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Evaluation of aquifers in an island situation: A case study of Mombasa, Kenya

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Abstract
Aquifers are dug very deep in many areas. When they are near the sea or an island the sea water tends to seep to the aquifers. This ultimately leads the aquifers to be salty and hence they don't achieve the objective of providing fresh water to the community. This study was conducted in 2020 with a name of coming up with preventive measures of the sea water seepage to the aquifers. This paper is only a review paper of how far the impact can be felt as the data is being analyzed.

Keywords: Aquifer, island, chloride levels

Introduction

Background of the Problem of Sea-water Intrusion

An aquifer is a geologic formation that stores and/or transmits water. An aquifer may be confined, semi-confined, perched or unconfined. A confined aquifer is bounded above and below by formations of impermeable or relatively impermeable material while unconfined aquifer is made up of loose material, such as sand or gravel that has not undergone settling, and is not confined on top by an impermeable layer (Utah, 2001) [10]. Semi-confined aquifers occur when water-bearing strata are confined, either above or below, by a semi permeable layer. When water is pumped from a semi-confined aquifer, it moves both horizontally within the aquifer and vertically through the semi permeable layer. Semi-confined aquifer is also called a leaky aquifer (Utah, 2001) [10]. An aquifer can therefore be defined as an underground bed or layer of earth, gravel, or porous stone that yields water, or formation or part of a formation capable of yielding a significant amount of groundwater to wells or springs.

The aquifer in Mombasa Island considered in this study is Semi-Confined also called Leaky aquifer. This aquifer is characterized by fractured limestone rock material which was formed by the direct lithification of coral reefs, marine organic shells, or marine organism skeletons. Limestone is a sedimentary rock formed from a chemical precipitate known as calcium carbonate while coral rock is a type of limestone rock which has not fully undergone lithification. Fractured limestone has a capability to allow both horizontal and vertical movement of water within and through an aquifer during pumping, thus a semi-confined or leaky aquifer (Caswell, 1953).

Aquifers typically consist of gravel, sand, sandstone, or fractured rock, like limestone or basalt. These materials are permeable because they have large connected spaces that allow water to flow through. Water in aquifers is brought to the surface naturally through a spring or can be discharged into lakes and streams. This water can also be extracted through a well or borehole drilled into the aquifer (Mogaka *et al*, 2004) [9].

Under natural conditions, groundwater is normally discharged into the ocean where the aquifers come into contact with the ocean due to the hydraulic gradient. With increased population and industrialization within the town precincts, groundwater demands in such coastal areas can also increase. With these increased demands the seaward flow of groundwater can be reduced or even reversed due to a landward hydraulic gradient, causing sea-water to penetrate inland in aquifers. This phenomenon is known as sea-water intrusion. Thus sea-water intrusion into groundwater formations is partly due to human activities such as abstraction and damage caused to natural barriers to sea-water intrusion. Groundwater supplies become less useful when the sea-water travels inland to well fields; moreover, the aquifer becomes contaminated with salts.

Protecting coastal aquifers from sea-water intrusion is important if future use of fresh groundwater is deemed necessary.

Mombasa Island has a high demand for groundwater due to frequent shortages of supply from Mombasa Water and Sewerage Company. The groundwater abstraction from the aquifer may exceed recharge and this has a serious consequence on future availability of groundwater and can also lead to land subsidence. Lowering of groundwater level due to large abstractions may motivate sea-water intrusion in the aquifer and hence lower the groundwater quality and usability for human consumption. Groundwater level fluctuations can result to reduction or reversal of hydraulic water gradients. There are many reasons which result to ground water level fluctuations and they include both natural and man made reasons. Sea-water intrusion into fresh groundwater formations results mostly due to fluctuations of groundwater levels which yields reversal of hydraulic gradients. It can also result inadvertently from excessive abstraction of groundwater by man which results to reduction in groundwater levels. A decrease in groundwater level due to excessive abstraction, causes permanent damage to the stability of the groundwater storage, and facilitates sea-water migration and the pollution of the aquifers with sea-water (Balek, 1983) [2].

Poor development of groundwater resources can lead to undesirable effects like; sea-water intrusion, lowering of water tables and reductions in well yields. Analyses of these problems are hampered by insufficient insight in the: Hydro-geological structures, groundwater flow systems and groundwater recharge.

It is thus essential to carry out hydrological study of sea-water intrusion and develop a model that represents sea-water intrusion into coastal aquifer. Thereby giving us the capability for simulating the behavior of the hydrologic and water quality components, that can be used for future planning and development.

Once an aquifer becomes contaminated with salt, it may take years to remove the salt even with adequate fresh water available to flush out the saline water. The determination of maximum and minimum water levels in order to regulate storage capacity is important. Reduction of water level to certain minimum values can lead to total reversal of hydraulic gradient resulting to sea-water intrusion which can make the whole aquifer contaminated with salt (Balek, 1983) [2]. This study aims to understand the dynamics of the groundwater aquifer in Mombasa Island and thus a more insight to hydro-geologic conditions which is important in regulating sea-water intrusion.

Ocean Salinity

There are many theories to explain the origin of ocean salinity (Barlow, 2003) [3]. Ocean salinity is due to presence of salts where sodium and chloride are the major components of salts in the ocean. The origins of salinity of sea-water may be due to the fact that salt and other minerals were carried into the sea by rivers, having been eroded out of the ground by rainfall runoff. Upon reaching the ocean, these salts would be retained and concentrated as the process of evaporation removes the water. In addition, sodium was leached out of the ocean floor when the oceans first formed. The presence of the other dominant element of salt, chloride, results from "outgassing" of chloride (as hydrochloric acid) with other gases from Earth's interior via volcanos and hydrothermal vents. The sodium and chloride subsequently became the most abundant constituents of salt in the sea (Barlow, 2003) [3].

Ocean salinity has been stable for millions of years, most likely as a consequence of a chemical/tectonic system which recycles the salt. Since the ocean's creation, sodium is no longer leached out of the ocean floor, but instead is captured in sedimentary layers covering the bed of the ocean. It is possible that plate tectonics result in salt being forced under the continental land masses, where it is again slowly leached to the surface (Barlow, 2003) [3].

Process of Sea-water Intrusion

Sea-water is not uniformly saline throughout the world. On average, sea-water in the world's oceans has a salinity of approximately 3.5%. This means that for every one liter of sea-water there are 35 grams of salts (mostly sodium chloride) dissolved in it. Water with this level of salinity is not potable. Sea-water has a density of about $1.025 g/cm^3$ while fresh groundwater has a density of approximately $1 g/cm^3$ (Utah, 2001) [10].

When groundwater is pumped from aquifers that are in hydraulic connection with the sea, the gradients that are set up may induce a flow of sea-water from the sea towards the aquifer. This migration of sea-water into freshwater aquifers under the influence of groundwater development is known as sea-water intrusion (Walton, 1970) [12].

The occurrence of sea-water intrusion is identified by increasing concentrations of sodium and chloride and by elevated specific conductivity and dissolved solids (Barlow, 2003) [3]. Typically, the concentration of chloride ions in water is used to identify seawater intrusion. Sea-water intrusion may occur in coastal areas, whereas up coning is the rise of freshwater sea-water interface below the well in response to draw down of the water table around the well, sea-water intrusion always increases the volume of sea-water stored underground and may finally lead to salinization of the entire aquifer (Barlow, 2003) [3].

Sea-water intrusion occurs through different ways, for example; whenever overdraft conditions occur in coastal aquifers connecting with the ocean, sea-water intrusion can result. By lowering the water table in unconfined aquifers, or the piezometric surface in confined aquifers, the natural gradient sloping downward toward the ocean is reduced or reversed resulting to sea-water intrusion. Because two fluids of different densities are involved, a boundary surface, or interface, is formed wherever the fluids are in contact. The more an interface moves landward, the more the sea-water intrusion. Fresh water has specific gravity of one and that of sea-water is 1.025 therefore at interface, you have a two layered system of fresh water floating on sea-water and the more the intrusion the more sea-water is pushed landward (Barlow, 2003) [3].

Up coning is the second mechanism through which sea-water intrusion occurs. When freshwater is underlain by saline water, pumping a well in the freshwater zone causes the freshwater sea-water interface to rise below the well in response to draw down of the water table around the well. Field studies have shown that if the bottom of the well is close to the saline water or the well discharge is relatively high, then the sea-water cone may reach into the well, causing the well discharge to be a mixture of fresh and saline groundwater (Barlow, 2003) [3].

The problem of sea-water intrusion has increased as population centers and concomitant water demands in localized coastal areas have developed. One of the earliest reports of intrusion was published in 1855 by Braithwaite in England (Walton, 1970) [12]. A study done by (Austre-mineral 1980) [1] in Tiwi and Ukunda in Coast Province in Kenya showed that there was a problem of sea-water intrusion.

Fresh Water on Oceanic Islands

Fresh groundwater is supplied entirely by rainfall hence only a limited amount is available. The close proximity of sea-water can introduce saline water into fresh groundwater even without overdraft unless care is exercised in developing underground water supplies (Kovar and Hrkal, 2002) [8]. This can occur for instance when drilling of wells exceeds saline-fresh groundwater boundary. An island well or borehole pumping water at a rate sufficient to lower the water table to sea-water level or below sea level disturbs the fresh salt-water equilibrium and can result to sea-water intrusion (Kovar and Hrkal, 2002) [8]. Thus, sea-water will rise as a cone to enter the well as illustrated in Figure 2.1. In practice a brackish mixture of the two waters would cause abandonment of the well before salt concentrations approaching sea-water are reached. To avoid this danger and to obtain the maximum groundwater yield, island wells should be designed for minimum draw down, just skimming fresh water from the top of the lens. Minimum drawdown can be achieved by calculating the maximum sustainable discharge from a well. The rise of sea-water cone below a well can be expressed according to (Bear and Dagan Fetter 1988) as:

$$Z = \frac{Q}{2\pi K' l} \quad \ldots\ldots 2.1$$

Where

Q is maximum sustainable discharge from a well

Z is rise of cone $\imath$ is depth of interface below the well bottom prior to pumping

$$K' = K \frac{\rho_s - \rho_f}{\rho_f}$$

Where

Q is maximum sustainable discharge from a well

Z is rise of cone

$\imath$ is depth of interface below the well bottom prior to pumping

$$K' = K \frac{\rho_s - \rho_f}{\rho_f}$$

Where

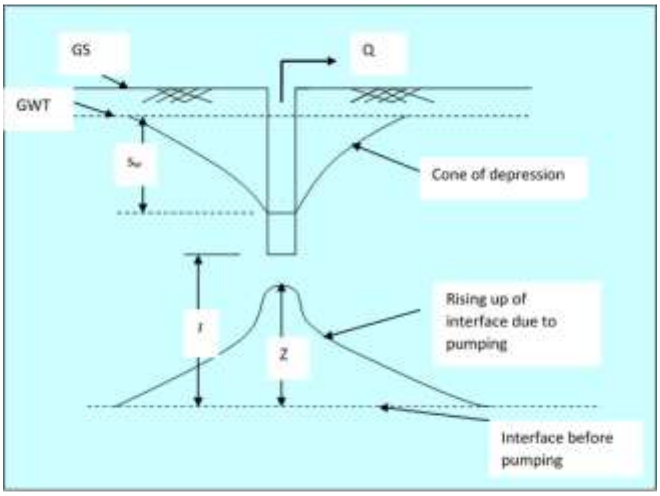
K is the permeability per day

ρ_s is density of sea-water

ρ_f is density of fresh water.

$$Z \leq \frac{1}{3} l \quad \ldots\ldots 2.2$$

In this study, the possibility of up coning was also analyzed. The aquifer within the Island is only seven meters below the sea level therefore up coning was not possible.



Source: Kovar and Hrkal, 2002)

Fig 1: Up coning of sea-water beneath a pumped well

Control of Sea-water Intrusion into Coastal Aquifers

Sea-water intrusion and land subsidence are induced by the decline of groundwater levels. The continuous decline in groundwater levels is caused by groundwater overexploitation. Therefore, in order to control further decline, groundwater abstractions should be reduced to maximum available yield. The strategy for controlling groundwater level decline is thus to determine the maximum groundwater abstraction rates under the constraint of no further decrease in groundwater levels.

There are several techniques for controlling sea-water intrusion, which include;Reduction or rearrangement of the pattern of groundwater pumping, for example, avoiding continuous pumping by pumping at certain times of the day only to allow water level to rise back to original level before further pumping, development of a fresh water ridge adjacent to the coast by means of a line of recharge wells parallel to the coastlines. In this case, some of the freshwater used to maintain the barrier is lost to the ocean but this amount is negligible compared to the fresh groundwater that would be lost by uncontrolled sea-water intrusion, artificial recharge of the intruded aquifer by use of recharge wells in order to raise the water level and avoid further intrusion, development of a pumping trough adjacent to the coast by means of a line of pumping wells parallel to the coastlines so as to lower the sea-water fresh-water interface and prevent further sea-water intrusion and construction of an artificial subsurface barrier to stop movement of sea-water towards the aquifer. (Barlow, 2003) [3].

Aspects of Groundwater Systems

Advantages of Groundwater

Groundwater is widely available therefore many semi-arid and arid parts can exploit groundwater where surface water is not available. This statement seems correct from Barlow, (2003) [3], but the exploitation of groundwater without proper planning can lower groundwater levels in coastal islands and result to sea-water intrusion. In this study, the model will focus on abstraction to see whether there is a problem of over exploitation, groundwater is available throughout the year. It is nearly a perfect storage medium. It can be used in drought conditions. This assures the population in Mombasa Island of constant supply of water but might pose a danger if the study of this storage medium and its sustainability is not put into consideration. This study analyses the storage medium of groundwater in Mombasa, thus the aquifer material and its characteristics together with its long-term sustainability, quality is superior, therefore, little or no sediments or suspended materials, smell, taste or color. Temperature is low and constant and is normally hygienically safe. This is a very good quality of groundwater but when groundwater gets contaminated with sea-water it can become very salty and its use becomes limited. Thus, this study is aimed at checking whether sea-water has contaminated the aquifer within Mombasa Island, treatment of groundwater is relatively simple. This can be challenged when groundwater is contaminated with sea-water and the treatment becomes very complex and uneconomical. (Barlow, 2003 and Kovar and Hrkal, 2002) [3].

Groundwater Modelling

Definition of Groundwater Model

A model is a tool that represents an approximation of a field situation. A mathematical model simulates groundwater flow indirectly by means of a governing equation thought to represent the physical processes that occur in the system, together with equations that describe heads or flows along the boundaries of the model (boundary conditions). Mathematical models can be solved analytically or numerically. Physical models such as laboratory sand tanks simulate groundwater flow directly (Van Der Gun, 1998) [11]. A numerical model in this study shall be used to investigate the possibility of sea-water intrusion in to Mombasa coastal aquifer. A numerical model has been used in this study to model both flow and chlorides. With numerical groundwater models, the groundwater flows and groundwater heads shall be computed at the cells of a model grid which covers the entire Mombasa Island. Cells of a model grid have been applied in computations of groundwater flows, groundwater heads and solute concentrations in modelling in this study because of their simplicity.

Definition of groundwater modelling

The proper way for determining the groundwater quality is by chemical analysis of water samples. However, other questions might arise such as understanding the different relations between the groundwater quality parameters and properties, prediction and simulation of future groundwater quality evolution, estimating the effects of groundwater flow and water quality management projects. Such problems cannot be solved by mere chemical analysis and it is here that models can have a significant and decisive role. Basically, groundwater modelling is a tool for analysing, extrapolating and predicting groundwater quality. Groundwater model may be considered as analysis instrument and is especially adapted to a certain application and a certain area and as such is unique (Van Der Gun, 1998) [11].

Processing MODFLOW, a ground water quality numerical model was used in this study

Importance of Groundwater Modelling

Models are important in the sense that they can be used in an interpretive manner to gain insight into the controlling parameters in a site setting or as a framework for assembling and organizing field data and formulating ideas about system dynamics. These ideas can be used in improving the performance of a system thus enhancing quality. Groundwater modelling is a vital tool in water management. It can be used for policy analysis, since very often the availability of groundwater is limited and there are many interfering or conflicting interests on groundwater use. The use of groundwater models may help to bring on board the effects of different scenarios and may guide the decision makers.

Groundwater modelling can assist the development of a fuller understanding of groundwater bodies and their interaction with the surface environment. Groundwater flow and transport modelling can play an important role in the development of the characterization of each groundwater body, through the bringing together of both quantitative data and qualitative information in a predictive framework. This framework allows the testing of the conceptual understanding of the groundwater system and the refinement of the conceptual model to maximize knowledge about the current state of the system and the possible future impacts of development (Van Der Gun, 1998) [11]. For example, a ground water model has been applied in areas adjacent to Australian shores to establish the extent to which equilibrium exists between groundwater and sea-water and the maximum abstraction for a given maximum permitted extend of saline intrusion. This maximum abstraction rate is a decision variable in the process of salinity control (Van Der Gun, 1998) [11]. Groundwater modelling aims at predicting the consequences of a proposed action therefore assisting in future planning, costing, and continuous improvement of a proposed action (Chiang and Kinzelbach, 2001) [7]. In this study, the groundwater model will be able to guide us in analyzing the possibility of sea-water intrusion.

Numerical Modeling

Numerical modelling is used to obtain an approximate solution when an analytical solution cannot be developed. Numerical solutions produce value at discrete points for one set of independent parameters, for example, they can compute the hydraulic heads and flows at discrete points in time and space. They are suitable for more complex systems, e.g. regional groundwater flow with changes of the properties of aquifers and semi-pervious layers, with wells, boreholes and rivers. Numerical modeling is the most appropriate in this study because the aquifer being studied is of irregular boundary and varying hydraulic heads therefore, cells of a model grid for example can compute different hydraulic heads at discrete points. There are three major types of numerical techniques: Groundwater models based on finite difference, groundwater models based on finite elements and boundary value. The two most widely used numerical techniques for solving mathematical models are finite difference and finite element models because they have the capability for the division of the model area into elements in common.

This finite difference numerical technique is the one to be applied in this study. MODFLOW is a well-known general finite difference groundwater flow model developed by the US Geological Survey as a modular and extensible simulation tool for modeling groundwater flow and is applied in this study. MODFLOW is a free software which 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, that is, flow and transport models, surface water and groundwater models and chemical reaction models, because of the simple, well documented nature of MODFLOW.

Boundary Conditions

In modeling it is important to determine the size and the boundary conditions of the model area so as to make sure that the effects of geo-hydrological measures in the area do not appear at the boundaries of the model for example abstractions should not appear at the boundaries as this will lead to wrong results.

There are three types of boundary conditions: The Dirichlet condition, prescribing a fixed head, the Neuman condition, prescribing a fixed flow, the Cauchy condition, prescribing a head dependent flow. Dirichlet condition occurs where constant-head boundary such as large lake or ocean represents an equipotential line and streamlines intersect at a right angle whereas Neuman condition may be encountered if the boundary is chosen at a groundwater divide. A cross- section perpendicular to the contour lines of the hydraulic head may also be considered as a fixed flow boundary. Boundary conditions in this study shall be specified as fixed head because the ocean surrounds the Island with a fixed hydraulic head of Zero. Cauchy condition normally deals with fluid flow problems where the boundaries are either impermeable, specified flux or specified values of hydraulic head.

Other models include; Analytic Element Method (AEM) and the Boundary Element Method (BEM), and 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, but cannot model solutes movement thus not applied in this study (Kovar and Hrkal, 2002) [8].

Mathematical Models

The tools available for the mathematical modelling of groundwater flow are:

I. Darcy’s law

Darcy's law states that;

$$Q_t = \frac{-KA_r}{\mu}\frac{(P_b - P_a}{L_p} \quad \text{……………………………… 2.3}$$

Where

Q_t is total discharge

K is the permeability

A_r is the cross sectional area

P_b is the pressure at point b

P_a is the pressure at point a

μ is the dynamic viscosity

L is the length the pressure drop is taking place

II. The mass balance equation.

The mass balance equation states that;

Inflow = outflow + change in storage.

$P – R + Q_{inf} + Q_{rech} + Q_{inn} = E + Q_{pump} + Q_{out} + S$……………………2.4

Where

P is precipitation

R is direct run off

Q_{inf} is infiltration from surface water

Q_{rech} is artificial recharge.

Q_{inn} is natural inflow

E is evapotranspiration

Q_{pump} is amount of water pumped out

Q_{out} is natural outflow.

S is change in storage, positive when storage increases.

By combining Darcy’s law with mass balance equation, the general groundwater differential equation can be generated.

Interpolation in Modelling

Gridding Methods

In MODFLOW the input data has to be interpolated in all grids either manually or using a computer program. For example if an area of hundred metres squared has three data points and grid are ten metres square, then each grid point has to have a value from the three data points interpolated well and assigned weights in reference to distance from the data points.

SURFER computer program was chosen in interpolating raw data to allow data to be input to MODFLOW in the right format

SURFER is a grid based contour program. Gridding is the process of using original data points (observations) in an XYZ data file to generate calculated data points on a regularly spaced grid. Interpolation schemes estimate the value of the surface at locations where no original data exists, based on the known data values (observations). SURFER then uses the grid to generate the contour map or surface plot (Kovar, 2002) [8].

The advantages of a grid based approach outweigh the disadvantages. Tasks such as drawing contour lines, volumetric calculations of map modifications are much faster with a grid based approach. Under most circumstances, there are few problems when using a grid file to produce a contour map versus using the original raw data to produce the contour map (Kovar, 2002) [8].

Most of the gridding methods in SURFER use a weighted average interpolation algorithm. This means that, with all other factors being equal, the closer a data point is to a grid node, the more weight it carries in determining the Z value at a particular grid node.

The gridding methods included in SURFER can be divided into two general categories: Exact Interpolators and Smoothing Interpolators (Kovar, 2002) [8].

Exact interpolators can honor data points exactly only when the data point falls directly on a grid node being interpolated. With weighted average interpolators this means that the coincident data point carries a weight of essentially one and all other data points carry a weight of essentially zero.

Smoothing interpolators or smoothing factors can be employed during gridding when one does not have strict confidence in his data measurements. Smoothing interpolators do not assign weights of one to any single data point, even when the data is exactly coincident with the grid node. This does not mean that the contour maps are not accurate representations of the data, only that smoothing interpolators modify the weighting factors in such a way that the surface is smoother; in other words, the weighting factors are spread out more evenly among the data points.

SURFER has a large list of gridding methods and options. Different gridding methods can have different results when interpreting data. The guidelines presented here were used as a first approach to deciding which gridding method was best to use in the collected data in this research.

The following list gives a quick overview of each gridding method and some advantages and disadvantages to selecting one method over another (Kovar, 2002) [8]. Inverse Distance to a power is fast but has the tendency to generate "bull's-eye" patterns of concentric contours around the data points. The Inverse Distance to a power gridding method is a weighted average interpolator. The Power parameter controls how the weighting factors drop off as distance from a grid node increases. For a larger power, closer data points are given a higher fraction of the overall weight; for a smaller power, the weights are more evenly distributed among the data points. The weight given to a particular data point when calculating a grid node is proportional to the inverse of the distance to the specified power of the observation from the grid node. When calculating a grid node, the assigned weights are fractions, and the sum of all the weights is equal to 1.0. Kriging is one of the more flexible methods and is useful for gridding almost any type of data set. With most data sets, Kriging with a linear variogram is quite effective. In general this is the method that is often recommended and has been applied in this study. It is discussed in detail in the next section.

Minimum Curvature generates smooth surfaces and is fast for most data sets. The interpolated surface generated by Minimum Curvature is analogous to a thin, linearly-elastic plate passing through each of the data values with a minimum amount of bending. Minimum Curvature generates the smoothest possible surface while attempting to honor all data as closely as possible. Minimum Curvature is not an exact interpolator however. This means that the data is not always honored exactly. Polynomial Regression processes the data so that underlying large scale trends and patterns are shown. This is used for trend surface analysis. Polynomial Regression is very fast for any amount of data, but local details in the data are lost in the generated grid. Polynomial Regression is not really an interpolator because it does not attempt to predict unknown values. Radial Basis Functions is quite flexible, and like Kriging, generates among the best overall interpretations of most data sets. This method produces results that are quite similar to Kriging. It has the ability to fit data and to produce a smooth surface. The functions to specify are analogous to variograms in Kriging.

The functions define the optimal set of weights to apply to the data points when interpolating a grid node. Shepard's Method is similar to Inverse Distance but does not tend to generate "bull's eye" patterns, especially when a Smoothing factor is used. Shepard's Method uses an inverse distance weighted least squares method. As such it is similar to the Inverse Distance to a Power interpolator but the use of local least squares eliminates or reduces the "bull's eye" appearance of the generated contours. Triangulation with Linear Interpolation is fast with moderately sized data sets. When you use small data sets triangulation generates distinct triangular facets between data points. One advantage of triangulation is that, with enough data, triangulation can preserve break lines defined in a data file. For example, if a fault is delimited by enough data points on both sides of the fault line, the grid generated by triangulation will show the discontinuity. The method works by creating triangles by drawing lines between data points. The original data points are connected in such a way that no triangle edges are intersected by other triangles. The result is a patchwork of triangular faces over the extent of the grid. Each triangle defines a plane over the grid nodes lying within the triangle, with the tilt and elevation of the triangle determined by the three original data points defining the triangle. All grid nodes within a given triangle are defined by the triangular surface. Because the original data points are used to define the triangles, all data is honored very closely.
(Kovar, 2002) [8].

Kriging Method

Kriging is a geostatistical gridding method which has been found to be very useful in many fields. Kriging has been applied in this study. It estimates from a user specified number of data values, the value of a model cell considering the interdependence expressed in a variogram. Kriging attempts to express trends that are suggested in any data, so that, for example, high points might be connected along a ridge, rather than isolated by bull's-eye type contours (Kovar, 2002) [8].

Variogram is used to determine the local neighborhood of observations used when interpolating each grid node, and how the weights are applied to the observations during the grid node calculation. A variogram is a plot of semi variance versus vector distance and it is normally used to estimate the distance over which measurements values are interdependent. The variogram is used to determine the local neighborhood of observations used when interpolating each grid node and the weights are applied to the observations during the grid node calculation. A sample data is used in the variogram analysis to derive a model to represent the population and the population model is then used in Kriging population (Kovar, 2002) [8].

Kriging method has several Variogram models in surfer and one is faced with an opportunity to choose which variogram model best suites his data. Mostly choosing the best variogram model is done by trial and error method. Available variogram models in SURFER ara stated below: Gaussian model, Hole Effect model, Linear model, Quadratic Model, Rational quadratic model and Spherical model

Linear Model

A linear model was chosen in this because kriging with a linear variogram is usually quite effective and simple. In this case the slope was given by the Scale/Radius, whereby, the scale is the semi variance and radius is the separation distance from one observation to the other. By allowing an anisotropic radius, it was possible to specify an anisotropic linear variogram slope.

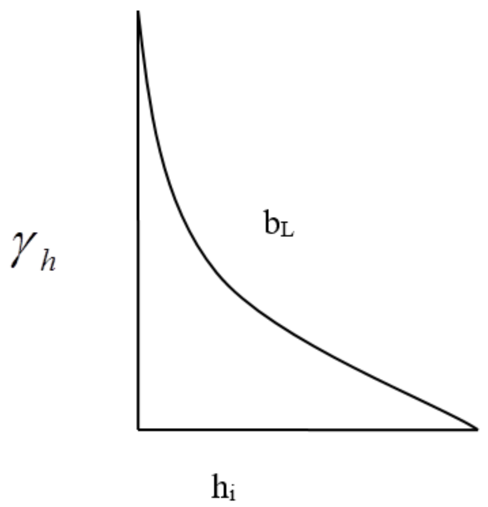
$$\gamma_h = b_L h_i \quad \text{..2.5.1}$$

And

$$b_L = {}^{\gamma_h}\!/_{h_i} \quad \text{..2.5.2}$$

Where
b_L is the slope of the variogram
h_i is the separation distance from one observation to the other.

γ_h Represents the variogram semi variance



Source: Kovar, 2002 [8].

Fig 2: Linear variogram

The above model was used only in interpolating collected data to be input in MODFLOW grid.

Mass Transport of Solutes
Processes of Mass Transport
There are three basic processes operating to transport solutes namely advection, diffusion and dispersion. Advection is the process by which moving groundwater carries with it dissolved solutes. As solutes are carried through porous media, the process of dispersion acts to dilute the solute and lower its concentration. These are discussed in detail in subsequent section. The process of seawater intrusion can be as a result of several processes. For example, after up coning, solutes can move by diffusion from the lower area of borehole which is highly concentrated with solutes to the upper part of the borehole with low solutes concentration. During diffusion, dispersion can as well take place and disperse the solutes laterally thus diluting them (Chapman, 1992) [4].

Diffusion
Diffusion is the process by which both ionic and molecular species dissolved in water move from areas of higher concentration (i.e., chemical activity) to areas of lower concentration. During sea-water intrusion there is transportation of solutes from the sea to the aquifer. In order to analyse seawater intrusion in Mombasa Island properly, it is essential for one to determine which process is mainly responsible for the transport of solutes from the sea to the aquifer.
The diffusion of a solute through water is described by Fick's law. Fick's first law describes the flux of a solute under steady state condition (Chapman, 1992) [4]:

$$F = -D\frac{dC}{dx} \quad \ldots\ldots 2.6$$

Where
F= mass flux of solute per unit area per unit time
D= diffusion coefficient (area/time)

$\frac{dC}{dx} =$

Concentration gradient (mass/ volume/distance)
For systems where the concentration may be changing with time, Fick's second law may be applied (Chapman, 1992) [4]:

$$\frac{dC}{dt} = D\frac{d^2C}{dx^2} \quad \ldots\ldots 2.7$$

Both Fick's first and second law as expressed above are for one dimensional situations. For three dimensional analyses, more general forms would be needed to include eddy diffusion which entails the presence or absence of turbulence. This depends on the *Reynolds Number*, which is a non-dimensional number which depends on velocity, width of the river or pipe, and the viscosity of the fluid.

Advection
Advection is a transport mechanism of a substance or a conserved property with a fluid in motion. The fluid motion in advection is described mathematically as a vector field, and the material transported is typically described as a scalar concentration of substance, which is contained in the fluid.

The rate of flowing groundwater can be determined from Darcy's law (Chapman, 1992) [4] as:

$$v_x = \frac{K}{n_e}\frac{dh}{dl} \quad \ldots\ldots 2.8$$

Where
v_x is Average linear velocity
K is Hydraulic conductivity
n_e Effective porosity
$\frac{dh}{dl}$ is Hydraulic gradient

Contaminants that are adverting are traveling at the same rate as the average linear velocity of the groundwater.

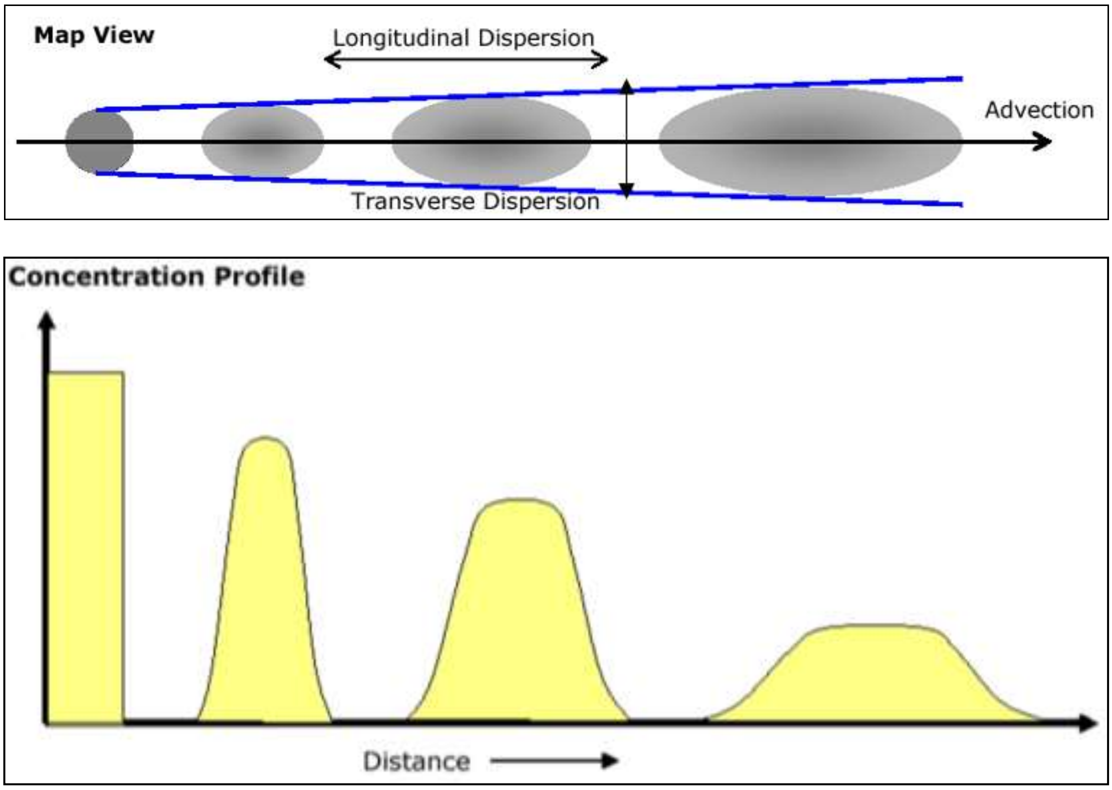


Fig 3: Explanation of pulse injection of a contaminant into groundwater

This can be explained from a pulse injection of a contaminant (Figure 2.3) into groundwater which yields a plume of solutes that migrate via advection both longitudinally and transversely. In the figure, the concentration of a contaminant is high at point of injection of the contaminant and covers a short stretch of a plume but the stretch of the plume increases with down flow of water whereas the concentration reduces with down flow of water.

Advection alone cannot explain the process of sea-water intrusion because it deals only with point source of contaminant.

Mechanical Dispersion

When a contaminant fluid flows through a porous medium, it mixes with non-contaminated water resulting to a dilution of the contaminant by a process known as dispersion. The mixing that occurs along the streamline of fluid flow is called longitudinal dispersion whereas dispersion that occurs normal to the pathway of fluid flow is lateral dispersion.

Mechanical dispersion is equal to the product of the average linear velocity and a factor called the dynamic dispersity (a_L) (Chapman, 1992) [4].

Mechanical dispersion=0020 $a_L v_x$ …………………………2.9

Mechanical dispersion is similar to turbulence, in that it is a result of variations in the movement of the water which carries a pollutant. In mechanical dispersion, these variations are the result of variations in the flow pathways taken by different fluid parcels that originate in the nearby locations near one another, and also variations in the speed at which fluid travels in different regions. This can be explained better from Figure 2.4.

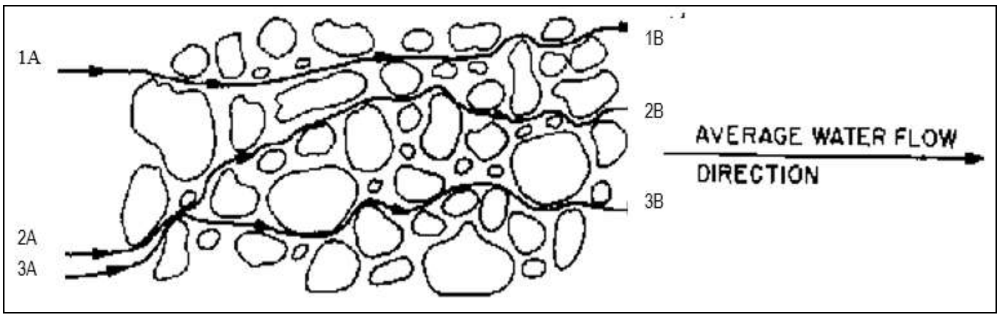


Fig 4: Process of mechanical dispersion

In Figure 2.4 two fluid parcels starting near each other at locations 2A and 3A are dispersed to locations farther apart 2B and 3B during transport through the soil pore, while parcels 1A and 2A are brought closer together to positions 1B and 2B, resulting in mixing of water from the two regions. Figure 2.5 shows the graphical presentation of the mechanical dispersion process.

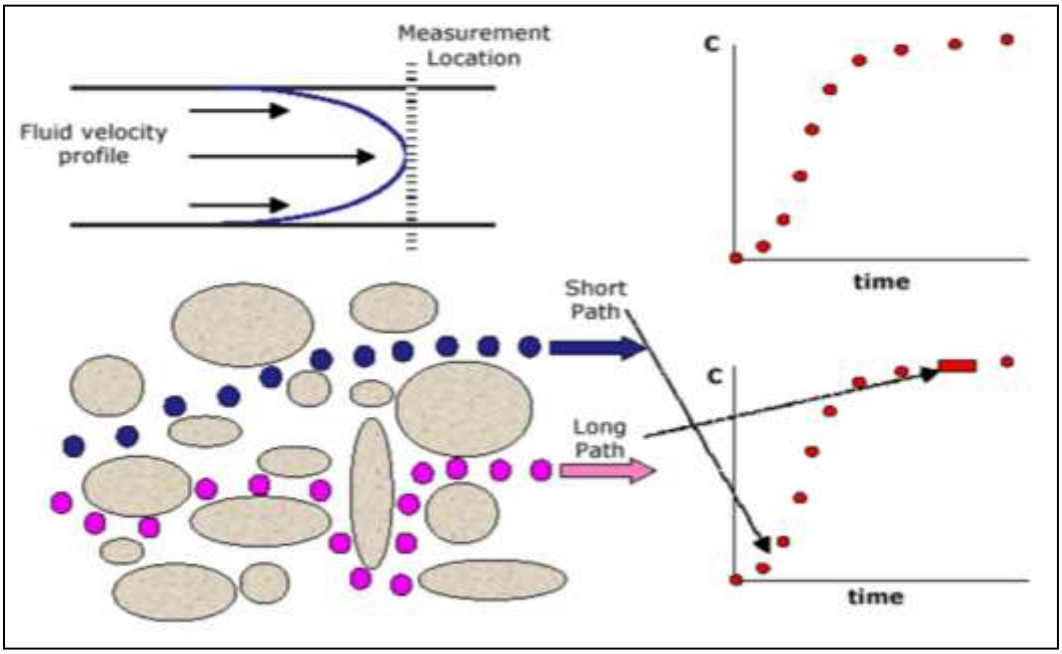


Fig 5: Graphical presentation of mechanical dispersion

Hydrodynamic Dispersion

Coefficient of dispersion is introduced to take into account both the Mechanical mixing and diffusion because the process of molecular diffusion and mechanical dispersity cannot be separated in flowing groundwater. For one dimensional flow it is represented by the following equation (Chapman, 1992) [4]:

$$D_L = a_L v_x + D^* \quad \text{...2.10}$$

Where

D_L = the longitudinal coefficient of dispersion

a_L = the dynamic dispersity

v_x = the average linear groundwater velocity

D^* = the molecular diffusion

The one dimensional equation for hydrodynamic dispersion is given by (Chapman, 1992) [4]:

$$D_L \frac{\partial^2 C}{\partial x^2} - v_x \frac{\partial C}{\partial x} = \frac{\partial C}{\partial t} \quad \text{..2.11}$$

Where

D_L is the longitudinal dispersion coefficient

C is solute concentration

v_x is the average groundwater velocity in the x-direction

t is time since start of solute invasion.

Mechanical dispersion causes mixing that occurs because the porous media forces some solute molecules to move faster than others while following a tortuous path through pores of different sizes. In Figure 2.6 and 2.7, the concentration profile increases when mechanical dispersion is combined with advection while it is lower when advection is combined with diffusion. This implies that there is higher concentration of solutes with time in mechanical dispersion.

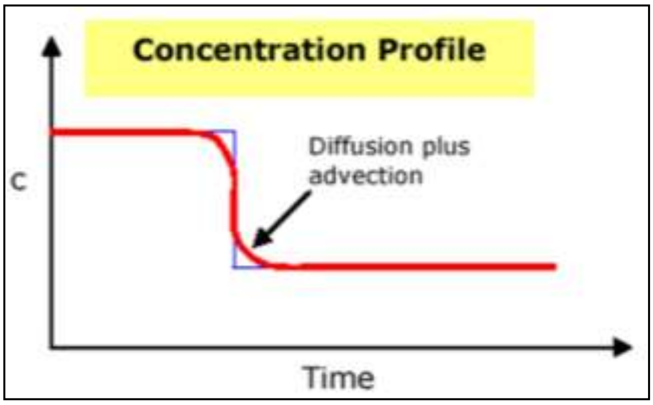


Fig 6: Diffusion plus advection

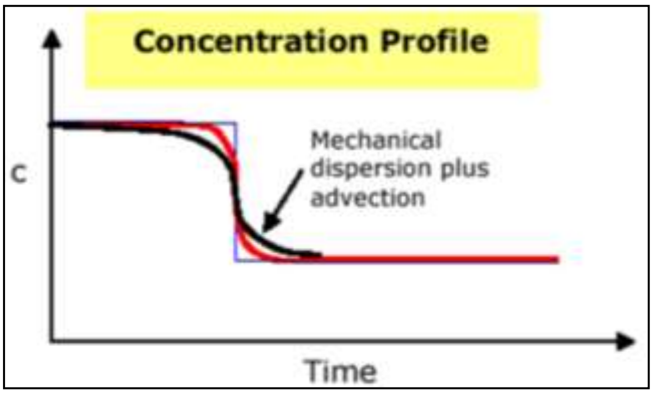


Fig 7: Mechanical dispersion plus advection

Key for Figure 2.6 and 2.7	
Mechanical dispersion plus advectio	
Diffusion plus advection	

Mass transport occurs during sea-water intrusion where by solutes are transported from the sea landwards.

The concentration, C, at some distance, L, from the source at concentration, C_0, at time, t, which is time of sampling, is given by the following expression, where erfc is the complementary error function (Chapman, 1992) [4].

$$C = \frac{C_o}{2}\left[erfc\left(\frac{L - v_x t}{2\sqrt{D_L t}}\right) + \exp\left(\frac{v_x L}{D_L}\right) erfc\left(\frac{L + v_x}{2\sqrt{D_L t}}\right)\right] \quad \text{……………………………2.12}$$

The complementary error function erfc in the x direction given by erfc (x) is defined as:

$$erfc(x) = \frac{2}{\sqrt{\pi}}\int_x^{\infty} e^{-t^2}\, dt = 1 - erf\,(x) \quad \text{…..……………..2.13}$$

Where

erf (x) is defined as

$$erf(x) = \frac{2}{\sqrt{\pi}}\int_0^x e^{-t^2} dt \quad \text{………………………… 2.14}$$

Error in the y direction was assumed to be zero due to the aquifer thickness which is so small relative to the study area. Equations 2.12 to equation 2.14 were used to check for error in the x direction. Importance of solute transport in groundwater is that geologic questions of ion migration and ore deposition are answered. Environmental problems such as contamination of drinking water by organic compounds and metals, radioactive waste disposal, saltwater intrusion can be prevented through the knowledge of solute transport.

Relative Importance of Advection versus Dispersion (Diffusion)

It is possible to evaluate the relative contribution of mechanical dispersion and diffusion relative to advection in mass transport in groundwater flow. A *Péclet number* is a dimensionless number that can relate the effectiveness of mass transport by advection to the effectiveness of mass transport by dispersion or diffusion. The Péclet number is given by:

$$p_e = \frac{v_x d_m}{D_d} \quad \text{……………….…………………………..2.15}$$

Where

v_x is the average linear groundwater velocity,

d_m is the mean grain size, and

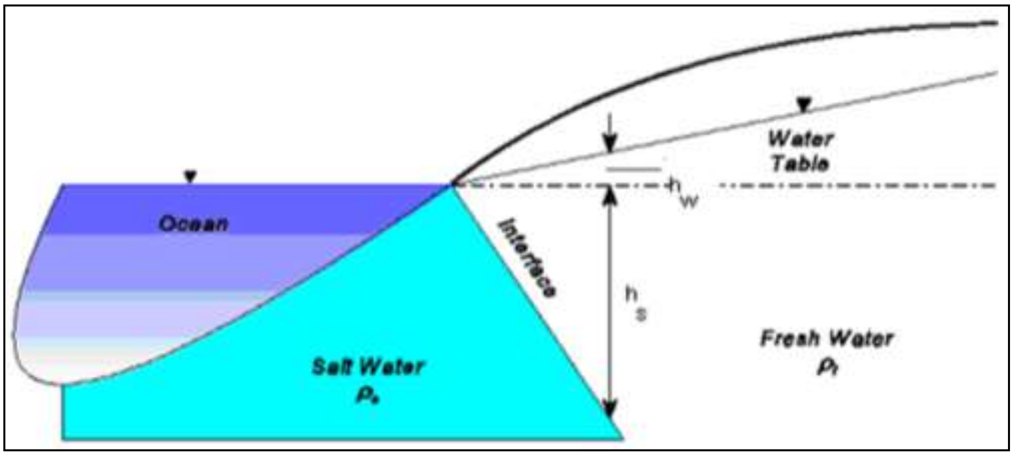
D_d is the effective diffusion coefficient.

At higher values of p_e, mechanical dispersion dominates, and the effect is proportional to velocity. Longitudinal dispersion dominates over transverse dispersion at values of p_e greater than one, which is reasonable, as the mechanical dispersion is mainly an effect along the direction of flow.

Sea-water Fresh Groundwater Interface

The Badon Ghyben / Herzberg Principle

Due to different concentrations of dissolved solids, the density of the saline water is greater than the density of fresh water. In aquifers hydraulically connected to the ocean, a significant density difference occurs which can discourage mixing of waters and result in an interface between fresh water and sea-water. The depth of this interface can be estimated by the Ghyben-Herzberg relationship (Figure 2.8)

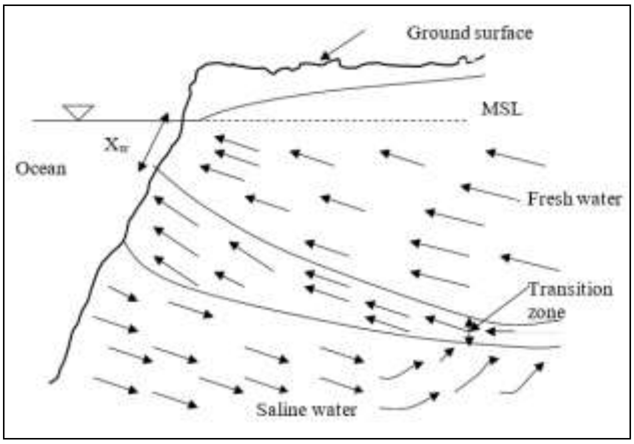


Source: Van der Eam; 1992

Fig 8: Representation of Hertzberg Principle at a shoreline

The Ghyben-Herzberg relationship assumes hydrostatic conditions in a homogeneous, unconfined aquifer. Additionally, it assumes a sharp interface between fresh water and salt water. In reality, there tends to be a mixing of salt water and fresh water in a zone of diffusion around the interface. If the aquifer is subject to hydraulic head fluctuations caused by tides, the zone of mixed water will be enlarged.

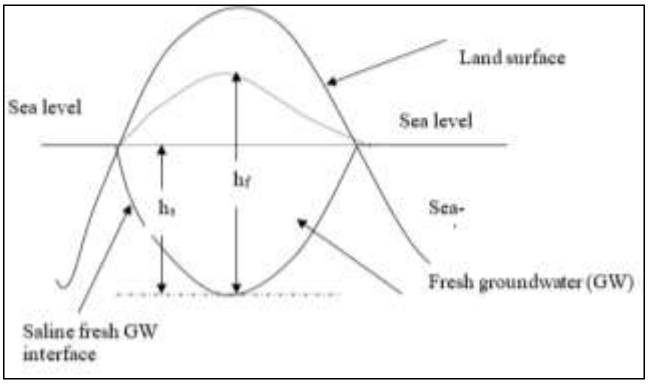
For visualization of the Island, Herzberg principle has been applied in this study. This principle explains that there is a fresh water lens in an oceanic Island under natural conditions. Mombasa Island cannot be an exception to this phenomenon even though it has a confined leaky aquifer. This principle was first formulated by (Badon Ghyben 1989), in the Netherlands and independently by (Herzberg 1990) in Germany (Van Der Gun, 1998) [11] Both of them concluded, from observations in boreholes that a lens of fresh groundwater, fed by natural recharge, was floating on heavier saline groundwater below it, in state of dynamic equilibrium with lateral discharge. The transition from fresh to saline groundwater was rather sharp, i.e. within 5 to 10 metres increase in depth and could be approximated by a sharp interface, though this is not usually the case as indicated in Figure 2.9 which shows that the interface is wide because two fluids mix within a distance not in a point. They found the principle that the depth of that interface below mean sea level was roughly 40 times the elevation of the phreatic groundwater table above mean sea level.



Source: Van der Eam; 1992

Fig 9: Flow patterns of fresh and saline water in a coastal aquifer

The principle can be understood theoretically as follows:



Source: Van der Eam; 1992

Fig 10: Representation of Hertzberg Principle in an Island

From Figure 2.10 the water pressure p in a point of the sharp interface can be expressed as;

$$\rho_f g h_f \text{ Or as } \rho_s g h_s \quad \text{……………………………….2.16.1}$$

Where;

ρ_f density of fresh groundwater

ρ_s density of saline groundwater

g acceleration due to gravity

h_f length of the column of freshwater

h_s length of the column of saline groundwater

The pressure p exerted at this point can be found from observed piezometric levels just above and below the sharp interface (in m)

$$p = \rho_f g h_f = \rho_s g h_s \quad \text{……………………………………..2.16.2}$$

Therefore

$$h_f = \frac{\rho_f}{\rho_s} h_s \quad \text{…………….……………………………..2.16.3}$$

This can further be written as:

$$h_f - h_s = \frac{\rho_s - \rho_f}{\rho_f} h_s = \Delta h_s \quad \text{………………………2.16.4}$$

Where

$$\Delta \quad \frac{\rho_s - \rho_f}{\rho_f} \quad \text{………………………………………..2.16.5}$$

ρ_f 1000kg/m^3

ρ_s 1025kg/m^3

Thus

$$h_f - h_s = \frac{(1025-1000)}{1000 h_s} = \frac{1}{40} h_s \quad \text{……………..……2.17}$$

The above principle was used to check for up coning which according to the depth of aquifer stated earlier could not be possible but it was deemed necessary to do a mathematical check for up coning.

Figure 2.9 shows the shape of a transition zone between fresh and saline water. Saline water in deeper sections of the sea normally flows landward whereas in shallow sections fresh water flows seawards. The gap X_w through which fresh water escapes to the ocean is approximated by:

$$X_w = \frac{q}{2K'} \quad \text{……………………………………………2.18}$$

Where;

q is freshwater flow into the sea in m^3/day,

$$K' = K \frac{\rho_s - \rho_f}{\rho_f} \quad \text{……………………………………..2.19}$$

Where:

K is the permeability per day

ρ_s is density of sea-water

ρ_f is density of fresh water.

Figure 2.9 is a good illustration of the flow of water in an aquifer in an Island and Mombasa Island is not an exception. Within this transition zone, salinity of ground water increases progressively with depth from that of fresh water to that of saline water. This principle of seawater-fresh water interface was applied in this study to allow for good modeling of solutes movement from the sea to the aquifer.

Conclusion

The purpose of calibration is to establish that the model can reproduce field-measured heads, flows and solute concentrations. During calibration a set of values for aquifer parameters and stresses is found that approximates field-measured heads and flows. Calibration is done by trial-and-error adjustment of parameters or by using an automated parameter estimation code. Calibration in this model was done by using estimated set of data based on aquifer of uniform thickness of eight meters within an ideal square Island. This data was estimated because there was no other available old data within the Mombasa Island on chemical concentrations in the boreholes.

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