smaller than in earlier years and this may have resulted in somewhat higher ground-water levels since that year, but the major trends of ground-water level apparently are produced by variations in precipitation. It is apparent, therefore, that the safe yield has not been exceeded in this area. This is in contrast to conditions in the west end of Long Island (Brooklyn) where ground-water levels are at present more than 40 feet lower than in 1903 (W. F. Laase, Subsurface water-supply of western Long Island and its utilization, Munic. Eng. J., first quarterly issue, 1934, map p. 24, profiles p. 26, and recent observations by U. S. Geological Survey).

U. S. Geological Survey,
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MAXIMUM GROUND-WATER LEVELS

Robert E. Horton

Introduction and synopsis

Aquifers are living, moving things. It is the application of hydraulic and hydrologic principles to geologic facts that enables one to interpret their behavior. The difficulties of this treatment which would result from variation of the phenomena with time are largely avoided by limiting the discussion to equilibrium-conditions. Geology and hydrology, respectively, bear much the same relation to ground-water phenomena that anatomy and physiology bear to biology.

In the absence of rainfall the flow of perennial streams without surface-storage in their basins is derived either from ground-water or channel-storage. It can readily be shown that channel-storage becomes exhausted in a relatively short time compared with that required to exhaust ground-water storage. The rate of ground-water inflow to a stream during a given dry period increases as the initial volume of available ground-water storage increases. Hence the question of the maximum amount of ground-water storage which can be retained by a given drainage-basin becomes immediately important. Hydrologic literature is filled with discussions of the economic value of ground-water storage but one does not find a single title dealing specifically with the question of the maximum or limiting volume of ground-water storage.

The expression "maximum ground-water levels" may be used in two senses: (1) With reference to the limiting height to which it is possible for the ground-water to rise in a given well; (2) with reference to the highest observed stage in any given period or, in the generic sense, to the highest annual stages over a series of years. It has not hitherto, so far as the author is aware, been pointed out that there is a natural maximum or limiting ground-water stage and, pari passu, a maximum or limiting volume of ground-water storage in every aquifer.

[C. C. Vermeule seems to have sensed the idea that there was a maximum volume of ground-water storage which could accumulate in a given drainage-basin. In "Geological survey of New Jersey" (v. 3, Water Supply, 1894, p. 83), he states: "The ground-water flow for full ground-water seems to range close to two inches per month and I have assumed this in every case." An examination of curves used by Vermeule for computation of ground-water flow indicates that for Atlantic Coast streams he found the highest rate of ground-water flow--usually in April--to be about two inches depth of runoff per month in most cases, running up to 2.5 inches in case of some streams with deep, sandy drainage-basins. Since the rate of ground-water flow is nearly--and in many cases exactly--in proportion to the volume of ground-water storage, it follows necessarily that there must be a maximum volume of ground-water storage and corresponding maximum or limiting ground-water stages in these drainage-basins.]

Every crest or peak on a phreatograph corresponds to a point at which the inflow to and outflow from the aquifer are equal. That this must be so is readily seen from the storage-equation

$$\text{Inflow} = \text{outflow} + \text{gain or} - \text{loss of storage}$$

As long as the inflow-rate is greater than the outflow-rate, the storage increases and the ground-water level at a given well continues to rise. If the outflow-rate exceeds the inflow-rate, the storage and stage decrease. The crest is a point of change from a rising to a falling stage. Hence at the crest the stage passes through a point where the inflow- and outflow-rates are equal.

Designating the infiltration-capacity of the soil by f and the transmission-capacity of the aquifer by K_t, the ratio (f/K_t) enters as an important factor in the analytical treatment of a great variety of ground-water problems. This ratio deserves a name and it will here be designated the "transmission-ratio," since it is the ratio of the rate at which water can enter or be trans-

mitted into the soil from rain or melting snow, to the rate at which it can flow out of the soil under unit-head.

Since the infiltration-capacity limits the rate of inflow, the limiting, maximum ground-water stage will be attained if and only if the ground-water has risen to such a stage that the ground-water outflow-rate equals the infiltration-capacity. As subsequently shown, this stage may never be reached in a given well or aquifer because rain-intensities exceeding infiltration-capacity are seldom of sufficient duration to bring the outflow-rate up to equality with f. In such cases a maximum stage or crest-stage occurs, less than the limiting stage. In such cases the crest-stage occurs when rain ends or when the rain-intensity falls below the infiltration-rate; the majority of ground-water crests are of this type.

For purposes of this discussion, surface-aquifers may be classified as (a) simple, (b) multiple, (c) compound.

In a "simple" aquifer the ground-water flows toward a single main outlet.

In many drainage-basins, especially in regions overlain with permeable glacial deposits, ground-water is continuous everywhere but each stream-valley receives ground-water from different slopes. There are no impervious divides between these sub-aquifers and such a system may be described as a "multiple" aquifer. The simple aquifers which make up a multiple aquifer may be called "sub-aquifers."

The flood-plain of a stream may contain an aquifer which is fed in part by direct infiltration and in part by ground-water drainage from adjacent hill-slope aquifers. This may be described as a "compound" aquifer. Pomperaug Valley, Connecticut, affords an excellent illustration of a compound aquifer.

While much of the discussion applies to other conditions, the treatment of this subject is primarily intended to apply to (a) surface or non-artesian aquifers, (b) aquifers in continuous porous media--not to aquifers in rock-fissures.

Following the analytical discussion of the subject, several long ground-water records are presented and examined to see if they afford evidence of behavior in accordance with the principles outlined.

Theory of limiting maximum ground-water stages

For the purpose of discussion of limiting ground-water stages, the case of a simple aquifer with an impervious horizontal sole will first be considered.

Let α be the rate of ground-water accretion, f the infiltration-capacity, and K_t the transmission-capacity of the soil, all in the same units, and with other notation as shown on Figure 1. Then for a strip of unit-width running from the stream or outlet to the phreatic divide, and for equilibrium-conditions, the inflow upstream from any point x must equal the flow past x or, measuring inflow and outflow both in terms of depth per unit of time on the entire phreatic drainage-area

$$\alpha[(\ell_0 - x)/\ell_0] = K_t(h/\ell_0)S \tag{1}$$

But the slope $S = (dh/dx)$. This is positive since the slope is upward in the direction in which x is measured. This gives

$$(\alpha/K_t)(\ell_0 - x)\,dx = h\,dh \tag{2}$$

and

$$(\alpha/K_t)(\ell_0 x - x^2/2) = h^2/2 + \text{constant} \tag{3}$$

where $x = 0$, $h = h_0$ and constant $= -(h_0^2/2)$. This gives

$$h = \sqrt{h_0^2 + 2(\alpha/K_t)(\ell_0 x - x^2/2)} \tag{4}$$

This is the equation of the ground-water profile for the assumed conditions with any rate of accretion α either equal to or less than the infiltration-capacity. [This equation is general and applies also where α is negative or where there is uniform abstraction of water from the water-table by vegetation. The equation has many applications which will not be discussed here.]

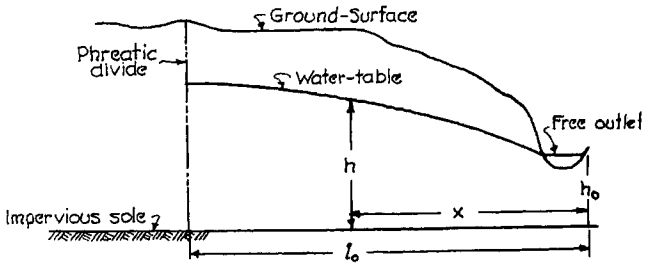


FIG. 1 - PROFILE OF SUB-AQUIFER WITH HORIZONTAL SOLE

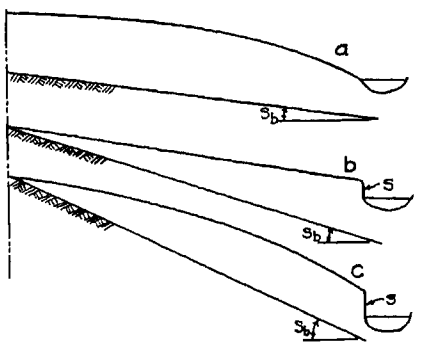


FIG. 2 - GROUND-WATER EQUILIBRIUM-PROFILES, INCLINED AQUIFERS (SCHEMATIC)
(a) $S_b < 2\sqrt{\frac{f}{K_t}}$ (b) $S_b = 2\sqrt{\frac{f}{K_t}}$ (c) $S_b = > 2\sqrt{\frac{f}{K_t}}$
Cases (b) and (c) are shown with vertical seep-face S at the outlet

Let h_g be the limiting maximum stage for a given infiltration-capacity f. Then

$$h_g = \sqrt{h_o^2 + 2(f/K_t)(\ell_o x - x^2/2)} \quad (5)$$

This is the equation of a parabola with the vertex at the point p, Figure 1, that is, at the intersection of the water-table and the phreatic divide.

It has been noticed that the profiles of surface-aquifers such, for example, as those between tile-drains, are not always precisely parabolic. One reason is that they are not taken under equilibrium-conditions. Since ground-water rises as long as inflow exceeds outflow, and falls whenever outflow exceeds inflow, it follows that at every maximum point on a phreatograph a condition exists temporarily where inflow and outflow are equal, that is, an equilibrium-condition, for which the formula applies.

Effect of various factors on limiting maximum ground-water stages

The manner in which the limiting maximum stage is affected by changes in the controlling variables is most easily seen by converting formula (5) into an expression for extreme range of ground-water fluctuation. Since the outlet level h_o represents the minimum possible stage where there is only a single outlet, it follows that

$$\text{Range} = h_g - h_o = \sqrt{h_o^2 + (f/K_t)(2\,\ell_o x - x^2)} - h_o \quad (6)$$

The equation was derived with reference to a strip of unit-width and hence applies strictly only to an aquifer having a rectangular catchment-area. Catchment-areas of simple aquifers are as varied in form as those of topographic drainage-basins, which they generally resemble.

If w_a is the average width of the phreatic drainage-basin upstream from a given point x, and w its width at x, then the flow past x bears the ratio (w_a/w) to that which would result for a rectangular phreatic basin of width w. It can readily be seen that for an irregular-shaped phreatic basin the flow past any point x is the same as it would be for a rectangular basin of width w but with an infiltration-capacity $(w_a/w)f$ instead of f. Making this correction, equations (5) and (6) can be written in the form

$$h_g = \sqrt{h_o^2 + (w_a/w)(f/K_t)(2\,\ell_o x - x^2)} \quad (7)$$

and

$$\text{Range} = \sqrt{(w_a/w)(f/K_t)}\ \sqrt{2\,\ell_o x - x^2} \quad (8)$$

Equation (8) shows that, other things equal:

(1) The limiting stage and range of fluctuation at a given well vary in proportion to the square root of the width-ratio and also in proportion to the square root of the transmission-ratio. If either ratio is doubled, the limiting stage and range are increased 41 per cent. If halved, the stage and range are decreased 29 per cent. If both are doubled, the range is doubled. If both are halved, the range is halved, and if one is doubled and the other halved, there is no change in the range or maximum stage.

(2) For a well at a fixed distance from a stream or outlet, the range and limiting stage increase with the length ℓ_o or the length of underground flow of the water approaching the well.

The relative values of the range for different ratios ℓ_0 to x, computed by formula (8), are as follows:

ℓ_0/x =	1	2	3	4	5	6	7	8	9	10
Range =	1.0	1.73	2.24	2.64	3.00	3.16	3.60	3.87	4.12	4.36

Wells at a given distance from the outlet in a long narrow aquifer have correspondingly larger ranges and higher maximum stages than similar wells in a shorter, wider aquifer of equal surface area.

(3) For an aquifer of a given length ℓ_0, the range and limiting stage increase with the distance of the well from the outlet or with the ratio (x/ℓ_0) as follows:

x/ℓ_0 =	0.0	0.1	0.2	0.4	0.6	0.8	1.0
Range =	0.0	0.19	0.36	0.64	0.84	0.96	1.0

For wells near the outlet the range increases rapidly with distance from the outlet. For wells near the phreatic divide, increase of range with distance from the outlet is relatively small.

Stratification

In this connection, stratification of an aquifer may be considered as any degree or arrangement of horizontal variations or diversity in the water-bearing medium which affects the hydraulic conditions. In addition to true stratification, this includes the important case of gradual variation, usually a decrease in such factors as porosity and transmission-constant proceeding downward from the surface. Extensive areas of Hudson River shale in eastern and central New York afford an excellent example. This rock weathers rapidly when exposed to the air or frost, but at a slight depth below surface it is wholly impervious. There is water in the fractured surface, decreasing in amount until at a depth, usually of a few feet, no water is found. Similar conditions exist in other classes of rocks and soils.

Let K_1 be the transmission-capacity of the aquifer at a height of one foot above the impervious level. Then at a height h the transmission-capacity is K_1h and the average below the height h is $K_1h/2$.

Considering a horizontal surface with the origin at the impervious level and at the outlet-end of the aquifer, then for equilibrium of inflow and outflow

$$(\ell_0 - x)\, f = (K_1h^2/2)\,(dh/dx)$$

Integration gives

$$(2f/K_1)\,(\ell_0x - x^2/2) = h^3/3 + \text{constant} \qquad (9)$$

When $x = 0$, $h = h_0$, therefore constant $= -h_0^3/3$. Inserting $(w_a/w)f$ in place of f to allow for form of the aquifer, the equation for the maximum limiting stage becomes

$$h_g = \sqrt[3]{h_0^3 + 3(w_af/wK_1)\,(2\,\ell_0x - x^2)} \qquad (10)$$

This is the equation of a parabola of two-thirds power. The profile-equation in other respects is identical with that of a homogeneous aquifer. The discussion of limiting ground-water stages and range, based on the equation for homogeneous aquifers, is therefore applicable to stratified aquifers of this type except that the maximum stages will not be the same in the two cases. It is also possible to obtain an equation for limiting stage applicable to an aquifer consisting of two or more horizontal strata with fixed lines of demarcation between them. Because of the great variety of conditions which this situation presents, it will not be discussed here.

Aquifers with inclined soles

The differential equation of the equilibrium-profile for an aquifer with inclined sole, typical of many hillside aquifers, is easily stated and has been integrated and studied at the Horton Hydrologic Laboratory. The resulting profile-equation is complicated and does not as readily show the effect of changes in individual factors as that given above. The profile-equation for horizontal aquifers has therefore been used for purposes of discussion.

While the conditions which fix limiting stages of hill-slope aquifers are essentially the same as for aquifers with horizontal soles, there are certain important differences. For horizontal aquifers both the limiting stage and the range increase proceeding upstream from the outlet. For sloping aquifers there are three principal cases and several sub-cases, dependent on the relations between the variables. The principal cases are outlined in Figure 2.

(1) Slope of sole $S_b < 2\sqrt{f/K_t}$--Underground back-water extends from the outlet to the divide and the depth at the divide is not zero. The maximum stage increases but the range generally decreases proceeding upstream from the outlet.

(2) Transition-stage, slope $S_b = 2\sqrt{f/K_t}$--It can readily be seen that for an aquifer with a sloping sole and uniform infiltration-rate, there is a certain constant slope for which the inflow upstream from a given well will equal the flow past the well. The profile of the water-table is nearly a straight line, extending from the sole at the divide to the outlet or to a minimum level adjacent to the outlet. In this case the limiting ground-water stage increases and the range decreases, both uniformly, proceeding upstream.

(3) Steep aquifers, slope $S_b > 2\sqrt{f/K_t}$--The depth at the divide is zero, and for a constant level at the outlet the stage increases proceeding upstream. The range is zero at the outlet and also at the divide and has a maximum at some intermediate point.

In some of these cases there is a minimum level at which the ground-water flow can arrive at the outlet. If the actual water-level in the outlet channel is below this minimum, there will be a "seep-surface" at the outlet, that is, ground-water will enter the stream at a level above the stream-level.

Surface and gully-outlets, bornes

In addition to the geologic and physical factors which fix the limiting ground-water stage in a given aquifer or well, there are certain topographic conditions to be considered. So far it has been assumed that the aquifer has only one outlet. Theoretically the water-table in an aquifer with a horizontal sole would intersect the ground-water if the aquifer extended back far enough from the outlet. For aquifers with inclined soles there are cases of this kind and also cases where the water-table would never intersect the ground-surface.

Swamp aquifers--In case of many swampy areas or in case of undissected flat or sloping areas with impervious floors at a slight distance below the surface, conditions of equilibrium may lead to a limiting ground-water stage h_g above the level of the soil-surface over part of the tributary area. On such areas the soil may be continually flooded or the ground-water may rise to the surface only at times. Swampy areas with deep muck-deposits underlain by impervious beds, such as the Everglades of Florida, provide excellent examples of aquifers of this type. As long as the surface is flooded, infiltration takes place at a constant rate equal to the infiltration-capacity and there is flow of ground-water to the outlet at the same constant rate. Rain tends to produce surface-runoff at an intensity approaching the rain-intensity as a limit, since in this case the ground-water outflow equals the infiltration-rate. Large reduction of runoff-intensity below rain-intensity is, however, usually provided by surface-storage. In case of swamp aquifers the maximum limiting stage is, of course, the ground-surface.

Ephemeral outlets, bournes--If the surface of the terrain is serrated by gullies, shallow relative to the main outlet of the aquifer, the conditions may be such that before reaching the limiting stage the water-table will begin to intersect the gullies. These will provide ephemeral or auxiliary outlets for ground-water. In general, the ground-water above the level of the bottom of the gully will flow a distance equal to one-half the spacing between gullies to reach an outlet. The great reduction in length of underground flow as compared with that required to reach the main outlet will limit the maximum ground-water stage to a height a little above that of the gullies. Conditions of this kind are fairly extensive. They are of particular interest because of the state of affairs to which they give rise in connection with flood-runoff. Until the water-table reaches the level of the auxiliary outlet, rain, however intense, cannot produce a surface runoff-rate exceeding the difference between the rain-intensity and infiltration-capacity. Rainfall not exceeding the infiltration-capacity, however prolonged, will therefore produce no surface-runoff. After the water-table rises to a point where gullies are intersected, and the limiting ground-water stage is attained, ground-water outflow-rate will equal the infiltration-capacity and, except for surface-storage, the runoff-intensity from then on will equal the rain-intensity. Thus it may happen that long-continued rains will produce little or no surface-runoff and no increase in ground-water flow until the field-moisture-deficiency becomes zero; then, somewhat abruptly, rain of no greater intensity will produce a relatively high runoff, substantially the same as if the infiltration-rate had been reduced to zero. Such high

runoff-intensities are commonly but inaccurately ascribed to the "soil becoming saturated." Actually the soil is not saturated above the water-table but the ground-water inflow and outflow have become equal, so that the total runoff-intensity approaches the rain-intensity as a limit.

If it happens that the water-table intersects gullies when the water-level is considerably below its limiting stage, the condition known in the chalk-downs of England as a bourne may result. A bourne is a stream fed by ground-water where the head of the stream or source of the flow travels up and down the stream-valley with the fluctuations of ground-water level. I have noticed this condition in certain shallow stream-valleys on the south slope of Long Island.

Local variation of ground-water regimen

It is sometimes found that in a group of adjacent wells the phreatographs are all closely similar except that fluctuation and range of stages in some wells is greater than in others. This is commonly attributed to local differences in permeability of the material comprising the aquifer, it being assumed that the wells showing the greatest range are in the least permeable beds. It is true, as formula (8) shows, that if permeability alone varied, then, since the range varies inversely as the square root of the permeability, it would be greatest for wells supplied by the least permeable aquifers. This, I think, is not generally the correct explanation. It will be noted that the range varies as the square root of the transmission-ratio. Permeability and infiltration-capacity, though not usually identical, are closely related. They may be practically identical in sandy soils containing little colloidal material. In general, if the surface of the soil and water-bearing beds are of the same or similar material, a change in transmission-capacity is accompanied by a similar and nearly proportional change of infiltration-capacity; the transmission-ratio may be changed but little and hence the fluctuations and range for a well may not in general vary to any great extent with change of transmission-capacity of the water-bearing beds.

There are other factors quantitatively adequate to explain differences in range of adjacent wells or wells in adjacent aquifers.

Summarized briefly, the factors which fix the maximum limiting stages and range for a given well and aquifer are: (1) Length of aquifer; (2) distance from well to outlet; (3) shape, as expressed in the width-ratio (w_a/w); (4) transmission-ratio (f/K_t); (5) slope of the sole, if not horizontal; (6) homogeneity and stratification; (7) topographic controls, gully-intersections, etc.

These factors are of more or less equal importance. Differences in any one or in any combination may produce considerable variation in range of one well as compared with other adjacent wells.

The effects of the first four factors are rigorously taken into account in equation (8) for homogeneous aquifers with horizontal soles. The others can be taken into account in similar analyses.

Keeping in mind the distinction between limiting ground-water stage and other maximum or crest-stages, it will be seen that the analysis thus far given does not show if or when the limiting maximum stage will be attained.

Conditions which determine times and frequency of occurrence of limiting ground-water stages

The limiting stage may be attained at frequent intervals; for example, it may be attained, as a rule, once each year, or it may be rarely attained or never attained. The frequency with which it will be attained depends on certain factors in addition to the seven described which govern its height. These are: (1) Initial ground-water storage; (2) initial ground-water level and field-moisture-deficiency; (3) duration of rainfall-excess. Since the limiting stage can only be attained at times when there is rainfall-excess over the infiltration-capacity, the criterion which determines the minimum duration t_e of rainfall-excess which will permit the ground-water to reach the limiting stage is

$$ft_e \geqq fmd + Q_g - Q_o$$

where fmd is the field-moisture-deficiency at the beginning of any given period of rainfall-excess and Q_o and Q_g are the volumes of ground-water storage at the beginning of rainfall-excess and at the limiting ground-water stage, respectively. It will be seen that, other things equal,

rainfall-duration and, more specifically, duration of rainfall-excess, is the principal controlling factor. Ground-water maximum stages occur at all seasons of the year but in general, particularly in middle latitudes, maxima in the spring months, usually just after the frost has gone out of the ground, tend to predominate.

Using the water-year beginning March 1, the 20-year record at the farm well at the Horton Hydrologic Laboratory shows that the maximum stage has occurred: Fifty per cent of the years in the spring months, March to May; 25 per cent of the years in the winter months, December to February; 15 per cent of the years in the fall months, September to November; 10 per cent of the years in the summer months, June to August.

Records of ground-water level in the chalk-downs in Sussex, England, compiled by D. Halton Thomson, and extending back continuously to 1836, show midwinter maximum levels strongly predominating. The excellent records of the Italian Hydrometric Service show midwinter maximum ground-water levels predominating at some stations, spring and early summer maxima at other stations. Several long ground-water-level records in Michigan show spring maxima strongly predominating, these usually occurring about the time the frost goes out of the ground or shortly after.

There are four factors which operate together to bring about the occurrence of ground-water maxima in the early spring: (1) Seasonal variation of rainfall-duration; (2) storage-deficiency; (3) melting of snow; (4) frost-removal. The manner in which these factors operate to produce maximum and limiting ground-water stages does not seem hitherto to have been described.

(A) Rainfall-duration--In many localities, including northern United States, rainfall-duration per shower or storm is much greater, and rain-intensity in inches per hour much less, in fall, winter, and spring than during midsummer months. Analysis of hourly rain-intensity records for ten years at representative regular United States Weather Bureau stations shows, in general, winter rainfall-duration per shower or storm one and one-half to two times as great as in summer, and summer average rain-intensity two to three times as great as for winter conditions. For a given infiltration-capacity the duration of rainfall-excess is therefore generally much greater in fall, winter, and spring than in midsummer and the chances of attainment of the limiting ground-water stage are correspondingly greater than in summer, provided the infiltration-capacity is the same. For rains sufficiently long-continued to produce limiting ground-water stages, the infiltration-capacity is usually at or near its minimum value for the given soil. If the ground is not frozen, this minimum value is nearly constant throughout the year except as it is affected by temperature. If, owing to recent cultivation of the soil, the infiltration-capacity tends to be much higher than the minimum, then the limiting stage and the volume of ground-water storage which must be replenished before the limiting stage can be reached are both increased. Nevertheless, high ground-water stages are sometimes attained in midsummer, as has happened twice at the author's well.

(B) Field-moisture-deficiency--Another reason why ground-water maxima are less frequent in summer than in early spring is because there is no abstraction of moisture from the soil by vegetation during the winter season; consequently the field-moisture-deficiency is usually less in winter than in summer--in fact, it is often relatively slight, especially if it has been replenished by fall rains. If the ground has been frozen for a long period, the ground-water storage-deficiency below the limiting level may, however, be nearly or quite as great at the end of the winter as in a summer dry-period. It appears, therefore, that as between the two factors, rainfall-duration and storage-deficiency, the former is the more important in relation to fixing the time of occurrence of ground-water maxima and limiting stages.

(C) Frost-removal--The author believes, although data for quantitative proof are not available, that an important factor in producing maximum ground-water stages in the spring is the rapid flushing of large volumes of water to the water-table when the frost goes out of the ground. This phenomenon has occasionally been noticed in connection with lysimeter-records. The author has observed two or three inches of water passing downward through a 3-foot depth lysimeter within a day or two after the frost went out of the ground.

The process involved seems to be as follows. The frost melts from the surface downward and, in the case of loams and heavier soils containing colloid material, most of the melt-water remains in the soil. It is true that, while there is usually some infiltration through frozen soils, yet, except in case of sandy soils, the infiltration-rate through frozen soil is relatively small--the rate is apparently measured in terms of thousandths rather than tenths of an inch. Expansion of ice in freezing produces a degree of porosity and a transmission-capacity at the time the frost goes out greatly in excess of the normal minimum values of these quantities. This condition persists until the soil has become settled. When the frost goes out there is in

the soil a column of water representing a fraction of the volume considerably in excess of the normal gravitational voids, all of which passes immediately downward to the water-table and often produces the maximum ground-water stage of the year. Incidentally this seems to be one of those wise provisions of nature, so often observed. This same volume of water, if it entered the water-table gradually over a period of two or three weeks while the soil-frost is melting, might be drained away as fast as it entered, leaving the water-table at about the same level as at the beginning of the melting period. Instead, all this water is flushed quickly into the water-table with little loss, just before the beginning of the growing season, and therefore provides the maximum available volume of ground-water storage for the subsequent dry period.

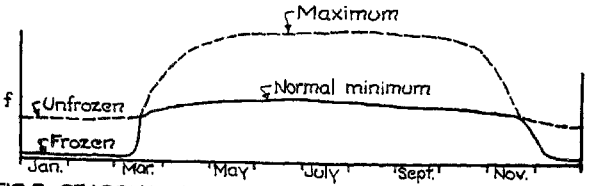


FIG. 3 - SEASONAL VARIATION OF INFILTRATION-CAPACITY, MEDIUM LOAM-SOIL (SCHEMATIC)

(D) Snow-cover--The complex phenomena taking place in the contact-zone between the soil-surface and a layer of melting snow are of great importance hydrologically but cannot be fully discussed here. It may be noted in passing that in case of heavy soils, where the process just described takes place, water derived from melting snow may contribute largely to the volume of water flushed into the water-table when the frost goes out, thus providing a high spring ground-water level in the absence of coincident rainfall. In case of sandy soils with low field-moisture-capacity, infiltration takes place at a relatively high rate even when the ground is frozen. Melting snow may provide a continuous supply of water for absorption by the soil of much longer duration than is ever provided by rainfall alone. For sandy regions in middle latitudes the occurrence of maximum ground-water stages in the early spring is likely to be due to this cause. There is, however, a limitation to the rate at which water can be supplied to the soil by this process, since snow can only melt at a rate dependent on the temperature-excess over 32° F.

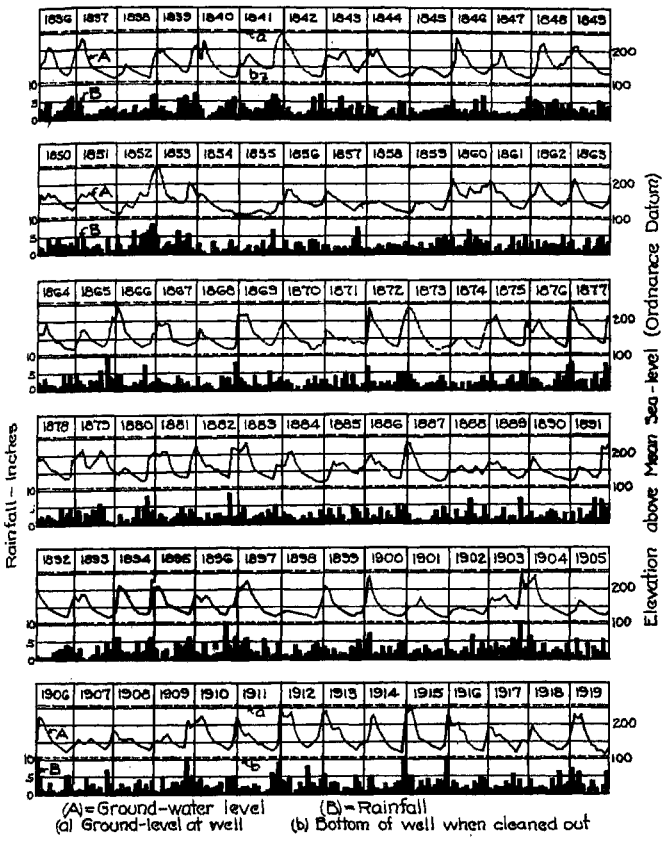


FIG. 4 - RAINFALL AND GROUND-WATER LEVEL IN CHALK AT CHILGROVE, W. SUSSEX, ENGLAND (DATA BY D. HALTON THOMSON)

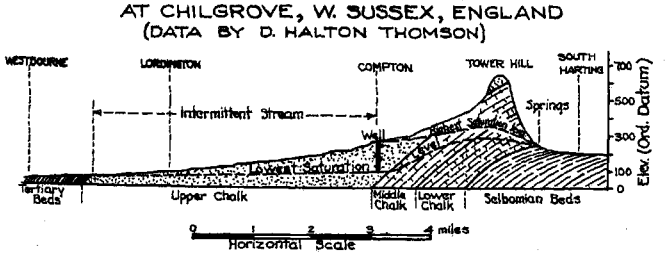


FIG. 5 - GENERAL SECTION ACROSS SOUTH DOWNS ALONG COMPTON AND EMS VALLEYS (DATA BY D. HALTON THOMSON)

Seasonal variation of ground-water range

For loams and heavier colloidal soils there is a somewhat irregular seasonal cycle of variation of infiltration-capacity. The transmission-capacity of the underlying aquifer remains constant except as affected by the variation of ground-water temperature, which is usually small.

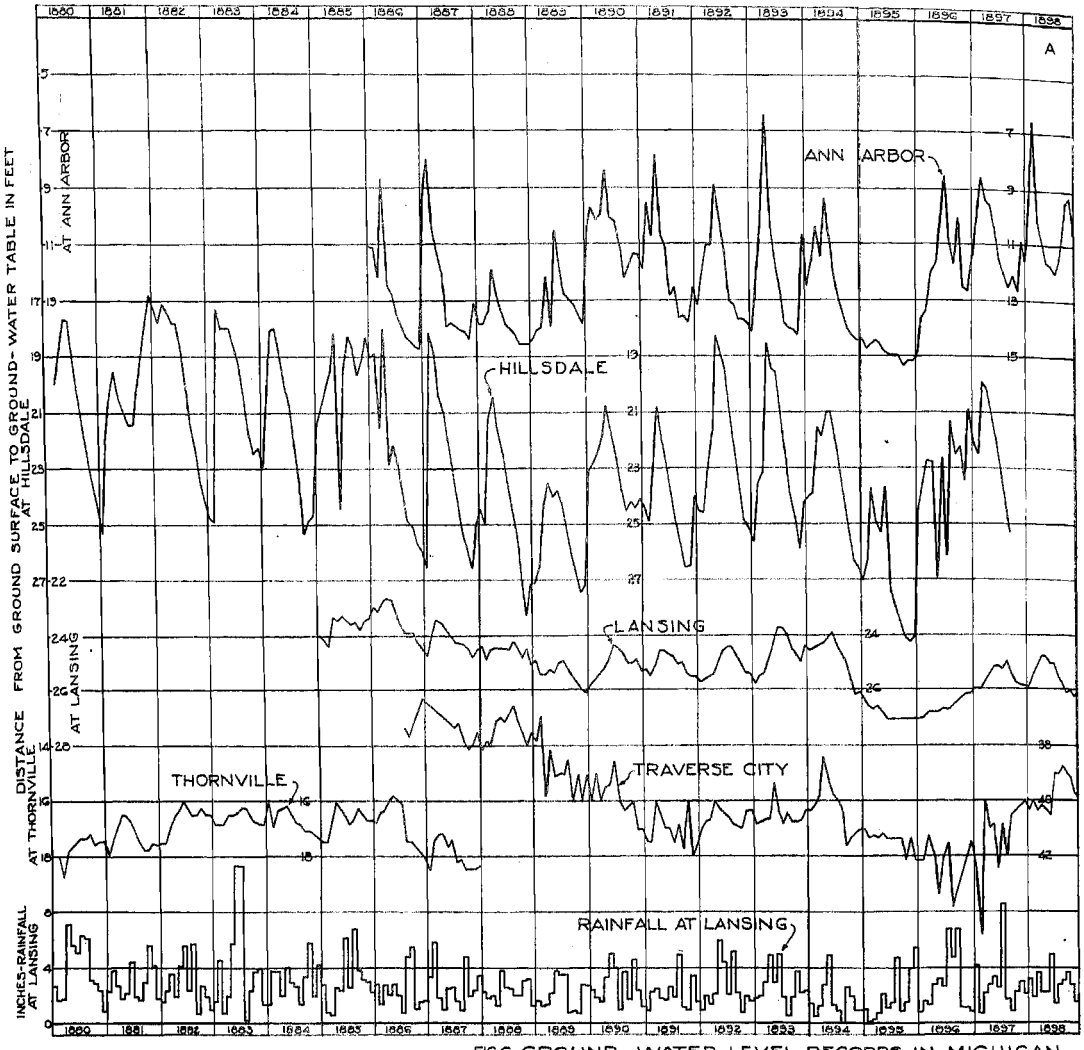
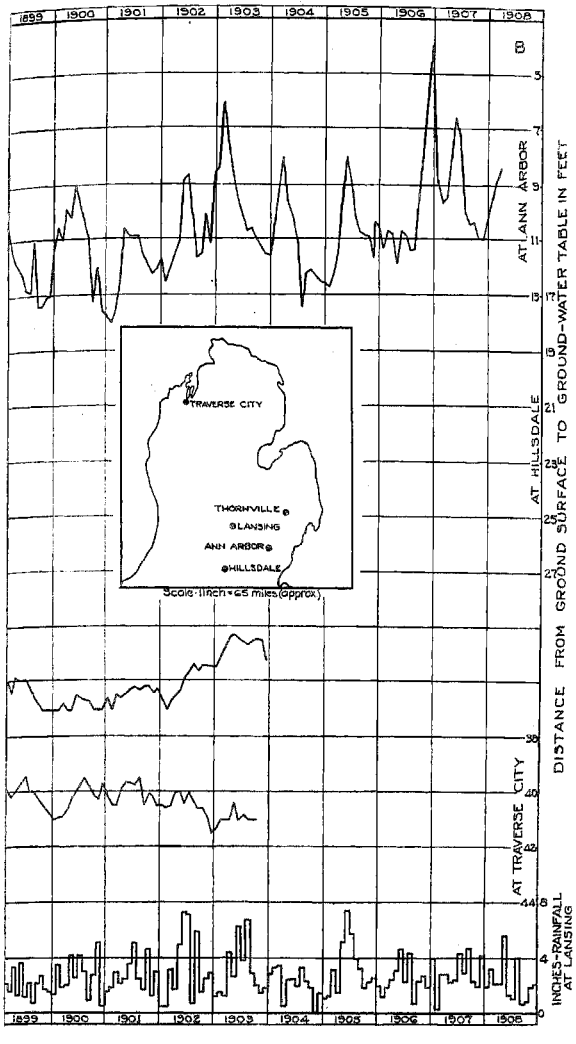


FIG.6-GROUND-WATER LEVEL RECORDS IN MICHIGAN
(DATA BY HENRY B. BAKER)

The relations of winter and summer infiltration-capacities, transmission-capacities, and transmission-ratios for a typical fertile loam soil are illustrated schematically by Figure 3. The upper of the two lines for infiltration-capacity in summer represents a condition with the soil newly cultivated, sun-checked or with numerous earthworm, insect, and root perforations. The lower line represents the minimum infiltration-capacity for a moist soil, packed by rain, and with the larger openings closed by swelling of colloids or inwashing of fine materials. The minimum infiltration-capacity seems to be nearly a constant and definite characteristic of the soil-type. Also the minimum line represents conditions usually prevailing during long-continued rains and hence corresponds to the infiltration-capacity at times when ground-water maximum stages are or may be attained. For summer conditions the minimum value of f is likely to be greater than the transmission-constant K_t because the soil-temperature is higher than the ground-water temperature. For example, in case of a sandy soil, where the transmission-ratio is unity at equal temperatures, and in a region with mean annual temperature of 50° F and ground-water temperature constant at 50° F, the transmission-ratio will be increased 15 per cent under mid-summer conditions with a ground-surface temperature of 70° F and decreased 15 per cent for winter conditions with a ground-surface temperature of 32° F.

Since the transmission-ratio is usually greater in midsummer than in the remainder of the year, it is evidently possible, other things permitting, for higher ground-water stages to be attained in midsummer than in the fall, winter, or spring.



The question naturally arises: Why are the maximum ground-water levels in mid-latitudes usually attained in the early spring or late fall? The answer is derived from a consideration of the seasonal relations of rain-intensity and duration already given. A rain-intensity > f and of sufficient duration to provide ground-water equilibrium or the limiting seasonal ground-water stage may be a common occurrence in fall and spring but only a rare occurrence in midsummer. Quantitative treatment of this subject by the use of curves of rainfall-intensity versus duration for different seasons of the year can easily be carried out. Space-limitations forbid full treatment here but it is evident that, while the limiting maximum ground-water stage for spring conditions is likely to be reached in the majority of years, the higher limiting maximum stage for midsummer will be reached only at rare intervals, if ever. Figures 4, 6, 8, and 9 serve well to illustrate these conditions.

Too much precision should not be expected in relation to the limiting maximum ground-water level at a given well. Even where the well-level is independent of the outlet-level, as it usually is in case of hillside aquifers not too close to streams, and even where it is controlled solely by the minimum infiltration-capacity, there is a variation in the latter, outside the frost-season, of at least 10 to 20 per cent due to temperature and other causes. There will be a corresponding variation in the limiting maximum ground-water stage of 5 to 10 per cent. If, for example, the highest observed stage at a given well is 20 feet below ground, the aquifer is 20 feet deep and stages 18 to 20 feet occur in some years, it may be fairly assumed that any stage less than about 20 feet below ground represents the limiting maximum stage for the conditions pertaining at the time it occurs.

Ground-water levels in the chalk-downs, Sussex, England--The longest continuous ground-water-level record known to the author is that compiled and published by D. Halton Thomson for a well in the chalk at Chilgrove, West Sussex, England [The effect of rainfall on the saturation-level in the chalk at Chilgrove, West Sussex, from 1836 to 1919, British Rainfall, 1919; also Hydrological conditions in the chalk at Compton, W. Sussex, Inst. Water Eng., 1921, (second paper) 1931]. The record extends back with short intermissions to 1834 and so covers more than a century. The record to 1919 is shown graphically on Figure 4. This is an excellent example of a well which overflows at the ground-surface. It will be noted that the maximum level is very definitely fixed but is only reached in occasional years. During dry years the ground-water level may remain continuously low for several years, as in 1887-91. In order that there may be ground-water accretion, the field-moisture-deficiency must first be eliminated. In most of the years shown on Figure 4 there was some rise of the ground-water level in the spring, showing that in spite of the long-continued drought, there was no field-moisture-deficiency at the beginning of the growing season. An important point in this connection is that the effects of a prolonged drought on the ground-water supply to streams is often much more prolonged than the effect of the drought on vegetation. As a rule, accretion to the water-table occurs and there is no field-moisture-deficiency in humid climates at the beginning of the growing season in years of drought. Vegetation suffers because of subsequent deficiency in rainfall. When this

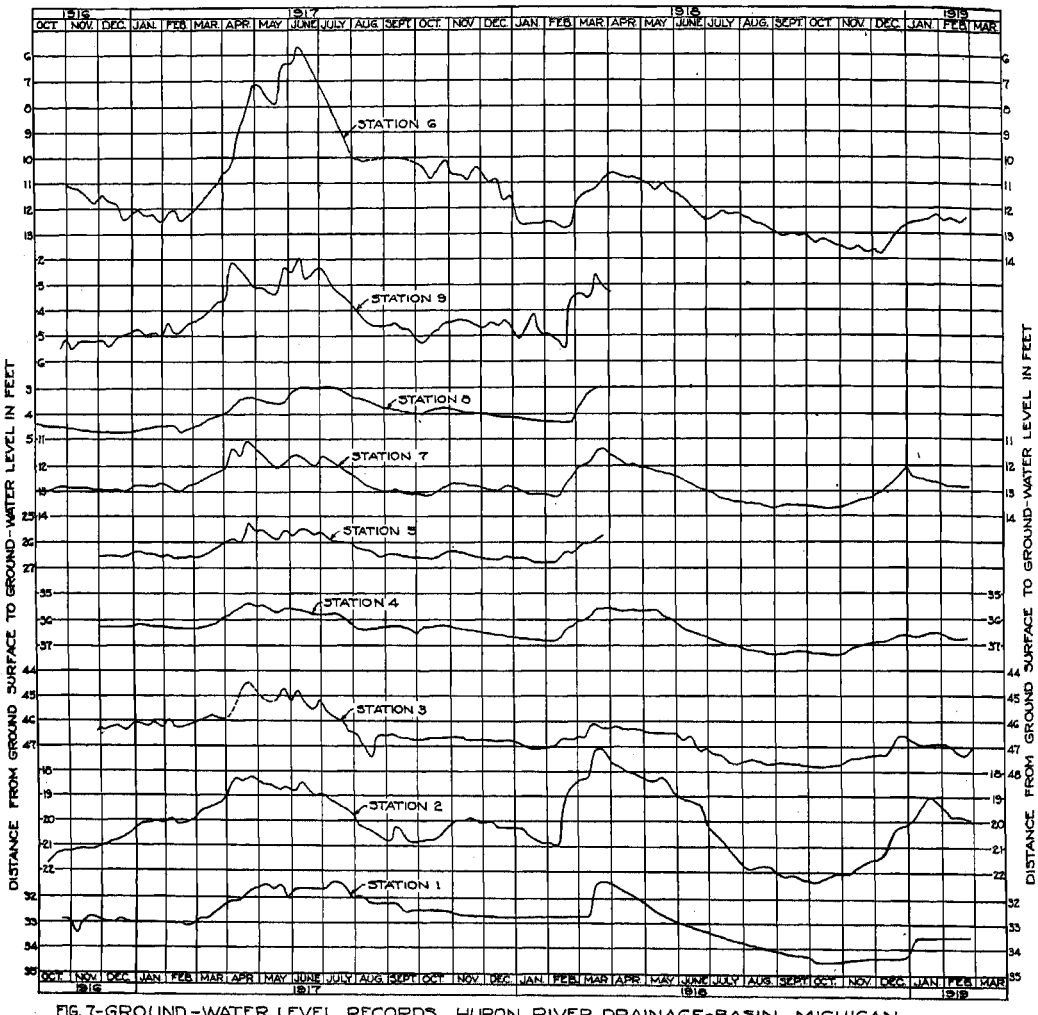


FIG. 7-GROUND-WATER LEVEL RECORDS, HURON RIVER DRAINAGE-BASIN, MICHIGAN
(DATA BY C.O. WISLER)

ends, normal soil-moisture conditions are almost invariably restored inside of a year, whereas several years may elapse before the water-table rises to its maximum or even to its normal level.

D. Halton Thomson has also kept and published an excellent record of ground-water levels in the chalk at Compton, West Sussex, beginning 1903. As large-scale phreatographs of this record have been published, it is not reproduced here. The Compton well is an excellent example of an aquifer with a bourne-outlet. Referring to the cross-section, Figure 5, it will be seen that the head of the bourne migrates throughout a range of more than four miles. This record is remarkable because of the wide range of fluctuation of the water-table. The extreme variation in height of the water-table at the Compton well is 126.2 feet. It appears that in this case the limiting stage is about 220 feet (Ordnance datum), but this stage is only attained at rare intervals.

Ground-water levels in Michigan--Dr. Henry B. Baker, long-time secretary of the State Board of Health of Michigan, believing that there was some relation between water-level in wells and the occurrence of typhoid fever of the virulent rural type, common in the middle west of his time, established well-records at several locations in southern Michigan, beginning in 1880. Some of the records were continued to 1908. Readings were taken in each well once a month, about the middle of the month. The longer and better records are shown graphically on Figure 6. In spite of the approximation due to single monthly readings, there is a close agreement between the curves for Ann Arbor and Hillsdale, although these places are 60 miles apart and in different drainage-basins. Corresponding stages are reflected, with greatly reduced range, in the records

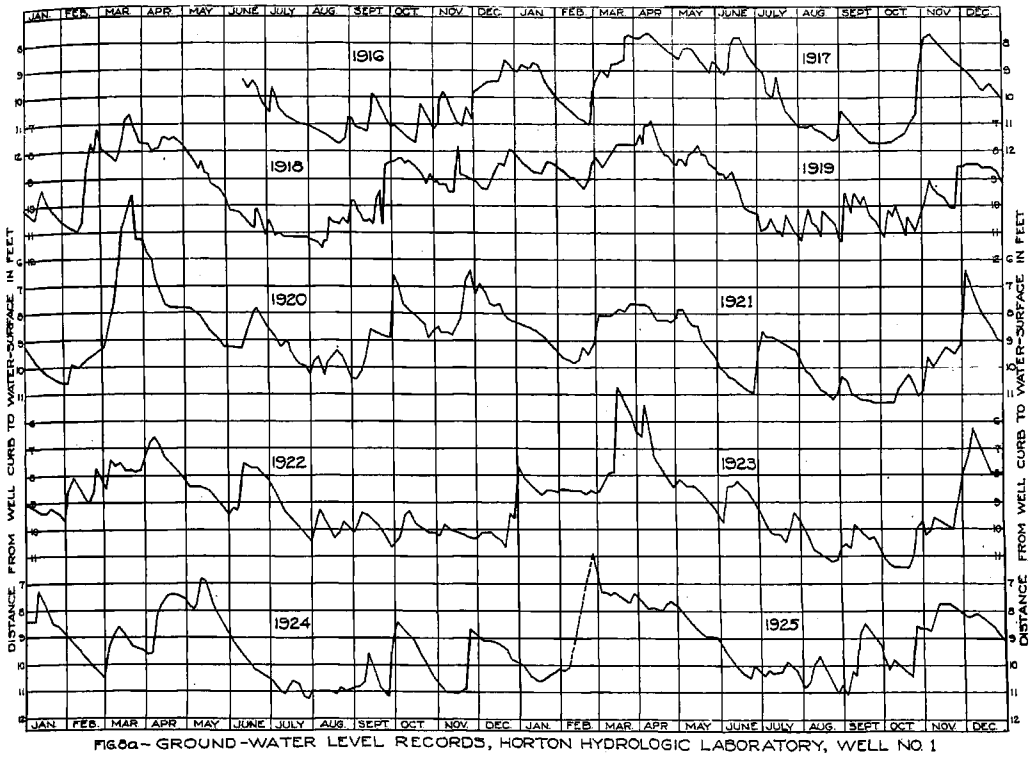
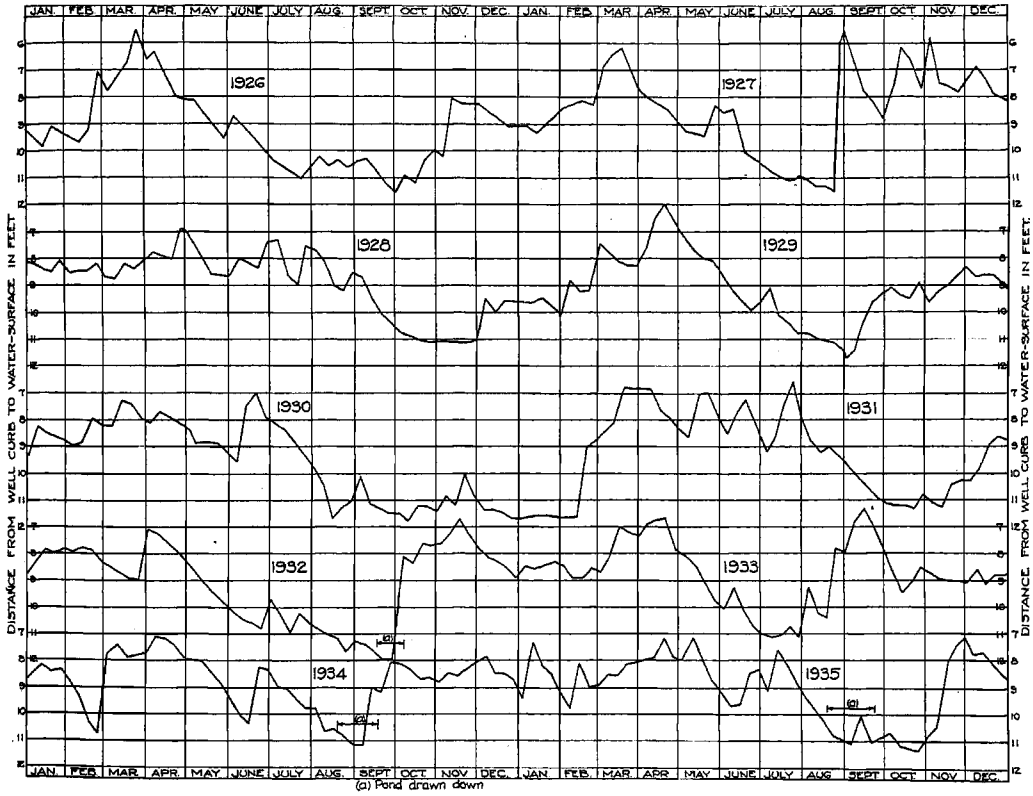


FIG.8a- GROUND-WATER LEVEL RECORDS, HORTON HYDROLOGIC LABORATORY, WELL NO. 1



(a) Pond drawn down

FIG.8b- GROUND-WATER LEVEL RECORDS, HORTON HYDROLOGIC LABORATORY, WELL NO. 1

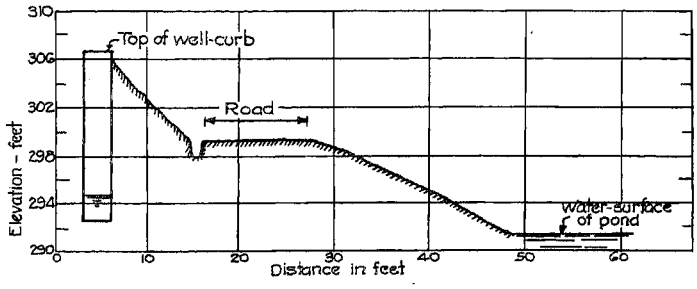


FIG.9— PROFILE OF GROUND-SURFACE ~ WELL NO.1 TO POND
HORTON HYDROLOGIC LABORATORY

for Lansing, Traverse City, and Thornville. All of the wells are located in regions of rather coarse-textured glacial sands, and just which of the reasons capable of producing the variations in range have operated to produce the marked distinction between the upper two and the lower three wells cannot be definitely stated.

Maximum stages of all wells most usually occur in the spring, with minimum in the late fall, although the highest rainfall is in general in the midsummer months. There are exceptions where the annual maximum occurred in midsummer, as in 1900.

The rock floor of lower Michigan is saucer-shaped and there is in places, in the southern part of the State, large artesian leakage into local aquifers. This may have the effect of equalizing the regimen of wells in these aquifers. As a rule, the region is poorly drained, so that the normal range of wells may vary widely as between those close to outlets and those more remote.

Huron River Basin--Figure 7 shows the results of ground-water-level observations taken continuously at different stations in the Huron River Drainage-Basin in southern Michigan. The data have been furnished by Professor C. O. Wisler. Here again we find close similarity of regimen as between different wells but a considerable difference in range, indicating the operation of some of the causes affecting range already described. These records are largely free from saw-tooth irregularities, common to some wells, and the phreatographs show evidence of the operation of some natural smoothing process. The Huron River Drainage-Basin contains numerous lakes, covering probably ten per cent or more of the area. [Well No. 1--Dexter Quadrangle; well on highway 1/4 mile east of east branch Mill Creek and 3/4 mile south of railway; also 3 miles southeast of Lima. Well No. 2--Stockbridge Quadrangle; 1/2 mile northeast from Sylvan, 5 miles southwest from Chelsea. Well No. 3--Dexter Quadrangle; well at first house north of end of north-south road, 1 mile east of small lake 5 miles northwest from Dexter. Well No. 4--Stockbridge Quadrangle; 1 mile east from Gregory. Well No. 5--Dexter Quadrangle; well at third house from south end of north-south road, 2 miles northeast from Pinckney. Well No. 6--Dexter Quadrangle; well at first house south from north end of cross-road, 1-1/2 miles southwest from Horseshoe Lake. Well No. 7--South Lyon Quadrangle; 1/2 mile east from southwest corner Oakland County, 2 miles south and a little west from South Lyon. Well No. 8--Milford Quadrangle; at house at south end of north-south road, 1/2 mile east from Ohio Mills school, 3/4 mile south from Beach Lake. Well No. 9--Milford Quadrangle; well at Wixon village. Well No. 10--Milford Quadrangle; well 1-1/4 miles east from Highland village.] Some of these have outlets to the streams, others have no surface outlets, but the surfaces of nearly all are continuous with the adjacent water-tables. Thus, as distinguished from ordinary aquifers, there is accretion to the ground-water in this region every time it rains, to an extent substantially equal to the amount of rain falling on the lake-surfaces. Thus the ground-water level rises and falls much more gradually than in cases where accretion to the water-table can take place only when and after the field-moisture-deficiency is reduced to zero.

Ground-water records at Horton Hydrologic Laboratory--Figure 8 shows a ground-water-level record covering approximately 20 years at the farm well at the Horton Hydrologic Laboratory. Readings are taken weekly or oftener. About five gallons of water per day are drawn from this well, the amount being kept practically constant. The well is in gravel underlain by Hudson River shale and is about 50 feet from the edge of a mill-pond, in which the water-level is nearly constant except during spring high-water. On a few occasions the pond has been drawn down temporarily, as noted on the diagram. This record affords an excellent example of a well where the maximum level is limited by an auxiliary outlet. When the water in the well rises to a certain elevation, water begins to appear in the ditch at the foot of the slope below the well, as

shown on Figure 9. On one or two occasions abnormally high levels have been recorded in this well in the early spring as the result of LaGrange Brook overflowing upstream from the well and flooding part of the area tributary to the well. Aside from this the maximum stage, as limited by the ephemeral outlet, is about five feet below ground.

Voorheesville, New York

FLUCTUATIONS IN GROUND-WATER AT WOODGATE, NEW YORK

E. S. Cullings

In an effort to improve the operation of river-regulating reservoirs in the Black River Regulating District, the observation of ground-water elevation at selected points within the District was begun in 1926, and has been continued to the present date at a limited number of stations. As yet no attempt has been made to draw conclusions, but some of the results obtained after nearly ten years of observation are submitted for discussion and suggestion.

The Black River drains the westerly slope of the Adirondack Plateau in northern New York and flows into Lake Ontario, draining an area of 1,918 square miles. Except in the lower reaches of the river and in the extreme westerly margin of the drainage-area, the underlying rock is a hard, impervious, crystalline rock of pre-Cambrian age, having the characteristics of granite. This rock is covered by beds of glacial drift of varying depth, dating from the Pleistocene glacial epoch. These sands provide a vast underground reservoir, or rather a system of reservoirs, covering the whole Adirondack Plateau, nearly 10,000 square miles in area.

The discharge of ground-water maintains the flow of the Adirondack streams at a relatively high rate after surface-runoff has ceased. These streams seldom fall below 0.25 sec-ft per square mile even in periods of drought. To supplement the natural storage and to equalize the flow through the year, the Black River Regulating District has in process of development a system of reservoirs, three of which have been in operation for some years. In an effort to coordinate the uncontrolled ground-storage with the regulated storage in the reservoirs, the more or less systematic observation of ground-water elevation was begun about ten years ago.

Owing to lack of funds it was impossible to put down observation-wells in the most favorable locations, from the geological point of view, and it was necessary to utilize abandoned or little-used domestic wells in the best available locations.

About 50 wells have been observed for periods varying from a few months to nearly ten years. Observations at most of these wells were discontinued after a few months, because for various reasons they did not yield significant results. Many of the wells appeared to be in regions of definite underground flow, and in these areas the rise and fall of the water-table does not indicate accurately the depletion and replenishment of ground storage.

Lack of funds has limited observations to occasional visits, except at one station, which seemed to offer the most satisfactory results and at which weekly observations have been made by a paid observer since 1927. The curves on Figure 1 show the rise and fall of ground-water at this station, known in our records as Well No. 6, for a period of nearly 10 years. Figure 2 is a continuous hydrograph of ground-water stage for the 10-year period. On Figure 2 is also shown the precipitation by calendar months for the same period.

Measurements are made from a bench-mark at the surface of the ground, the elevation of which is about 1480. The well is in a level area of glacial sand at Woodgate (White Lake Corners), shown on the McKeever sheet, United States Geological Survey map. This area appears to be a delta-deposit formed in an ancient glacial lake. The ground-surface slopes gently in all directions from the well. The maximum fluctuation during the 10-year period is slightly in excess of 16 feet; that is, the water-table has varied from 13.8 feet to 30.1 feet below the ground surface. The fluctuation of the 10-year mean stage, shown by the heavy line, is about 8.6 feet. The water-table, therefore, is well below the root-zone at all times.

Observations are made weekly, with a steel tape. A recording gage has been operated during parts of two summers, but the results have not been entirely satisfactory.

Two other wells, distant about one-quarter mile on either side of Well No. 6, have been observed as a check, and have been found to follow the regimen of No. 6 very closely. Another well, similarly situated in a level area of glacial sand and distant about eight miles north-