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HYDROLOGY FOR ENGINEERS

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U.S. NUCLEAR REGULATORY COMMISSION
In the Matter of LOUISIANA ENERGY SERVICES, LP
Docket No. 70-3103-ML Official Exhibit No. 25
OFFERED by: Applicant/Licensee Intervenor NIRS/PC
NRC Staff Other
IDENTIFIED on ______ Witness/Panel G. Rice
Action Taken: ADMITTED REJECTED WITHDRAWN
Reporter/Clerk ______

of the soil zone encompassing the sample sites. Since the elements actually measure soil-moisture tension they have about the same resistance at field capacity in any type of soil. The same is true at the wilting point.

6-5. Movement of soil moisture. *Infiltration* is the movement of water through the soil surface into the soil and is distinguished from *percolation* which is movement of water through the soil. When water is first applied to the soil surface, gravity water begins to move down through the larger soil openings while the smaller surface pores take in water by capillarity. The downward-moving gravity water is also taken in by capillary pores. As the capillary pores at the surface are filled and the intake capacity is reduced, the infiltration rate decreases. In homogeneous soil, as pores at lower levels are filled, infiltration decreases gradually until the zone of aeration is saturated. Normally, the soil is stratified and frequently the subsoil layers are less permeable than the surface soil. In this case, the infiltration rate is eventually limited to the rate of percolation through the least pervious subsoil stratum.

Two cases of infiltration must be recognized. Infiltration from rainfall is distinguished by very shallow depths of water on the soil surface but is extensive over large areas. Quantities of water infiltrated are usually very small (a few inches per day maximum) and rarely sufficient to saturate a great depth of soil. At the termination of rain, gravity water remaining in the soil continues to move downward and, at the same time, is taken up in capillary pore spaces. Usually the infiltrated water is distributed within the upper few feet of soil, with little or no contribution to groundwater unless the soil is highly permeable or the zone of aeration very thin. Infiltration from rainfall is discussed in greater detail in Sec. 8-2.

For irrigation and artificial recharge of the groundwater (Sec. 6-16), water is ponded to a considerable depth over limited areas for long periods of time. The aim of recharge operations is to saturate the soil down to the water table. Under these conditions the time variation of infiltration is complex, with temporary increases in rate superimposed on a gradually declining trend. Escape of soil air around the infiltration basin, bacterial action, changes in water temperature, changes in soil structure, and many other factors appear to influence these variations.

Movement of moisture within the soil is governed by the moisture potential following the equation

$$q = -K_w \frac{\partial \Lambda}{\partial x} \qquad (6\text{-}2)$$

where q is the flow per unit time through unit area normal to the direction of flow, x is distance along the line of flow, K_w is conductivity, and Λ is potential. After gravity water has left the soil, the principal

component of total potential is the capillary potential. Equation (6-2) states that flow is from a region of high potential to a region of lower potential. Quantitative determination of the conductivity is difficult although it has been shown to increase with moisture content and decrease with pore size. Thus capillary movement decreases as soil dries and is least in fine-grained soils. Fortunately, a qualitative understanding of these phenomena is normally sufficient for engineering hydrology.

Transport of water vapor in the soil is controlled by temperature differences. Vapor movement is from high temperature (high vapor pressure) to low temperature. Vapor transport is an important factor in moisture movement when the moisture content is lowered to the point where capillary moisture is discontinuous. Under this condition, however, moisture-content and temperature gradients are usually so small that the quantity of moisture moved is negligible. When the surface soil is frozen, the vapor-pressure gradient is upward and is accentuated by the lower vapor pressure of ice relative to water at the same temperature. Thus when frozen soil thaws, its moisture content may be greater than at the time of freezing. Conversely, during summer, vapor-pressure gradients would be downward were it not for evaporation and transpiration.

MOISTURE IN THE ZONE OF SATURATION

Within the zone of saturation all pore spaces are filled with water, and the different states of moisture, moisture tension, etc., are of little concern. Interest is centered on the amount of water present, the amount which can be removed, and the movement of this water.

6-6. Aquifers. A geologic formation which contains water and transmits it from one point to another in quantities sufficient to permit economic development is called an *aquifer*. In contrast, an *aquiclude* is a formation which contains water but cannot transmit it rapidly enough to furnish a significant supply to a well or spring. An *aquifuge* has no interconnected openings and cannot hold or transmit water.

The ratio of the pore volume to the total volume of the formation is called *porosity*. The *original porosity* of a material is that which existed at the time the material was formed. *Secondary porosity* results from fractures and solution channels.

Secondary porosity cannot be measured without an impossibly large sample. Original porosity is usually measured by oven-drying an undisturbed sample and weighing it. It is then saturated with some liquid and weighed again. Finally, the saturated sample is immersed in the same liquid and the weight of displaced liquid is noted. The weight of liquid required to saturate the sample divided by the weight of liquid displaced is the porosity as a decimal. If the material is fine-grained,

the liquid may have to be forced into the sample under pressure to assure complete saturation.

High porosity does not necessarily indicate a productive aquifer, since much of the water may be retained in small pore spaces under capillary tension as the material is dewatered. The *specific yield* of an aquifer is the ratio of the water which will drain freely from the material to the total volume of the formation and must always be less than the porosity. The relation between specific yield and porosity is dependent on the size of the particles in the formation. Specific yield of a fine-grained aquifer will be small whereas coarse-grained material will yield a greater amount of its contained water. Table 6-2 lists approximate average

TABLE 6-2. Approximate Average Porosity, Specific Yield, and Permeability of Various Materials

Material	Porosity, %	Specific yield, %	Permeability [K_p, Eq. (6-4)], gpd/sq ft
Clay	45	3	1
Sand	35	25	800
Gravel	25	22	5000
Gravel and sand	20	16	2000
Sandstone	15	8	700
Dense limestone and shale	5	2	1
Quartzite, granite	1	0.5	0.1

values of porosity and specific yield for some typical materials. Large variations from these average values must be expected. Note that clay, although having a high porosity, has a very low specific yield. Sand and gravel which make up most of the more productive aquifers in the United States will yield about 80 per cent of their total water content.

6-7. Movement of groundwater. In 1856 Darcy confirmed the applicability of principles of fluid flow in capillary tubes, developed several years earlier by Hagen and Poiseuile, to the flow of water in permeable media. *Darcy's law* is

$$v = ks \tag{6-3}$$

where v is the velocity of flow, s is the slope of the hydraulic gradient, and k is a coefficient having the units of v (usually ft/day). The discharge q is the product of area A and velocity. The effective area is the gross area times the porosity p of the media. Hence

$$q = 7.48kpAs = K_pAs \tag{6-4}$$

where 7.48 converts to gallons when the other terms are in feet. The

coefficient of permeability K_p is usually expressed in *Meinzer units*, the discharge in gallons per day through an area of one square foot under a gradient of one foot per foot at 60°F (Table 6-2). Values at other temperatures can be found by multiplying by the ratio of the kinematic viscosities, i.e.,

$$K_{pT} = K_p \frac{\nu_{60}}{\nu_T} \tag{6-5}$$

It is convenient to use the *transmissibility* T to represent the flow in gallons per day through a section 1 ft wide and the thickness of the aquifer under a unit head (slope of 1 ft/ft):

$$T = K_pY \tag{6-6}$$

where Y is the saturated thickness of the aquifer. With this coefficient Eq. (6-4) becomes

$$q = TBs \tag{6-7}$$

where B is the width of the aquifer.

Note that the equations of groundwater flow are analogous to the electrical equation

$$i = \frac{1}{R}E \tag{6-8}$$

where the current i is equivalent to q, the voltage E is comparable to s, and the reciprocal of resistance R is equivalent to permeability. This similarity is sometimes used to advantage in electrical models of groundwater-flow problems.

6-8. Determination of permeability. Laboratory measurements of permeability are made with instruments called *permeameters* (Fig. 6-6). A sample of the material is subjected to water under a known head, and the flow through the sample in a known time is measured. Such tests have limited practical value because of the difficulty of placing samples of unconsolidated materials into the permeameter in their natural state and the uncertainty as to whether a sample is truly representative of the aquifer. Flow in solution cavities or rock fractures and the effect of large boulders in gravel aquifers cannot be duplicated in a permeameter.

The earliest field techniques for determining permeability involved introducing salt into the aquifer at one well and timing its movement to a downstream well.[1] Fluorescein dye,[2] detectable at a concentration of 0.03 ppm by the unaided eye and in concentrations as low as 0.0001 ppm

[1] C. S. Slichter, Field Measurements of Rate of Movement of Underground Water, *U.S. Geol. Survey Water-supply Paper* 140, 1905.

[2] R. B. Dole, Use of Fluorescein in Study of Underground Waters, *U.S. Geol. Survey Water-supply Paper* 160, pp. 73–85, 1906.

under ultraviolet light, has also been used as a tracer. More recently, radioactive materials have been tested. Tracer techniques have encountered numerous difficulties.[1] Chemical reactions between the tracer elements and the formation sometimes occur. Because of diffusion, tests must be conducted over short distances in order to have detectable concentrations at the downstream well, and even then it is difficult to

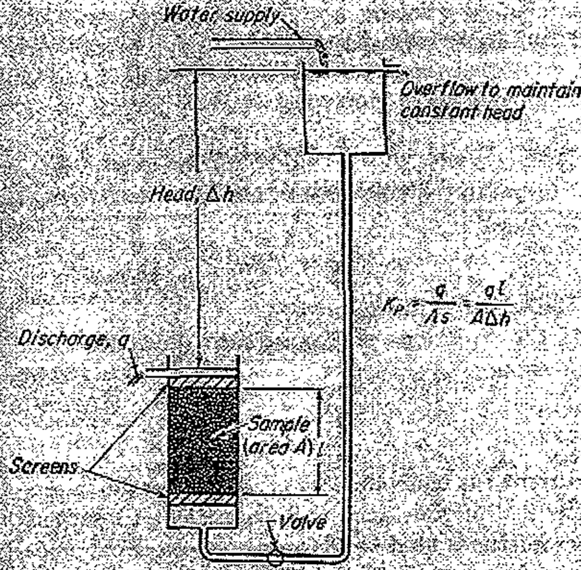

FIG. 6-6. A simple upward-flow permeameter.

determine a representative time of arrival. Tracers are most useful for determining path of flow as, for example, when it is necessary to locate a source of pollution.

Today permeability is most commonly determined by pumping tests. By using the principles of well hydraulics (Sec. 6-12) it is possible to estimate average permeability of an aquifer for a large distance around the test well.

6-9. Sources of groundwater. Almost all groundwater is *meteoric water* derived from precipitation. *Connate water*, present in the rock at its formation and frequently highly saline, is found in some areas. *Juvenile water*, formed chemically within the earth and brought to the surface in intrusive rocks, occurs in small quantities. Connate and juvenile waters are sometimes important sources of undesirable minerals

[1] W. J. Kaufman and G. T. Orlob, An Evaluation of Ground-water Tracers, *Trans. Am. Geophys. Union*, Vol. 37, pp. 297–306, June, 1956.

in the groundwater. Groundwater in the San Joaquin Valley, California, contains considerable boron brought to the surface from great depths.

Water from precipitation reaches the groundwater by the process of infiltration and by percolation from streams and lakes. Direct percolation cannot yield large quantities of groundwater except where the soil is highly permeable or the water table is close to the surface. Large quantities of direct percolation occur through the permeable basaltic lavas of northern California, eastern Oregon, southern Idaho, and Hawaii, and in the southern Appalachian region where thin soil overlies cavernous limestone.

In the southern part of the Central Valley of California groundwater is hundreds of feet below the surface, and little or no recharge from rain can occur. In such cases most of the recharge is from stream channels. Streams contributing to the groundwater are called *influent streams*. Such streams frequently go dry during prolonged dry spells when percolation absorbs all the available flow. Streams are rarely influent throughout their entire length. Often the channel crosses strata of varying permeability, with most of the percolation loss in short reaches of high permeability. Considerable percolation often occurs from stream channels crossing coarse gravels in an alluvial fan. In areas of artesian groundwater, the overlying aquiclude prevents appreciable direct recharge; the recharge area may be far removed from the artesian area.

6-10. Discharge of groundwater. Without interference by man, a groundwater basin fills with water and discharges its excess by several routes. Streams intersecting the water table and receiving flow from the groundwater are known as *effluent streams*. Perennial streams are generally effluent through at least a portion of their length and may flow in impermeable formations elsewhere so that there is little or no loss by seepage.

Where an aquifer intersects the earth's surface, a spring may form. There may be a concentrated flow constituting the headwater source of a small stream or merely effluent seepage which evaporates from the ground surface. Figure 6-7 illustrates several types of springs. The flow of most springs is small and usually of little hydrologic significance, although even a small spring may provide water for a single farmstead. Meinzer[1] classified springs from first to eighth magnitude with respect to flow. First-magnitude springs discharge 100 cfs or more while eighth-magnitude springs have a flow less than 1 pt/min. According to Meinzer[2]

[1] O. E. Meinzer, Outline of Ground-water Hydrology, *U.S. Geol. Survey Water-supply Paper* 494, 1923.

[2] O. E. Meinzer, Large Springs in the United States, *U.S. Geol. Survey Water-supply Paper* 557, 1927.