4 Groundwater: The Water-Budget Myth

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INTRODUCTION

Water-resource scientists are concerned that some basic principles are being overlooked by water managers. Rather than discuss the scope of groundwater hydrology, we have chosen to focus on a common misconception to illustrate the point.

Perhaps the most common misconception in groundwater hydrology is that a water budget of an area determines the magnitude of possible groundwater development. Several well-known hydrologists have addressed this misconception and attempted to dispel it. Somehow, though, it persists and continues to color decisions by the water-management community. The laws governing the development of groundwater in Nevada as well as several other states are based on the idea that pumping within a groundwater basin shall not exceed the recharge. It is the intent of this paper to re-examine the issue.

HISTORICAL PERSPECTIVE

Theis (1940) addressed the subject:

> Under natural conditions . . . previous to development by wells, aquifers are in a state of approximate dynamic equilibrium. Discharge by wells is thus a new discharge superimposed upon a previously stable system, and it must be balanced by an increase in the recharge of the aquifer, or by a decrease in the old natural discharge, or by loss of storage in the aquifer, or by a combination of these.

Brown (1963) attempted to illustrate these points by demonstrating that (1) under virgin conditions the height of the water table is a function of the recharge and transmissivity, and recharge is balanced by discharge from the aquifer; (2) the effects of groundwater development are superimposed upon these virgin conditions; and (3) the rate at which the hydrologic system reaches a new steady state depends on the rate at which the natural discharge (in his example to a stream) can be captured by the cone of depression. Brown's argument, which was highly technical, was essentially ignored by many hydrologists.

Bredehoeft and Young (1970) re-examined the issue and restated Theis's conclusions:

> Under virgin conditions, steady state prevails in most groundwater systems, and natural re-

charge is equal to the natural discharge. We can write the following expression for the system as a whole

$$R_0 - D_0 = 0, \tag{4.1}$$

where R_0 is the mean recharge under virgin conditions and D_0 is the mean discharge under virgin conditions.

Some disturbance of the system is necessary to have a development. At some time after the start of pumping we can write the following expression:

$$(R_0 + \Delta R_0) - (D_0 + \Delta D_0) - Q + \frac{dV}{dt} = 0, \tag{4.2}$$

where ΔR_0 is the change in the mean recharge, ΔD_0 the change in the mean discharge, Q the rate of withdrawal due to development, and dV/dt the rate of change in storage in the system. From Eqs. (4.1) and (4.2) we can obtain

$$\Delta R_0 - \Delta D_0 - Q + \frac{dV}{dt} = 0. \tag{4.3}$$

Assuming water-table conditions we can them compute an average drawdown for the system as a whole in the following manner:

$$s_a = \Delta V_t/(S_y \cdot A_b), \tag{4.4}$$

where s_a is average basinwide drawdown, ΔV_t the volume removed from storage at time t, S_y the specific yield of the aquifer, and A_b the area of the basin. Such an input-output analysis treats the system just as we would treat a surface water reservoir. The response of the system is assumed to take place rapidly with effects equally distributed throughout the basin. In most groundwater systems the response is not equally distributed.

RESPONSE OF GROUNDWATER SYSTEMS

In groundwater systems the decline of water levels in a basin because of withdrawal will occur over a period of years, decades, or even centuries. Some water must be taken from storage in the system to create gradients toward a well. There are two implications to be gathered from these facts: (1) some water must always be mined to create a development, and (2) the time delays in a groundwater system differ from those in surface-water systems.

It is apparent from Eq. (4.3) that the virgin rates of recharge R_0 or discharge D_0 are not of paramount importance in groundwater investigations. For the system to reach some new equilibrium, which we define as $dV/dt = 0$, there must be some change in the virgin rate of recharge and/or the rate of discharge D_0. It is these changes, ΔR_0 and ΔD_0, that are interesting.

The response of groundwater systems depends on the aquifer parameters (transmissivity and storage coefficient), the boundary conditions, and the positioning of the development within the system.

Lohman (1972a), referring to the High Plains of Texas and New Mexico, made the point again. The following discussion is a synopsis of Lohman's argument taken from Bulletin 16 of the U.S. Water Resources Council (1973):

> Withdrawals cannot exceed the rates of recharge or discharge for a prolonged period of time without resultant "mining" of ground water. Adjustments in recharge and discharge rates as a result of pumping can be referred to as capture, and, inasmuch as sustained yield is limited by capture and cannot exceed it, estimates of capture are fundamentally important to quantitative groundwater analysis and planning for long-term water supply.
>
> Decline of water levels in response to sustained withdrawal may continue over a long period of time. At first, some water must be taken from storage in the system to create gradients toward pumping wells. Two important implications of these statements concerning a long-term water supply are that (1) some water must be removed from storage in the system to develop a groundwater supply, and (2) time delays in areal distribution of pumping effects in many groundwater systems demonstrate that balanced (equilibrium or steady-state) conditions of flow do not ordinarily exist. In the clearest examples, water levels decline drastically, and some wells go dry long before the system as a whole reaches a new equilibrium balance between replenishment and natural and imposed discharge rates.
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> The most well-known example of such a condition of nonequilibrium is the major groundwater development of the southern High Plains of Texas and New Mexico. Water is contained in extensive deposits (the Ogallala formation) underlying the plains (Figure 4.1). Average thickness of these deposits is about 300 feet. They consist of silt, sand, and gravel and form a groundwater reservoir of moderate permeability. The reservoir rests on relatively impermeable rock and constitutes the only large source of groundwater available to the area.
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> The southern High Plains slope gently from west to east, cut off from external sources of water upstream and downstream by escarpments,

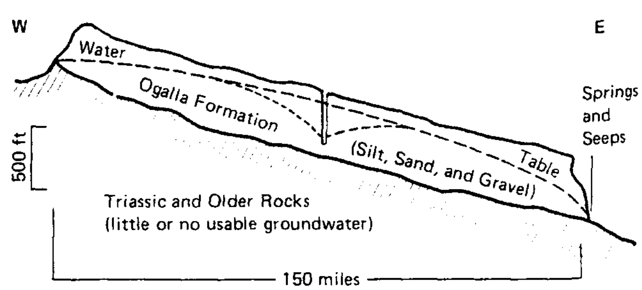

FIGURE 4.1 Development of groundwater in the southern High Plains of Texas and New Mexico. Withdrawal has resulted in a pronounced decline of water levels in the middle of the southern High Plains, but it has had little effect on the gradient to the east (natural discharge) or on natural recharge.

as illustrated in Figure 4.1. Replenishment is dependent on the scanty precipitation, and total recharge in the southern High Plains is extremely small in comparison with the enormous imposed discharge (pumping for irrigation). Total recharge is equivalent to only a fraction of an inch of water per year over the whole of the High Plains. The natural discharge, of the same order as the recharge, continues from seeps and springs along the eastern escarpment.

Withdrawal by pumping has increased rapidly in the past 50 years and at present amounts to about 1.5 trillion gallons per year (4.6 million acre-feet per year). The withdrawal has resulted in a pronounced decline of water levels in the middle of the Plains, where pumping is heaviest (and where the increase in cost of pumping has been greatest). Little additional natural recharge can be induced into the system because the water table lies 50 feet or more beneath the land surface in most of the area, the unsaturated volume of aquifer available for possible recharge is more than ample.

Nor has natural discharge been salvaged by the lowered water levels. As may be noted in Figure 4.1, the hydraulic gradient, or water-table slope, toward the eastern escarpment has been virtually unchanged. Even if all discharge could be salvaged by pumping, however, the salvaged water would be only a small percentage of present pumpage.

THE CIRCULAR ISLAND

Perhaps the easiest way to illustrate our point further is to consider pumping groundwater on an island situated in a freshwater lake. The situation is shown schematically in Figure 4.2. An alluvial aquifer overlies bedrock of low permeability on the island. Rainfall directly on the island recharges the aquifer. Under virgin conditions, this recharge water is discharged by outflow from the aquifer into the lake. The height of the water table beneath the island is determined by the rate of recharge, the area of the island, and the transmissivity.

Under virgin conditions, we can determine a water balance for the island. From our previous notation, recharge to the island is

$$\int^{A} r_a \, dA = R_0,$$

where r_a is the average rate of recharge and A is the total area of the island. Discharge from the island is and

$$\int^{L} kh \left.\frac{\partial h}{\partial s}\right|_s dL = D_0,$$

where k is the hydraulic conductivity of the aquifer, h the height of the water table defined to be equal to the hydraulic head), and

$$\left.\frac{\partial h}{\partial s}\right|_s$$

the gradient in hydraulic head taken at the shoreline of the island (defined to be normal to the shoreline), L the total length of the shoreline, and $R_0 = D_0$.

We drill a well and begin to pump water from the aquifer on the island. A cone of depression develops and expands outward from the well. Figure 4.3 shows this cone of depression a short time after pumping has begun.

If we look at the periphery of the island, we see that until the pumping causes a significant change in the gradient in head at the shoreline the discharge continues unchanged. Gradients in hydraulic head, or saturated thickness, must be changed at the shoreline in order to change the discharge.

If we write the system balance for the entire island, at some time before the cone expands to the shoreline, we see that

$$R_0 - D_0 - Q \neq 0,$$

where Q is the rate of pumping. As neither the recharge, R_0, nor the discharge, D_0, has changed from its initial value, the water pumped, Q, is balanced by the water removed from storage.

The cone of depression will eventually change the gradients in hydraulic head at the shoreline significantly. At this time, discharge from the system begins to change. This is shown schematically in Figure 4.4. The discharge can be changed by pumping so that the system is brought into balance. At some time

$$R_0 - \int^{L} kh \left.\frac{\partial h}{\partial s}\right|_s dL = Q.$$

Since the virgin rate of recharge, R_0, equals the virgin rate of discharge, D_0, we can write

$$D_0 - \int^{L} kh \left.\frac{\partial h}{\partial s}\right|_s dL = Q,$$

where the quantity

$$R_0 - \int^{L} kh \left.\frac{\partial h}{\partial s}\right|_s dL = D_0 - \int^{L} kh \left.\frac{\partial h}{\partial s}\right|_s dL = \Delta D,$$

which we define as the "capture." The system is in balance when the capture is equal to the water pumped, i.e., $Q = \Delta D$.

The term capture is defined and discussed in *Definitions of Selected Ground-Water Terms--Revisions and Conceptual Refinements* (Lohman, 1972b):

Water withdrawn artificially from an aquifer is derived from a decrease in storage in the aquifer, a reduction in the previous discharge from the aquifer, an increase in the recharge, or a combination of these changes. The decrease in discharge plus the increase in recharge is termed capture. Capture may occur in the form of decreases in the groundwater discharge into streams, lakes, and the ocean, or from decreases in that component of evapotranspiration derived from the saturated zone. After a new artificial withdrawal from the aquifer has begun, the head in the aquifer will continue to decline until the new withdrawal is balanced by capture.

For the island system chosen, we can induce flow from the lake into the aquifer. In fact, the capture can be greater than the virgin recharge of discharge

$$\int^{L} kh \left.\frac{\partial h}{\partial s}\right|_s dL > D_0$$

or

$$\int^{L} kh \left.\frac{\partial h}{\partial s}\right|_s dL > R_0.$$

In fact, the magnitude of pumpage that can be sustained is determined by (1) the hydraulic conductivity of the aquifer and (2) the available drawdown, which are independent of other factors (Figure 4.5).

At first glance, this island aquifer system seems much too simple for general conclusions; however, the principles that apply to this system apply to most other aquifer systems. The ultimate production of groundwater depends on

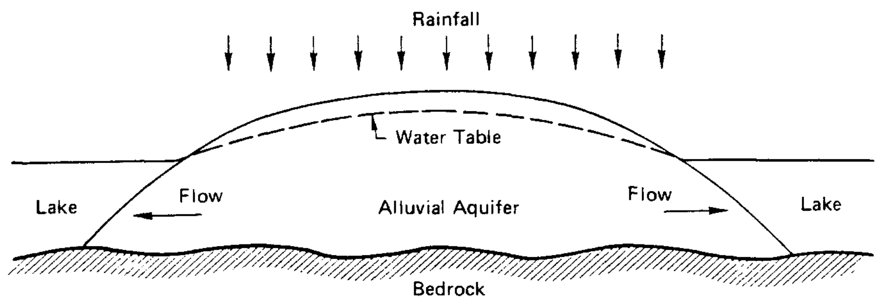

FIGURE 4.2 Cross section of an alluvial aquifer, underlain by bedrock of low permeability, on an island in a freshwater lake.

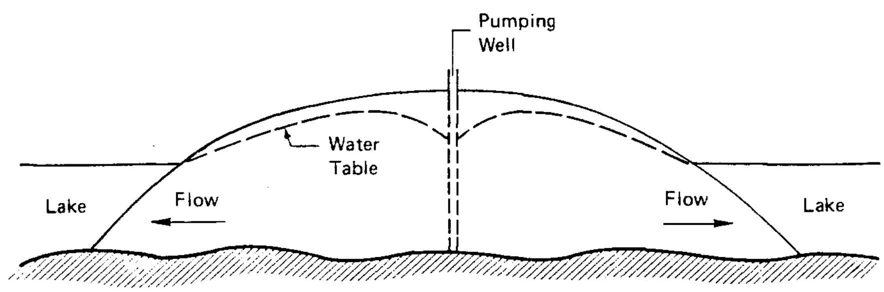

FIGURE 4.3 Cross section of the island depicting the cone of depression soon after pumping has begun.

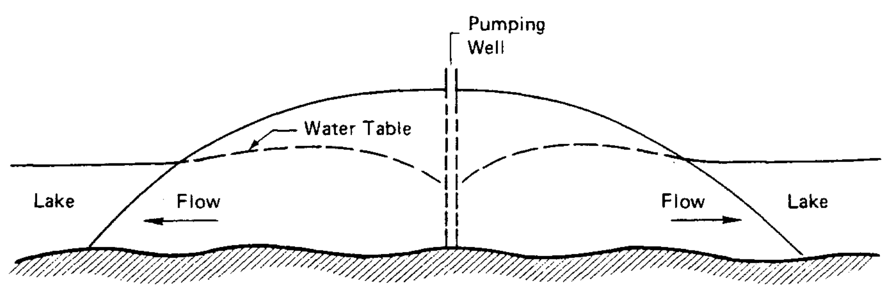

FIGURE 4.4 Cross section of the island aquifer system when the influence of pumping has reached the shoreline.

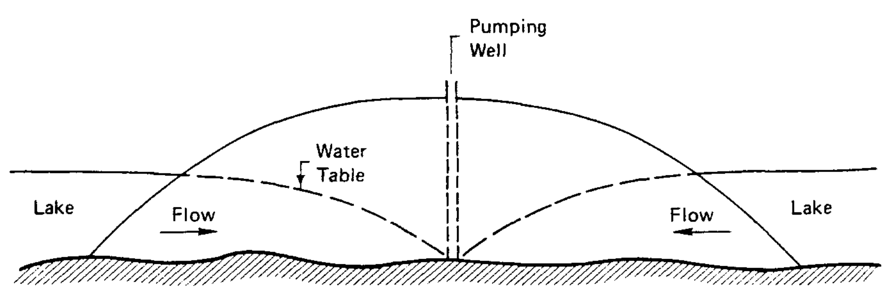

FIGURE 4.5 Schematic cross section of the island aquifer system, which illustrates that the magnitude of pumpage from this system is dependent on the available drawdown, aquifer thickness, and the hydraulic conductivity of the aquifer.

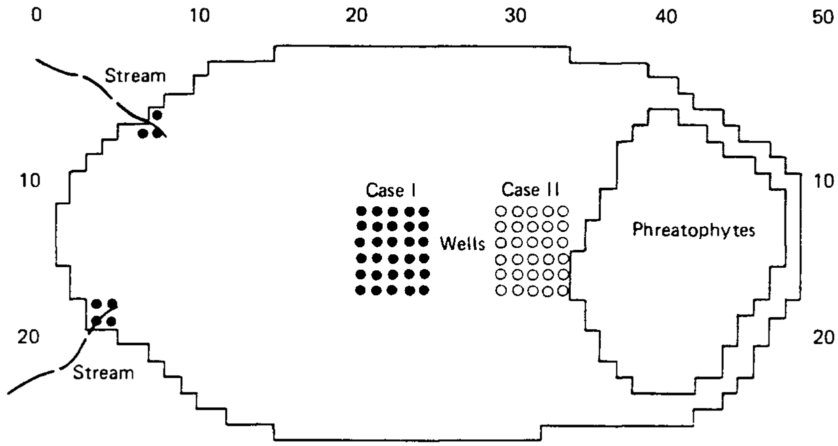

FIGURE 4.6 Schematic map of an intermontane basin showing areas of recharge, discharge, and two hypothetical water-development schemes, Case I and Case II.

how much the rate of recharge and (or) discharge can be changed--how much water can be captured. Although knowledge of the virgin rates of recharge and discharge is interesting, such knowledge is almost irrelevant in determining the sustained yield of a particular groundwater reservoir. We recognize that such a statement is contrary to much common doctrine. Somehow, we have lost or misplaced the ideas Theis stated in 1940 and before.

RESPONSE TIME

Groundwater systems generally respond much slower than other elements of the hydrologic cycle. It can take long periods of time to establish a new steady state. For this reason, groundwater hydrologists are concerned with the time-dependent dynamics of the system.

To illustrate the influence of the dynamics of a groundwater system, we have chosen a rather simple system for analysis. Consider a closed intermontane basin of the sort one might find in the western states. Under virgin conditions the system is in equilibrium: phreatophyte evapotranspiration in the lower part of the basin is equal to recharge from the two streams at the upper end (Figure 4.6).

Pumping begins in the basin, and, for simplicity, we assume the pumpage equals the recharge. The following two assumptions regarding the hydrology are made:

1. Recharge is independent of the pumping in the basin, a typical condition, especially in the arid west.

2. Phreatophyte use decreases in a linear manner (Figure 4.7) as the water levels in the vicinity decline by 1-5 ft. Phreatophyte use of water is assumed to cease when the water level is lowered 5 ft below the land surface.

The geometry and pertinent hydrologic parameters assumed for the system are shown in Table 4.1.

The system was simulated mathematically by a finite-difference approximation to the equations of flow. The equations are nonlinear and of the following form:

$$\nabla \cdot (kh\nabla h) = S \frac{\partial h}{\partial t} + W (x \cdot y \cdot t)$$

where k is the hydraulic conductivity, h the hydraulic head (which is equal in our case to the saturated aquifer thickness), S the storage coefficient, and W the source function (time dependent). In essence, this is a two-dimensional water-table formulation of the problem in which the change in saturated thickness within the aquifer is accounted for.

One-thousand years of operation were simulated. Stream recharge, phreatophyte-water use, pumping rate, and change in storage for the entire basin were graphed as functions of time. Two development schemes were examined: Case I, in which the pumping was more or less centered within the valley, and Case II, in which the pumping was adjacent to the phreatophyte area.

The system does not reach equilibrium until the phreatophyte-water use (the natural dis-

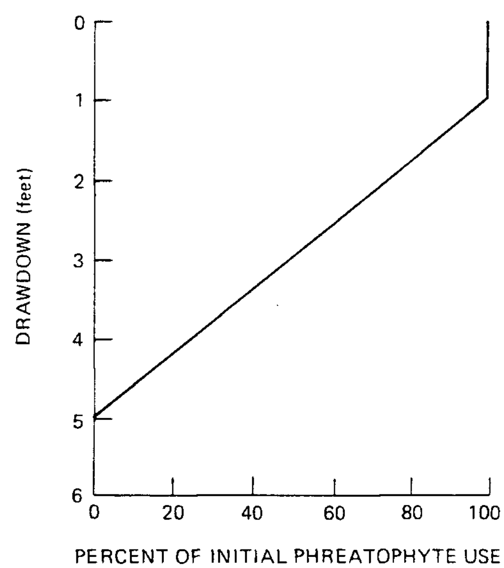

FIGURE 4.7 Assumed linear function relating phreatophyte-water use to drawdown.

charge) is entirely salvaged (captured) by pumping, i.e., phreatophyte water use equals zero (we define equilibrium as $\delta V/\delta t = 0$). In Case I, phreatophyte-water use (Figure 4.8) is still approximately 10 percent of its initial value at year 1000. In Case II it takes 500 yr for the phreatophyte-water use to be completely captured.

We can illustrate the same point by looking at the total volumes pumped from the system, along with the volume taken from storage "mined" (Figure 4.9).

In both cases, for the first 100 yr, nearly all of the water comes from storage. Obviously, as the system approaches equilibrium, the rate of change of the volume of water removed from storage also approaches zero. If the aquifer was thin, it is apparent that wells could go dry long before the system could approach equilibrium.

This example illustrates three important points:

1. The rate at which the hydrologic system can be brought into equilibrium depends on the rate at which the discharge can be captured.
2. The placement of pumping wells in the system significantly changes the dynamic response and the rate at which natural discharge can be captured.
3. Some groundwater must be mined before the system can be brought into equilibrium.

CONCLUSIONS

We have attempted to make several important points:

1. Magnitude of development depends on hydrologic effects that you want to tolerate, ultimately or at any given time (which could be dictated by economics or other factors). To calculate hydrologic effects you need to know the hydraulic properties and boundaries of the aquifer. Natural recharge and discharge at no time enter these calculations. Hence, a water budget is of little use in determining magnitude of development.
2. The magnitude of sustained groundwater pumpage generally depends on how much of the natural discharge can be captured.
3. Steady state is reached only when pumping is balanced by capture ($\Delta R_0 + \Delta D_0$), in most cases the change in recharge, ΔR_0, is small or zero, and balance must be achieved by a change in discharge, ΔD_0. Before any natural discharge can be captured, some water must be removed from storage by pumping. In many circumstances the dynamics of the groundwater system are such that long periods of time are necessary before any kind of an equilibrium condition can develop. In some circumstances

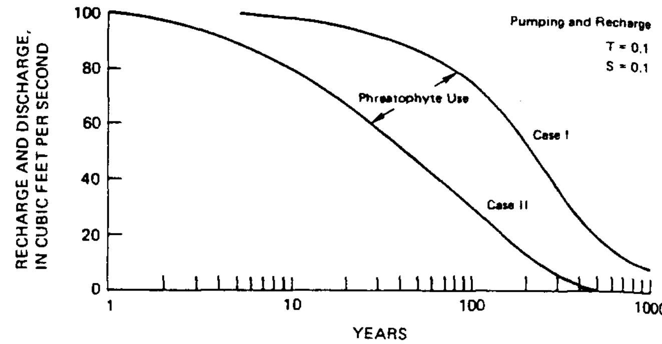

FIGURE 4.8 Plot of the rate of recharge, pumping, and phreatophyte use versus time.

TABLE 4.1 Aquifer Parameters

Basin dimensions	50 x 25 miles
Aquifer	
Hydraulic conductivity (k)	0.5×10^{-3} ft^2/sec
Storage coefficient (S)	0.1
Initial saturated thickness (h)	2000 ft
Phreatophytes	
Area	172 miles^2
Average use (annual)	100 ft^3/sec
Recharge	
Area	7 miles^2
Average recharge rate	100 ft^3/sec
Development	
Area	30 miles^2
Average pumping rate	100 ft^3/sec

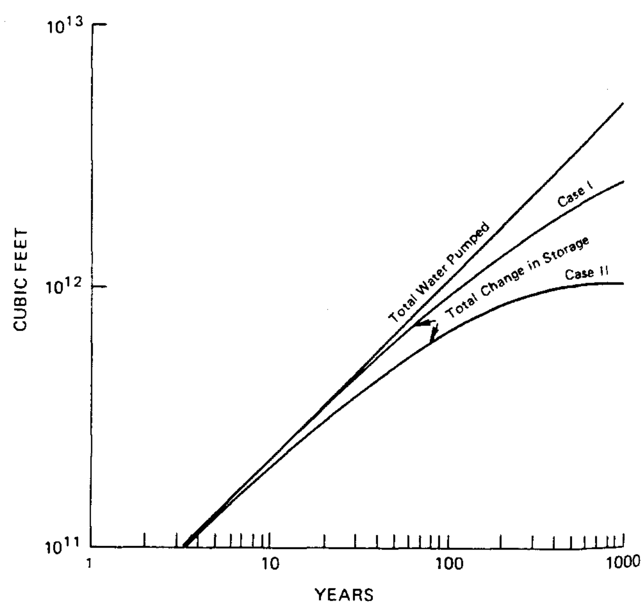

FIGURE 4.9 Total volume pumped and the change in storage versus time.

the system response is so slow that mining will continue well beyond any reasonable planning period.

These concepts must be kept in mind to manage groundwater resources adequately. Unfortunately, many of our present legal institutions do not adequately account for them.

REFERENCES

Bredehoeft, J. D., and R. A. Young (1970). A temporal allocation of ground-water--A simulation approach, *Water Resour. Res. 6*, 3-21.

Brown, R. H. (1963). The cone of depression and the area of diversion around a discharging well in an infinite strip aquifer subject to uniform recharge, *U.S. Geol. Surv. Water-Supply Pap. 1545C*, pp. C69-C85.

Lohman, S. W. (1972a). Ground-water hydraulics, *U.S. Geol. Surv. Prof. Pap. 708*, 70 pp.

Lohman, S. W. (1972b). Definitions of selected ground-water terms--Revisions and conceptual refinements, *U.S. Geol. Surv. Water-Supply Pap. 1988*, 21 pp.

Theis, C. V. (1940). The source of water derived from wells: Essential factors controlling the response of an aquifer to development, *Civil Eng. 10*, 277-280.

U.S. Water Resources Council (1973). Essentials of ground-water hydrology pertinent to water-resources planning, *U.S. Water Resour. Council Bull. No. 16*, 48 pp.