Transmissivity, specific yield, and storativity

Transmissivity (T) is another concept that is commonly used to describe an aquifer’s capacity to transmit water. It represents the amount of water that can be transmitted horizontally by the entire saturated thickness of the aquifer under a hydraulic gradient of one. It is equal to the multiplication product of the aquifer thickness (b) and the hydraulic conductivity (K).Commonly used units for T are m2/d.

T= K b eq. 1

An aquifer typically serves two functions: (1) a conduit through which flow occurs and (2) a storage reservoir. This is accomplished by the openings in the aquifer matrix. If a unit of saturated formation is allowed to drain by gravity, not all of the water it contains will be released. The ratio of water that can be drained by gravity to the entire volume of a saturated soil is called specific yield, while the part retained is the specific retention.

Table 1 lists typical porosity, specific yield, and specific retention of soil, clay, sand, and gravel. The sum of the specific yield and the specific retention of a formation is equal to its porosity. The specific yield and the specific retention are related to the attraction between water and the formation materials. Clayey formations usually have a lower hydraulic conductivity. This often leads to an incorrect idea that clayey formations have a lower porosity.

As shown in Table 1, clay has a much higher porosity than sand, and sand has a higher porosity than gravel. The porosity of clay can be as high as 50%, but its specific yield is extremely low at 2%. Porosity determines the total volume of water that a formation can store, while specific yield defines the amount that is available to pumping. The low specific yield explains the difficulty of extracting groundwater from clayey aquifers.

When the head in a saturated aquifer changes, water will be taken into or released from storage. Storativity or storage coefficient (S) describes the quantity of water taken into or released from storage per unit change in head per unit area. It is a dimensionless quantity. The response of a confined aquifer to the change of water head is different from that of an unconfined aquifer.

Table 1: Typical porosity, specific yield and specific of selected Materials

|  |  |  |  |
| --- | --- | --- | --- |
|  | Porosity (%) | Specific Yield (%) | Specific retention (%) |
| soil | 55 | 40 | 15 |
| clay | 50 | 2 | 48 |
| sand | 25 | 22 | 3 |
| Gravel | 20 | 19 | 1 |

When the head declines, a confined aquifer remains saturated; the water is released from storage by the expansion of water and compaction of aquifer. The amount of release is exceedingly small. On the other hand, the water table rises or falls with change of head in an unconfined aquifer. As the water level changes, water drains from or enters into the pore spaces. This storage or release is mainly due to the specific yield. It is also a dimensionless quantity. For unconfined aquifers the storativity is practically equal to the specific yield and ranges typically between 0.1 and 0.3. The storativity of confined aquifers is substantially smaller and generally ranges between 0.0001 and 0.00001, and that for leaky confined aquifers is in the range of 0.001. A small storativity implies that it will require a larger pressure change (or gradient) to extract groundwater at a specific flow rate. The volume of groundwater (V) drained from an aquifer can be determined from the following:

(2)

where S is the storativity, A is the area of the aquifer, and ∆h is the change in head. Example 1.

Estimate loss of storage in aquifers due to change of head An unconfined aquifer has an area of 5 square miles. The storativity of this aquifer is 0.15. The water table falls 0.24 m during a drought. Estimate the amount of water lost from storage. If the aquifer is confined and its storativity is 0.0005, what would be the amount lost for a decrease of 0.24 m in head?

Mile = 1600 m

The volume of water drained for the unconfined aquifer as in eq.2

V = 0.15\* 5\*1600\*1600\*0.24=460800m3

For the confined aquifer: V = (0.0005)[(5)(1600)2 (0.24 m) = 1536 m3

Discussion. For the same amount of change in head, the water lost in the unconfined aquifer is 300 times more, which is the ratio of the two storativity values (0.15/0.0005 = 300).

**Determine groundwater flow gradient and flow direction**

Having a good knowledge of the gradient and direction of groundwater flow is vital to groundwater remediation. The gradient and the direction of flow have great impacts on selection of remediation schemes to control plume migration, such as location of the pumping wells and groundwater extraction rates, etc. Estimates of the gradient and direction of groundwater flow can be made with a minimum of three groundwater elevations. The general procedure is described below and an example follows.

Step 1: Locate the three surveyed points on a map to scale.

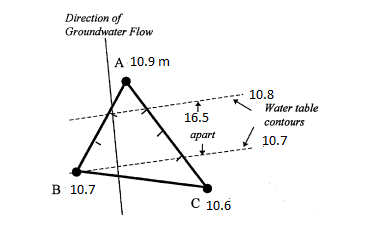
Step 2: Connect the three points and mark their water table elevations on the map. Step 3: Subdivide each side of the triangle into a number of segments of equal size. (Each segment represents an increment of elevation.)

Step 4: Connect the points of equal values of elevation (equipotential lines), which then form the groundwater contours.

Step 5: Draw a line that passes through and is perpendicular to each equipotential line. This line marks direction of flow.

Step 6: Calculate the groundwater gradient from the formula, i = dh/dl.

Example 2: Estimate the gradient and direction of groundwater flow from three groundwater elevations Three groundwater monitoring wells were installed at a contaminated site. Groundwater elevations were determined from a recent survey of these wells and the values were marked on a map. Estimate the flow gradient and direction of the groundwater flow in the underlying aquifer.



Solution: a. Water elevations (10.9m, 10.7m, and 10.6m) were measured at three monitoring wells and marked on the map.

b. These three points are connected by straight lines to form a triangle.

c. Subdivide each side of the triangle into a number of segments of equal intervals. For example, subdivide the line connecting point A (10.9) and point B (10.7) into three intervals. Each interval represents a 0.07m increment in elevation.

d. Connect the points of equal values of elevation (equipotential lines), which then form the groundwater contours. Here, we connect the elevations of 10.8 and 10.7 to form two contour lines. e. Draw a line that passes through and is perpendicular to each equipotential line and mark it as the groundwater flow direction.

f. Measure the distance between two contour lines, 16.5 m in this example. Calculate the groundwater gradient from the formula, i = dh/dl

i= (10.8-10.7)/16.5=0.0061

Discussion. The groundwater elevations, especially those of the water table aquifers, may change with time. Consequently, the groundwater flow gradient and direction would change. Periodic surveys of the groundwater elevation may be necessary if fluctuation of the water table is suspected. Off-site pumping, seasonal change, and recharge are some of the reasons that may cause the fluctuation of the water table elevation.

**Groundwater pumping**

Steady-state flow in a confined aquifer

The equation describing steady-state flow of a confined aquifer (an artesian aquifer) from a fully penetrating well is shown below. A fully penetrating

well means that the groundwater can enter at any level from the top to the bottom of the aquifer.

where Q = pumping rate or well yield (m3/d), h1, h2 = static head measured from the aquifer bottom (in m), r1, r2 = radial distance from the pumping well (in m), b = thickness of the aquifer (in ft or m), and K = hydraulic conductivity of the aquifer (m/d). Many assumptions are made to derive the above equation.to calculate hydraulic conductivity of a confined aquifer, if two steady-state drawdowns, flow rate, and aquifer thickness are available

Another parameter, specific capacity, can also be used to estimate the hydraulic conductivity of an aquifer. Let us define the specific capacity as

where Q = the well discharge rate (extraction rate),in gpm , and sw = draw-down in the pumping well, in ft.

For example, if a well-produced 50 gpm and the drawdown in the well is 5 ft, the specific capacity of this pumping well is 10 gpm/ft; it will produce 10 gpm for each foot of available drawdown. A rough estimate on transmissivity (in gpd/ft) can be obtained by multiplying the specific yield (in gpm/ft) by 2000 for confined aquifers and 1550 for unconfined aquifers. The hydraulic conductivity (in gpd/ft2) can then be determined by dividing the transmissivity with the aquifer thickness (in ft).

1gpm=3.75e-3m3/min

Example: A Steady-state drawdown from pumping a confined aquifer A confined aquifer 30 ft (9.1 m) thick has a piezometric surface 80 ft (24.4 m) above the bottom confining layer. Groundwater is being extracted out from a 4-in (0.1 m) diameter fully penetrating well. The pumping rate is 40 gpm (0.15 m3 /min). The aquifer is relatively sandy with a hydraulic conductivity of 200 gpd/ft2 . Steady-state drawdown of 5 ft (1.5 m) is observed in a monitoring well 10 ft (3.0 m) from the pumping well. Estimate a. The drawdown 30 ft (9.1 m) away from the well

b. The drawdown in the pumping well.

a. Determine h1 at r1 =10ft; h1= 80-5=75ft; or (24.4-1.5=22.9m)

= 22.9 m

So drawdown at 9.1 m away= 24.4-23=1.4m

b) to determine the drawdown at the pumping well, set r at the well=well radius=

2/12ft= 0.05 m

; h2= 22.89=23m

Drawdown in the extraction =24.4-23=1.4

Example B:Estimate hydraulic conductivity of a confined aquifer from steady-state drawdown data Use the following information to estimate the hydraulic conductivity of a confined aquifer: Aquifer thickness = 30.0 ft (9.1 m) thick Well diameter = 4-in (0.1 m) diameter Well perforation depth = fully penetrating Groundwater extraction rate = 20 gpm Steady-state drawdown = 2.0 ft observed in a monitoring well 5 ft from the pumping well = 1.2 ft observed in a monitoring well 20 ft from the pumping well

Example: Estimate hydraulic conductivity of a confined aquifer using specific capacity, take the data from example B

Aquifer thickness=9m; pumping rate =0.15 m3/min, steady state drawdown in the well=3.4m

The transmissivity of the aquifer=(0.044)(600)=26.4m3/d/m

The hydraulic conductivity

Steady state flow in an unconfined aquifer

The equation describing steady state flow of an unconfined aquifer (water table aquifer from fully penetrating well may be written as follows:

It can be easily modified to calculate the hydraulic conductivity of an unconfined aquifer if data of two steady-state drawdowns and flow rate are available.

The specific capacity can also be used to estimate the hydraulic conductivity of an unconfined aquifer.

Example III.2.2B Estimate hydraulic conductivity of an unconfined aquifer from steady-state drawdown data Use the following information to estimate the hydraulic conductivity of an unconfined aquifer: Aquifer thickness = 30.0 ft (9.1 m) thick Well diameter = 4-in (0.1 m) diameter Well perforation depth = fully penetrating Groundwater extraction rate = 20 gpm Steady-state drawdown = 2.0 ft observed in monitoring well 5 ft from the pumping well = 1.2 ft observed in a monitoring well 20 ft from the pumping well Solutions: First we need to determine h1 and h2

Solution: h1= 9.1-0.6=8.5 m; h2=9.1-0.36=8.74m

= 0.137

Example: Estimate hydraulic conductivity for an unconfined aquifer using a specific capacity

Aquifer thickness=24 m; pumping rate =40 gpm, steady state drawdown in the well=2.25m