Presented at SDG Short Course II on Exploration and Development of Geothermal Resources,
organized by UNU-GTP, GDC and KenGen, at Lake Bogoria and Lake Naivasha, Kenya, Nov. 9-29, 2017.

BASIC HYDROGEOLOGY IN GEOTHERMAL SYSTEMS

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ABSTRACT

Hydrogeology is the part of hydrology that deals with the occurrence, movement and quality of water beneath the Earth's surface. Hydrogeology deals with water in a complex subsurface environment, but much of its basic terminology and principles can be understood readily by non-hydrogeologists.

This paper presents basic terms and principles of hydrogeology in geothermal systems. The first section introduces many key terms and concepts in definition form. The following section introduces principles of ground water movement, using these terms. The last section gives an overview of the hydrogeology of the Kenyan rift, with special reference to geothermal system.

1. INTRODUCTION

Hydrogeology (hydro- meaning water, and geology meaning the study of the Earth) is the area of geology that deals with the occurrence, 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 an engineer applying himself to geology (geohydrology), and a geologist applying himself to hydrology (hydrogeology).

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.

When rain falls to the ground, some water flows along the land surface to streams or lakes, some water evaporates into the atmosphere, some is taken up by plants, and some seeps into the ground. As water begins to seep into the ground, it enters a zone that contains both water and air, referred to as the unsaturated zone or vadose zone. The upper part of this zone, known as the root zone or soil zone, supports plant growth and is crisscrossed by living roots, holes left by decayed roots, and animal and worm burrows. Below lies an intermediate zone, followed by a saturated capillary fringe, which results from the attraction between water and rocks. As a result of this attraction, water clings as a film on the surface of rock particles. Water moves through the unsaturated zone into the saturated zone, where all the interconnected openings between rock particles are filled with water. It is within this saturated zone that the term "ground water" is correctly applied.

2. TERMINOLOGIES

Ground water is water held within the interconnected openings of saturated rock beneath the land surface. Figure 1 illustrates some of the terms used.

Artesian well - A well whose source of water is a confined (artesian) aquifer. The water level in artesian wells stands at some height above the water table because of the pressure (artesian pressure) of the aquifer. The level at which it stands is the potentiometric (or pressure) surface of the aquifer. If the potentiometric surface is above the land surface, the well is a flowing artesian well.

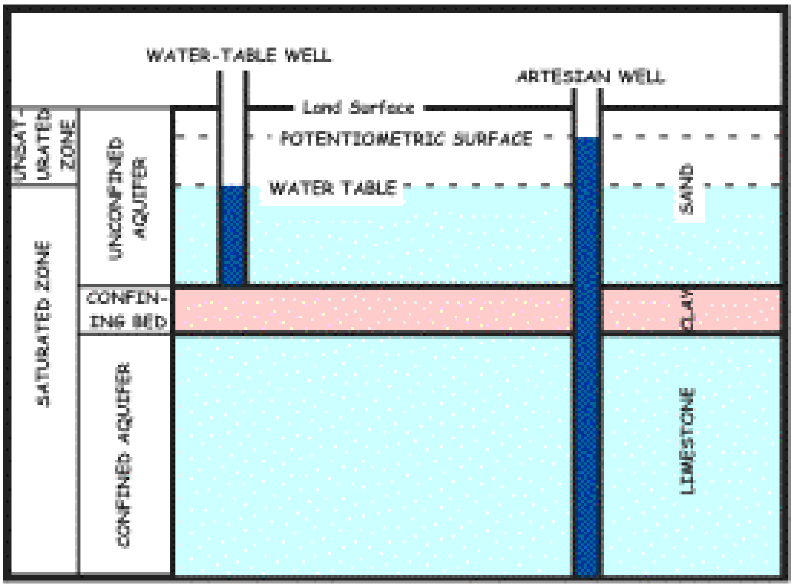

FIGURE 1: An illustration of groundwater terms, adapted from Heath (1993)

Water table - The top of an unconfined aquifer below which the pore spaces are generally saturated; the level in the saturated zone at which the pressure is equal to the atmospheric pressure.

Capillary fringe - The area of the saturated zone just above the water table in which water is held in the soil by surface tension.

Confined aquifer - An aquifer in which ground water is held under pressures greater than atmospheric pressure by upper and lower confining layers, forcing water to rise in wells to heights above the top of the aquifer (artesian wells). Also known as artesian aquifer.

Confining layer - A layer of geologic material which hampers the movement of water into and out of an aquifer. Examples are unfractured igneous rock and metamorphic rock, and shale, or consolidated sediments such as clays. This is also known as a confining bed.

Consolidated rock/ Bedrock - A general term for the solid rock that underlies soils or other surficial material; consists of mineral and/or rock particles of different sizes and shapes that have been welded into a mass by heat and pressure or by chemical reaction. This rock must contain interconnected pores or fractures to serve as an aquifer.

Discharge - The movement of ground water to the surface into a spring, lake, river, or other surface water body; or outflow of ground water from a pumping or flowing well.

Groundwater aquifer - A water-bearing layer of rock or sediment capable of yielding usable quantities of water; composed of unconsolidated materials such as sands and gravel, or fractured consolidated rock such as sandstone or fractured limestone and lava. Note: Although the term *aquifer* is used to mean *ground water aquifer,* there are also oil aquifers, natural gas aquifers, and saltwater aquifers.

Perched aquifer - An aquifer in which a ground water body is separated from the main ground water below it by an impermeable layer (which is relatively small laterally) and an unsaturated zone. Perched aquifers are common in glacial outwash, where lenses of clay formed in small glacial ponds are present. They are also common in volcanic depositional sequences where weathered ash layers of low permeability are sandwiched between high permeability basalts. Water moving downward through the unsaturated zone will be intercepted and accumulate on top of the lens before it moves laterally to the edge of the lens and seeps downward to the regional water table or forms a spring on the side of a hillslope.

Permeability - The capacity of a porous rock, sediment, or soil to transmit ground water. It is a measure of the inter-connectedness of a material's pore spaces and the relative ease of fluid flow under unequal pressure.

Pores - The spaces between particles within geological material (rock or sediment) occupied by water and/or air.

Porosity is defined as the ratio of the volume of voids to the volume of aquifer material. It refers to the degree to which the aquifer material possesses pores or cavities which contain air or water (compare permeability).

Drawdown - The vertical drop of the water level in a well caused by ground water pumping; also, the difference between the water level before pumping and the water level during pumping, also water rest level.

Overwithdrawal - Withdrawal of ground water from an aquifer at a rate that exceeds the recharge rate of that aquifer. This can lead to lowered water table, saltwater intrusion and sinkholes.

Recharge area/zone - Recharge is the process that allows water to replenish an aquifer. This process occurs naturally when rainfall filters down through the soil or rock into an aquifer. Artificial recharge is achieved through the pumping (called injection) of water into wells or by spreading water over the surface where it can seep into the ground. The land area where recharge occurs is called the recharge area or recharge zone.

Saturated zone - The subsurface zone in which all pores in the aquifer are filled with water.

Unsaturated zone - The subsurface zone in which the geological material contains both water and air in pore spaces. The top of the unsaturated zone typically is at the land surface, otherwise known as the vadose zone.

Spring - A place where ground water naturally comes to the surface at the intersection of the water table and land surface.

Subsidence - The sinking or depression of the land surface as a result of too much ground water withdrawal (or overwithdrawal of any mined fluid such as petroleum, gas).

Watershed - All the land area and water within the confines of a drainage divide in which all surface runoff will pass through an identifiable outlet, such as a stream or river.

Cone of depression - As water is withdrawn from a well, the water level in the well begins to decline as water is removed from storage in the well. The head in the well will fall below the level of the surrounding aquifer and water begins moving from the aquifer into the well. The water level will continue to decline and the flow rate of water into the well will increase until the inflow rate is equal to withdrawal rate. Water from the aquifer must converge on the well from all directions and the hydraulic gradient must get steeper near the well. For this reason the resultant 3-D shape of water withdrawal is a called a cone of depression.

3. PRINCIPLES OF A GROUND WATER MOVEMENT

Hydrogeology is an interdisciplinary subject; it can be difficult to account fully for the chemical, physical and biological interactions between soil, water and nature. The study of the interaction between groundwater movement and geology can be quite complex. Groundwater does not always flow in the subsurface down-hill following the surface topography; groundwater follows pressure gradients (flow from high pressure gradient to low) often following fractures and conduits in circuitous paths. 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.

3.1 Governing equations

Darcy's law - Darcy's law is a constitutive equation (empirically derived by Henri Darcy, in 1856) that states the amount of groundwater discharging through a given portion of aquifer is proportional to the cross-sectional area of 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 analogies 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.

3.2 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, while numerical methods solve it under more general conditions to an approximation.

3.2.1 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. Analytic solutions typically are simply an equation that can give a quick answer based on a few basic parameters. The Theis equation is a very simple analytic solution to the groundwater flow equation, typically used to analyze the results of an aquifer test or slug test.

3.2.2 Numerical methods

The topic of numerical methods is quite large, obviously being of use to most fields of engineering and science in general. There are two broad categories of numerical methods used in hydrogeology: 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.

3.3 Hydraulic head

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

Hydraulic head is a directly measurable property that can take on any value; pressure head can be measured with a pressure transducer, and elevation head 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.

Darcy's law & hydraulic conductivity (K). In 1856 the French engineer Henry Darcy successfully quantified several factors controlling ground water movement. These factors are expressed in an equation that is commonly known as Darcy's Law.

$$Q = KA\left(\frac{d\boldsymbol{h}}{dl}\right) \tag{1}$$

where Q = Discharge (volume of water per unit time);
K = Hydraulic conductivity (depends upon size and arrangement of pores, and fluid dynamics such as viscosity, density and gravitational effects);
A = Cross-sectional area (at a right angle to ground water flow direction); and
dh/dl = Hydraulic gradient (the common notation for a change in head per unit distance).

By rearranging Darcy's Law and solving for hydraulic conductivity (K) in common units we can get a sense of what hydraulic conductivity really represents:

$$K = \frac{Qdl}{Adh}\ or, K = \frac{Q = volume\ in\ ft^3\ per\ time\ (day)}{A = a\ cross - sectional\ area\ in\ ft^2}\left(\frac{dl = distance\ in\ ft}{d\boldsymbol{h} = head\ c\boldsymbol{h}ange\ in\ ft}\right)$$

Thus, in practical terms hydraulic conductivity is the volume of water flowing through a 1 ft. x 1 ft. cross-sectional area of an aquifer under a hydraulic gradient of 1 ft. / 1 ft. in a given amount of time (usually a day). If we cancel out units, we see that hydraulic conductivity is usually expressed in ft/day.

Hydraulic conductivity ranges approximately 12 orders of magnitude depending upon differing water transmitting characteristics of aquifer materials (Figure 2). As one may conclude from the chart below, the volume of water that can flow from sandy and cavernous carbonate aquifer materials of the coastal plains are in most cases far greater than igneous and metamorphic rocks.

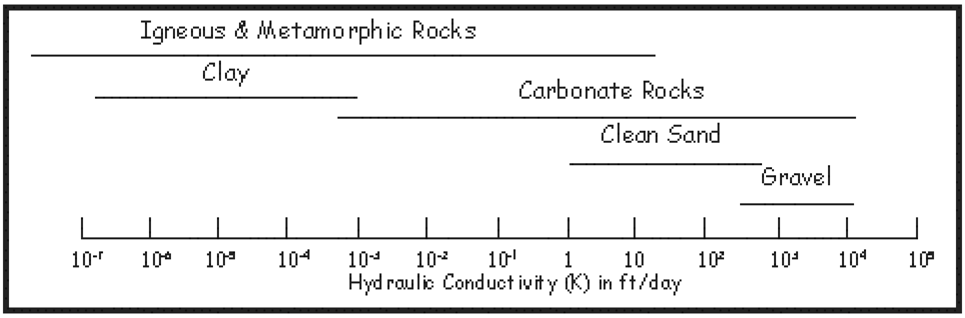

FIGURE 2: Hydraulic conductivity in different aquifer material

There are also some other terms used when regarding the spatial distribution of hydraulic conductivity within an aquifer. If the value of hydraulic conductivity is the same in all directions, the aquifer is said to be isotropic. If hydraulic conductivity is different in different directions, the aquifer is said to be anisotropic. The terms homogeneous and heterogeneous are used when comparing hydraulic conductivity at separate locations of the aquifer, regardless whether the value is the same or different in all directions. The aquifer is homogeneous if hydraulic conductivity is the same and heterogeneous if different. The diagram below shows the four possible combinations when describing the hydraulic conductivity of aquifers.

Another important factor controlling ground water movement is its velocity. The ground water velocity equation can be derived from a combination of the velocity equation of hydraulics and Darcy's law, with the velocity equation as:

$$Q = Av \tag{2}$$

where Q = Discharge (volume of water per unit time); and
v = Velocity.

Solving for ground water velocity gives

$$v = k\left(\frac{d\boldsymbol{h}}{dl}\right) \tag{3}$$

In practical terms we know that ground water does not move in open space, rather it moves through aquifer materials that impede ground water velocity. To account for this we need to include effective porosity (specific yield) to accurately quantify ground water velocity. Remember that hydro-geologists use effective porosity because this value better represents water flowing through an aquifer under the forces of gravity. Thus, the velocity equation is modified as:

$$v = \left(\frac{Kd\boldsymbol{h}}{n_e dl}\right) \tag{4}$$

where n_e = Effective porosity.

3.4 Transmissivity (T)

Transmissivity (T) is the volume of water flowing through a cross-sectional area of an aquifer that is 1 ft. x the aquifer thickness (b), under a hydraulic gradient of 1 ft./ 1 ft. in a given amount of time (usually a day). Thinking about the definition of hydraulic conductivity, we can conclude that transmissivity (T) is actually equal to hydraulic conductivity (K) times aquifer thickness (b). Or otherwise denoted as T = Kb. We can also conclude that transmissivity is expressed as ft2/day because if T = Kb, then T = (ft./day)(ft./1).

3.5 Storage coefficient (S)

The "S" is used to represent the storage coefficient of an aquifer which is the volume of water released from an aquifer per 1 foot surface area per 1 foot change in head. Notice that we are not speaking of water flowing through an aquifer, rather we are referring an aquifer's ability to store water. Mathematically, the storage coefficient is dimensionless as the equation below illustrates:

$$S = \frac{Q = volume\ of\ water\ in\ ft^3}{(surface\ area\ in\ ft^2)(\boldsymbol{h}ead\ c\boldsymbol{h}ange\ in\ ft)} \tag{5}$$

The size of the storage coefficient is dependent whether the aquifer is unconfined or confined. In regard to a confined aquifer, water derived from storage is relative to the expansion of water as the aquifer is depressurized (pumped) and compression of the aquifer.

4. DISCUSSION

The aquifer parameters of transmissivity (T) and storage coefficient (S) are variables that can dictate the shape of the cone of depression. Relative to a confined aquifer, the expansion of water in response to depressurizing (pumping) is very small, yet compression of the rock skeleton is great. This permits the cone of depression to expand and deepen rapidly when pumped. Thus, lower "S" values create deeper and wider cones than higher "S" value aquifers that do not deepen and expand as readily. In regards to transmissivity, aquifers of low "T" develop deep and narrow cones of depression and aquifers of high "T" area characterized by shallow and wide cones.

Utilizing the heating and cooling demand for a geothermal project, the best source water for the geothermal system will be determined based on an understanding of the local geology and hydrogeology. When an underground source is to be utilized, there are several source "development" tools available to help achieve a higher probability for success. This includes the use of fracture trace analysis, coupled with very low frequency (VLF) surveys and magnetometer readings. A review of well completion records for a region can also be conducted to determine aquifer (bedrock) geology and yield potential. If necessary, a pumping test can be performed on both surface and groundwater sources for the determination of a long-term safe yield for the water source, and to determine what effects (changes in water chemistry, temperature, level, etc.) the pumping is predicted to have. These data are incorporated into the thermal model to "fine tune" the predicted capability of the heat/cooling source capabilities.

Study of a possible geothermal system would include geologic mapping, hydrogeochemical analyses, and geophysical interpretation. Geologic and geophysical methods emphasized structure, particularly faults, reservoir and heat sources. Hydrogeochemical analyses shows the geochemistry of the expected geothermal fluids and whether they are of meteoric origin, so as to create a geothermal model predicting reservoir temperatures, including location and trend of heat flow anomalies. For example the proposed model may call for deep circulation of meteoric water along fault zones. Surface waters from nearby rivers, lakes or catchment areas percolating down the faults and being heated to high temperatures. The thermal waters then migrate laterally, then rises back up the fault.

In geothermal exploration and exploitation the application of reactive flow modelling provides a means to lower risk, costs and time during reservoir management, remediation or exploration targeting. In order to deal with the complexity of natural systems, simplified models are employed to illustrate the principal and regulatory factors controlling a chemical system. Following the aphorism of Albert Einstein: "Everything should be made as simple as possible, but not simpler", models need not to be completely realistic to be useful, but need to meet a successful balance between realism and practicality. Properly constructed, a model is neither so simplified to be unrealistic nor too entailed so that it cannot be readily evaluated and applied to the problem of interest. The results of a model have to be at least partially observable or experimentally verifiable.

Reactive transport modelling is extremely useful in understanding the spatial and temporal distributions of solute concentrations and mineral assemblages in the environment. The main target of reactive flow modelling is the simulation on a real time scale with real spatial coordinates, but generally this goal is only partially achievable. The limitations of reactive transport simulation are embedded in the conceptualized set of equations used to best approximate the real situations. But models provide a tool for critical analysis. They are a means to organize our thinking, test ideas and indicate which the sensitive parameters are. They direct further studies and help to design new experiments and to critically test hypotheses. Particularly surprising model outputs often provide new insight otherwise inaccessible.

Studies of hydrothermal systems should begin by considering the following five questions:

(1) What is the architecture of the system?
(2) What is its geodynamic history?
(3) What processes are driving fluid flow on the scale of the system?
(4) What is the nature of the fluids in the hydrothermal system?
(5) What are the mechanisms of alteration?

Qualitative answers to these questions lead the way to conceptual models, which in turn provide a framework for quantitative, numerical models. Answering these questions as good as possible will help to conceptualise the scientific problem.

Key steps in conceptualising the models are:

(1) Simplifying assumptions must be made;
(2) Appropriate boundary conditions and initial conditions must be identified; and
(3) The range of applicability and limits of prediction of the models need to be understood.

Models for coupled simulation of reactive transport processes consist of five physical and chemical processes to be solved numerically:

(1) mechanical deformation;
(2) fluid flow;
(3) heat transfer;
(4) multi species transport; and
(5) chemical reactions.

5. HYDROGEOLOGY OF THE CENTRAL KENYAN RIFT

The hydrogeology of central to southern portion of the rift valley is mainly controlled by the rift flanks faults, the grid faulting and the tectono-axis along the rift floor (Clark et al., 1990). Fluids are recharged laterally from the high rift flanks and axially along the rift floor southwards. The grid faulting act as channels for ground water or they provide permeable barriers to lateral flow. A microseismic study has shown that the grid faulting unlike the escarpment faulting is quite active suggesting they are open (Aller et al., 1987). Thus the faulting causes the ground water to flow from the escarpment to the center and then follow longer flow paths reaching greater depths, and aligning their flow within the rift along its axis.

In Olkaria and Eburru where drilling has been carried out, the geothermal reservoirs are hosted by the faulted trachytes and basalts and from contacts between these lavas and pyroclasts, which are common within the floor of the southern Kenya Rift valley (Clarke et al., 1990). It is therefore probable that the reservoirs of Suswa and Longonot are hosted in the same formation. Groundwater flows are generally directed towards the lakes. Some flow away from the area to the North West, North East from the Mau Escarpment, South West from the Bahati escarpment and North Ward from Eburru. It is also probable that there is some southerly flow from Menengai towards Lake Nakuru (Becht et al., 2005). However at depth it is likely that flow occurs to the northeast away from the Nakuru-Elementaita catchment towards the lower lying catchment around Lake Bogoria (Clarke et al., 1990).

Tectonic movements of the rift valley have important effects on aquifer properties, both on a small scale by creating the local fracture system which comprise many aquifers. In such settings fractures provide conduits for fluid flow through such rocks. Therefore, the fracture system of the Kenyan rift, are important for successful operation of the geothermal reservoir. Groundwater flow to depths through faults and conduits where its heated by a heat source, later manifesting as High enthalpy steam vents, through hot and boiling springs Eburru and Olkaria. Scattered warm springs are also seen rising from underneath lavas and fissures along the foot of escarpments to the east of the rift valley axis.

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