Contents

Ground Water

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Pressure in the saturated zone experiences positive pressure. 30% of the world's fresh water and 99 % of liquid fresh water is groundwater. 1/4 of domestic, irrigation and industrial water use in the US comes from groundwater.

1. Basic Principles of Ground-Water flow

Bulk flow of water in the saturated zone is governed by Darcy's Law:

$$V_x = -K_{hx}\frac{dh}{hx} \tag{1}$$

where V_x is the specific discharge at right angle to the x-direction. K_{hx} is the saturated hydraulic conductivity in the x-direction, h is the total hydraulic head:

$$h = z + \frac{p}{\gamma_w} \tag{2}$$

The head at any "point" can be measured as the height above the selected arbitrary datum to which water rises in a **piezometer**. Remember that the actual velocity is found by dividing the Darcy Velocity by the the porosity $v = V_x/\phi$. Remember that Darcy only applies to laminar porous media flows.

2. Aquifers

An aquifer is a geologic unit that can store water and transmit it at a rate that is hydrologically significant.

2a. *Unconfined Aquifer*

The upper boundary of the ground-water flow is a water surface at atmospheric pressure ($p = 0$) where the water table can be determined as the water surface in an open well. Recharge occurs by percolation.

2b. *Confined Aquifer*

Confined aquifers are bounded above and below by formations with lower hydraulic conductivity (confining layers or aquicludes). Water level in an observation well will rise above the upper boundary of the aquifer to coincide with the **potentiometric surface** which is an imaginary surface analogous to water table. . Recharge occurs generally upstream where water is not confined. However, most confining layers can transmit some water (leaky aquitards).

2c. *Aquifer Properties*

Specific Storage S_s is the volume of water that a unit volume of aquifer beneath the are releases (takes up) in response to a unit decrease (increase) in head:

$$S_s = \frac{\text{Volume of water released (taken) from storage}}{\text{Volume of aquifer} \times \text{Change in head}} \tag{3}$$

Storage Coefficient S defined for two-dimensional flow in the horizontal plane.

$$S = \frac{\text{Volume of water released (taken) from storage}}{\text{Surface area of aquifer} \times \text{Change in head}} \tag{4}$$

Consequently $S = H \times S_s$ where H is the saturated thickness of aquifer beneath its unit surface area.
For an *unconfined aquifer* a change is head is reflected by a change in the water table and the water content change is characterized by the **specific yield** $S_y = S$. The volume of water retained by the portion of the aquifer experiencing a water table decline is the **specific retention** S_r where $\phi = S_y + S_r$.
For an *confined aquifer* a change is head is reflected by a change in the piezometric surface. The change in storage is due to compression or expansion of the aquifer when there are changes in the weight of liquid and solid weight, and also to the change in volume due to changes in pressure. The storage coefficient is at least an order of magnitude less than the specific

yield for unconfined aquifers.

3. Transmission Properties of Porous Media

The saturated hydraulic conductivity for small samples can be determined in permeameters , where A_s and dx are known, Q and dh are measured and K_{hx} is calculated with Darcy Law, but this may not be reflective of the field. Hydraulic conductivity has significant spatial variability, in general a medium can be isotropic, anisotropic, homogeneous or heterogeneous.

If H doesn't vary much and flow paths are approximately horizontal we use the concept of **transmissivity**.

$$T = H \times K_h \tag{5}$$

4. Groundwater Flow Equations

Using the conservation equation of water and Darcy's Law, we can obtain the 3D continuity equation for porous media flow.

$$\frac{\partial}{\partial x}\left(K_{hx}\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_{hy}\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_{hz}\frac{\partial h}{\partial z}\right) = S_s\frac{\partial h}{\partial t} \tag{6}$$

For an isotropic homogeneous aquifer, K_h doesn't vary with location, so:

$$\frac{\partial^2 h}{\partial x^x} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = \frac{S_s}{K_h}\frac{\partial h}{\partial t} \tag{7}$$

And furthermore, if there is steady-state flow:

$$\frac{\partial^2 h}{\partial x^x} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = 0 \tag{8}$$

Which is **Laplace's Equation**, where we need *boundary conditions*, and when we don't have steady state, we need the initial conditions. While some analytical solutions exist, most of the time we use numerical approximations and iterative procedures.

Equipotential lines are lines of equal head.

Streamlines or flowlines give the direction of groundwater flow and are at right angles to the equipotential lines.

Streamtube is the space between adjacent streamlines.

Flow net is a diagram that shows the equipotential lines and streamlines.

4a. *Dupuit Approximation for Unconfined Flow*

We can use Dupuit formulation when:

1. the control volume extends from a horizontal impermeable base in the xy plane up to the water table.

2. at any point in the xy plane the total head h is constant in the vertical direction so that $V_z = 0$.

3. the head gradients are assumed equal to the slope of the water table.

Dupuit approximation leads to the governing equation for a homogeneous isotropic aquifer:

$$\frac{2R_I}{K_h} + \frac{\partial^2 h^2}{\partial x^2} + \frac{\partial^2 h^2}{\partial y^2} + \frac{\partial^2 h^2}{\partial z^2} = \frac{2S_y}{K_h}\frac{\partial h}{\partial t} \tag{9}$$

Where R_I is the net rate of recharge. For steady state:

$$\frac{2R_I}{K_h} + \frac{\partial^2 h^2}{\partial x^2} + \frac{\partial^2 h^2}{\partial y^2} + \frac{\partial^2 h^2}{\partial z^2} = 0 \tag{10}$$

5. Regional Groundwater Flow

Recharge Area portion of a drainage basin in which groundwater flow is directed away from the water table.

Discharge Area region in which the ground-water flow is directed toward the water table. Water table is usually at or near the surface in discharge areas which are usually surface-water bodies like streams, lakes, wetlands, or strings.

Hinge Line Line separating recharge and discharge areas.

Average rate of discharge = average rate of recharge.
Topography and geology determine the spatial distribution, not the absolute rates. Depending on topography we can identify a local flow system in which water moves from a recharge area to the next adjacent discharge area or a regional flow system in which the flow is from the recharge area farthest from the main valley, and intermediate systems in between.

Figure 1: Figure 8-10

6. Groundwater-Surface Water Interactions

Water can enter streams promptly in response to individual water-input events (event flow, direct flow, storm flow or quick flow). This is different from **base flow** which is water that enters from persistent, slowly varying sources and maintains streamflow between water-input events. Streams that receive large proportions of their flow as groundwater base flow tend to have relatively low temporal flow variability.

Gaining or effluent Stream Stream or reach that occurs in a discharge area and receives groundwater flow- discharge increases downstream.

Losing or influent Stream where discharge decreases downstream, may occur in a recharge zone and be connected or "perched" above the general groundwater flow.

Flow-through Stream simultaneously receives and loses groundwater.

Streams that flow all year are **perennial streams**, those that flow only during wet seasons are **intermittent sreams**. These are almost always gaining streams that are sustained by groundwater flow between water input events. Ephemeral streams flow only in response to water input events and are usually losing.

At a more local scale the river can be locally permeable.

Hyporheic zone is the zone of down-river groundwater flow in the bed.

Bank storage is the lateral exchange of water between the channel and banks during high flows.

Figure 2: Figure 8-18

Figure 3: Figure 8-21