.08	46.9	67.5	120.3	270.4
29	30.55	47.5	68.5	121.9	274.1
30	30.94	48.2	69.4	123.4	277.6
36	34.1	53.2	76.7	136.3	306.6
48	39.1	61.0	88.0	156.5	352.1
60	43.8	68.4	98.6	175.2	394.3
72	48.2	75.2	108.0	192.9	434.0
84	51.9	81.0	116.8	207.6	467.0
96	55.6	86.7	125.0	222.2	500.0
108	58.9	92.0	132.6	235.9	530.8
120	62.2	98.0	139.9	248.7	559.5
132	65.1	102.6	146.5	260.4	585.9
144	68.0	106.4	153.1	272.2	612.5

B. Flow from horizontal pipes.

Horizontal length of jet.	Flow in gallons per minute. 1-inch pipe. 6-inch level.	1-inch pipe. 12-inch level.	2-inch pipe. 6-inch level.	2-inch pipe. 12-inch level.
Ins.				
6	7.01	4.95	27.71	19.63
7	8.18	5.77	32.33	22.90
8	9.35	6.60	36.94	26.18
9	10.51	7.42	41.56	29.45
10	11.68	8.25	46.18	32.72
11	12.85	9.08	50.80	35.99
12	14.02	9.91	55.42	39.26
13	15.19	10.73	60.03	42.54
14	16.36	11.56	64.65	45.81
15	17.53	12.38	69.27	49.08
16	18.70	13.21	73.89	52.35
17	19.87	14.04	78.51	55.62
18	21.04	14.86	83.12	58.90
19	22.21	15.69	87.74	62.17
20	23.37	16.51	92.36	65.44
21	24.54	17.34	96.98	68.71
22	25.71	18.17	101.60	71.98
23	26.88	18.99	106.21	75.26
24	28.04	19.82	110.83	78.53
25	29.11	20.64	115.45	81.80
26	30.38	21.47	120.07	85.07
27	31.55	22.29	124.69	88.34
28	32.72	23.12	129.30	91.62
29	33.89	23.95	133.92	94.89
30	35.06	24.77	138.54	98.16
31	36.23	25.59	143.16	101.43
32	37.40	26.42	147.78	104.70
33	38.57	27.25	152.39	107.98
34	39.64	28.08	157.01	111.25
35	40.45	28.64	161.63	114.52
36	41.60	29.46	166.25	117.79
Continue by adding for each inch—	1.15	0.82	4.62	3.27

NOTE.—To convert results into cubic feet, divide the number of gallons by 7.5, or, more accurately, by 7.48.

The flow in pipes of diameters not given in the table can easily be obtained in the following manner:

For ½-inch pipe, multiply discharge of 1-inch pipe by 0.25
For ¾-inch pipe, multiply discharge of 1-inch pipe by 0.56
For 1¼-inch pipe, multiply discharge of 1-inch pipe by 1.56
For 1½-inch pipe, multiply discharge of 1-inch pipe by 2.25
For 3-inch pipe, multiply discharge of 2-inch pipe by 2.25
For 4-inch pipe, multiply discharge of 2-inch pipe by 4.00
For 4½-inch pipe, multiply discharge of 2-inch pipe by 5.06
For 5-inch pipe, multiply discharge of 2-inch pipe by 6.25
For 6-inch pipe, multiply discharge cf 2-inch pipe by 9.00
For 8-inch pipe, multiply discharge of 2-inch pipe by 16.00

Whenever fractions occur in the height or horizontal distance of the stream, the number of gallons can be obtained by apportioning the difference between the readings in the table for the nearest whole numbers, according to the size of the fraction. For example, if the distance from the top of the pipe to the top of the stream in the first case is $9\frac{1}{3}$ inches, one-third of the difference between the reading in the table for 9 and 10 inches must be added to the former to give the correct result.

In case one measures the flow of a well by both methods he may think that the results should agree, but such is not the case. In the vertical discharge, there being less friction, the flow will be larger; so also in the second method differences will be found according to the length of the horizontal pipe used.

As pipes are occasionally at an angle, it is well to know that the second method can be applied to them if the first measurement is taken strictly vertically from the center of the opening, and the second measurement from that point parallel with the axis of the pipe to the center of the stream, as before. The measurements can then be read from the table.

FAILURE OF ARTESIAN WELLS.

It is not uncommon to find that the flow of an artesian well is gradually growing less or failing altogether. The cause of the failure may be purely local and pertain to that particular well, or it may be general and associated with some depreciation of the artesian basin as a whole. If the failure is due to local causes it may not be duplicated in neighboring wells, while causes affecting the artesian basin as a whole must of course be evidenced by a diminution of flow in all of the wells.

Perhaps the most common cause of failure of single wells is improper casing. Some wells are not properly cased from the beginning. Or the trouble may be due to poor jointing or packing where the casing meets the uncased rock, or to poor packing where a reduction in the size of the casing occurs. Casing is sometimes omitted where a porous stratum really requires it, or where a friable or fractured rock is subject to constant caving. The leaks around the casing or into porous strata may easily cause the total destruction of a well, or so injure it that it can be repaired only at great expense.

The filling of the lower portion of a well with sand and débris falling from an upper stratum, or brought into the well by the water, is a common cause of the falling off in the flow of an artesian well. This difficulty is easily remedied by cleaning out the débris by means of a sand pump.

Sometimes the water-bearing rock is so friable and the pressure so great that the well must be cased to the bottom and the end of the tube be driven fast into the bottom rock, the water being admitted into the well through numerous holes drilled in the lower sections of the casing, which act on the same principle as the cylindrical screens in a common driven well.

A cause of the partial failure of wells, which is not yet thoroughly understood, is the apparent clogging of the water-bearing rock in the immediate neighborhood of the well. Whether this clogging is due to the deposit of fine silt in the pores of the rock or to a growth of microscopic plant life, or to a gelatinous deposit of iron, etc., has not, to the writer's knowledge, been ascertained in any of the cases in which the phenomenon has been observed. In the old artesian wells at Savannah, Ga., it was found that the explosion of dynamite at the bottom of the wells did not remove the difficulty. Later, in the new wells at the same city, it was found that strong back flushing of the wells—that is, the forcing of large quantities of water into them—caused a very marked restoration of the flow. This matter will be especially referred to on a later page (p. 100), where the wells at Savannah are more fully discussed.

Failure due to causes pertaining to the entire artesian basin is also rather common. It is probably true that in nearly all artesian basins the original pressure gradient in the water-bearing rock is appreciably lowered by the artificial drafts made upon the subterranean supply, with a consequent actual decrease in the capacities of the wells. A small decrease of this kind is to be expected, and it does not necessarily indicate approaching disaster to the wells. Even in the Dakota basin, which seems thus far to be quite adequate to the enormous demands made upon it, it is thought that there is evidence of a slight depreciation of the wells on higher ground, and even some wells of low pressure have ceased to flow, as at Scotland, Tripp, and elsewhere. Most wells, however, have failed but little.[a]

It must be kept well in mind that there is a limit to the amount of water that can be withdrawn from an artesian basin. There is no such thing as an inexhaustible supply in this connection. The amount of water available is limited on the one hand by the amount of rainfall upon the catchment area and the facility with which the rainfall can obtain entrance to the porous stratum, and on the other hand by the capacity of the water-bearing rock to transmit the water over long distances and diminution through leakage and seepage. These two limiting conditions are usually of sufficient magnitude to render the overdrawing of the supply a practical and present danger which should be constantly kept in mind.

One of the most striking examples of a general and gradual failure of the wells of an entire basin is presented by those at or near the city of Denver.[b] This basin was discovered in 1884, and in a few

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[a] Geology and water resources of a portion of southeastern South Dakota, by J. E. Todd: Water-Supply and Irrigation Paper U. S. Geol. Survey No. 34, 1900, p. 31.

[b] See Artesian wells of Denver Basin, by G. H. Eldridge: Mon. U. S. Geol. Survey Vol. XXVII. Also see The artesian wells of Denver; a report by a special committee of the Colorado Scientific Society: Proc. Colo. Sci. Soc., Denver, Vol. I, 1883-84, pp. 76-108.

years about 400 wells had been drilled within an area extending a distance of 40 miles along South Platte River in a strip about 5 miles wide on both sides of the stream. Most of the wells were within the limits of the city itself. Many of the wells had a good pressure and strong flow when first constructed. In 1886 it was not thought that any general decrease in the flow of the wells could be detected. Between 1888 and 1890, however, a continuous decrease in the flow of the city wells took place, and by the end of the latter year all but six of the city wells had to be pumped, while numerous wells in the basin were permanently abandoned.[a]

It is thought that the cause of the remarkable failure of this basin is not lack of rainfall upon the catchment area or the size or absorption power of the latter, but the low porosity and transmission power of the water-bearing strata. It is believed that during the early years in the history of the wells the water withdrawn represented a supply stored in the rocks but not readily transported by the strata to meet the enormous draft. Van Diest estimated in 1890 that if all the wells in Denver were plugged it would be forty years before the water-bearing strata of the Tertiary of the Denver Basin would be again in the condition of saturation existing when the first well was sunk.[b]

MUTUAL INTERFERENCE OF ARTESIAN WELLS.

It is a common experience to find that the yield of an artesian well is noticeably influenced by the construction of a new well in the same neighborhood. When this phenomenon is associated with a general lowering of the water pressure throughout the region, it is more properly considered a case of partial failure of the artesian basin than a case of interference. One well may interfere with another well without the basin being subject to general depreciation. The interference of two wells, in the technical sense, is determined by a comparison of the flow of one of two wells when both are flowing freely with the flow of each well when the other is shut off. In a similar way we may determine the interference of several wells with one well.

To illustrate the distinction between the general depreciation of an artesian area and the interference of wells with one another we will consider the case presented by the Chicago wells. The large drafts made upon the various zones of supply in that district have caused the head of wells constructed in 1864 to fall about 100 feet. This would be referred to as depreciation or partial failure of the basin. But if at the time mentioned two wells had been constructed 50 feet apart, the flow of each well during the entire period of gradual depreciation would fluctuate with the shutting down or starting up of the other well. This is what is meant by interference.

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[a] Artesian wells of Denver Basin, by G. H. Eldridge: Mon. U. S. Geol. Survey Vol. XXVII, p. 428.

[b] Ibid., pp. 426–427. See also The artesian wells of Denver: Proc. Colo. Sci. Soc., Denver, Vol. I, 1883–84, pp. 76–108.

It is more surprising to find cases in which no interference between neighboring wells can be detected than to observe cases in which it is very pronounced. The enormous 12-inch Ponce de Leon well at St. Augustine, Fla., seems to suffer no interference from a neighboring 6-inch well. The Ponce de Leon well is 1,400 feet deep and flows 10,000,000 gallons per day under a closed head of 17 pounds per square inch.[a] While a dynamo was being operated by the 12-inch well, the 6-inch well (capacity 400,000 gallons daily) was suddenly turned on and off, but the closest observation did not detect the slightest trembling of the pressure gage on the 12-inch well. The latter well, however, is much deeper than the 6-inch well, and furnished 3,000,000 gallons daily at the depth reached by the 6-inch well. This fact and the disparity in their total yield may sufficiently account for the failure to detect any interference.

At Savannah the flow from the first 7 of 12 wells arranged 300 feet apart in a straight line was observed in 1892 to be 6,500,000 gallons per day, an average of 930,000 gallons for each well, at an elevation of 8 feet above the sea. The water stood in wells Nos. 10, 11, and 12 at altitudes of 14.93, 15.96, and 16.20 feet, respectively. The total yield of the 12 wells under the same head was 9,500,000 gallons per day, an average of 792,000 gallons for each well. In 1897 a measurement of the mutual interference of these wells and the Springfield well, 2,000 feet east, was made, and it was ascertained that when the Springfield well was shut off the total flow of the 12 wells was 6,910,000 gallons per day, and that when the Springfield well was in use the flow was 7,258,000 gallons per day. The net gain from that well was, therefore, 348,000 gallons per day, which would be increased to 468,000 gallons per day if the leakage between the well and the works be allowed for. The latter quantity (468,000 gallons) is therefore the net gain due to the use of the Springfield well. The latter well, however, was actually flowing at the rate of 1,000,000 gallons per day. Therefore the 12 wells suffered a decrease of 552,000 gallons per day, which is an interference of 8 per cent due to the Springfield well. The Savannah wells are discussed more fully on the following pages.

What is reported to be interference of wells sometimes proves to be a fault largely due to defective construction. Such was in part the case at Rockford, Ill.[b] The first well drilled was 8 inches in diameter and 1,530 feet deep. The water rose about 28 feet above the ground and flowed about 1,000,000 gallons per day. Three more wells were sunk, and when all were connected with the pump the total flow was 3,000,000 gallons per day. This would indicate a mutual interference of 25 per cent. This flow was not permanent, however, and soon diminished to 2,500,000 gallons per day, while a paper mill 1 mile south of the works and at 17 feet lower elevation drilled a well

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[a] Engineering News, Vol. CCLXXI, pp. 183, 328.

[b] Engineering Record, Vol. XXII, p. 7.

which affected the original yield still more. An additional well was then drilled, but, the supply failing to increase as much as was expected, an investigation was made. Well No. 1 was found to be flowing only 70,000 gallons per day at the surface and only 210,000 gallons at the place of connection. The well was recased and repacked, the result being a flow of 177,000 gallons at the surface and 400,000 gallons at the point of connection. Similar work on well No. 4 increased the delivery at the surface from 142,000 gallons to 254,000 gallons and at the point of connection to 700,000 gallons. The wells were faulty in construction, leaking considerably in the upper strata.

ARTESIAN WELLS OF SAVANNAH, GA.

One of the most carefully planned artesian water supplies of this country is that of the city of Savannah, Ga. A brief account of these wells is of much interest, not only as illustrative of an intelligently designed and well-constructed engineering work, but also because of the valuable lessons to be learned from the way in which difficulties were met and overcome as they arose. The works were planned by Mr. Thomas T. Johnston, consulting engineer, of Chicago, Ill., and the following description is taken largely from his paper in the Engineering News.[a]

Prior to 1887 the water supply of Savannah was taken from the river, but in that year 14 artesian wells were put in use (a fifteenth well being nearly ready) near the old pumping station (see fig. 50). Of these wells, 2 were 10-inch, 12 were 6-inch, and 1 was 4-inch, and not one was more than 400 feet deep. In 1888 the entire supply of the city was drawn from these wells, the total for the year being 2,135,842,000 gallons, or about 5,850,000 gallons per day. At the close of 1889 5 new wells had been put in use, the 20 wells being distributed very irregularly within a 10-acre lot. About this time, however, it became necessary to open the river supply to some extent. Experience and observation showed conclusively that the water supply was materially affected by the tide, and it was supposed that the shallowness of the wells caused a filling by sand to a greater or less extent, thus affecting the supply. But little better results were had, however, from deepening some of the wells; one of them was deepened to 1,009 feet. Two additional 10-inch wells, 502 and 505 feet deep, were connected with the old works on December 29, 1890, but in November, 1891, it was necessary to again draw water from the river. Nothing resulted from dynamiting the wells.

In June, 1891, it was found that water in well No. 1, close to the old pumping station, was 9 feet lower than in well No. 2, distant from well No. 1 about 1,600 feet, and it was determined to abandon the old wells.

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[a] Vol. XXIX, p. 527.

Twelve new wells in a new location were finally decided upon, and 7 were put in use on December 9, 1892, and the others on March 21, 1893. The relative locations are shown on the accompanying map, fig. 50. The wells are arranged along a highway which runs nearly parallel to the ocean and some distance from the old pumping sta-

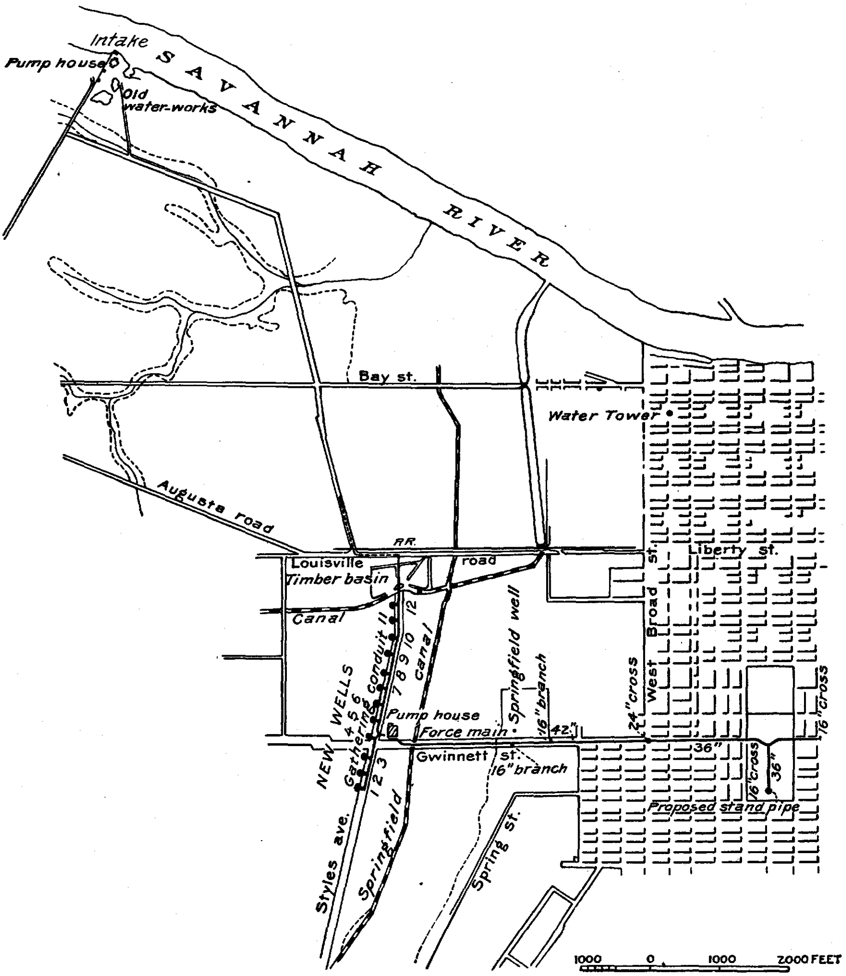

FIG. 50.—Map showing location of artesian wells at Savannah, Ga. The old wells are near the old waterworks, shown in upper left-hand corner of map. The new wells are on Styles avenue and are numbered 1-12, inclusive. (From Journal of Western Society of Engineers, Vol. II, No. 6.)

tion. They are 300 feet apart and flow into a brick and concrete conduit on the opposite side of the highway. The strata through which they are bored belong to the Cretaceous. Sands and clays were met to a depth of about 250 feet, where a cherty lime rock was found. The rock varies in character at greater depths. The first water was found at 325 feet, and the supply continually increased until the depth exceeded 500 feet. The water-bearing rock is uniformly a porous limestone, or rather one full of cavities, very similar to the rock outcrop from which flow the mammoth Suwanee springs in northern Florida. Water is occasionally found below 500 feet, but not in sufficient quantity to justify deeper wells. The 12 wells of the new waterworks average 500 feet in depth. They are cased to the rock (about 250 feet) with casing of 12 inches internal diameter. Below this there is a 12-inch clear opening. The normal position of water in the wells, as determined by the position in the old wells when first dug, showed that it reached the static level at an elevation of 41 feet above mean low tide. Subsequent wells, bored in a wide range of country, show about the same elevation. After a number of wells had been bored the elevation of this static level was lowered. At the time of an examination in 1890 about 6,000,000 gallons per day were being pumped at the old works. The static level was then 7 feet; at the site of the new works it was at an elevation of 28 feet, while 9 miles south of the old works it was at the original elevation, 41 feet. The influence of the heavy pumpage was noticed to be greater in wells in the same line perpendicular to the shore of the ocean than in wells in a line parallel to the ocean. To predetermine the probable flow from the new wells, the Springfield well, 2,000 feet east of the new station, was examined, with the following results: [a]

TABLE X.

Flow of Springfield well, Savannah, Ga.

Elevation of static level above mean low tide.	Equivalent head.	Year.	Flow.
Feet.	*Feet.*		*Gallons per 24 hours.*
22.67	0	1890	----------
25.67	2.00	1890	482,000
24.40	3.27	1891	623,000
22.77	4.90	1890	1,000,000
22.68	4.99	1891	977,000
21.80	5.87	1891	1,133,000
21.79	5.88	1891	1,172,000
20.57	7.10	1890	1,360,000
13.00	14.67	1890	2,500,000

It was estimated that about 3,000,000 gallons per day could be obtained with the static head reduced to an elevation of 7 feet, which reduction is practicable at the pump house, as constructed. The old wells spread over an area about 1,700 feet wide at its largest part, and their flow at an elevation of 7 feet was about 6,000,000 gallons per day.

The first seven wells embraced a line 1,800 feet long. Reducing the level to 8 feet, the flow was 6,500,000 gallons per day. The level in wells Nos. 10, 11, and 12 was at this time 14.93, 15.96, and 16.20

[a] Section of this well is shown in Engineering News, Vol. XXX, p. 4.

feet, respectively. On March 21, 1893, the 12 wells were put into service, with elevation of water at 8 feet, and the flow was 9,500,000 gallons per day. At the beginning of 1897 the flow of the 12 wells had decreased to 6,900,000 gallons per day, and the fear was entertained that the history of the old plant was to be repeated. Mr. Johnston was again called in consultation. The results of his measurements of the flow of each of the 12 wells were as follows:

TABLE XI.

Yield of Savannah wells in 1897.

Well.	Flow.	Well.	Flow.
	Gallons per 24 hours.		*Gallons per 24 hours.*
No. 1	713,460	No. 9	399,740
No. 2	460,460	No. 10	468,050
No. 3	543,950	No. 11	371,910
No. 4	361,790	No. 12	485,760
No. 5	485,760		
No. 6	293,480		5,490,100
No. 7	452,870	Springfield well	1,019,590
No. 8	452,870		

From this table it is seen that wells Nos. 4, 6, and 11 showed a very small flow, entirely out of proportion to that of the other wells.

It having been determined that the flow of water in some wells was more obstructed than in others, the work of attempting to remove the obstructions was undertaken. The method used was back-flushing, or the forcing of a strong reverse flow back into the well. The first effort was made on well No. 2 by means of fire hose attached to a neighboring hydrant. Only a slight improvement resulted. The next attempt was made on well No. 10, fire engine No. 2 being used. But little improvement resulted. A more powerful fire engine (No. 3) was next employed, and the process was tried on well No. 6. A radical improvement was the result, the net gain in the yield of the well being 100,000 gallons per day. Encouraged by this result, the next effort was made on well No. 4, fire engines Nos. 1 and 3 being used. The result was even more radical, the net gain in yield for this well being 200,000 gallons per day.[a] The actual results on well No. 6 were as follows:

TABLE XII.

Results of flushing well No. 6, Savannah waterworks, in 1897.

	Gallons per 24 hours.
Flow before flushing:	
May 12	295,000
May 18	304,000
Flow after flushing:	
May 18	433,000
May 19	395,000
May 20	400,000
May 24	389,620

[a] Report of Thos. T. Johnston to chairman of water commission, Savannah, Ga., dated May 24, 1897.

After these experiments Mr. Johnston recommended that a special 8-inch water main of the city system be laid along the row of 12 wells, and that it be connected to each well by means of a 6-inch pipe and gate. By opening a gate any well can now be flushed at any time, and Superintendent Kinsey of the waterworks states that all of the wells have been flushed and tested regularly every year since Mr. Johnston's tests in 1897. The following is the record of the test for the year 1900:

Results of flushing Savannah wells in 1900.

	Gallons per 24 hours.
Yield before flushing	5,104,275
Yield after flushing	5,850,878
Gain	746,603

The yield before flushing given above is 385,825 gallons less than the total yield before flushing in 1897.

The results at Savannah are especially valuable on account of their reliability and the fulness of the information. In the first place, the yield of the wells can be closely relied upon, for they were determined by a special current meter designed by Mr. Johnston, which can at any time readily be lowered into one of the wells. In the second place, it was found by actual measurements at the old waterworks that there had been no lowering of the water table during the years 1892 to 1897, as the measurements made in the two years were in substantial accord. Thus the depreciation of the wells must be referred to the clogging of the pores in the rock in some unknown way, and not to a general depreciation of the basin.

Reliable data upon which to base discussion of artesian phenomena are not common. Most of the round-number estimates of yield, such as 500,000 or 1,000,000 gallons per day, are not even approximations of the actual facts.

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