Groundwater

U.S. Groundwater Law

Groundwater can either be privately owned or publicly owned. Groundwater owned by the State is usually distributed through an appropriation system. Privately owned groundwater may allow unlimited production or limited production rights based on land ownership or liability rules. It is possible to regulate the spacing of wells and groundwater production under any of these systems, but the methods, effectiveness and results of that regulation varies greatly from one system to the next depending on the type of aquifer to be regulated. Effective regulation is tailored to both the hydrology and economics of the region to be regulated. Below is a short discussion of each system and the advantages and disadvantages of each.

The Rule of Capture provides each landowner the ability to capture as much groundwater as they can put to a beneficial use, but they are not guaranteed any set amount of water. As a result, well-owners are not liable to other landowners for damaging their wells or taking water from beneath their land. The Rule of Capture allows capture of groundwater only to the extent the use is beneficial and not malicious. The advantage of this system is that it encourages economic development and maximum utilization of the resources. Another advantage of this system is that it leads to minimal government involvement in the operations of water

wells. The primary disadvantage of this system is the potential for overproduction of the aquifer system that may result when each landowner attempts to protect the water right by drilling bigger, deeper wells. Because no landowner is given a quantifiable or set amount of production capacity, all landowners are encouraged to capture as much water as they can as quickly as they can.

Correlative groundwater rights represent a limited private ownership right similar to riparian rights in a surface stream. The amount of groundwater right is based on the size of the surface acreage where each landowner gets a corresponding amount of the available water. Once adjudicated, the maximum amount of the water right is set, but the right can be decreased if the total amount of available water decreases as is likely during a drought. Landowners may sue others for encroaching upon their groundwater rights, and water pumped for use on the overlying land takes preference over water pumped for use off the land. This system benefits those who have low demand for water but own large expanses of property--such as ranchers--and harms those who have a high demand for water without correspondingly large tracts of land--such as cities and some irrigators. Only California follows the correlative right system for groundwater, although many states follow a similar system for oil and gas production. Because water is a rechargeable resource and the amount of the water right may be reduced, marketing the groundwater right can be

difficult. The preference for water uses on the land makes it difficult to market the water or water rights.

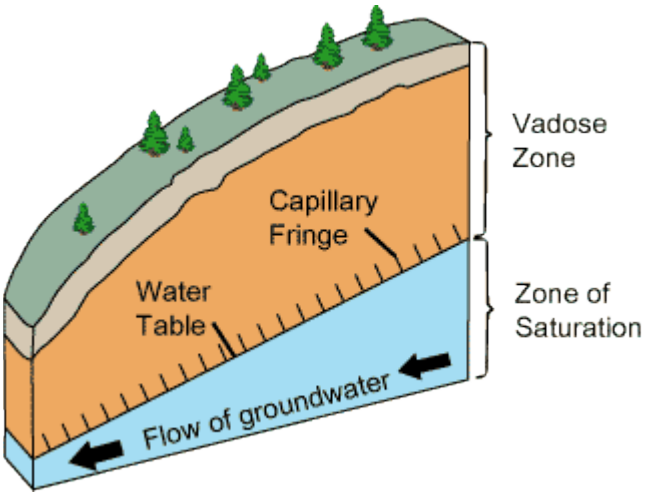
The third system involving private ownership rights is the liability rule known as the American Rule or Reasonable Use Rule. This rule does not guarantee the landowner a set amount of water, but allows unlimited pumpage as long as the result does not unreasonably damage other wells or the aquifer system. Usually this rule gives great weight to historical uses and prevents new uses that would interfere with the prior use. The determination of who gets a well and how much water may be pumped is usually made by a court unless the state creates a regulatory agency to perform that function, and the primary issue is the "reasonableness" of the use. The advantage to this system is its flexibility in adjudicating competing uses of an aquifer system. Unfortunately, this same flexibility can lead to excessive litigation because well owners may sue at any time to determine if a competing use is "reasonable," a standard that may change with time. The reasonableness standard is also highly dependent on the location of the suit and who ends up in the jury pool. Marketing water rights does not take place until the system is fully adjudicated; new users generally do not purchase groundwater rights until they are sure they cannot obtain “free” water through litigation.

Many states, especially in the western United States, claim ownership of groundwater and allocate the resource through an appropriative system

just as they would any surface right. Typically water rights are appropriated based on each aquifer's sustainable yield, and once all the rights are granted no further permits will be issued. Some states allow the permits to be marketed and some do not. Where the water is not owned by the state and the tort law proves to be an inadequate means to prevent overproduction, states have created administrative regulatory agencies to allocate groundwater rights between competing landowners. In those cases the administrative law essentially supplants the tort law, making the tort remedy (or lack thereof) irrelevant.

Vadose zone

Cross-section of a hill slope depicting the vadose zone, capillary fringe, water table, and phreatic or saturated zone.



The **vadose zone**, also termed the **unsaturated zone**, is the portion of Earth between the land surface and the phreatic zone or zone of saturation ("vadose" is Latin for "shallow"). Water in the vadose zone has a pressure head less than atmospheric pressure, and is retained by a combination of adhesion (*funiculary groundwater*), and capillary action (*capillary groundwater*). If

the vadose zone envelops soil, the water contained therein is termed soil moisture.

Movement of water within the vadose zone is studied within soil physics and hydrology, particularly hydrogeology, and is of importance to agriculture, contaminant transport, and flood control. The Richards equation is often used to mathematically describe the flow of water, which is based partially on Darcy's law. Recharge, which is an important process that refills aquifers, generally occurs through the vadose zone from precipitation.

In speleology, cave passages formed in the vadose zone tend to be canyon-like in shape, as the water dissolves bedrock on the floor of the passage. Passages created in completely water-filled conditions are called phreatic passages and tend to be circular in cross-section.

Hydrogeology

Hydrogeology (*hydro-* meaning water, and *-geology* meaning the study of rocks) is the part of hydrology that deals with the distribution and movement of groundwater in the soil and rocks of the Earth's crust, (commonly in aquifers). The term **geohydrology** is often used interchangeably. Some make the minor distinction between a hydrologist or engineer applying them to geology (geohydrology), and a geologist applying them to hydrology (hydrogeology).

Introduction

Hydrogeology (like most earth sciences) is an interdisciplinary subject; it can be difficult to account fully for the chemical, physical, biological and even legal interactions between soil, water, nature and society. Although the basic principles of hydrogeology are very intuitive (e.g., water flows "downhill"), the study of their interaction can be quite complex. Taking into account the interplay of the different facets of a multi-component system often requires knowledge in several diverse fields at both the experimental and theoretical levels. This being said, the following is a more traditional (reductionist viewpoint) introduction to the methods and nomenclature of saturated subsurface hydrology, or simply hydrogeology.

Hydrogeology in relation to other fields

Hydrogeology, as stated above, is a branch of the earth sciences dealing with the flow of water through aquifers and other shallow porous media (typically less than 450 m or 1,500 ft below the land surface.) The very shallow flow of water in the subsurface (the upper 3 m or 10 ft) is pertinent to the fields of soil science, agriculture and civil engineering, as well as to hydrogeology. The general flow of fluids (water, hydrocarbons, geothermal fluids, etc.) in deeper formations is also a concern of

geologists, geophysicists and petroleum geologists. Groundwater is a slow-moving, viscous fluid (with a Reynolds number less than unity); many of the empirically derived laws of groundwater flow can be alternately derived in fluid mechanics from the special case of *Stokes flow* (viscosity and pressure terms, but no inertial term).

The mathematical relationships used to describe the flow of water through porous media are the diffusion and Laplace equations, which have applications in many diverse fields. Steady groundwater flow (Laplace equation) has been simulated using electrical, elastic and heat conduction analogies. Transient groundwater flow is analogous to the diffusion of heat in a solid; therefore some solutions to hydrological problems have been adapted from heat transfer literature.

Traditionally, the movement of groundwater has been studied separately from surface water, climatology, and even the chemical and microbiological aspects of hydrogeology (the processes are uncoupled). As the field of hydrogeology matures, the strong interactions between groundwater, surface water, water chemistry, soil moisture and even climate are becoming clearer.

Definitions and material properties

One of the main tasks a hydro geologist typically performs is the prediction of future behavior of an aquifer system, based on analysis of

past and present observations. Some hypothetical, but characteristic questions asked would be:

- Can the aquifer support another subdivision?
- Will the river dry up if the farmer doubles his irrigation?
- Did the chemicals from the dry cleaning facility travel through the aquifer to my well and make me sick?
- Will we need to start drinking bottled water?

Most of these questions can be addressed through simulation of the hydrologic system (using numerical models or analytic equations). Accurate simulation of the aquifer system requires knowledge of the aquifer properties and boundary conditions. Therefore a common task of the hydro geologist is determining aquifer properties using aquifer tests.

In order to further characterize aquifers and aquitards some primary and derived physical properties are introduced below. Aquifers are broadly classified as being either confined or unconfined (water table aquifers), and either saturated or unsaturated; the type of aquifer affects what properties control the flow of water in that medium (e.g., the release of water from storage for confined aquifers is related to the storativity, while it is related to the specific yield for unconfined aquifers).

Hydraulic head

Changes in hydraulic head (h) are the driving force that causes water to move from one place to another. It is composed of pressure head (ψ) and elevation head (z). The head gradient is the change in hydraulic head per length of flow path, and appears in Darcy's law as being proportional to the discharge.

Hydraulic head is a directly measurable property, which can take on any value (because of the arbitrary datum involved in the z term); ψ can be measured with a pressure transducer (this value can be negative, e.g., suction, but is positive in saturated aquifers), and z can be measured relative to a surveyed datum (typically the top of the well casing). Commonly, in wells tapping unconfined aquifers the water level in a well is used as a proxy for hydraulic head, assuming there is no vertical gradient of pressure. Often only *changes* in hydraulic head through time are needed, so the constant elevation head term can be left out ($\Delta h = \Delta\psi$).

A record of hydraulic head through time at a well is a hydrograph or, the changes in hydraulic head recorded during the pumping of a well in a test are called drawdown.

Porosity

Porosity (n) is a directly measurable aquifer property; it is a fraction between 0 and 1 indicating the amount of pore space between

unconsolidated soil particles or within a fractured rock. Typically, the majority of groundwater (and anything dissolved in it) moves through the porosity available to flow (sometimes called effective porosity).

Porosity does not directly affect the distribution of hydraulic head in an aquifer, but it has a very strong effect on the migration of dissolved contaminants, since it affects groundwater flow velocities through an inversely proportional relationship.

Water content

Water content (θ) is also a directly measurable property; it is the fraction of the total rock that is filled with liquid water. This is also a fraction between 0 and 1, but it must also be less than or equal to the total porosity.

The water content is very important in vadose zone hydrology, where the hydraulic conductivity is a strongly nonlinear function of water content; this complicates the solution of the unsaturated groundwater flow equation.

Hydraulic conductivity

Hydraulic conductivity (K) and transmissivity (T) are indirect aquifer properties (they cannot be measured directly). T is the K integrated over the vertical thickness (b) of the aquifer ($T=Kb$ when K is constant over the entire thickness). These properties are measures of an aquifer's ability to

transmit water. Intrinsic permeability (κ) is a secondary medium property which does not depend on the viscosity and density of the fluid (K and T are specific to water); it is used more in the petroleum industry.

Specific storage and specific yield

Specific storage (S_s) and its depth-integrated equivalent, storativity ($S=S_s b$), are indirect aquifer properties (they cannot be measured directly); they indicate the amount of groundwater released from storage due to a unit depressurization of a confined aquifer. They are fractions between 0 and 1.

Specific yield (S_y) is also a ratio between 0 and 1 ($S_y \leq$ porosity) that indicates the amount of water released due to drainage, from lowering the water table in an unconfined aquifer. Typically S_y is orders of magnitude larger than S_s. Often the porosity or effective porosity is used as an upper bound to the specific yield.

Contaminant transport properties

Often we are interested in how the moving groundwater water will move dissolved contaminants around (the sub-field of contaminant hydrogeology). The contaminants can be man-made (e.g., petroleum products, nitrate or Chromium) or naturally occurring (e.g., arsenic, salinity). Besides needing to understand where the groundwater is flowing, based on the other hydrologic properties discussed above, there are

additional aquifer properties that affect how dissolved contaminants move with groundwater.

Dispersivity (α_L, α_T) is an empirical factor that quantifies how much contaminants stray away from the path of the groundwater that is carrying it. Some of the contaminants will be "behind" or "ahead" the mean groundwater, giving rise to a longitudinal dispersivity (α_L), and some will be "to the sides of" the pure advective groundwater flow, leading to a transverse dispersivity (α_T).

Dispersivity is actually a factor that represents our *lack of information* about the system we are simulating. There are many small details about the aquifer that are being averaged when using a macroscopic approach (e.g., tiny beds of gravel and clay in sand aquifers); they manifest themselves as an *apparent* dispersivity. Because of this, α is often claimed to be dependent on the length scale of the problem — the dispersivity found for transport through 1 m^3 of aquifer is different than that for transport through 1 cm^3 of the same aquifer material.

Hydrodynaic dispersion (D) is a positive physical parameter which describes the molecule-scale movement of solute away from the mean flow; it is a result of Brownian motion. This is the same mechanism as dye uniformly spreading out in a still bucket of water. The dispersion coefficient is typically quite small (typically orders of magnitude smaller

than α), and can often be considered negligible (unless groundwater flow velocities are extremely low, as they are in clay aquitards).

It is important not to confuse hydrodynamic dispersion with dispersivity, as the former is a physical phenomenon and the latter is an empirical factor that is cast into a similar form as dispersion, because we already know how to solve that problem.

Governing equations

Darcy's Law

Darcy's law is a Constitutive equation (empirically derived by Henri Darcy, in 1856) which states the amount of groundwater discharging through a given portion of aquifer is proportional to the cross-sectional area to flow, the hydraulic head gradient and the hydraulic conductivity.

Groundwater flow equation

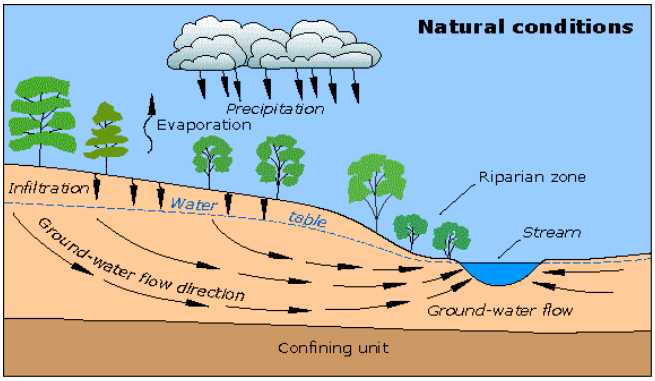
The groundwater flow equation, in its most general form, describes the movement of groundwater in a porous medium (aquifers and aquitards). It is known in mathematics as the diffusion equation, and has many analogs in other fields. Many solutions for groundwater flow problems were borrowed or adapted from existing heat transfer solutions.

It is often derived from a physical basis using Darcy's law and a conservation of mass for a small control volume. The equation is often

used to predict flow to wells, which have radial symmetry, so the flow equation is commonly solved in polar or cylindrical coordinates.

The Theis equation is one of the most commonly used and fundamental solutions to the groundwater flow equation; it can be used to predict the transient evolution of head, due to the effects of pumping one or a number of pumping wells.

The Thiem equation is a solution to the steady state groundwater flow equation (Laplace's Equation). Unless there are large sources of water nearby (a river or lake), true steady state is rarely achieved in reality.



Calculation of groundwater flow

To use the groundwater flow equation to estimate the distribution of hydraulic heads, or the direction and rate of groundwater flow, this partial differential equation (PDE) must be solved. The most common means of analytically solving the diffusion equation in the hydrogeology literature are:

- Laplace, Hankel and Fourier transforms (to reduce the number of dimensions of the PDE),

- similarity transform (also called the Boltzmann transform) is commonly how the Theis solution is derived,
- separation of variables, which is more useful for non-Cartesian coordinates, and
- Green's functions, which is another common method for deriving the Theis solution — from the fundamental solution to the diffusion equation in free space.

No matter which method we use to solve the groundwater flow equation, we need both initial conditions (heads at time (t) = 0) and boundary conditions (representing either the physical boundaries of the domain, or an approximation of the domain beyond that point). Often the initial conditions are supplied to a transient simulation, by a corresponding steady-state simulation (where the time derivative in the groundwater flow equation is set equal to 0).

There are two broad categories of how the (PDE) would be solved; either analytical methods, numerical methods, or something possibly in between. Typically, analytic methods solve the groundwater flow equation under a simplified set of conditions *exactly*, while numerical methods solve it under more general conditions to an *approximation*.

Analytic methods

Analytic methods typically use the structure of mathematics to arrive at a simple, elegant solution, but the required derivation for all but the simplest domain geometries can be quite complex (involving non-standard coordinates, conformal mapping, etc.). Analytic solutions typically are also simply an equation, which can give a quick answer based on a few basic parameters. The Theis equation is a very simple (yet still very useful) analytic solution to the groundwater flow equation, typically used to analyze the results of an aquifer test or slug test.

Numerical methods

The topic of numerical methods is quite large, obviously being of use to most fields of engineering and science in general. Numerical methods have been around much longer than computers have (In the 1920s Richardson developed some of the finite difference schemes still in use today, but they were calculated by hand, using paper and pencil, by human "calculators"), but they have become very important through the availability of fast and cheap personal computers. A quick survey of the main numerical methods used in hydrogeology, and some of the most basic principles are below.

There are two broad categories of numerical methods: gridded or discretized methods and non-gridded or mesh-free methods. In the

common finite difference method and finite element method (FEM) the domain is completely gridded ("cut" into a grid or mesh of small elements). The analytic element method (AEM) and the boundary integral equation method (BIEM — sometimes also called BEM, or Boundary Element Method) are only discretized at boundaries or along flow elements (line sinks, area sources, etc.), the majority of the domain is mesh-free.

General properties of gridded methods

Gridded Methods like finite difference and finite element methods solve the groundwater flow equation by breaking the problem area (domain) into many small elements (squares, rectangles, triangles, blocks, tetrahedra, etc.) and solving the flow equation for each element (all material properties are assumed constant or possibly linearly variable within an element), then linking together all the elements using conservation of mass across the boundaries between the elements (similar to the divergence theorem). This results in a system that overall approximates the groundwater flow equation, but exactly matches the boundary conditions (the head or flux is specified in the elements which intersect the boundaries).

Finite differences are a way of representing continuous differential operators using discrete intervals (Δx and Δt), and the finite difference methods are based on these (they are derived from a Taylor series). For

example the first-order time derivative is often approximated using the following forward finite difference, where the subscripts indicate a discrete time location,

$$\frac{\partial h}{\partial t} = h'(t_i) \approx \frac{h_i - h_{i-1}}{\Delta t}.$$

The forward finite difference approximation is unconditionally stable, but leads to an implicit set of equations (that must be solved using matrix methods, e.g. LU or Cholesky decomposition). The similar backwards difference is only conditionally stable, but it is explicit and can be used to "march" forward in the time direction, solving one grid node at a time (or possibly in parallel, since one node depends only on its immediate neighbors). Rather than the finite difference method, sometimes the Galerkin FEM approximation is used in space (this is different from the type of FEM often used in structural engineering) with finite differences still used in time.

Application of finite difference models

MODFLOW is a well-known example of a general finite difference groundwater flow model. It is developed by the US Geological Survey as a modular and extensible simulation tool for modeling groundwater flow. It is free software developed, documented and distributed by the USGS. Many commercial products have grown up around it, providing graphical user interfaces to its input file based interface, and typically incorporating

pre- and post-processing of user data. Many other models have been developed to work with MODFLOW input and output, making linked models which simulate several hydrologic processes possible (flow and transport models, surface water and groundwater models and chemical reaction models), because of the simple, well documented nature of MODFLOW.

Application of finite element models

Finite Element programs are more flexible in design (triangular elements vs. the block elements most finite difference models use) and there are some programs available (SUTRA, a 2D or 3D density-dependent flow model by the USGS; Hydrus, a commercial unsaturated flow model; FEFLOW, a commercial modeling environment for subsurface flow, solute and heat transport processes; and COMSOL Multiphysics (FEMLAB) a commercial general modeling environment), but unless they are gaining in importance they are still not as popular in with practicing hydrogeologists as MODFLOW is. Finite element models are more popular in university and laboratory environments, where specialized models solve non-standard forms of the flow equation (unsaturated flow, density dependent flow, coupled heat and groundwater flow, etc.)

Other methods

These include mesh-free methods like the Analytic Element Method (AEM) and the Boundary Element Method (BEM), which are closer to analytic solutions, but they do approximate the groundwater flow equation in some way. The BEM and AEM exactly solve the groundwater flow equation (perfect mass balance), while approximating the boundary conditions. These methods are more exact and can be much more elegant solutions (like analytic methods are) but have not seen as widespread use outside academic and research groups

