# Groundwater Recharge Simulator

**M. Tech. Thesis**

by

**Dharmvir Kumar**

**Roll No: 07305902**

**Guide: Prof. Milind Sohoni**

**Co-guide: Prof. T. I. Eldho**

**Department of Computer Science and Engineering**

**Indian Institute of Technology, Bombay**

**Powai, Mumbai-400076**

June 19, 2009

Acknowledgment

I would like to express sincere gratitude and thanks to my guide, Prof. Milind Sohoni. I am grateful to him for the opportunity given me to explore this real world application. He was always approachable and spent ample time with me and maintained enthusiasm and interest to explore this problem domain. I also thank Prof. T. I. Eldho from Civil Engineering Department for his keen interest and invaluable suggestions and support. His guidance is actually proving very useful in understanding of different aspects of problem domain. I would also like to thank Prof. Avijit Chatterjee from Aerospace Engg. Dept. for his interest and suggestions. I would also like to thank Prof. Haresh Patel from M. S. University, Baroda for his valuable inputs.

Dharmvir Kumar

# Contents

# List of Figures

# List of Tables

# Chapter 1

# Introduction

Groundwater is water located beneath the ground surface in soil pore spaces and in the fractures of geological formations. Ground water is an important resource to meet our needs. It allows people to live in places where surface water is scarce. It is the major source of drinking water in both urban and rural area. Besides, it is an important source of water for the agricultural and the industrial sector. The bottled water so many of us drink is actually groundwater. The demand for water has increased over the years and this has led to water scarcity in many parts of the world. The number of wells drilled for irrigation of both food and cash crops have rapidly and indiscriminately increased. Countries like India's rapidly rising population and changing lifestyles has also increased the domestic need for water. The water requirement for the industry also shows an overall increase. Intense competition among users such as agriculture, industry, and domestic sectors is driving the groundwater table lower. Because groundwater is a hidden resource, both its availability and purity are usually taken for granted. Being an important and integral part of the hydrologic cycle, its availability depends on the rainfall and recharge conditions. An uncontrolled use of the borewell technology has led to the extraction of groundwater at such a high rate that often recharge is not sufficient. We can run out of groundwater if more water is withdrawn by pumping (discharge) than is fed by recharge from the surface. This is a problem considering that in many areas groundwater is being withdrawn in increasing quantities to meet growing demands and development pressures. The human demands compete with natural uses. Unsustainable groundwater extraction won't only mean less water for human use; it will also have an impact on the environment.

Hence, it is very necessary to manage groundwater resources intelligently and effectively. For this purpose, we need to understand how these systems operate, and establish a sustainable groundwater yield for each system. Overall groundwater is an integral part of the environment, and hence cannot be looked upon in isolation. In the process of groundwater resource management, any water harvesting system has an important role to maintain water table. Typical groundwater resource management system includes:

- Ground Improvement
- Watershed Management
- Contour Bunding

- Check Dam
- Percolation Tanks

These type of system should be simulated before implementation so that degree of its success could be measured.

## Check Dam as Harvesting System

We all know, precipitation is the principle source of replenishment of moisture in the soil through the infiltration process and subsequent recharge to the groundwater through deeper percolation. The amount of infiltrated moisture that will eventually reach the water table is accounted as the natural groundwater recharge. The natural recharge depends on intensity, duration, amount of rainfall, infiltration capacity of the topsoil zone, antecedent soil moisture conditions and water-table depth. The natural recharge takes place in pulses in semi-arid regions as the area experiences monsoon rainfall in pulses. Percolation ponds are water harvesting structures constructed across water courses of very small watershed area to impound rain water and to retain it for a longer time to increase the opportunity time for infiltration. In the harvesting condition, water is continuously available till the pond becomes dry, making infiltration and percolation active till the availability of water in the pond. It is therefore expected that the responses of the water table in pond and non-pond areas are to be different and the comparative analysis will yield accrued recharge due to recharge structure. This type of ponds are cost effective solutions for water problems in rural areas where discharge is far more than recharge. Wells, and pumping are on increase for all type of water requirements like irrigation etc in these rural areas. Hence, uncalculated extractions and no replenishment of groundwater may lead to a deeper water table in such areas. So, one of the cost effective solution is construction of percolation tanks in such areas. These percolation tank may serve some of water requirements of such areas, and also replenishment of groundwater level.

At first sight percolation tanks or check dams, as a solution for water problems in rural areas looks very effective. But success stories of already made tanks is not upto expectation. One of the major reason behind this is the lack of study of simulated results before making a pond in a vicinity. This type of study is considered costly because these construction projects are smaller in size and budget too. Our aim is to develop a software tool which provides cost effective and more accurate simulations for varoius type of structures. Although a lot of software tools are available but niether they are easy to use to use nor cost effective. Even these tool have not sufficient features to design a simulation model for percolation tank or reservoir in a given regime In this report, we have explored one widely used software tool, GMS which has good graphical user interface and costs around $5000 USD. We found that a lot of features necessary to simulate reservoir/dam structures do not work as desired in GMS. We will discuss these later.

This present study is intended to understand the domain of groundwater recharge-discharge problem towards development of a software utility tool which is easy to use, cost effective and gives more accurate simulation based results, mainly for recharge purposes. As this software will be open source, any utility as per requirement could be added in future which is not being possible with GMS like softwares. Although development of this proposed tool is a long term goal and may take years to evolve, this study provides a protocol for groundwater modelling and as a result of many simulation efforts on existing software. It also explores main issues to be referred in proposed tool.

### Objective of the Study

1. **to understand the a typical GUI layer over MODFLOW (in our case, GMS), its features and its limitations, for a typical ground water scenarios.**

2. **to extract specifications and develop a test suite which will drive the development of an independent low-cost public domain simulator**

Groundwater is water located beneath the ground surface and certain hydraulic principles govern the movement of groundwater beneath the ground surface. This chapter explains these principles and terminology from groundwater hydrology. These terminologies are used throughout all chapters. Using these principles we will get a groundwater water flow equation which is basically a partial differential equation. This chapter also explains finite difference approach of solving groundwater flow equation. Later part of this chapter explains features and working of a software tool, MODFLOW, which is basically a solver for groundwater flow equation.

## 1.1 Groundwater Theory

### 1.1.1 Darcy's Law

Groundwater theory starts with discussion of the basic law by French hydraulic engineer, Henry Darcy, which governs the groundwater flow [11]. In his report on the water supply of the city Dijon, France, he described an experiment to analyze the flow of water. The result of his experiment is generalized as an empirical law known as Darcy's law [11].

Consider a lab set up as shown in figure 1.1. A circular cylinder of cross-section $A$ filled with a porous medium(e.g. sand), stoppered at each end and outfitted with inflow and outflow tubes and a pair of manometers. Water is introduced such that inflow rate $Q$ is equal to the outflow rate. We set an arbitrary datum at elevation $z = 0$. All elevation and distances described in figure 1.1 are observed.

We will define a term $v$ as *specific discharge* through the cylinder as

$$v = \frac{Q}{A} \tag{1.1}$$

This experiment of Darcy showed

$$v \propto \Delta h$$

and

$$v \propto \frac{1}{\Delta l}$$

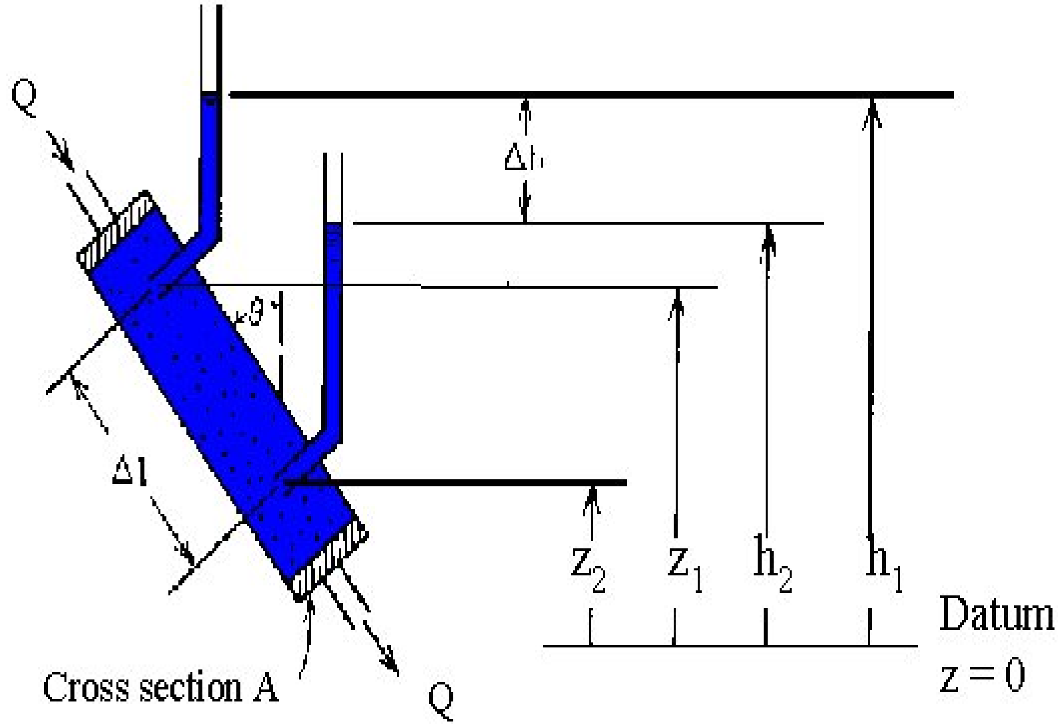



Figure 1.1: Experimental apparatus and measurements of Darcy [6]

Now, Darcy's law can be written as

$$v = -K\frac{\Delta h}{\Delta l} \tag{1.2}$$

$h$ is called *hydraulic head* and $\frac{\Delta h}{\Delta l}$ is the *hydraulic gradient.* K is the constant of proportionality , known as, *hydraulic conductivity* of porous medium. Now let us discuss these terms and their significance in groundwater flow one by one.

**Hydraulic head and gradient**

Every physical process which involves flow usually requires the recognition of potential and its flow occurs in the direction of higher to lower potential. It seems true also in case of groundwater flow. Our task is to determine the physical quantity involved in groundwater flow system which serves as potential in general. Hubbert(1940) clarifies the concept of groundwater potential and its relationship with Darcy's head by deriving it from basic physical principles. Groundwater potential at a given point is the energy required to transport a unit mass of water from a standard reference state to that point. Using this concept, he calculates that the fluid potential $\phi$ at any point $P$ in a porous medium is simply the hydraulic head at that point multiplied by $g$(almost constant). It concludes that potential term in the groundwater flow system is basically head term in Darcy experiment. In the process of this derivation it exposes that Hydraulic head term has two components elevation head $z$, and pressure head $\psi$. Figure 1.2 displays the

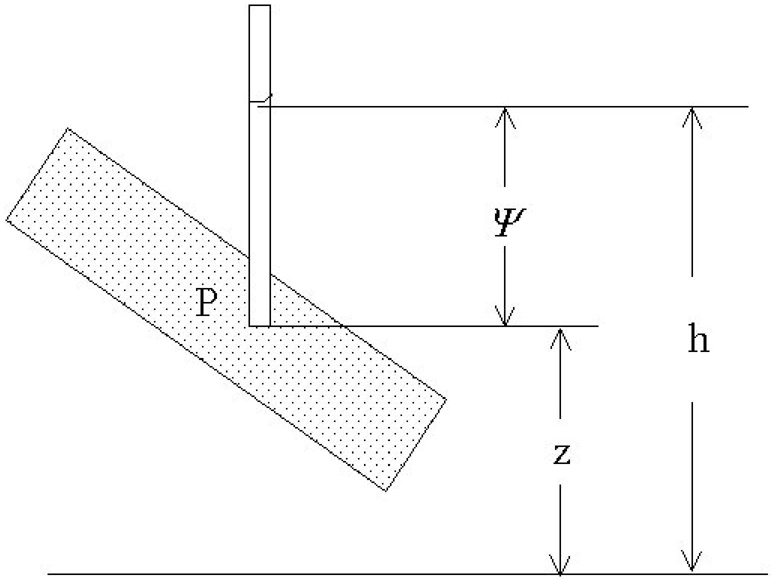



Figure 1.2: Hydraulic head $h$, pressure head $\psi$, elevation head $z$ in Darcy experiment [4]

relationship for all these terms in Darcy experiment. Now, hydraulic head

$$h = z + \psi \tag{1.3}$$

As a result of this discussion groundwater flows from higher $h$ to lower $h$ values in 3-D space. Piezometer is a device using which we measure head values at any point in system. If a large number of piezometers could be distributed throughout three dimensional hydrological system, it would be possible to contour the position of equal hydraulic head. In 3-D the locus of such points forms *equipotential surfaces.* In any two-dimensional cross section, the traces of equipotential surfaces on the section are called equipotential lines. If pattern of heads is known in a cross section, *flowlines* can be constructed perpendicular to the equipotential lines (in the direction of maximum potential gradient). The resulting diagram of equipotential lines and flowlines is called *flow net.* Figure 1.3 shows an example of flow net diagram.

**Hydraulic Conductivity**

As described earlier in this section, constant of proportionality involved in Darcy's law is hydraulic conductivity. It depends on the size and arrangements of porous medium and on dynamic characteristics of fluid such as dynamic viscosity, density, and strength of gravitational Field. Hence, hydraulic conductivity term is a function of not only the porous medium but also of fluid. We express hydraulic conductivity ($K$) term as

$$K = \frac{Cd^2\rho g}{\mu} \tag{1.4}$$

Where
$C$ is to include medium properties that affect flow, apart from grain diameter ( for example distribution of grain sizes, roundness, nature of packing );

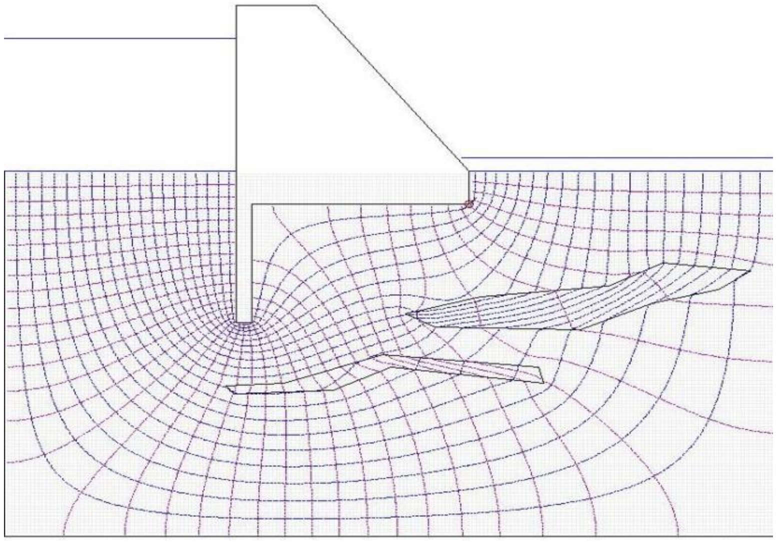


Figure 1.3: An example of flow net [1]

$d$ is diameter of grains of porous medium;
$\rho$ is fluid density;
$\mu$ is dynamic viscosity of fluid;

Sometimes hydraulic conductivity term is defined as

$$K = \frac{k\rho g}{\mu} \tag{1.5}$$

where $k = Cd^2$ is known as specific permeability of porous media.
Hydraulic conductivity values show variations through space within geologic formation. If hydraulic conductivity is essentially the same in any area, the area is said to be *homogeneous*. If it differs from one part of the area to another, then said to be *heterogeneous*. Hydraulic conductivity may also differ in different directions at any place in an area. If the hydraulic conductivity is essentially the same in all directions, it is called *isotropic* medium. If it is different in different directions, it is called *anisotropic* .
Darcy's law is valid for groundwater flow in any direction in space. For three-dimensional flow, in a medium that may be anisotropic, it is necessary to generalize the one-dimensional form of Darcy law. The simplest generalization would be

$$v_x = -K_x \frac{\partial h}{\partial x}$$

$$v_y = -K_y \frac{\partial h}{\partial y}$$

$$v_z = -K_z \frac{\partial h}{\partial z}$$

### 1.1.2 Unsaturated Flow and the Water Table

Most people know that the water table has something to do with ground water. The word table provides an image of a flat surface, like a tabletop, and it is commonly assumed that when a well is drilled it strikes water once it reaches below the water table. Underground water occurs in two different zones. One zone, which occurs immediately below the land surface in most areas, contains both air and water is known as unsaturated zone. Below this unsaturated zone, there is a zone in which all interconnected openings are full of water. This zone is known as saturated zone. We usually think of water table as boundary between these two zones.

The *water table* is known as the surface on which the fluid pressure $p$ in the pores of porous medium is exactly atmospheric. This surface reveals when we dig a well up to a depth just enough to encounter standing water. If $p$ is measured then on the water table $p = 0$. This implies $\psi = 0$, and then the hydraulic head $h$ equals to the elevation $z$ of the water table at that point. In the terms of pressure head component $\psi$ of hydraulic head $h$, saturated zone is defined as where $\psi > 0$, whereas unsaturated zone as where $\psi < 0$.

Unsaturated and saturated zone can also be explained in terms of moisture content and porosity terminology. If the total unit volume $V_t$ of a soil is divided into the volume of the solid portion $V_s$ and the volume of the voids $V_v$, then *porosity* $n$ is $V_v/V_t$. Similarly, if the volume of water content in the soil is $V_w$ then *moisture content* $\theta$ is defined as $\theta = V_w/V_t$. Now we can say that $\theta = n$ for saturated zone, and $\theta < n$ for unsaturated zone. One more distinction between saturated and unsaturated zones is in term of hydraulic conductivity. The hydraulic conductivity $K$ is constant; it is not a function of the pressure head $\psi$ in saturated zones. But in case of unsaturated zones, both hydraulic conductivity $K$ and the moisture content $\theta$ are functions of the pressure head $\psi$.

Both the pressure head and the hydraulic conductivity are nonlinear functions of the moisture content [2]. The pressure head becomes increasingly negative as the moisture content decreases as shown in Figure 1.4. Note that the sandy soil (with larger pore spaces) has a lower moisture content for a given pressure head then clayey soil. At low moisture contents, only the smallest pore spaces, with the highest surface tension, will hold onto water. At higher moisture contents, larger pores, with lower surface tensions, will fill. The shape of the relationship between pressure head and water content depends on the pore size distribution. Furthermore, the shape of the curve actually depends on whether the soil is wetting or drying out. This occurs because drainage is controlled by the size of the smallest pore spaces, whereas wetting is controlled by the size of the largest pore spaces.

As moisture content decreases, hydraulic conductivity decreases. This has been determined by flow measurements at varying moisture contents. It occurs because the water fills only part of the pore space (with air filling the rest) and so effectively the flow takes place through smaller, more poorly-connected channels. As the moisture content approaches the porosity, the hydraulic conductivity will approach the saturated hydraulic conductivity.

Unsaturated zone flow equations are developed similarly to those for the saturated zone, using Darcy's law and the conservation of mass. Solution of the flow equations is more complex than for the saturated zone because of the dependence of pressure head and hydraulic conductivity on the moisture content.

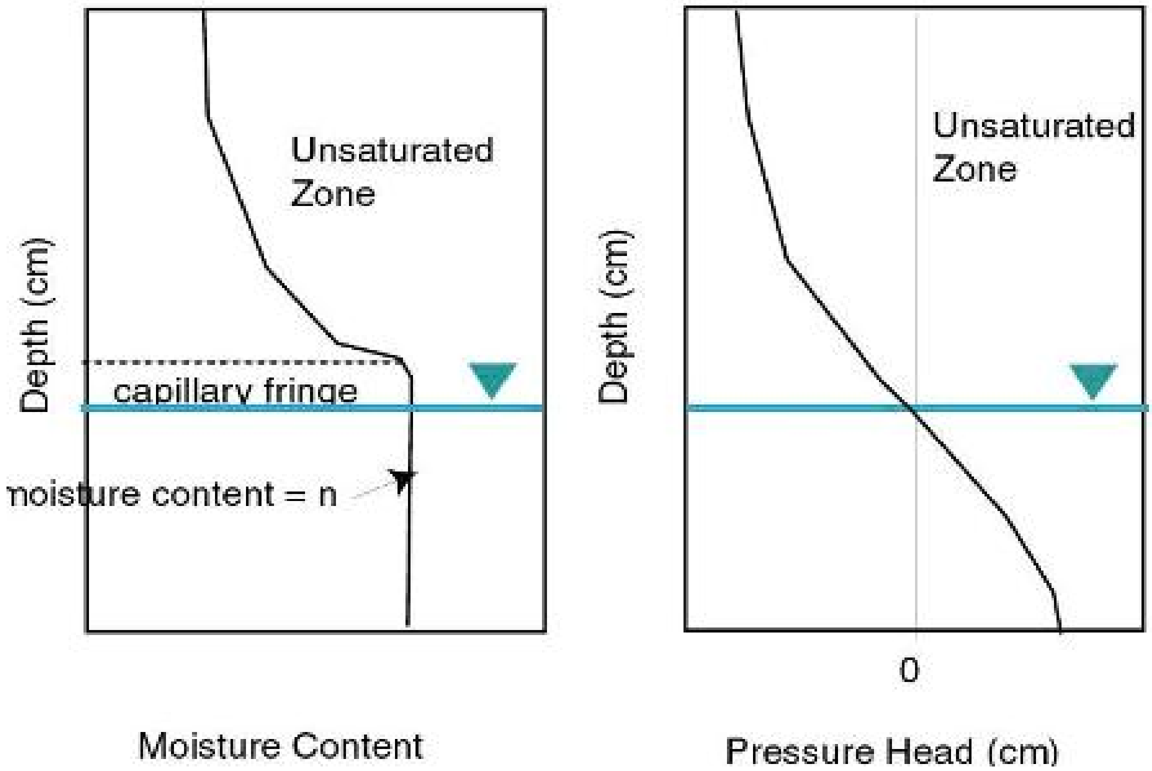



Figure 1.4: Moisture content and pressure head [2]

### 1.1.3 Some more Terminology

In this section, we will list some important terms from hydrologic vocabulary.

#### Aquifers, Aquitards, and Aquicludes

An aquifer is defined as a saturated permeable geologic unit that can transmit significant quantities of water under ordinary hydraulic gradients. An aquiclude is incapable of of this type of transmission. In other word, an aquifer is sufficient to allow the completion of production wells. A middle classification between aquifer and aquicludes is aquitards which is permeable enough to transmit water but do not allow completion of production wells.

#### Confined and Unconfined Aquifers

A confined aquifer is an aquifer between two aquitards. An unconfined aquifer is an aquifer in which the water table forms the upper boundary. Confined aquifers occurs at depth, unconfined at near the ground surface. If the water level elevation in wells tapping a confined aquifer are plotted on a map and contoured, the resulting surface, which is actually a map of the hydraulic head in the aquifer, is called a *potentiometric surface.* as shown in Figure 1.5. In genearl, we use unconfined aquifers for the purpose of extraction using wells.

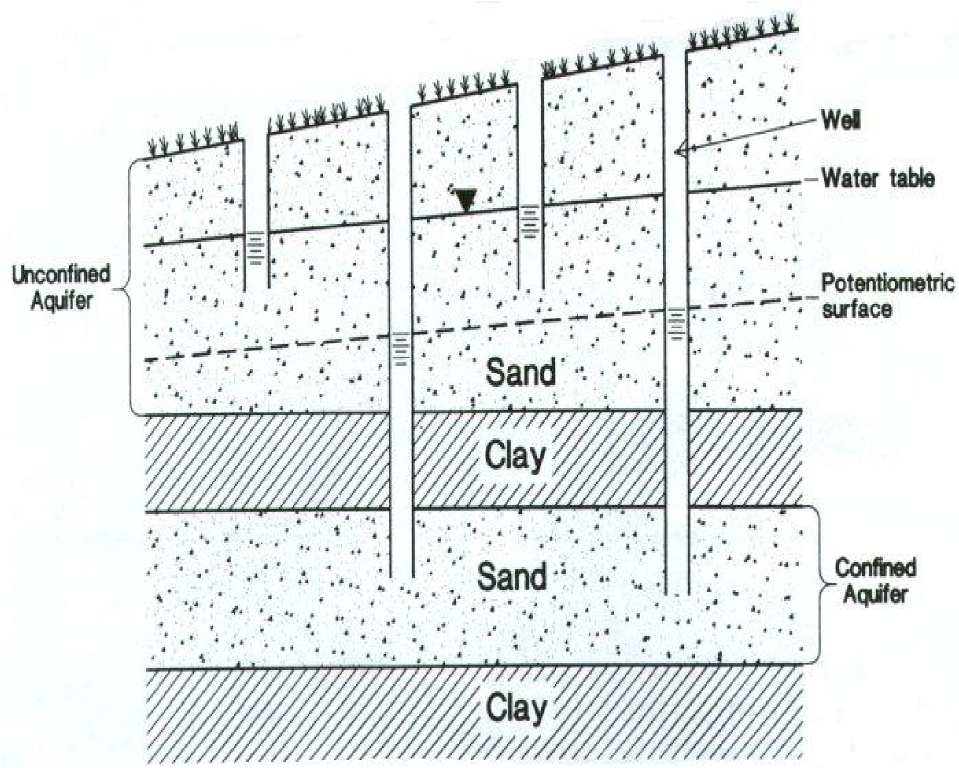



Figure 1.5: Confined and unconfined aquifer and potentiometric head [3]

**Steady-State flow and Transient Flow**

Steady-state flow occurs when at any point in a flow field the magnitude and direction of the flow velocity is constant with time whereas in transient flow it changes with time. One major difference between steady and transient system lies the relationship between their flowlines and pathlines. Flowline must be orthogonal to the equipotential lines throughout the region of flow at all times. Pathlines map the route that an individual particle of water follows through a region. In steady state pathline will be same as flow line, but in case of transient sate pathline will follow flow line which is changing; as a result flow line and pathline do not coincide.

**Specific Storage**

The specific storage is the amount of water that a portion of an aquifer releases from storage, per unit mass or volume of aquifer, per unit change in hydraulic head, while remaining fully saturated. According to mechanism worked under condition of decreasing head the specific storage $S_s$ is expressed as

$$S_s = \rho g(\alpha + n\beta) \tag{1.6}$$

where $\alpha$ is fluid compressibility, $\beta$ is a fluid compressibility, and $\rho$ is fluid density.

**Transmissivity and Storativity of Confined Aquifer**

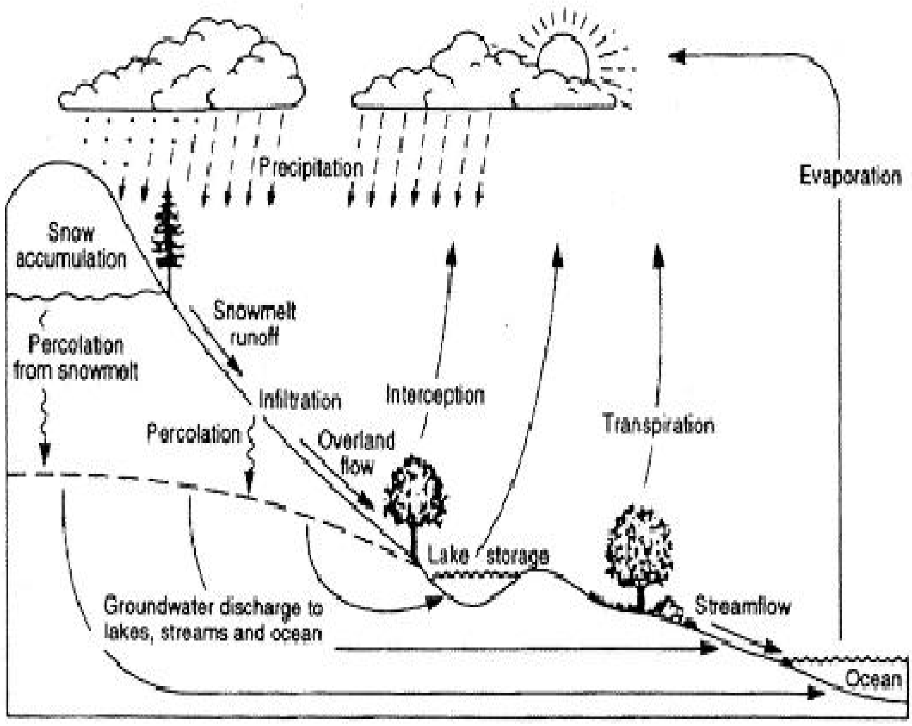



Figure 1.6: Hydrologic cycle [5]

For a confined aquifer of thickness b, the transmissibility $T$ is defined as

$$T = Kb \tag{1.7}$$

and the Storativity $S$ is defined as

$$S = S_s b \tag{1.8}$$

**Specific Yield in Unconfined Aquifer**

The storage term for unconfined aquifer is called as specific yield. It is defined as the volume of water that an unconfined aquifer releases from storage per unit surface area of aquifer per unit decline in the water table. Porosity of an aquifer is equal to sum of specific retention and specific yields of the aquifer. Specific retention is the amount of water retained by the sample. Specific yield value ranges from 0.05 percent to 35 percent.

### 1.1.4 Hydrologic Cycle

The term hydrologic cycle refers to the constant movement of water table above, on, and below the Earth's Surface as shown in Figure 1.6. The concept of hydrological cycle is very important to understand of the occurrence of water and the development and management of water resources. Although the hydrologic cycle has nither a start nor an end, we start discussion with evaporation from vegetation, from exposed moist surfaces including land, and ocean. This moisture forms clouds which returns the water to the land surface, in form of precipitation. The first rain wets vegetation and other surfaces and then begins to infiltrate into the ground. Infiltration rates vary widely, depending on land use, the character and moisture content of soil and the intensity and duration of precipitation. When the rate of precipitation exceeds the rates of infiltration, overland flow occurs. The first infiltration replaces soil moisture, and thereafter the excess percolates slowly across the intermediate zone to the zone of saturation. Water in the zone of saturation moves downward and laterally to the sites of ground water discharge such as springs, seeps in the bottoms of streams and lakes or beneath oceans

### 1.1.5 Groundwater Flow Equation

Groundwater flow can also be described mathematically. When we put Darcy's law with an equation of continuity that describes the conservation of fluid mass during flow through a porous medium, a partial differential equation gets derived. In this section we will describe these equations for steady and transient state for saturated flow.

**Steady State Saturated Flow**

Consider a unit volume of porous media. The law of conservation of mass for steady state flow through a saturated medium requires that the rate of fluid mass flow into unit volume be equal to the rate of mass flow out of the unit volume. This sentence can be coded into mathematical equation as [11]

$$-\frac{\partial(\rho v_x)}{\partial x} - \frac{\partial(\rho v_y)}{\partial y} - \frac{\partial(\rho v_z)}{\partial z} = 0 \tag{1.9}$$

If the fluid is incompressible then $\rho$ is a constant, in that too differential term containing $\rho$ is negligible. Hence equation may be simplified to

$$-\frac{\partial v_x}{\partial x} - \frac{\partial v_y}{\partial y} - \frac{\partial v_z}{\partial z} = 0 \tag{1.10}$$

and after applying Darcy's law, we get below equation for anisotropic medium.

$$\frac{\partial}{\partial x}\left[K_x \frac{\partial h}{\partial x}\right] + \frac{\partial}{\partial y}\left[K_y \frac{\partial h}{\partial y}\right] + \frac{\partial}{\partial z}\left[K_z \frac{\partial h}{\partial z}\right] = 0 \tag{1.11}$$

For isotropic medium ($K_x = K_y = K_z$) equation can be written as

$$\frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = 0 \tag{1.12}$$

This is also known as Laplace's equation.

**Transient Saturated Flow**

The law of conservation of mass for transient flow in saturated porous medium requires that the net rate of fluid mass flow into any unit volume be equal to the time rate change of fluid mass storage within the element. Thus, equation of continuity here takes the form as

$$-\frac{\partial(\rho v_x)}{\partial x} - \frac{\partial(\rho v_y)}{\partial y} - \frac{\partial(\rho v_z)}{\partial z} = \frac{\partial(\rho n)}{\partial t} \tag{1.13}$$

And after expanding,

$$-\frac{\partial(\rho v_x)}{\partial x} - \frac{\partial(\rho v_y)}{\partial y} - \frac{\partial(\rho v_z)}{\partial z} = n\frac{\partial \rho}{\partial t} + \rho\frac{\partial n}{\partial t} \tag{1.14}$$

After applying some physical properties, it becomes

$$-\frac{\partial(\rho v_x)}{\partial x} - \frac{\partial(\rho v_y)}{\partial y} - \frac{\partial(\rho v_z)}{\partial z} = \rho S_s\frac{\partial h}{\partial t} \tag{1.15}$$

After inserting Darcy law it becomes

$$\frac{\partial}{\partial x}\left[K_x\frac{\partial h}{\partial x}\right] + \frac{\partial}{\partial y}\left[K_y\frac{\partial h}{\partial y}\right] + \frac{\partial}{\partial z}\left[K_z\frac{\partial h}{\partial z}\right] = S_s\frac{\partial h}{\partial t} \tag{1.16}$$

## 1.2 Numerical Solvers and Boundary Conditions

The mathematical model here consists of a set of differential equations, as mentioned earlier which are known to govern the flow of groundwater. The reliability of all analysis from a groundwater model depends on how well the model approximates the field situation. One another aspect is simplification of assumptions made, which is necessary to construct a model because the field situation is too complex to be simulated exactly. To deal with more realistic situations, it is usually necessary to solve the mathematical model approximately using numerical techniques. In this section, we will study basic ways of solving a set of partial equation involved in groundwater flow system.

### 1.2.1 Boundary Value Problem

In mathematics, in the field of differential equations, a boundary value problem is a differential equation together with a set of additional constraints, called the boundary conditions. A solution to a boundary value problem is a solution to the differential equation which also satisfies the boundary conditions. Boundary value problems arise in several branches of physics as our flow system. To be useful in applications, a boundary value problem should be well posed. This means that given the input to the problem there exists a unique solution, which depends continuously on the input. In general ,there are three types of the boundary condition

- If the boundary gives a value to the function involved then it is a *Dirichlet boundary condition.(First type).*
- If the boundary gives a value to the normal derivative of the function then it is a *Neumann boundary condition(Second type)*

- If the boundary has the form of a curve or surface that gives a value to the normal derivative and the problem itself then it is a *Cauchy boundary condition(Third type)*

In terms of groundwater modelling these boundary condition can be said as

- Head is known quantity for surfaces bounding the flow region(Dirichlet)
- Flow is known quantity across surfaces bounding the region(Neumann)
- Some condition of either type is known for surfaces bounding the region

.

### 1.2.2 Examples of Boundary Value Problem in Groundwater Flow System

As an example [11] of a groundwater model involving Laplace's equation ( Equation 1.12) and suitable boundary condition, we represent a problem described by Toth(1962). This problem as shown in Figure 1.7, consists of a small watershed bounded on one side by a topographic high, which marks a regional groundwater divide, and on the other side by a major stream, which is a groundwater discharge area and marks another regional ground water divide. The aquifer is assumed to consist of homogeneous, isotropic, porous material underlain by impermeable rock. We first consider the boundary conditions, the left and right groundwater divides can be represented mathematically as impermeable, no-flow boundaries. Although no physical barrier exists, a groundwater divide has the same effect as an impermeable barrier because no groundwater crosses it. Groundwater to the right of the valley bottom discharges at point $A$, and Groundwater on either side of the topographic high flows away from the point $B$. The lower boundary is also a no-flow boundary because the impermeable basement rock forms a physical barrier to flow. The upper boundary of the mathematical model is the horizontal line $AB'$ even though the water table of the physical system lies above it. Thus the rectangular problem domain of mathematical model (shown in Figure 1.8) is an approximation to the actual shape of the saturated flow region. Along the boundary $AB'$, the head is taken to be equal to the height of the water table, and the water table configuration is considered to be a straight line.

We must express the boundary condition in mathematical terms. An equation is required for each boundary. Consider the upper boundary first. The boundary first, located at $y = y_0$ for $x$ ranging from 0 to $s$. The distribution of head along this boundary is assumed to be linear. The equation for a linear variation such that $h(0, y) = y_0$ is $h(x, y_0) = cx + y_0$ for $0 \leq x \leq s$ where c is the slope of the water table. The specification of head along the upper boundary makes it a Dirichlet boundary condition.The other three boundary conditions are no-flow boundary. According to Darcy's law on these boundaries hydraulic gradient should be zero.

Now boundary condition can be summarized as
Top($0 \leq x \leq s$):

$$h(x, y_0) = cx + y_0$$

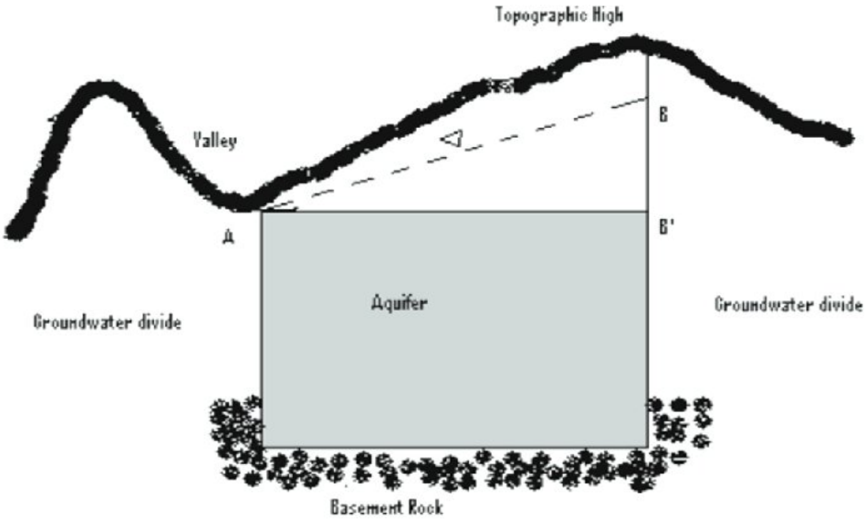



Figure 1.7: Schematic representation of the boundaries of a two-dimensional regional groundwater flow system

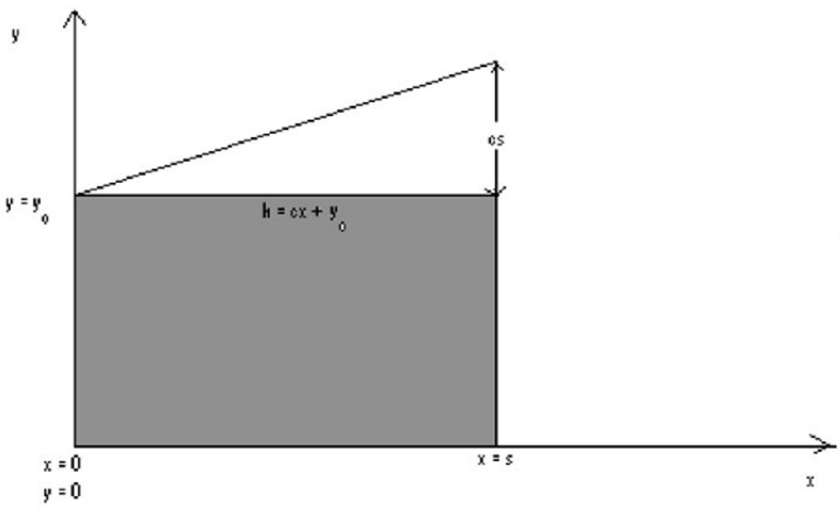



Figure 1.8: Mathematical model the two-dimensional regional groundwater flow system

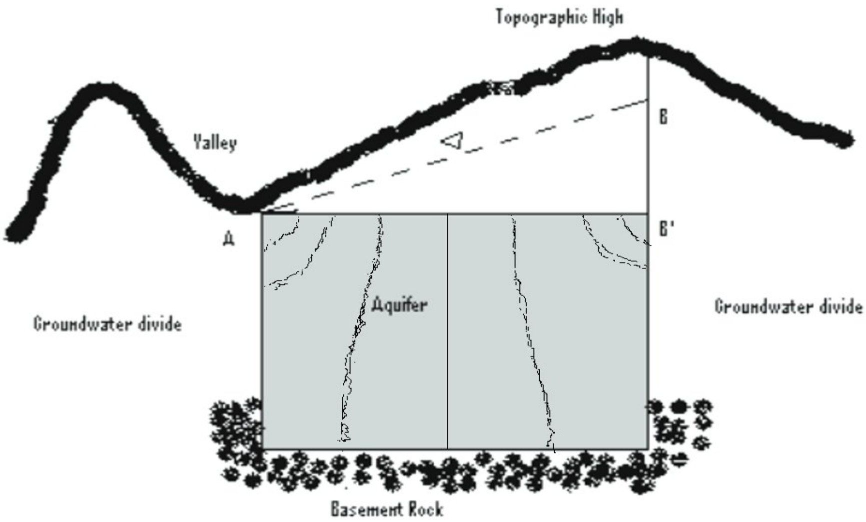



Figure 1.9: Flow net of above problem

Bottom($0 \leq x \leq s$):

$$\left.\frac{\partial h}{\partial y}\right|_{y=0} = 0$$

Left($0 \leq x \leq s$):

$$\left.\frac{\partial h}{\partial x}\right|_{x=0} = 0$$

Right($0 \leq x \leq s$):

$$\left.\frac{\partial h}{\partial x}\right|_{x=0} = 0$$

Now we consider the governing equation. We know that Laplace equation simulates groundwater flow in homogeneous, isotropic aquifer if there is no accumulation or loss of water within the system. Above model suits this case. So, the governing equation in this case is two dimensional Laplace equation as

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

The analytical solution, obtained by separation of variable, is

$$h(x,y) = y_0 + \frac{cs}{2} - \frac{4cs}{\pi^2} \sum_{m=0}^{\infty} \frac{\cos\left[(2m+1)\,\pi x/s\right] \cosh\left[(2m+1)\,\pi y/s\right]}{(2m+1)^2 \cosh\left[(2m+1)\,\pi y_0/s\right]} \tag{1.17}$$

This equation satisfies both the flow and boundary conditions. When plotted and contoured, its equipotential net leads to Figure 1.9 In next section, we will provide a way to get solution of this problem by numerical methods.

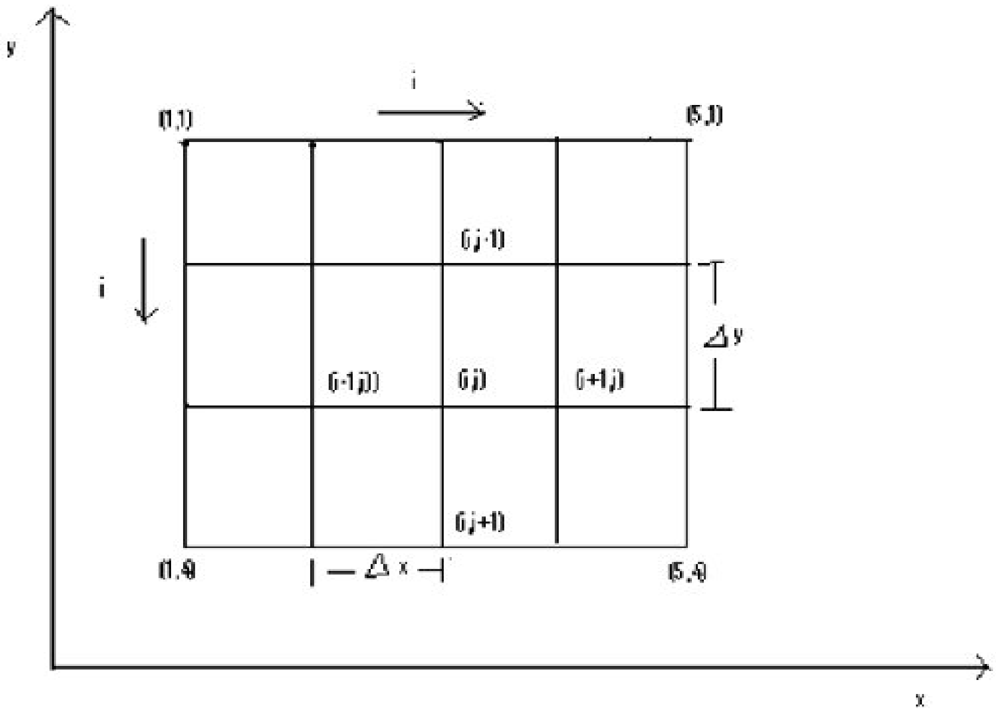



Figure 1.10: Finite difference grid

### 1.2.3 Finite Difference Method

For many problem, the assumptions that must be made to obtain an analytical solution (by calculus techniques) will not be realistic. In these cases, we must rely on approximate methods using numerical techniques to solve the mathematical model. One of this numerical technique is finite difference method. This method provides a rationale for operating on the differential equations that make up a model and for transforming them into algebraic equations. Numerical solution yield values for only a predetermined finite number of points in domain. By limiting our need to know the head to a reasonable number of points $N$, we can convert a partial differentiation equation into a set of $N$ algebraic equations involving $N$ unknown potentials. In this section, we will describe the finite difference method formulation for steady flow equation in two dimension as an example and then we will discuss some simple example based upon this formulation.

Consider a finite set of points on a regularly space grid as described in Figure 1.10. The distances $\Delta x$ and $\Delta y$ are spacing distances between points along horizontally and vertically respectively. To locate any intersection point in the grid, we specify an integer ordered pair $(i, j)$. The value of hydraulic head at point is $h_{i,j}$. Let the Cartesian coordinates $(x_0, y_0)$ be represented by (i,j). Consider a profile of head which has in succession the values $h_{i-1,j}$,$h_{i,j}$, and $h_{i+1,j}$. In the finite difference method, derivatives are replaced by differences taken between nodal point. A central approximation to $\partial^2 h/\partial x^2$ at $(x_0, y0)$ is obtained by approximating the first derivative at $(x+x_0/2, y_0)$ and at $(x-x_0/2, y_0)$, and then obtaining the second derivative by taking a difference between the first derivative at those points. We can write it as

$$\frac{\partial^2 h}{\partial x^2} = \frac{\frac{h_{i+1,j} - h_{i,j}}{\Delta x} - \frac{h_{i,j} - h_{i-1,j}}{\Delta x}}{\Delta x} \tag{1.18}$$

and it simplifies to

$$\frac{\partial^2 h}{\partial x^2} = \frac{h_{i+1,j} - 2h_{i,j} + h_{i-1,j}}{\Delta x^2} \tag{1.19}$$

Similarly,

$$\frac{\partial^2 h}{\partial y^2} = \frac{h_{i+1,j} - 2h_{i,j} + h_{i-1,j}}{\Delta y^2} \tag{1.20}$$

And according to laplace equation sum of preceding two equation should be zero. If we choose $\Delta x = \Delta y$ (square grid). Then equation becomes as

$$h_{i+1,j} + h_{i-1,j} + h_{i,j+1} + h_{i,j-1} - 4h_{i,j} = 0 \tag{1.21}$$

In case of Dirichlet boundary condition, as we have given head value of cells located at boundaries, we can formulate set of equations according to above equation. And we get values of unknown head values of interior cells. Now, what will happen in case of mixed boundary condition (Problem Figure 1.8 ). Top boundary conditions will be mapped as above said Dirichlet condition. Left, right and bottom conditions (Neumann condition), where there is no flow, that is hydraulic gradient is zero, can be handled with adding one row or columns of cells as fictitious cells and taking their head values same to its adjacent in the direction of real cells. After that we proceed as Equation 1.21 to form set of linear equations.

## 1.3 MODFLOW

MODFLOW is the U.S. Geological Survey modular finite-difference flow model, which is a set of computer programs that solves the groundwater flow equation [12]. The program is used by hydrogeologists to simulate the flow of groundwater through aquifers. The code is free software, written primarily in Fortran. MODFLOW has become the worldwide groundwater flow model and used to simulate systems for water supply solutions, containment remediation etc. MODFLOW is most appropriate in those situations where a relatively precise understanding of the flow system is needed to make a decision. It also shares all advantages of open source code programs, as its features can be customized by changing codes as per needs. And that is the reason that here are several graphical interfaces to MODFLOW, which often include the compiled MODFLOW code with modifications. These programs aid the input of data for creating MODFLOW models.

The modular structure of MODFLOW consists of a Main Program and a series of subroutines. The subroutines are grouped in packages. Each package deals with a specific feature of the hydrologic system which is to be simulated such as flow from rivers or drains. A groundwater flow within the aquifer is simulated in MODFLOW using a block-centered finite-difference approach. Layers can be simulated as confined, unconfined, or a combination of both. Flows from external stresses such as flow to or from wells, areal recharge, evapotranspiration, drains, riverbeds can also be simulated.

MODFLOW has the ability to manage the large data sets required when running large simulation problems. Large amount of information and a complete description of the flow system is always required to make the most efficient use of MODFLOW. To use its program, the region to be simulated must be divided into cells with a rectangular grid resulting in layers, rows and columns. Files must be prepared that contain hydraulic parameters such as hydraulic conductivity, Transmissivity, specific yield, etc., boundary conditions such as impermeable boundaries and constant heads, and stresses like pumping wells, recharge from lake precipitation, rivers, drains, etc.

### 1.3.1 As Solver

Partial Differential equation which represents three dimensional movement of ground water is

$$\frac{\partial}{\partial x}\left[K_{xx}\frac{\partial h}{\partial x}\right]+\frac{\partial}{\partial y}\left[K_{yy}\frac{\partial h}{\partial y}\right]+\frac{\partial}{\partial z}\left[K_{zz}\frac{\partial h}{\partial z}\right]+W=S_S\frac{\partial h}{\partial t} \tag{1.22}$$

where

- $K_{xx}$, $K_{yy}$ and $K_{zz}$ are the values of hydraulic conductivity along the $x$,$y$, and $z$ coordinate axes and may be function of space.
- $h$ is the potentiometric head (hydraulic head)
- $W$ is a volumetric flux per unit volume representing sources and/or sinks of water, where negative values are water extractions, and positive values are injections. It may be a function of space and time ( i.e. $W = W(x, y, z, t)$)
- $S_s$ is the specific storage of the porous material and may be function of space.
- $t$ is time.

For simple systems, analytical solutions of this partial differential equation are possible. But there are a lot of realistic problems, for which various numerical methods must be employed to get an approximate solution . One such approach is finite difference method, upon which MODFLOW is based.The process leads to system of simultaneous linear equation and solution yields values at specific points and times.

A spatial discretization of an aquifer system (as in Figure 1.11 with a grid of blocks called cells, the locations of which are described in terms of rows, columns, and layers (i=1,2..NROW; j=1,2..NCOL; K=1,2..NLAY). Within each cell a point is taken as representing node at which head is to be calculated. Figure 1.11 shows block centered formulation of these nodes.The grid in MODFLOW is assumed to be rectangular horizontally and vertically. MODFLOW handles discretization of space in the horizontal direction by reading the number of rows, the number of columns, and the width of each row and column. Discretization of space in the vertical direction is handled in the model by specifying the number of layers to be used, and by specifying the top and bottom elevations of every cell in each layer. Vertical discretization can be viewed as an effort to represent individual aquifers or permeable zones by individual layers of the model.

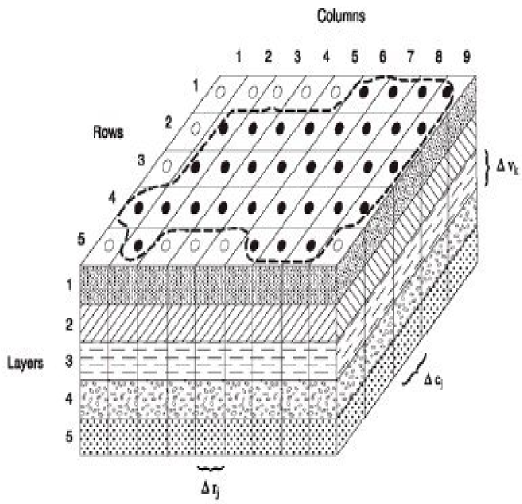



Figure 1.11: Discretization [12]

## Finite Difference Discretization Equation

Continuity equation expressing the balance of flow for a cell i,j,k is

$$\sum Q_i = S_s \frac{\Delta h}{\Delta t} \Delta v \tag{1.23}$$

where
$Q_i$ is flow rate into cell;
$S_s$ is specific storage ;
$\Delta v$ is the volume of the cell; and
$\Delta h$ is the change in head over time interval of length $\Delta t$.
This equation sums all flows from adjacent nodes for cell i,j,k as shown in Figure 1.12.

And using Darcy's law, flow between cell i,j,k and cell i,j-1,k can be written as

$$q_{i,j-1/2,k} = KR_{i,j-1/2,k} \Delta c_i \Delta v_k \frac{h_{i,j-1,k} - h_{i,j,k}}{\Delta r_{j-1/2}} \tag{1.24}$$

Again we can write this equation as

$$q_{i,j-1/2,k} = CR_{i,j-1/2,k} h_{i,j-1,k} - h_{i,j,k} \tag{1.25}$$

where CR term (Known as hydraulic conductance) is

$$CR_{i,j-1/2,k} = \frac{KR_{i,j-1/2,k} \Delta c_i \Delta v_k}{\Delta r_{j-1/2}} \tag{1.26}$$

Similarly, we can write equations of q's for each face as in Figure 1.12( top-bottom, front-back and right using conductance terms CV, CC, CR ,similar to CR term of the cell.

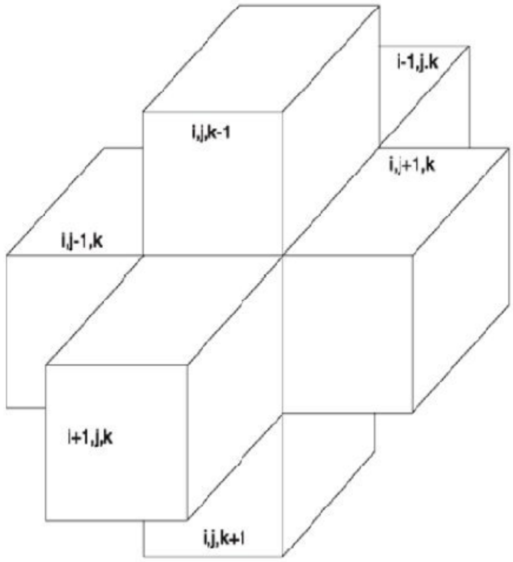



Figure 1.12: Adjacent cells corresponding to cell i,j,k [12]

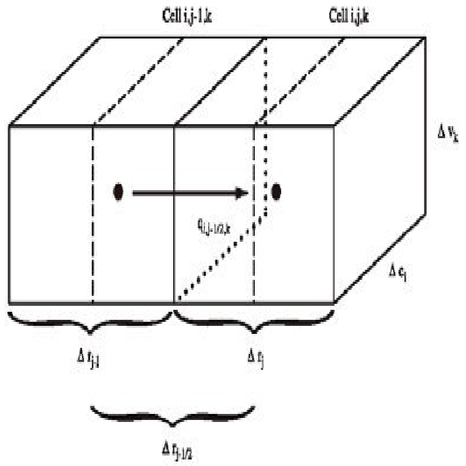



Figure 1.13: Adjacent conductivity term [12]

Flow from outside the aquifer like well, river,drains,recharge may be dependent on the head in receiving cells or may be entirely independent of receiving cells, so it is represented as

$$a_{i,j,k} = p_{i,j,k}h_{i,j,k} + q_{i,j,k} \tag{1.27}$$

where $a_{i,j,k}$ represents flow from the external sources into cell i,j,k;and $p_{i,j,k}$ and $q_{i,j,k}$ are constants, respectively. And finally, involved time derivative of head at time $t^m$ in main governing equation is obtained as

$$\frac{\Delta h_{i,j,k}}{\Delta t} \cong \frac{h^m_{i,j,k} - h^{m-1}_{i,j,k}}{t^m - t^{m-1}} \tag{1.28}$$

The resulting equation

$$\begin{aligned}CV_{i,j,k-1/2}h_{i,j,k-1} + CC_{i-1/2,j,k}h_{i-1,j,k} + CR_{i,j-1/2,k}h_{i,j-1,k} + \\ (-CV_{i,j,k-1/2} - CC_{i-1/2,j,k} - CR_{i,j-1/2,k} \\ -CR_{i,j+1/2,k} - CC_{i+1/2,j,k} - CV_{i,j,k+1/2} \\ +HCOF_{i,j,k})h_{i,j,k} \\ +CR_{i,j+1/2,k}h_{i,j+1,k} + CC_{i+1/2,j,k}h_{i+1,j,k} + CV_{i,j,k+1/2}h_{i,j,k+1} &= RHS_{i,j,k}\end{aligned} \tag{1.29}$$

where
$HCOF_{i,j,k} = P_{i,j,k} - \dfrac{SS_{i,j,k}\Delta r_j \Delta c_i \Delta v_k}{t - t^{m-1}}$, and
$RHS_{i,j,k} = -Q_{i,j,k} - SS_{i,j,k}\Delta r_j \Delta c_i \Delta v_k \dfrac{h^{m-1}_{i,j,k}}{t - t^{m-1}}$
The system of equation of the form of above equation, which includes one equation for each variable head cell in the grid, may be written in matrix form as

$$[A]h = \{q\} \tag{1.30}$$

### 1.3.2 Procedures and Packages

The period of simulation is divided into a series of stress periods within which specified stress data are constant. Each stress period, in turn, is divided into a series of time steps. The system of finite-difference equations of the form of equation is formulated and solved to yield the head at each node at the end of each time step.

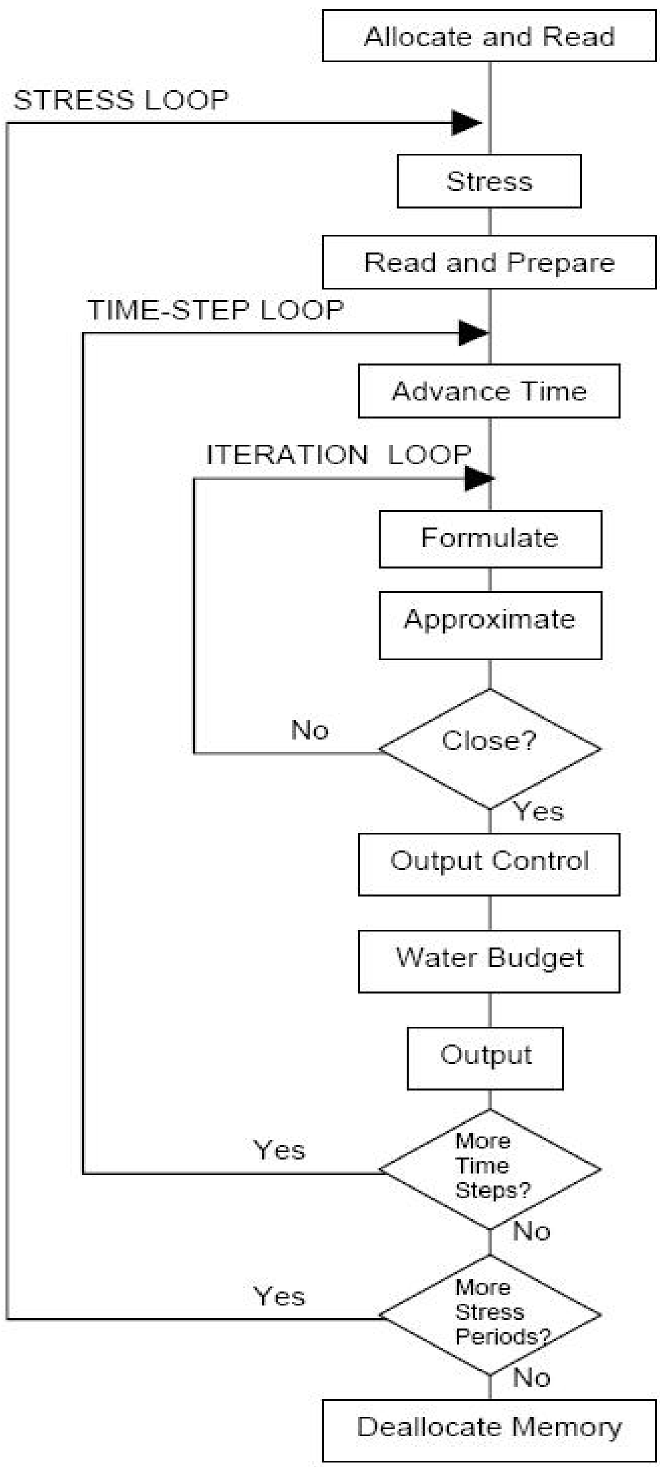



Figure 1.14: Flow chart [12]

Iterative solution methods are used to solve for the heads for each time step. Thus, the program includes three nested loop as in Figure 1.14. All the features that is needed to support this flow chart is facilitated with the help of packages, mainly of two types:

1. Internal Flow packages :An Internal flow packages helps in simulation of flow between adjacent node. It calculates the CV, CR, and CC conductance coefficients and the ground-water storage terms in the finite-difference flow equation which simulates flow between adjacent cells. It also does storage calculation for cells. It takes specific yield or specific storage term into account for storage calculation on the basis of whether cell is partially saturated (i.e. unconfined) or fully saturated (i.e. confined). Internal Flow package also supports calculation for these two conditions:

    - Dewatered Condition: The vertical flow calculation is modified if a cell is unconfined (head is below its top elevation) while the cell directly above is fully or partially saturated(e.g. under perched condition).
    - Dry Cell to Wet Cell: When head being less than the bottom elevation of a cell (i.e. saturated thickness = 0), it is clear that a variable-head (wet) cell should convert to dry. Although, there is no straight-forward way to know when a dry cell should convert to wet. A dry cell converts to wet based upon the head in an adjacent cell compared to a wetting threshold for the cell. We will discuss it later with a simulation example.

2. Stress Packages: The stress packages add term to the groundwater flow governing equation representing external inflow or outflow to the system. Stress packages provide modelling of stress conditions like well, recharge, rivers, lake, drain etc. These stress packages can also serve purpose of modelling regime based boundary conditions. Well and recharge are head independent stresses but others such as drain and lake are head dependent. Head dependent boundary conditions are basically IF-ELSE type boundary conditions. We will discuss these later in subsequent chapters with examples.

### 1.3.3 Boundary Conditions

In mathematics, in the field of differential equations, a boundary value problem is a differential equation together with a set of additional constraints, called the boundary conditions. A solution to a boundary value problem is a solution to the differential equation which also satisfies the boundary conditions. These boundary conditions are of equally importance in groundwater modelling for their physical and mathematical aspects of a system. Boundary conditions available to apply in MODFLOW are:

1. **Default Boundary condition**: In formulating the finite-difference equations, cell-to-cell conductance terms are omitted for the exterior of cells on the outer surface of the rectangular grid. Thus, considering flow along a row, a cell-to-cell conductance term is developed for the interval between column 1 and column 2, but not for the interval to the opposite side of column 1; similarly, a conductance term is developed for the interval between column (NCOL-1) and column (NCOL), but not for the interval beyond column (NCOL). Similar conventions are established in the other

two directions, so that, in effect, the grid is bounded externally by planes across which no cell-to-cell flow occurs. If these boundaries of the model grid, which are actually embedded in the program, coincide with impermeable boundaries in the aquifer, they can be relied upon to simulate the no-flow condition along those aquifer boundaries without further specification by the user.

2. **Constant Head**: Constant-head or specified head cells may be used to represent features such as surface-water bodies of constant level that are in full contact with the aquifer.

3. **Specified Flow**: Boundaries that are characterized by a constant rate of flow into or out of the aquifer may be simulated using a no-flow boundary in conjunction with the Well Package, by assigning appropriate withdrawal or recharge rates to nodes just inside the boundary.

Constant-head cells, no-flow cells, and variable-head cells are distinguished from one another in the model through the IBOUND variable, which contains one value for each cell in the grid.

## 1.4 Groundwater Modelling System(GMS)

There are several graphical interfaces to MODFLOW, which often include the compiled MODFLOW code with modifications. Non-commercial MODFLOW versions are free, however, their licensing usually limit the use to non-profit educational or research purposes. Commercial MODFLOW programs are typically used by governments and consultants for practical applications of MODFLOW to real-world groundwater problems. Professional versions of MODFLOW are generally priced at a minimum of around $1000 and typically range upward to $7000 USD. Although there are no standards in this respect. GMS(Groundwater Modeling System) is one of the widely used groundwater modelling software among hydrologist due to its GUI feature.

Two approaches can be used to construct a MODFLOW simulation in GMS: the grid approach and the conceptual model approach. The grid approach involves working directly with the 3D grid and applying sources/sinks and other model parameters on a cell-by-cell basis. The conceptual model approach involves using the GIS tools in the *Map module* to develop a conceptual model of the site being modeled. The data in the conceptual model are then copied to the grid. GMS provides tools for pre and post processing of data over MODFLOW solver. We used GMS to run MODFLOW for example simulations and encountered some serious problems. We will explore these problems in different chapters later. This report proposes an interface over MODFLOW, much likely GMS, but with more encapsulated and intelligent features for groundwater modelling.

## 1.5 Chapter Organization

This section shows the organization of remainder of the report. Basics of groundwater simulations is explained in Chapter 2. The chapter explains about basic input data requirement to design a groundwater model for a system. The chapter presents results obtained after simulation and issues encountered during these simulations. Lake/reservoir modelling is described in Chapter 3. The main issue in this chapeter is on quantification of augmented water in groundwater due to lake like structures. The chapter also presents inabilities of GMS which are encountered during different simulation examples. Chapter 4 is about modelling of well in a given regime. The chapter also explains a regime description of a dam having high conductivity channel underneath on slopping topology and some of its initial results. Chapter 5 concludes the thesis by summarizing main issues. The chapter also discusses related issues and directions for future work.

# Chapter 2

# Basics of Groundwater Simulation

Groundwater models are powerful techniques to record the effects on a system created by different elements like stress elements such as infiltration/extractions or like input parameters such as hydraulic conductivity etc. Simulation of a groundwater flow regime involves designing a groundwater model that involves the inherent characteristics of the physical system. These models are used to understand the behavior of groundwater system and responses to any external stresses. This chapter describes basic groundwater simulation results of a plain topology. As mentioned earlier, groundwater flow is dependent on hydraulic conductivity of medium and other inputs. This chapter is aimed to elaborate groundwater flow and its simulation results on varying conductivity. It seems obvious that more natural recharge (i.e. rain) to a system improves its water table. We will see this with the help of these basic simulations. In groundwater hydrology, confined or unconfined aquifer is normally considered for the groundwater balance. A groundwater balance involves various components which may contribute to or take from the system being studied. This chapter explains some of these basic components with the help of a simulation. Here, we create a hypothetical but realistic groundwater flow model of a regime having plain topology and then use this model for different results. This chapter may be seen as protocol for preparing groundwater flow models.

Overall, this chapter introduces an example groundwater flow regime which explains

- Designing Issues and Results:
  - Typical input data set required to prepare a model of any groundwater flow regime.
  - Available results/plots after simulation and its interpretation.
- Experiments:
  - Effect of hydraulic conductivity on groundwater flow system.
  - Effect of conductivity on time to get stabilization in groundwater flow.
  - Effect of different recharge rate on the system.

The regime described in this chapter will be carried forward in subsequent chapters for further studies of different other factors. These results serve as a comparison base

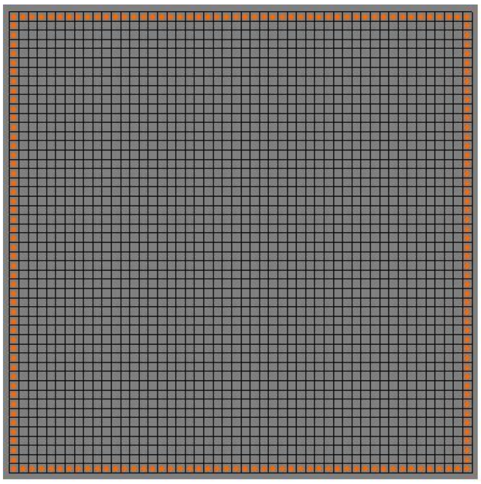


Figure 2.1: Grids on X and Y axis

for subsequent chapters, which are simply based on elaborating more stresses on example case study of this chapter.

## 2.1 Regime Description

A hypothetical but realistic regime of plain topology is considered here for groundwater simulation example. The simulation method is based on finite difference based solver (as MODFLOW). The method requires various input parameters to be set.
These inputs and description are basic input requirements to prepare a groundwater simulation model for any regime. This can be viewed as protocol for groundwater modelling too.

Inputs/Description required to be set with respect to our example model:

1. **Area**: Area of our regime is $500m \times 500m$.

2. **Layering and Griding**: A discretization of regime with grids and layers is required to prepare a model. In general, layering is done according to geology of area and highly dependent on layering of equi-conductivity zone of physical system. Finer result and numerical instability are trade offs in choosing number of grids and layering. Here, we create three layers in Z -axis, and regional area is divided as X and Y-axis on griding of $50 \times 50$ as shown in Figure 2.1. We have thus created $50 \times 50 \times 3 = 7500$cells.

3. **Elevations**: Each layers specification is supplemented and in particular each cell by its top and bottom elevation. Layers elevations for our example are 0-4m, 4-8m and 8-22m in order of bottom to top as shown in Figure 2.2.

4. **Boundary Condition**:We have already mentioned the significance of boundary conditions in groundwater flow regime in Section 1.2.2. Section 1.3.3 explains different boundary conditions available in MODFLOW. All boundaries are at constant

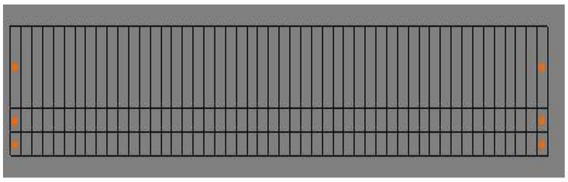


Figure 2.2: Elevations and layers in crossectional view

head at 9m in this example. Default boundary condition available in GMS is no flow boundary which is not a good choice for recharge studies and it does not map conditions according to physical system i.e. no flow boundary condition will accumulate all recharge water in the regime which does not happen in reality. Although for real world regimes appropriate boundary condition is time variant specified head dependent on ambient head values but for the sake of simplicity, we have taken specified constant head boundary for this regime( as in Fig 2.3).

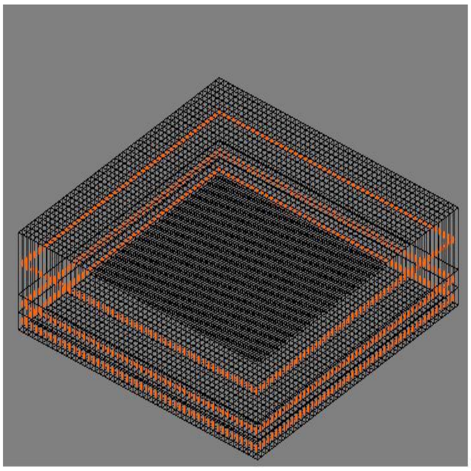


Figure 2.3: Three dimensional view of grids, layers and boundary at constant head 9m

5. **Specific Storage and Specific yield**: As mentioned in Section 1.1.3, specific storage term is defined for saturated zone simulation. Specific storage value in this case is set to .0001. Specific yield (as mentioned in Section 1.1.3) is defined for unsaturated or partially unsaturated zone. Specific yield value is set to 20 percent in our case.

6. **Conductivity**: Conductivity (i.e. $K$ term involved in Darcy Law) is next input to be defined. This value is to be defined for each cells of regime. We have created three input sets of conductivities values. Table 2.1 is the first set of conductivity values.

7. **Starting Head values**: The next input required is initial head values for each cell. Starting head values of all cells are fixed at 9m in this example. It means that our

| Layer | Conductivity | Soil Class |
|---|---|---|
| Layer 1 | 2.00 m/d | Very fine sand |
| Layer 2 | 2.00 m/d | very fine sand |
| Layer 3 | 0.05 m/d | Silt |

Table 2.1: Conductivity input set 1

first layer from top is partially saturated (i.e. unconfined aquifer system). Most of the hydrological projects and systems in India are unconfined aquifer based. So it looks more realistic. Starting head value of a topology should lie in between top and bottom elevation of the cell. Although there are option to break this limitation but this may lead to some instability in system. We will refer this issue later in this chapter.

8. **Stress Periods and Recharge Rates**:Recharge rate defines the amount of rainfall that reaches the water table. The amount of rainfall that reaches the water table depends upon several factors such as soil type, its thickness, surface topography and temperature. These factors should be kept in mind by user. Therefore, some percentage of actual rainfall is taken as recharge rate for a given area.

   Typically, stress periods are divided based on discretization of recharge rates and extraction rates to/from the different stress system. In this example, we are trying to simulate recharge term due to rain factor only. In this case stress periods are typically divided into rain and non-rainy season. One stress period is of having rain from June to August, and other is from September to May each year for the duration of three years. Further, each stress periods is to be divided into time steps. Table 2.2 describes different stress periods, no. of time steps and associated recharge rates.

| Stress Period | No. of time steps | Recharge Rate |
|---|---|---|
| 1/05/2007 to 31/05/2007 | 10 | 0 m/d |
| 1/06/2007 to 31/08/2007 | 10 | 0.003 m/d |
| 1/09/2007 to 31/05/2008 | 30 | 0 m/d |
| 1/06/2008 to 31/08/2008 | 10 | 0.003 m/d |
| 1/09/2008 to 31/05/2009 | 30 | 0 m/d |
| 1/06/2009 to 31/08/2009 | 10 | 0.003 m/d |
| 1/09/2009 to 31/05/2010 | 30 | 0 m/d |

Table 2.2: Stress periods, number of time steps and recharge rates

9. **Effective Recharge Zone**: Recharge rates are only applicable to cells of $44 \times 44$ grids($440 \times 440$ sq. m.) in middle. This is only to record more stabilized system.

## 2.2 Simulation Results

After designing model with suitable set of inputs, we run MODFLOW solver and as results :

1. We can obtain head value vs time plot of any cell( as shown in Figure 2.4)

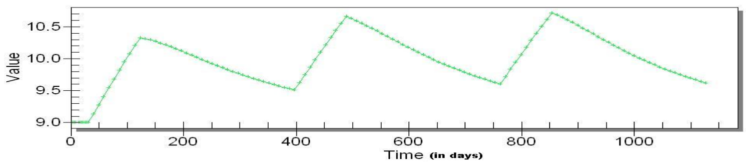



Figure 2.4: Head time series

2. We obtain a mass balance plot which describes inflow and outflow of the regime or selected area of regime. We describe plots relative to basic groundwater simualation. We discusses mass balance plot having data items( in terms of in/out both) like; Storage, Constant Head, Recharge. Other items will be disscussed in coming chapters.

   (a) Storage: Storage In as shaded area in Figure 2.5 always is always projected on positive y-axis. It shows quantity of water contribution from groundwater storage aginst some storage that needs some extraction or outflow from sytem. In simple words this is the amount of water extracted from groundwater storage. Storage Out as shaded in Figure 2.5 signifies quanity of water augumented in groundwater storage. Difference of these two items may lead towards depletion/improvement of water table of regime taken for study during the time taken for calculation.

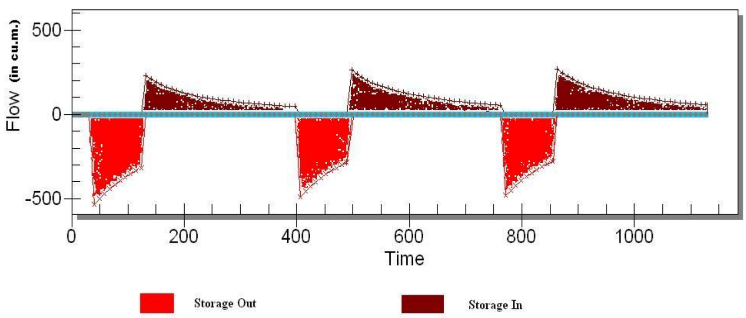



Figure 2.5: Storage term in mass balance plot

   (b) Recharge: As shaded in Figure 2.6, the recharge term is always plotted on positive side of y-axis and it means the quantity of water coming into groundwater flow due to recharge. In other words, this quantity is product of area of recharge zone by recharge rate at each time steps.

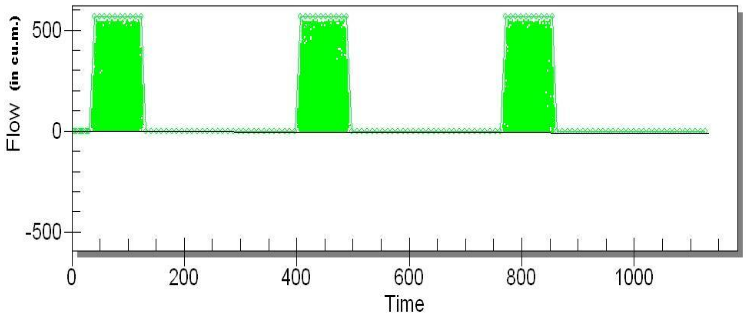



Figure 2.6: Recharge term in mass balance plot

(c) Constant head: Constant head of cells signifies that the cell cannot gain/loose water itself. In overall, flow budget/mass balance, Constant Head Out term is plotted on -ve side of y-axis. and it signifies the quantity of water going outside of regime through constant head cells. Similarly Constant Head In term is plotted on positive side of y-axis and it means the quantity of water coming in towards regime through constant head cells. As Figure 2.7 shows the amount of water going out through constant head cells. In this figure, amount of water coming in regime through constant head cells is zero.

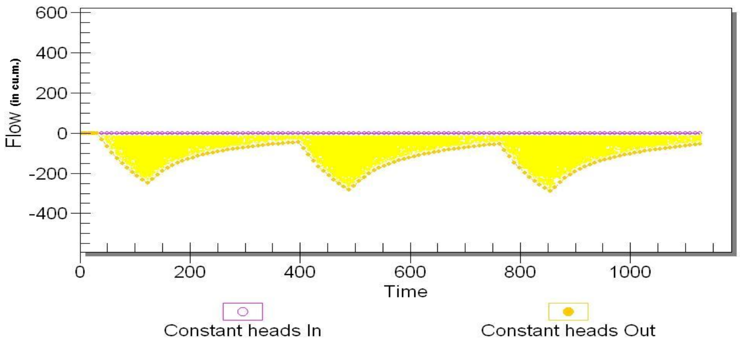



Figure 2.7: Constant Head Out term in mass balance plot

(d) In addition to these plots, we have countour of equipotential lines for every time steps. We can see resultant flow vectors for each time steps too (as shown in Figure 2.8).

## 2.3 Effect of Varying Conductivity and Rain

We have created three sets of conductivity input set (as one set described above). Using these sets, we will get a flavor of groundwater flow systems changes as a response of conductivities values. Each set of these conductivities and its results are described with the help of several plots.

One another aspect of varying recharge rate is also combined with each input set. So that response of system due to more or better rain could be also possible to visualize along with. We have experimented with different recharge rates on this set of conductivities. At first we experimented with a recharge rat of 0.0003 m/d as mentioned in regime description. Then we multiplied this recharge rate by three and five respectively.

### 2.3.1 Conductivity Input Set 1

Followings are the lists of results/plots to describe the resultant groundwater flow system based on the input set (as Table 2.1) and different recharge rates.

1. **Contour Map and Flow Vector**: Contour map shows contour of equipotential lines. Flow vectors are normal to equipotential lines. Figure 2.8 shows contour map and flow vector of this regime on 31 Aug, 2007(i.e. Just after end of rain).

2. **Flow Budget:** Flow budget explains mass balance for whole regime over all the time steps. In this case, it quantifies water going out of regime through constant head boundary and water being stored in groundwater against recharge. A set of flow budget plots is shown in Figure 2.9 with different recharge values(i.e. 0.003m/d, 0.009m/d, 0.015m/d). We can understand these flow budget by putting some lining over selected area (not in original flow budget plot) as follows:

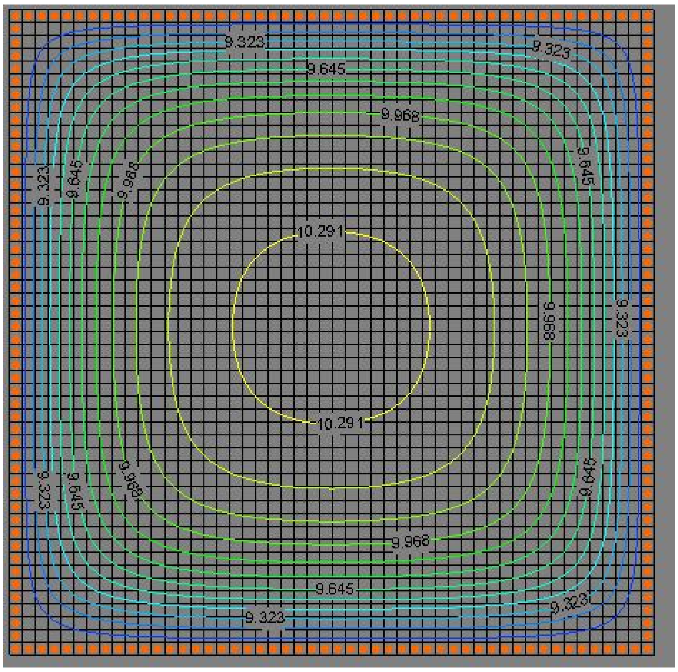
   (a) Quantity of water augmented in regime over period of simulation = Total Recharge quantity - Quantity of water went off through constant head boundary( i.e. green lined area - yellow lined area as shown in Figure 2.9a)

   (b) Quantity of water added during rainy season instantly against recharge happened is red lined area (as shown in Figure 2.9b)

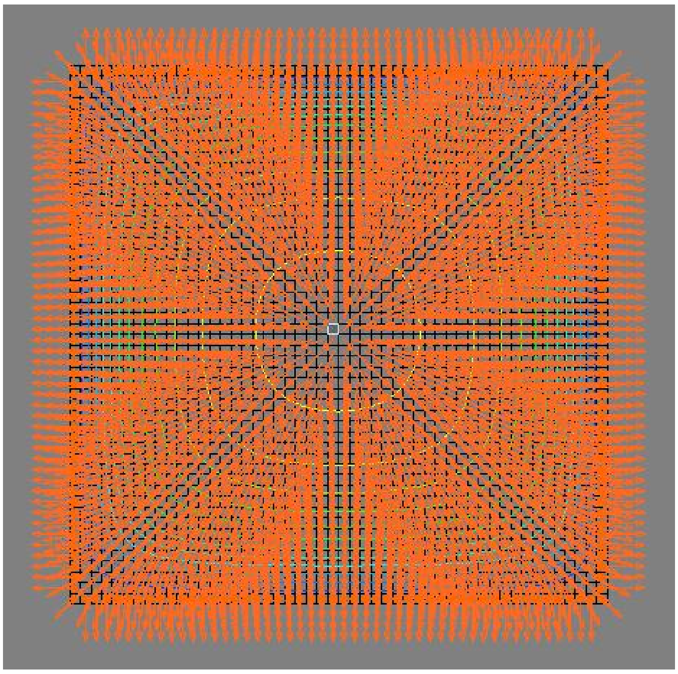
   (c) Quantity of water going outside through constant head boundary after rainy season is Grey lined area (as shown in Figure 2.9c)

3. **Water table after rain**: Figure 2.10 shows crossectional views of water table from middle row of regime just after finishing first rain (i.e. at time step 31 August, 2007). The figure shows higher water table formation on higher recharge rate.

4. **Time series for head values of some selected cells**: Head values of the system increases due to recharge in rainy season and it decreases due to constant head boundary condition after that. We have selected some of the cells from top layer and plotted their respective head time series with different recharge rate in Figure 2.11, Figure 2.12 and Figure 2.13. Next two sets of plots are also of same head vs time plot but represents plots for set of selected cells. The next two figures (Figure 2.14, Figure 2.15) give idea about contour formation over all time steps.

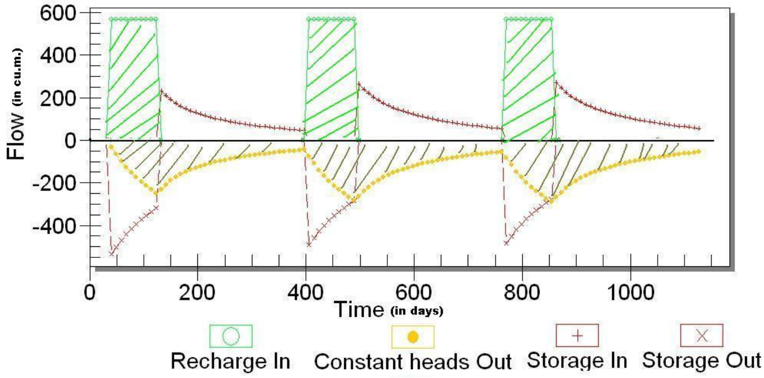



(a) Contour Map

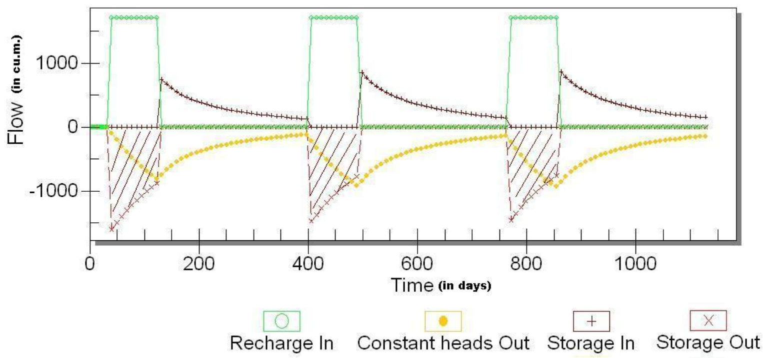


(b) Flow Vector

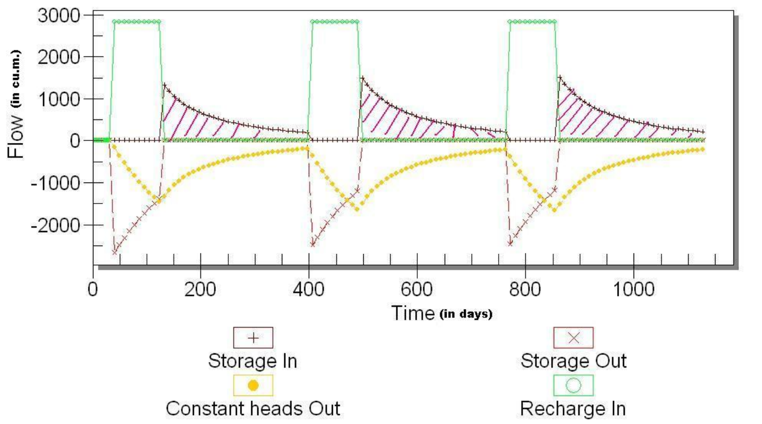
Figure 2.8: Contour and flow vector map of top layer on 31 Aug, 2007(i.e. just after rain)




(a) Mass balance plot when recharge rate is 0.003m/d

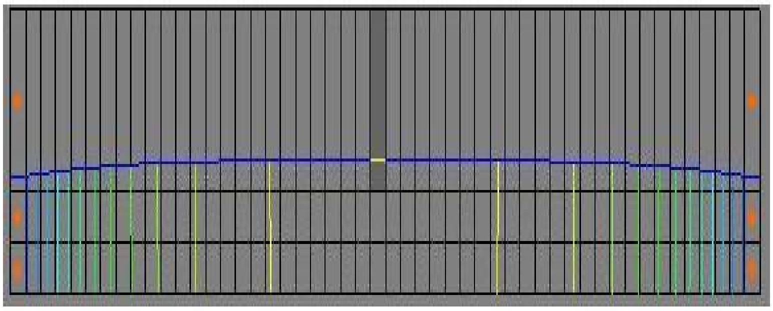



(b) Mass balance plot when recharge rate is 0.009m/d

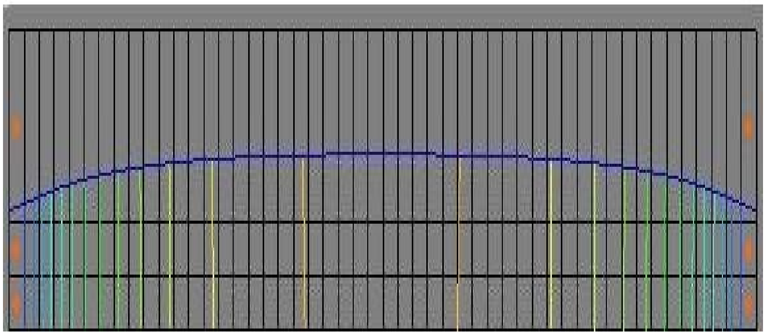



(c) Mass balance plot when recharge rate is 0.015m/d

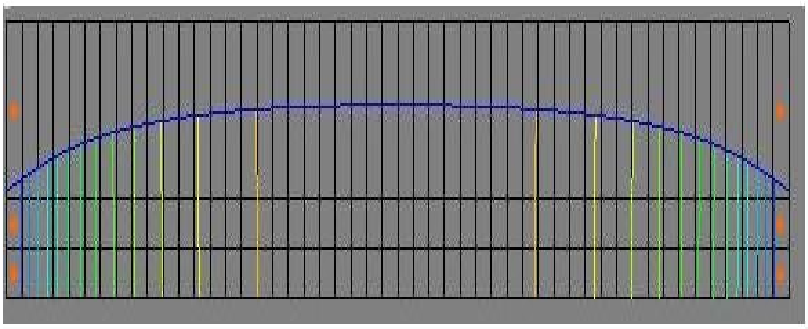
Figure 2.9: Mass balance plots with different recharge rate



(a) Water table when recharge rate is 0.003m/d

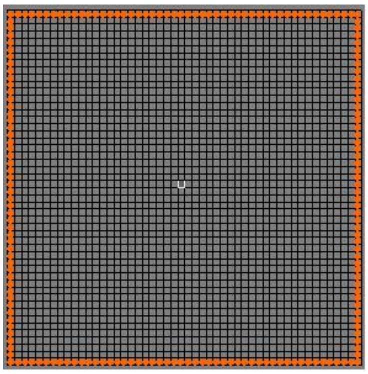


(b) Water table when recharge rate is 0.009m/d

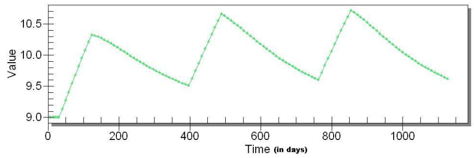


(c) Water table when recharge rate is 0.015m/d

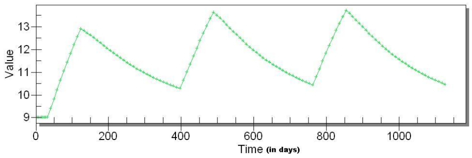
Figure 2.10: Water table at the end of rain(i.e. 31st Aug, 2007) with different recharge rate



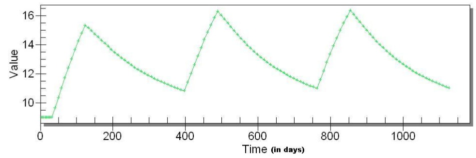
(a) Position of selected cell at $25 \times 25$ from top layer, Id:= 1225




(b) Head time series of this cell when recharge rate is 0.003m/d

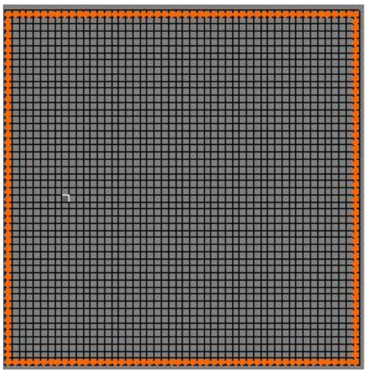



(c) Head time series of of this cell when recharge rate is 0.009m/d

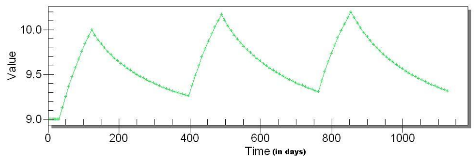



(d) Head time series of of this cell when recharge rate is 0.015m/d

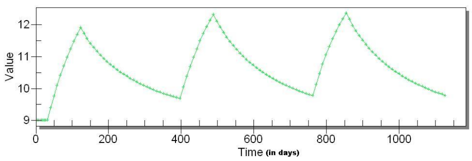
Figure 2.11: Head vs Time plot of selected cell (cell id = 1225) at different recharge rate



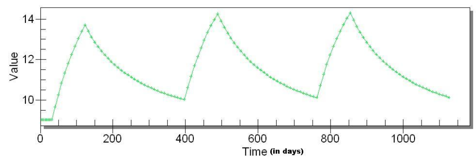
(a) Position of selected cell at $27 \times 9$ from top layer, Id:= 1309




(b) Head time series of this cell when recharge rate is 0.003m/d

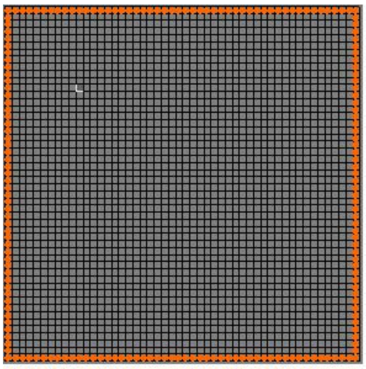



(c) Head time series of of this cell when recharge rate is 0.009m/d

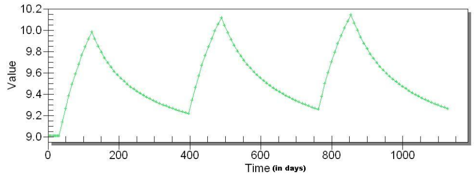



(d) Head time series of of this cell when recharge rate is 0.015m/d

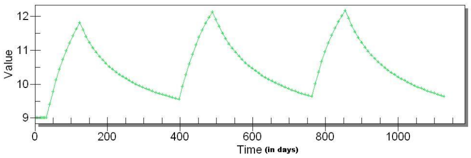
Figure 2.12: Head vs Time plot of selected cell (cell id = 1309) at different recharge rate



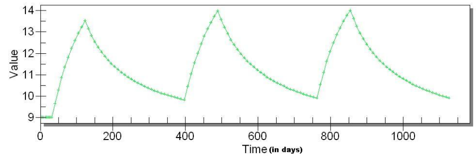
(a) Position of selected cell at $12 \times 11$ from top layer, Id:= 561




(b) Head time series of this cell when recharge rate is 0.003m/d

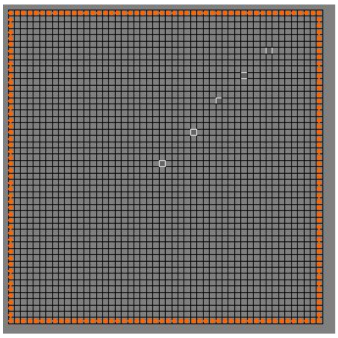



(c) Head time series of of this cell when recharge rate is 0.009m/d

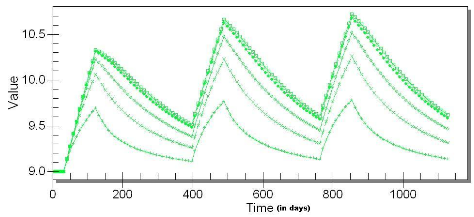



(d) Head time series of of this cell when recharge rate is 0.015m/d

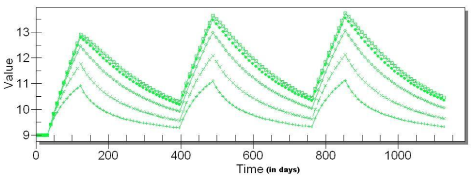
Figure 2.13: Head vs Time plot of selected cell (cell id = 561) at different recharge rate



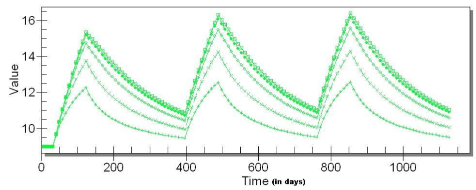
(a) A set of selected cell almost on diagonal




(b) Head time series of this cell when recharge rate is 0.003m/d

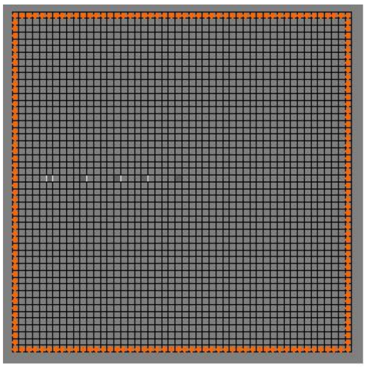



(c) Head time series of of this cell when recharge rate is 0.009m/d

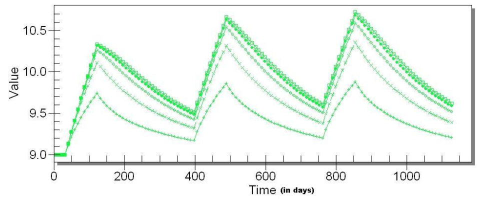



(d) Head time series of of this cell when recharge rate is 0.015m/d

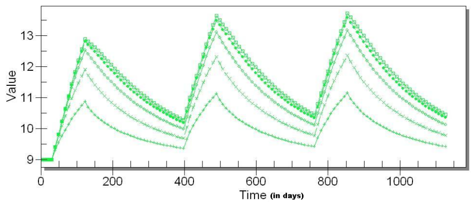
Figure 2.14: Head vs Time plot of a group of selected cell (cell id = 561) at different recharge rate



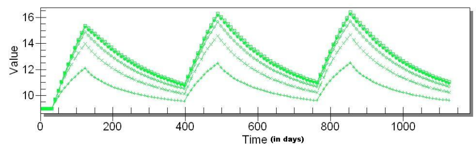
(a) A set of selected cell from middle cell towards west




(b) Head time series of this cell when recharge rate is 0.003m/d




(c) Head time series of of this cell when recharge rate is 0.009m/d




(d) Head time series of of this cell when recharge rate is 0.015m/d

Figure 2.15: Head vs Time plot of another group of selected cell at different recharge rate

### 2.3.2 Conductivity Input Set 2

Table 2.3 is second set of conductivity values. As previous results set, we plotted same type of results set for this input set too (See Appendix A).

| Layer | Conductivity | Soil Class |
|---|---:|---|
| Layer 1 | 1m/d | Sand stone |
| Layer2 | .01m/d | Clay |
| Layer3 | .05m/d | Silt |

Table 2.3: Conductivity input set 2

### 2.3.3 Conductivity Input Set 3

Table 2.4 is third set of conductivity values. As previous results set, we plotted same type of results set for this input set too( See Appendix B)

.

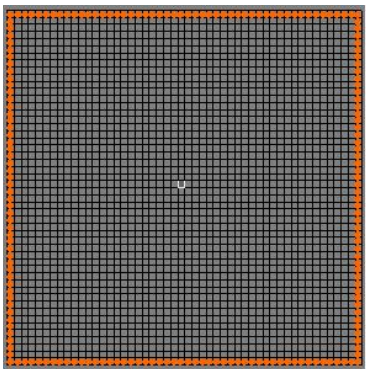
| Layer | Conductivity | Soil Class |
|---|---:|---|
| Layer 1 | 10m/d | Fine sand |
| Layer 2 | 2m/d | Very fine sand |
| Layer 3 | 0.05m/d | Silt |

Table 2.4: Conductivity input set 3

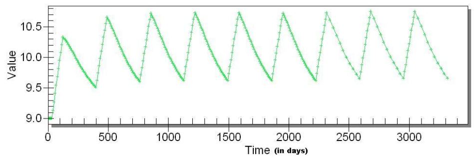
## 2.4 Stability Time and Conductivity

To get an impression of stability time versus conductivity we have experimented with all three previous input sets of conductivity. We have incremented no. of stress periods up to 10 in all the cases. Mass balance plots are shown in Figure C.1a, Figure C.1b and Figures C.1c for input set 1, 2 and 3 respectively. Figure 2.16, Figure C.2, and Figure C.3 show head time series for selected cells as earlier plots in previous section.

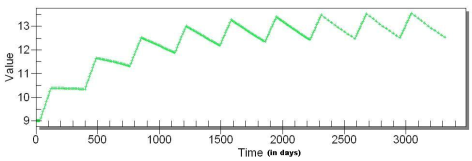
Observation: Stabilization time in plots can be observed as point of time steps after which graph is periodic. It is observed that lesser values of conductivity set take more time to get stabilized groundwater flow. As in our example set, second input set (Table 2.3) takes more time.



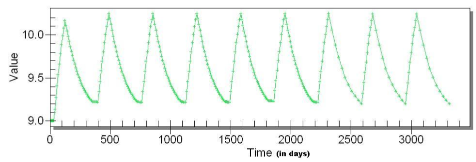
(a) Position of selected cell at $25 \times 25$, Id:= 1225




(b) Head time series of this cell when recharge rate is 0.003m/d




(c) Head time series of of this cell when recharge rate is 0.009m/d




(d) Head time series of of this cell when recharge rate is 0.015m/d

Figure 2.16: Head vs Time plot of selected cell (cell id = 1225) at different recharge rate

## 2.5 Related Issues

1. **Boundary condition**: In real physical system, typical boundary conditions for a regime are decided using ambient head value observations at boundaries. And these head values are transient. So more accurate boundary conditions for such type of studies are transient specified constant head boundaries. Moreover, the information on any regional system is often isolated and the quality obtained by the numerical simulation mainly depends on the experience and intuition of the engineer. The model idealizations or calibration and the usage of a chosen discretization have to be justified by a calibration process. Geological properties (layer properties) as well as boundary and initial conditions are subject to special investigation in a process of adjusting the mathematical model of the given flow system. For example, we have 92 observations of last eight years of Thane District (see Figure C.4). These readings may serve as base to calibrate a model and these reading may provide us boundary conditions to design a model for any hydrological project in Thane district.

2. **Number of grids**: Number of grids and layers decides number of head values to be computed by solver. More refined results can be obtained by putting more grids on system. On the other hand, if we increase it by large then numerical instability occurs.

3. **Number of layers**: Number of layers are typically decided by geological formation of regime (based on hydraulic conductivity).

4. **Layers and Starting Head**: Cells which have starting head value less then its bottom elevation are considered as dry cells. This situation may occur after any iteration during simulation too. By default MODFLOW solver cannot be run when any cell has starting head value is below its bottom elevation. But there is a feature which can be used to run this forcefully. It is WET/DRY option. Using wet dry/option for any cell, cell can be converted from dry to wet status depending on their adjacent cells head values. Using this option, we tried to run our previous example with starting head values at 7m everywhere. Obviously top layers cells are dry, second layer cells are partially saturated, and third layer is fully saturated at start of simulation (as shown in Figure 2.17). Constant head boundaries are all fixed at 9m as previous. When rain happens, head values of second layers increases gradually beyond 8m (its top elevation) and this is the situation where dry cells of top layer should becomes wet. When these cell becomes wet, their head values are assigned as using equation

$$h_{i,j,k} = BOT_{i,j,k} + WETFCT(h_n BOT_{i,j,k}) \tag{2.1}$$

where $WETFCT$ is a user-specified constant, generally between 0 and 1, and $h_n$ is the head at the neighboring cell that causes cell i,j,k to convert to wet.
For example, some cells from top layer becomes wet and active on 31st, Aug, 2007 (i.e. just after first rain ends) as in Figure 2.18. Due to this dry/wet switching of a cell, time series of head values of a cell may occur as Figure 2.19. Although wet/dry option simulates some realistic condition from field, it may lead to some

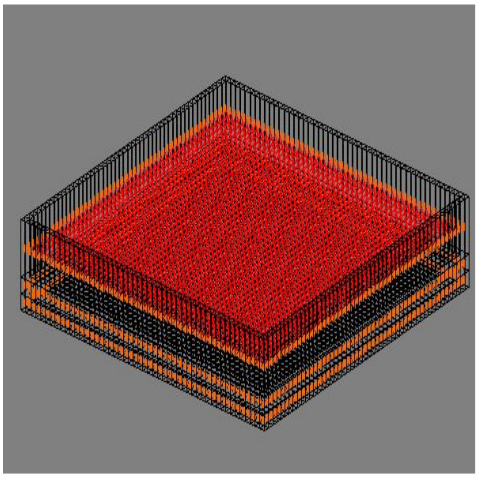
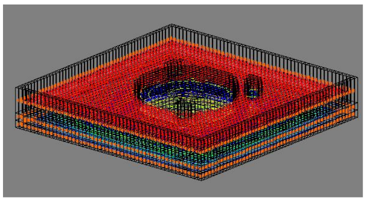
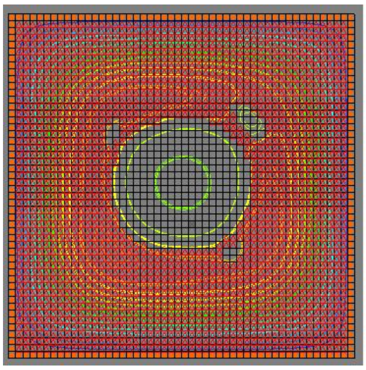


Figure 2.17: Three dimensional view of regime where top layer's cells are dry cells(red marked)

instabilities in results. As our contour plot has always concentric equipotential lines in previous example, but we get some distorted contour plots at some time steps by introducing wet/dry option to this example (as shown in Figure 2.20). Choosing right $WETFCT$ value which gives more accurate results is an important issue.



(a) Three dimensional view of wet cells



(b) Top layer view of wet cells

Figure 2.18: Wet cells on 31 Aug, 2007 (just after rain ends)

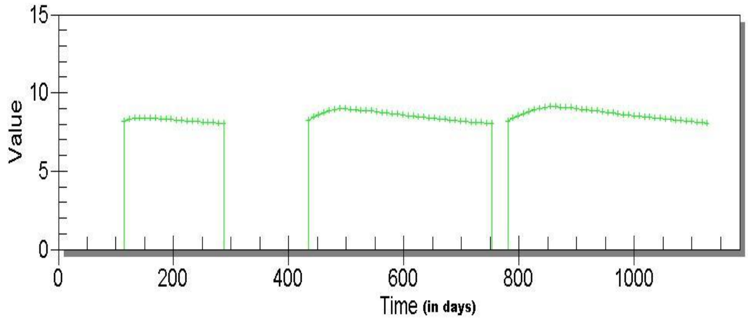



Figure 2.19: Head value variation of a cells (at postion $25 \times 25$) from middle of top layer(absence of line indicates dryness of this cell)

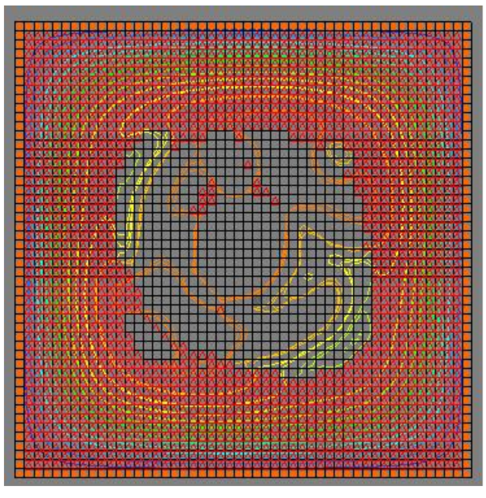


Figure 2.20: Grids on X and Y axis

# Chapter 3

# Reservoir/Lake Modelling

Groundwater is the major source of drinking water in both urban and rural India. During the past two decades, the water level in several parts of the country has been falling rapidly due to an increase in extraction. The number of wells drilled for irrigation have rapidly and indiscriminately increased. These all causes water table depletion in India and other part of world too. Therefore, groundwater planning and management is required. Very obvious solution to water table depletion problem is construction of artificial recharge structures.

The artificial recharge to ground water aims at augmentation of ground water by modifying the natural movement of surface water utilizing suitable civil construction techniques. Construction of check dam like structures are one of them. In rural and semi-arid area, these dams/ reservoirs serve for household needs and improve water table locally. In theory, it looks simple, but success stories of these types of dam structure is not satisfactory. In practice, no groundwater simulation is used to get certainties for success of project. Cost of simulation and unavailability of simple and intelligent tool are main reason for this. In this chapter, we tried to model a reservoir/lake in a simple topology using GMS interface. This chapter examines improvements of water table in regime due to dam. One major aspect in construction of dam is the dimension of dam because its cost of construction is dependent on dimension. Anyone would like to get cost benefit data for varying dimension of dam before constructing it. We will try to get these answers and results in this chapter.

This chapter mainly explains:

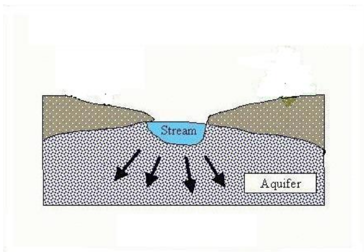
- Designing and conceptualization of lake and simulation results:
  - Lake and groundwater interaction concepts
  - How to model a lake/reservoir and regime description
  - Results/plots interpretation afetr simulation
- Experiments to examine:
  - Improvement of head values in regime due to construction of a lake

– Quantification of water augmented to ground water with respect to different dimensions of reservoir/dam.

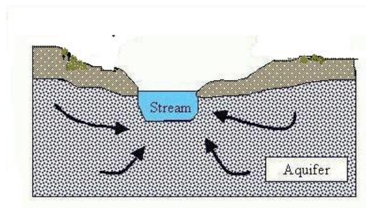
During preparation of lake/dam simulations of this chapter, we encountered lot of issues related to lake/reservoir simulation. We will discuss these issues in last section of this chapter.

## 3.1 Lake and Groundwater Interaction

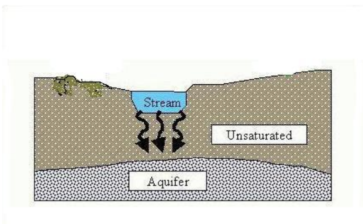
The aquifers are often partially fed by seepage from streams and lakes. In other locations, these same aquifers may discharge to feed the streams, rivers, and lakes. This section explores both situations of recharge and discharge from surface to groundwater or vice versa. In some situations, lake seepage (losses to the aquifer) may be affected by ground water pumping and natural variations in aquifer water level. When the aquifer water level is near land surface, seepage from the river is partially controlled by the height of the aquifer water level (as Figure 3.1a). Activities or events that result in a lowering of the water table, such as ground water pumping, induce more seepage from the river. Conversely, events that cause the aquifer water level to rise (recharge events) will result in a decrease in river seepage. If aquifer water levels rise above the level of the river, what was previously a losing lake will become a lake that is gaining water from the aquifer (as Figure 3.1b). Another hydrologic condition exists that is very important in understanding surface and ground water interaction. A surface water body is perched above an aquifer when aquifer water levels are well below the bed of the lake (as in Figure 3.1c). Under these conditions, water will seep from the surface water body to the ground water, but the surface water body will not be affected by aquifer water levels and consequently does not change in response to ground water pumping. Nearby ground water pumping will cause a lowering of the water table, but will not affect surface water supplies.




(a) Seepage case of water from lake to aquifer




(b) Gaining case of water from aquifer to lake



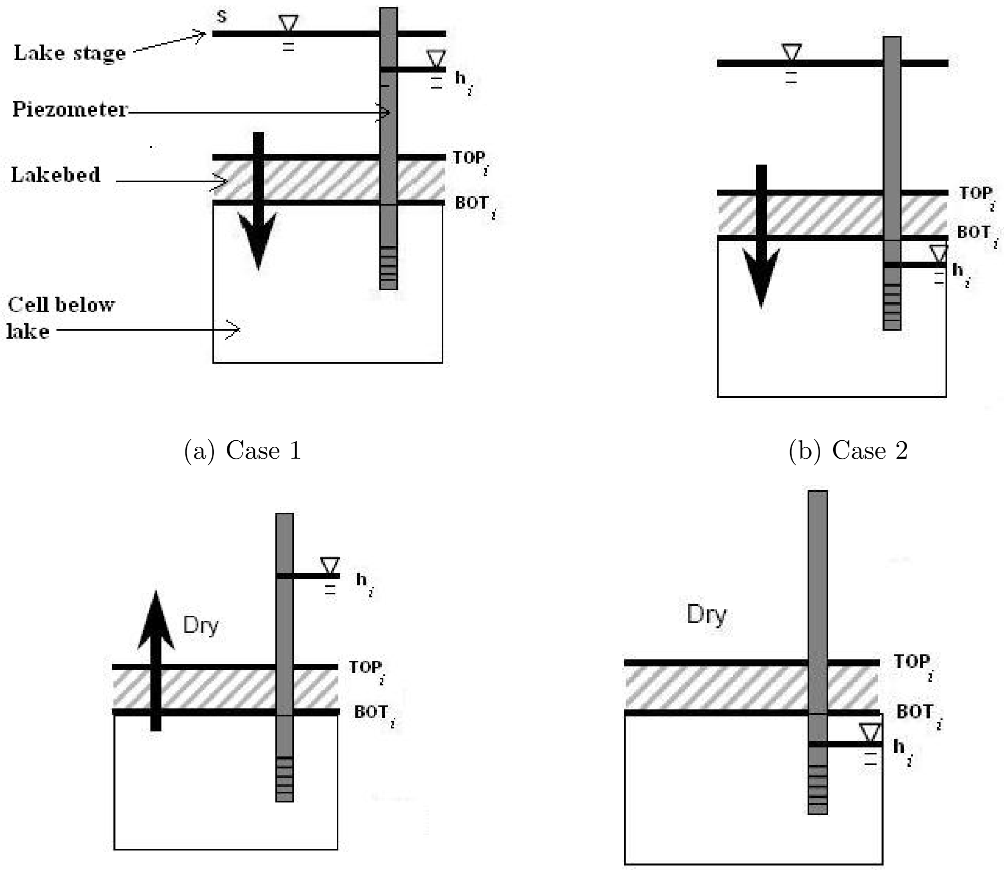

(c) Perched condition

Figure 3.1: Conditions of interaction between surface water body and groundwater [7]

Modelling of all these situation in case of any surface water body is done by adding or removing water from flow budget of groundwater regime. The quantification of added or removed water depends on situation as described above. And we can summarized these calculations as:

- If $S \geq TOP_i$ and $h_i \geq BOT_i$ then $Q_i = CD_i(S - h_i)$. (as in Figure 3.2a)
- If $S \geq TOP_i$ and $h_i \leq BOT_i$ then $Q_i = CD_i(S - BOT_i)$ (as in Figure 3.2b)
- If $S < TOP_i$ and $h_i > TOP_i$ then $Q_i = -CD_i(h_i - TOP_i)$ (as in Figure 3.2c)
- If $S < TOP_i$ and $h_i < TOP_i$ then $Q_i = 0$ (as in Figure 3.2d)

where $S$ is water stage in lake, $Q_i$ is the amount of water flow from/to lake to/from one of adjacent cell (i.e. $i$th) one of the interacting cell of the model, $TOP_i$ and $BOT_i$ are to define lakebed thickness adjacent to $i$th cell, $h_i$ is head value of $i$th cell.




(a) Case 1

(b) Case 2

(c) Case 3

(d) Case 4

Figure 3.2: Computation of lake-to-groundwater flow

## 3.2 Regime description and lake modelling

Regime taken here is almost same as previous chapter's regime with some minor changes which are only modeling related. These changes are only with respect to input parameters but overall physical conceptualization of regime is same as previous. It is like constructing a dam/reservoir in previous example. The changes and some augmented parameters are only to include a reservoir in model.

These parameters are:

1. **Layers, grids and elevations:** Regional area and number of grids are same as previous example but number of layers is increased to 4. Top layer( elevation 8-22m ) of earlier example has been broken into two layers of elevation 8m to 15m and 15m to 22m (as in Figure 3.3).

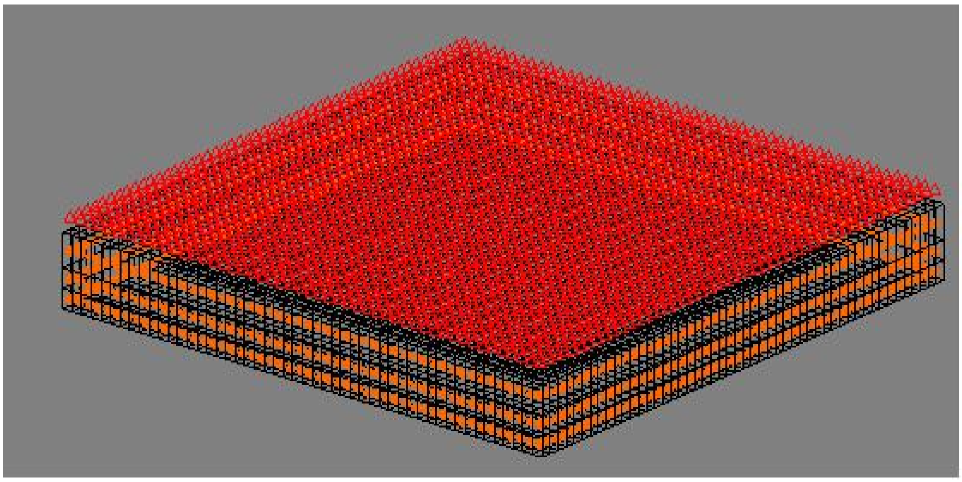


Figure 3.3: Layers, grids and elevations of regime. Red marked cells are dry cells

2. **Hydraulic conductivity**: Hydraulic conductivities in layers are as table 3.1.

| Layer | Conductivity | Soil Type |
|---|---|---|
| Layer 1 | 2.00 m/d | Very fine sand |
| Layer 2 | 2.00 m/d | Very fine sand |
| Layer 3 | 0.5 m/d | More finer sand |
| Layer 4 | 0.05 m/d | Silt |

Table 3.1: Hydraulic conductivity of layers

3. **Starting head**: Starting head value is 9m everywhere as previous example. As we have split top layer of previous example into two layers here, new top layer becomes completely dry (i.e. starting head is less than bottom elevation of cells). Second layer is partially saturated. Dry cells are shown as red marked cells in Figure 3.3.

4. **Wet and Dry**: As starting head values of top layer cells are not in domain of the cell elevation (as they are dry cells), we cannot run this model as it is mentioned in last chapter. To run this model we have to set wet/dry flag as on. After setting this flag when solver runs, it checks for wet/dry switching of the cells after each iteration.

5. **Boundary Conditions**: All boundaries are at constant head except top layer's boundary. It cannot be a constant head boundary as these cells are dry cells (i.e. head is not greater than cell's bottom elevations). Dry cells are taken as inactive cells and solver does not consider these cells for calculation.

6. **Recharge Zone and Rate**: These settings are same as previous.

7. **Lake Modeling**: These are the settings particular to lake modeling.

   (a) **Lake Dimension**: Lake has dimension $5 \times 5$ cells in middle of regime (i.e. $area = 50 \times 50 sq.m$). These lake cells are blue marked cells in Figure 3.4. Lake cells should be inactive because its not a part of groundwater. and no head values are to be calculated for lake cells. Only water level and storage are calculated during different iteration based on other lake package parmeters.

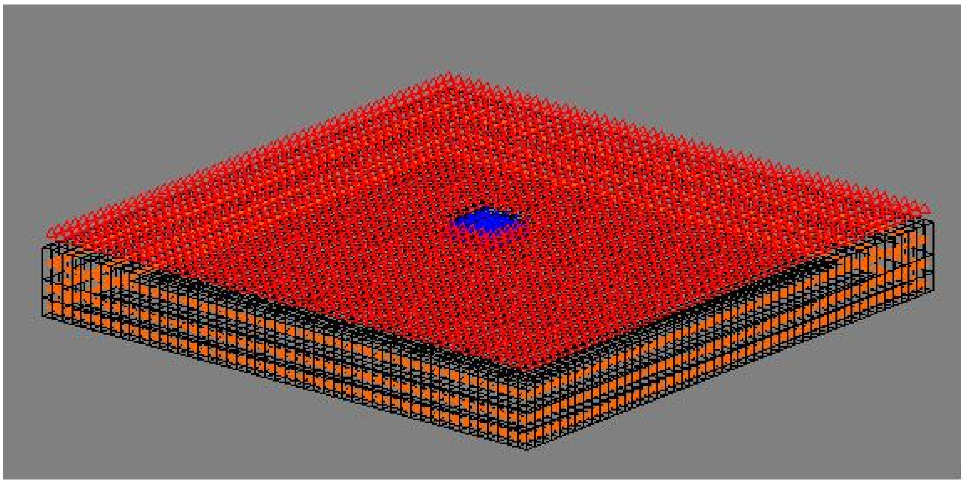


Figure 3.4: Regime with lake cells (Blue marked cells)

   (b) **Lake Depth and Layering**: Depth of this lake is 7 meter and that is the reason we have broken our top layer into two layers into 8-15m and 15-22m. Lake formation in this example is from 15-22m with 1m of assumed lakebed. In GMS, there is no direct implementation of lakebed and it will consider lake formation from 15m to 22m for water level variation (not 16to 22m) in this case. Thus whenever a lake is to be defined in a given regime, layering should be done according to lake depth.

   (c) **Lake Package Parameters** (as shown in Figure 3.5):

      - **Initial Stage**: Initial stage defines initial water level before any stress periods. For our case, we take it as 19m.

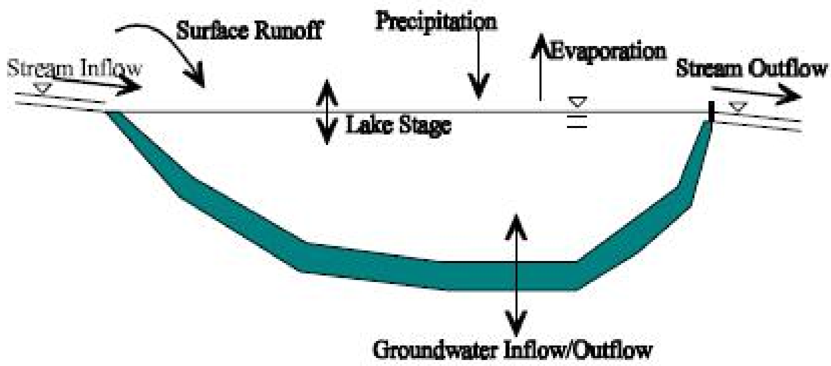



Figure 3.5: Input parameters for lake modeling [10]

- **Leakance**: Leakance term defines the quantity of water seepage through lake. Leakance term involves conductivity of lakebed and lakebed thickness both. In GMS, both factors (lakebed conductivity and width) are incorporated in term of leakance.
  Leakance = Conductivity of Lakebed/Lakebed Thickness
  We take it as 0.01 per day for our case.
- **Precipitation**: Precipitation term defines the rate of water added to lake due to rain. It is just like recharge rate in basic groundwater modelling but it is defined as same value as precipitation rate for the regime of lake. In this example, we take it as .03 m/d for stress periods (same as previous example) of rainy season (June to August). We are adding precipitation term to lake cells for rainy seasons. Therefore , recharge rate to these cells should be equal to 0 for all stress periods. It means that recharge rate should not be assigned to lake cells or cells below lake cells.
- **Run off**: This term defines amount of water added into lake due to surrounding surface run off in rainy season. In this case we take it as 70 cubic meter per day during rain stress periods.
- **Evaporation**: This term defines rate of extraction from lake due to evaporation. In this example we set it to .00005m/d throughout all the stress periods.
- **Outflow/Withdrawal**: Outflow terms defines quantity of natural or artificial withdrawl from lake. We are not using this term in this simulation.

There are terms in lake package which are not functioning in the expected way in GMS with MODFLOW grid approach. These are:

- Maximum stage
- Minimum stage
- Parent lake

- Sill elevation

Issues related to these terms will be described later in this chapter.

## Simulation Results

In addition to basic simulation results described in Section 2.2, we may plot some resultant plots particularly in case of lake modeling. These are

1. Lake stage Plots: It is a plot of water level of lake. It is similar to head time series plot.

2. Water budget of a regime: Water budget of a regime having a lake can be visualized as Figure 3.6. The figure shows different data items involved in water budget for a given regime.

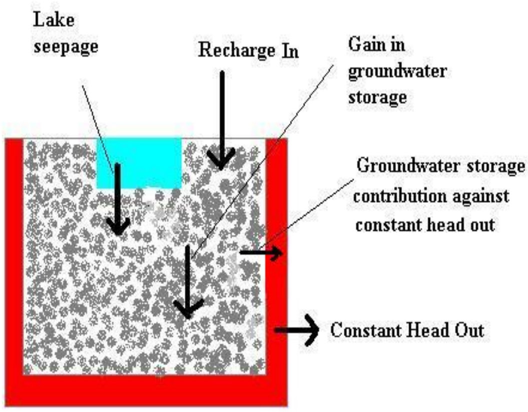



Figure 3.6: A typical water budget for regime having lake

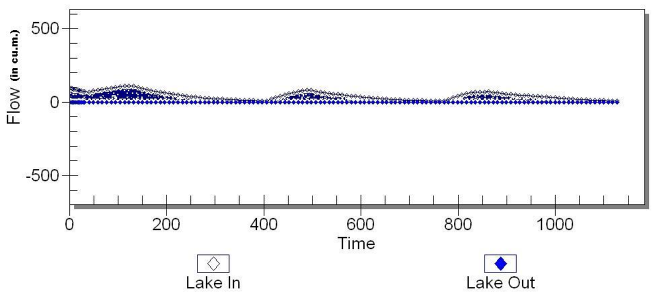



Figure 3.7: Lake term in mass balance plot

3. Mass Balance Plots: Lake in/out data item is plotted in addition to basic resultant data item as explained in Section 3.7. Figure 3.7 is a typical plot of quantification of

water added in flow due to lake seepage. Lake In term signifies gain in groundwater due to lake seepage whereas Lake Out term signifies amount of groundwater which goes out in form of gaining water level in lake (in Figure it is zero all the time). Lake out term is plotted on negative side of y-axis.

## 3.3 Improvements in Groundwater due to Lake

Now we can run this model having lake. For comparison purpose we simulated this model without lake too. Here we lists results as comaprisons between system without lake and with lake so that improvement in groundwater due to lake can be visualized. These results are:

1. Table 3.2 describes cumulative data over complete stress periods. Noticeable thing is lake seepage quantity and storage out quantity. Output categorization here is done according to GMS. It is little bit confusing in terms of storage in and storage out terminology. In actual, Storage In term is quantity contributed from storage towards constant head out quantity and storage out term is actual quantity of water augmented into groundwater. As a summarizing of this table, we can say that constructing lakes in a regime will be advantageous in terms of getting lake seepage and water augmented in groundwater storage and both can be quantified.

| Volume (in cu.m.) | Without Lake | With lake |
|---|---:|---:|
| **In** | | |
| Storage | 90331.8281 | 90357.3594 |
| Recharge | 160300.7969 | 158230.7969 |
| Lake Seepage | 0 | 49335.0234 |
| Total In | 250632.6250 | 297923.1875 |
| **Out** | | |
| Storage | 124563.0078 | 139207.8281 |
| Constant head | 126073.7500 | 158716.4219 |
| Total Out | 250636.7500 | 297924.2500 |

Table 3.2: Cumulative water balance comparison between a system without lake and with lake

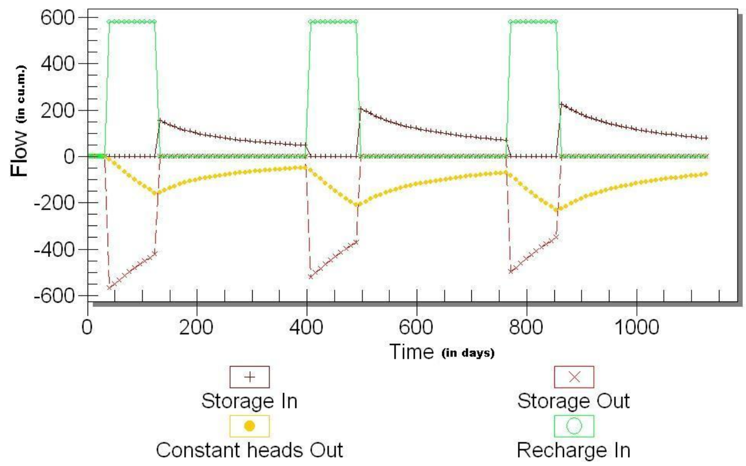
2. The next very obvious question is about water level in lake. Figure 3.9 shows time series of water stage of lake for this example case.

3. Figure 3.8 describes mass balance plots of a system without lake and with lake both. In Figure 3.8b, blue lined area quantifies lake seepage into groundwater.

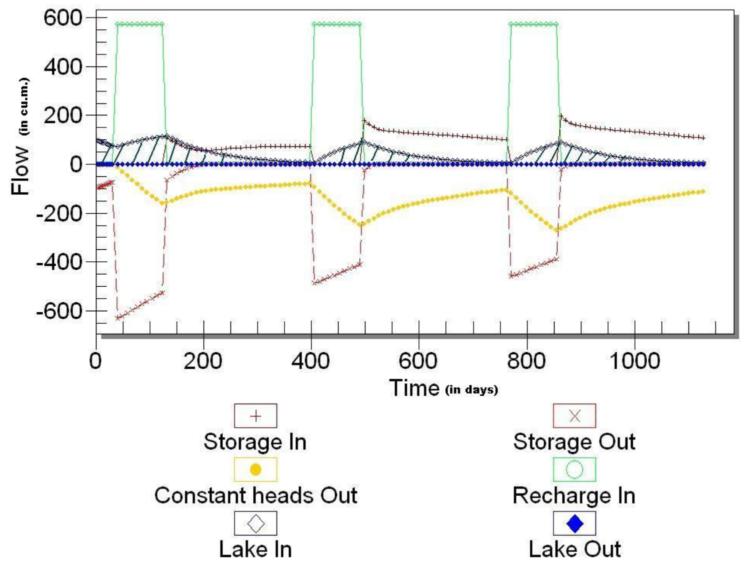
4. Figure 3.10 and Figure 3.11 are head plots of some selected cells from second layer for both cases( with/without lake). With combination of both figures we can imagine contour lines for both the cases. In both cases as we go towards boundary in any direction head value decreases to maintain 9m constant head at boundary. Head values of the regime without lake rises up to 11m in middle during rainy season and

it comes down to 10.5m in summer (as in Figure 3.10b and Figure 3.11b). Whereas head values in the system with lake rises up to 14m in rain and comes down to 12m approximately in summer (as in Figure 3.10c and Figure 3.11c). See Appendix D for some other individually selected cell's head improvements.

5. To record stabilization time for the regime with lake we have extended same pattern of stress periods, recharge rate, and stress based parameters of lake for next 7 years. And results plots of lake stage and head variation of a selected cell from middle of second layers are shown in Figure 3.12.




(a) Mass balance plot without a lake in system




(b) Mass balance plot with lake

Figure 3.8: Mass balance plots of a system without lake and with lake

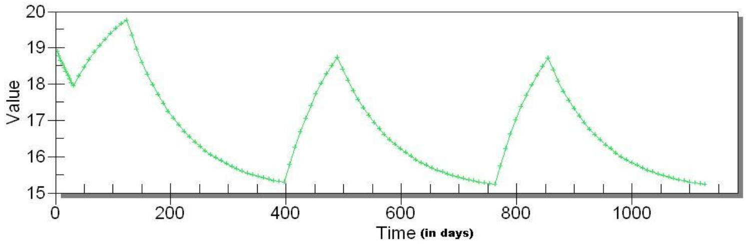
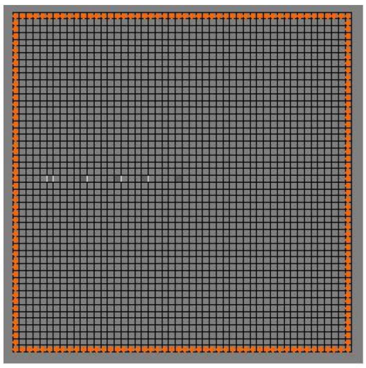
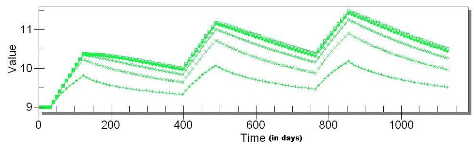


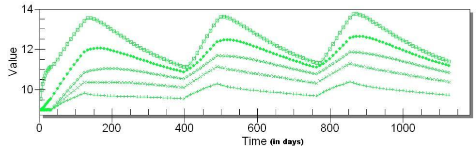

Figure 3.9: Lake water level





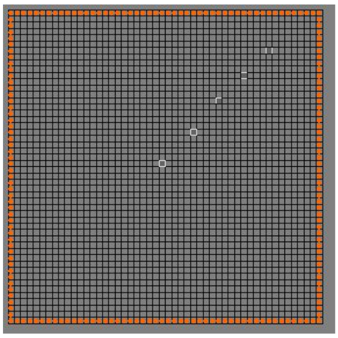

(a) Position of selected cell from middle towards west on second layer

(b) Head time series of westside selected cells without lake

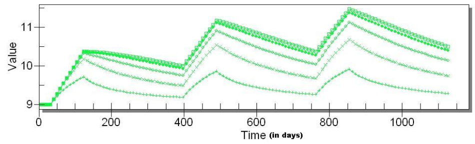



(c) Head time series of westside selected cells with lake

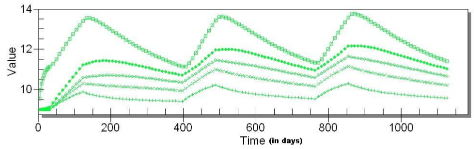
Figure 3.10: Improvement of head levels at different points



(a) Position of diagonally selected cell on second layer

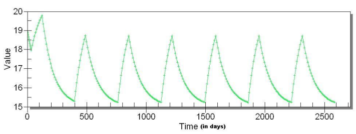



(b) Head time series of these cells without lake

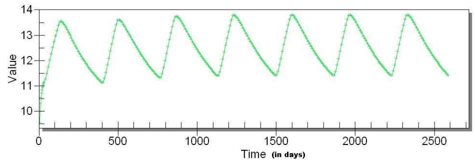



(c) Head time series of these cells with lake

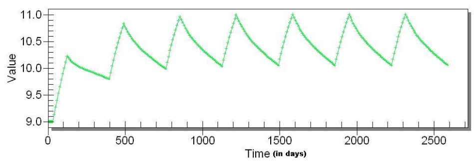
Figure 3.11: Improvement of head levels at different points



(a) Stabilization of lake stage




(b) Stabilization of head values at cell $25 \times 25$ of second layer




(c) Stabilization of head values at cell $12 \times 11$ of second layer

Figure 3.12: Stabilization of system with lake

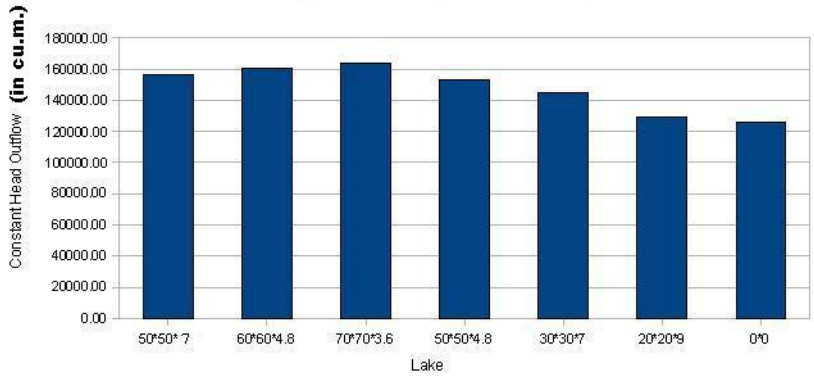
## 3.4 Experiments with Dimension of Lake

Cost of constructing lake in a regime varies with its dimension. Quantity of water recharged due to a lake as seepage may also vary with its dimension. For the purpose of cost-benefit analysis, simulation based results of lakes having different dimension may serve as a good analytical data.

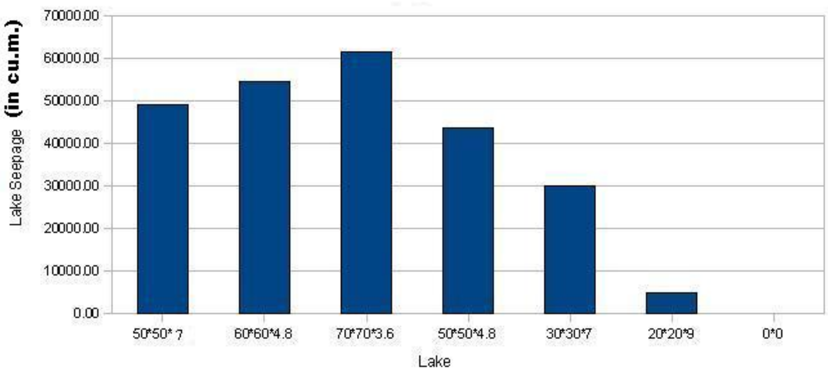
In this section, we discuss effects on regime of having lake of different dimension. At first we experimented with lakes of having same volumes (i.e. $50m \times 50m \times 7m$, $60m \times 60m \times 4.8m$ and $70m \times 70m \times 3.6m$). Then we experimented with depth (i.e. $50m \times 50m \times 7m$ and $50m \times 50m \times 4.8m$) and then with area (i.e $50 \times 50 \times 7$ and $30m \times 30m \times 7m$). In all these experiments all parameters are same. Further we have simulated a lake of having $20m \times 20m \times 9m$ which has different values for parameters of lake inflow (i.e. less runoff), so that its lake stage at any time steps could not cross topmost layers elevation. We will describe this restriction in last section of this chapter. All the plots similar to earlier results are plotted (See Appendix E) for these different sized lake. Here we have created a table as summarized results in terms of cumulative water balances of regimes (Table 3.3). We have plotted (As Figure 3.13) some charts using some of decision making data from this table.

| Dimension | Storage In | Recharge | Lake Seepage | Constant Head | Storage Out | Total In/out balance |
|---:|---:|---:|---:|---:|---:|---:|
| $50 \times 50 \times 7$ | 88397.77 | 158230.80 | 49222.59 | 156644.38 | 139207.83 | 295852.20 |
| $60 \times 60 \times 4.8$ | 87999.08 | 157320.00 | 54659.64 | 160634.52 | 139346.27 | 299980.78 |
| $70 \times 70 \times 3.6$ | 86261.09 | 156243.59 | 61663.13 | 163587.75 | 140581.67 | 304169.42 |
| $50 \times 50 \times 4.8$ | 87501.02 | 158230.80 | 43743.20 | 153443.75 | 136033.25 | 289477.00 |
| $30 \times 30 \times 7$ | 88645.55 | 159555.61 | 29953.54 | 145223.64 | 132932.73 | 278156.38 |
| $20 \times 20 \times 9$ | 90257.40 | 159969.56 | 4845.42 | 129243.31 | 125833.30 | 255076.61 |
| No lake | 90331.83 | 160300.80 | 0.00 | 250632.63 | 124563.01 | 250636.76 |

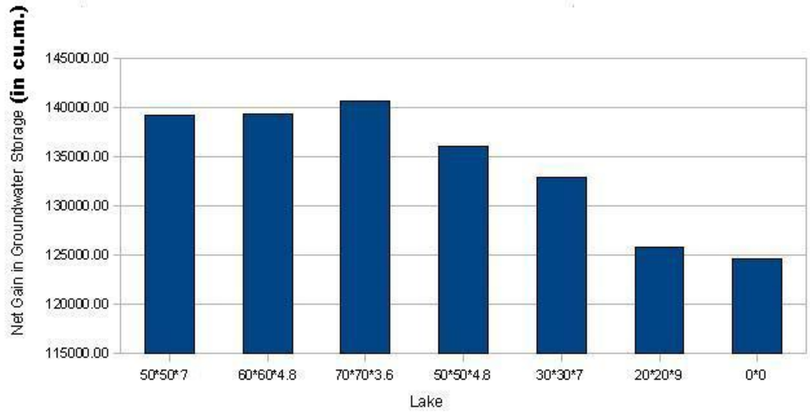
Table 3.3: Water balances of lakes of different dimensions( all figures in cu.m.)




(a) Constant head out quantiy with respect to different dimension of Lake




(b) Lake seepage quantiy with respect to different dimension of Lake



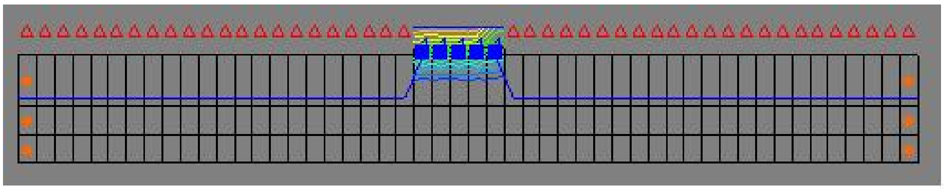

(c) Net gain in groundwater storage with respect to different dimension of Lake

Figure 3.13: Plots of different contributions flow with dimension of lakes

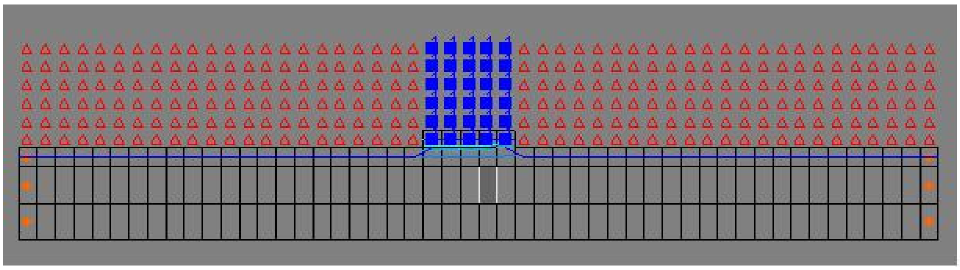
## 3.5 A technical Question: Lateral Seepage

Till now whatever examples we have done in this chapter, there are only vertical seepage is under calculation. In all above examples there is no lateral seepage but lateral seepage is always possible in real physical system. Reason behind no lateral seepage in our examples is dryness of surrounding cells (as in Figure 3.14a) of lakes over all duration. It seems unrealistic but dry cells are considered as inactive cell in modflow solver as mentioned earlier in last chapter. Therefore in all these examples there is no lateral seepage.

Here we manipulated the earlier case such that lateral seepage happens at some time steps. Regimes are same as previous but number of layers are increased to 9. Bottom two layers are the same but top layers (from elevation 8-22m) are redistributed into seven layers, each having elevation of 2m. Starting head is same as earlier 9m. Therefore, initial water table exists into lake seventh layer from top. All layers above this layers are dry initially. Now, we have formed a lake of dimension $50m \times 50 \times 12$ with same parameters of precipitation, run off but initial stage at 12m. Now crossectional view of regime is as Figure 3.14b. After running this simulation, we have a crossectional view as shown in



(a) Our previous examples

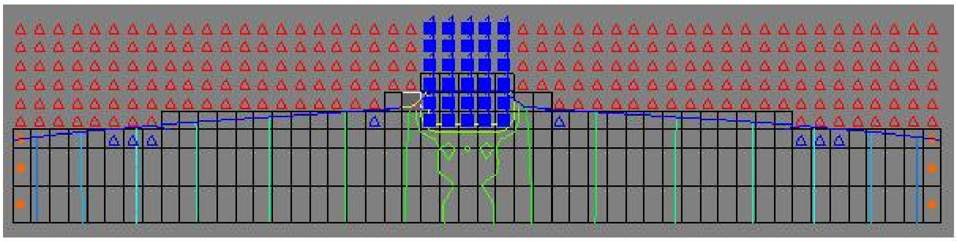


(b) Lake with more layers

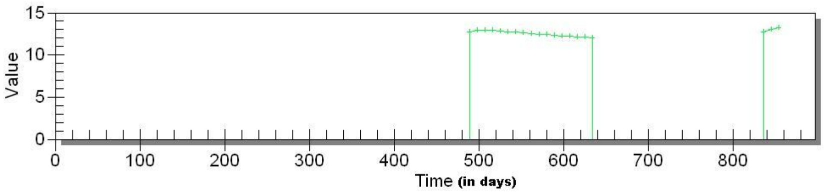
Figure 3.14: Crossectional view earlier regimes and newer with more layers

Figure 3.15a at some time steps. In this figure we see that some cells lateral to lake are active. We have selected one of these cell (marked in Figure 3.15a) and plotted its flow budget plot (Figure 3.15c). Clearly this plot shows some quantity of lake seepage when it becomes wet (As in figure 3.15b)

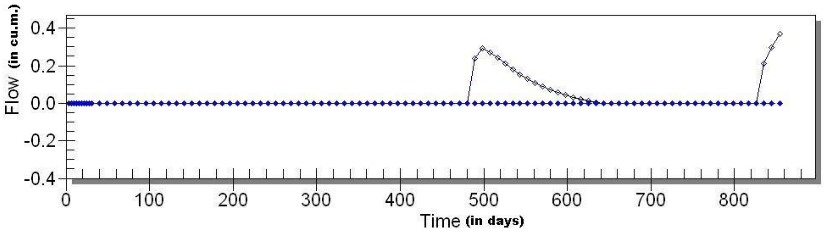
Therefore, we conclude that lateral seepage from a lake is possible only when its lateral surrounding cells becomes wet (active) or remains wet.



(a) Some of surrounding cell of lake becomes wet at some time steps during simulation; one cell of these cells is selected for plots




(b) Active time periods (i.e. wet periods) and head values of selected cell




(c) Lake seepage plots of the selected cell

Figure 3.15: Lateral seepage

## 3.6 Issues Related to Lake Modelling

Although we have managed to simulate a basic lake model in regime but many features of GMS did not work properly as expected, and many real life situation are unsupported. We discuss these issues in this section. So far, we found that simulation of lake is most problematic subdomain in groundwater modeling. Here are the list of issues:

1. **Maximum stage**: In last example of lake $50m \times 50 \times 7m$, elevation surrounding lake is 22m. This means that lake stage should not rise above this. If we change runoff parameter of lake package from 70 cu. meter per day to 200 cu. meter per day in rainy season then we are allowing more inflow to lake which can rise water stage of lake. We only need to verify whether lake goes beyond 22m or just stops increasing after 22m. Figure 3.16 shows resultant plot of lake stage after increasing lake inflow. Red line( manually drawn) in this figure shows expected maximum level of lake. Clearly, our expectation is not met here. Although there is a field for specifying maximum stage in lake package of GMS but it never works accordingly. People using GMS generally manually check when lake stage is crossing maximum stage and then they put some withdrawl from lake so that lake should not cross its maximum stage.

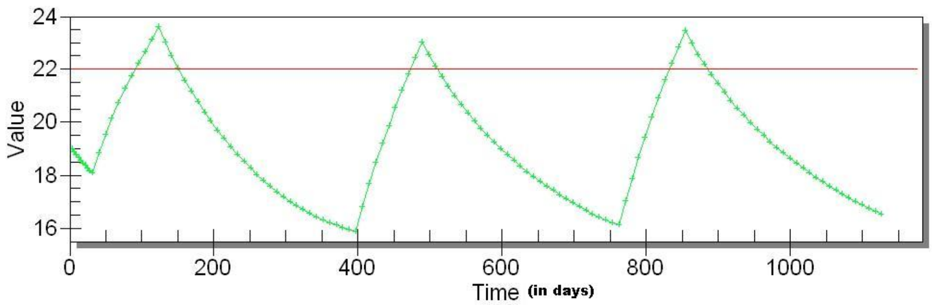



Figure 3.16: Lake water level crossing maximum stage

2. **Minimum Stage**: Minimum stage of lake is always defined by elevation of cells which are just below lake cells. It is not followed by input of minimum stage defined by lake package of GMS.

3. **Lake dry/regain**: In our example of lake $50m \times 50 \times 7m$, lake bottom elevation is 15m (as it is top elevation of layers below the lake cells). It means if lake stage goes below it then it is completely dry. We would like to mention an experimentation with example case. We have just increased two stress periods; one of two years without any rain (Lake precipitation = 0 Runoff =0 , Recharges =0) and second of three months rain (precipitation = .03m/d, recharge = 0.0003m/d , Runoff = 200cu m/d) as usual. Now it is expected that lake would become dry in newly dry stress period and would regain in rain of last stress periods. Figure 3.17 shows resultant lake stage plot of this experiments. It is clear that once lake becomes dry (i.e. if lake stage less than 15m in this case), it never regains.

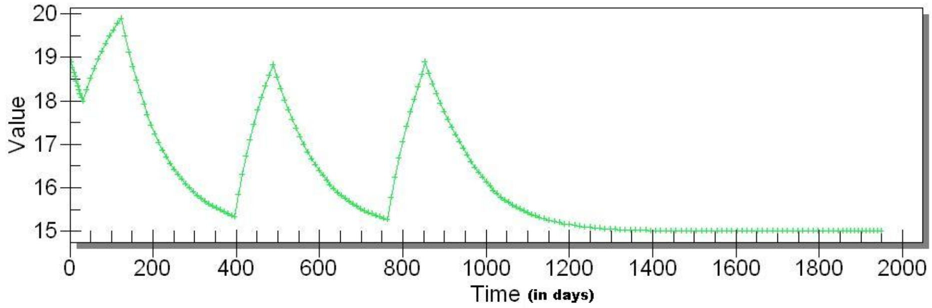



Figure 3.17: Dry lakes never regains

4. **Lake sill elevation**: Figure 3.18 explains a situation of double dam system where if one dam fills completely and reaches to its sill elevation then water overflows into a second dam which is just below it. This type of simulation is also not being possible into GMS grid approach. Overall GMS models interaction of lake and

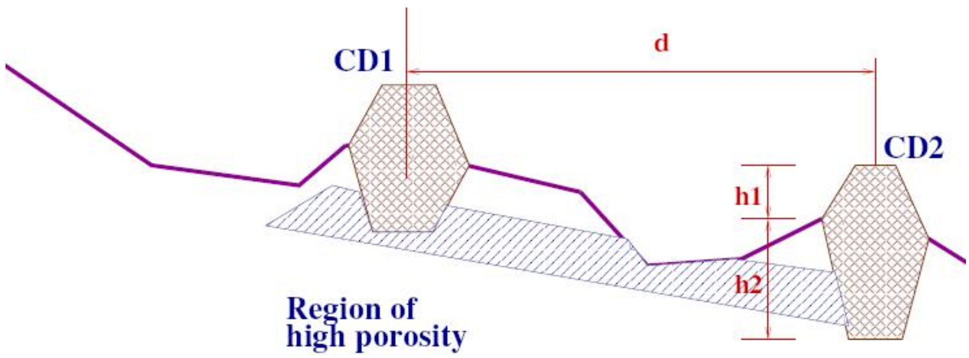



Figure 3.18: Sill elevation problem [8]

groundwater regime but it does not model lake stages properly. Simulation of lake stage is controlled manually in this tool. And situation like double dam system can not be modelled due to its inabilities of incorporating surface water flow.

# Chapter 4

# Wells, Drains and Other Examples

Groundwater is generally extracted, either for domestic use or for irrigation, using wells. These may be either surface or bore, and with or without pumps. In this chapter, we explain how wells may be simulated in GMS and its features and limitations.

The usual objectives for well simulation could be

1. to understand the carrying capacity of a region as far as water demand goes.
2. to understand the effect of an intervention (such as a percolation tank) on the carrying capacity.

A typical well, serving about 30 households would cause a well discharge of about 30 cubic meters per day. We use this or similar figures for our sample simulations. The irrigation use of well water is far more varied. Roughly 1500 cubic meters per acre are used for a single crop.

The chapter has four parts. We first quickly review the theory and explain the key technical terms. Next, we do an elementary simulation and analyze the output, stressing on the features and limitations. After that, we do a comparative simulation which illustrates a typical prescriptive use of simulation.

Last part of the chapter presents an example of a relatively complex and more realistic regime of groundwater flow. This parts attempts to model a sequential structure, in this a lake with an underground high conductivity channel is first investigated and then we consider a plug for the channel and its effect on the properties of the lake. Meanwhile we have attempted to get more corrected simulation for this case using drain specification. The use of drain is also explained. This example is just a typical test case for further developments. Although we have not been able to elaborate this example completely but this will give a visualization of complex cases of groundwater flow.

## 4.1 Well Hydraulics in Literature

Most of the wells in India are in unconfined aquifer region. Here we describe drawdown relationship for a steady radial flow towards a well in an unconfined aquifer. Pumping

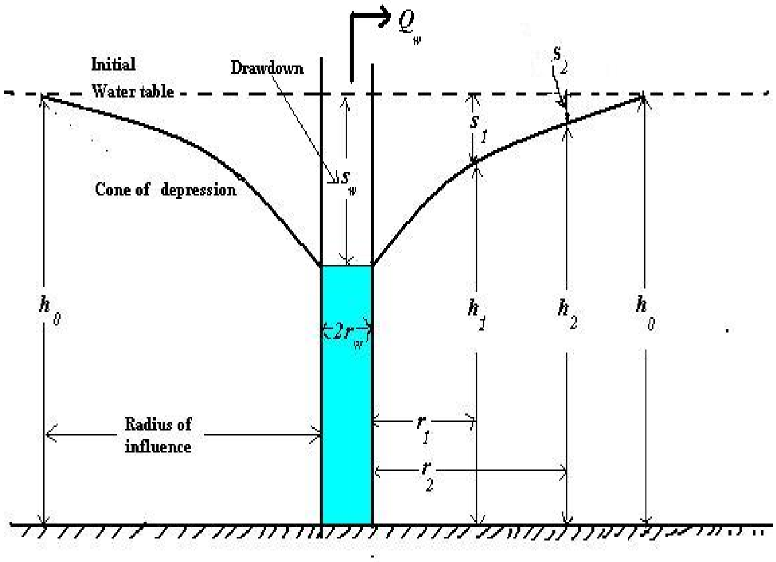



Figure 4.1: Steady state well drawdown in a unconfined aquifer [13]

of groundwater is always accompanied by depletion of water table, This is referred to drawdown. Drawdown is maximum at well face and decreases as the distances from well increases and becomes almost zero at large distances. The distance at which drawdown becomes zero is called radius of influence of pumping well. This distance varies depending upon pumping rate, duration of pumping rate. Drawdown creates a hydraulic gradient in the direction of well due to which the groundwater flows towards the well in the radial directions.

For theoretical analysis the usual assumptions are [13]:

1. The aquifer is infinite in extent.
2. The aquifer is homogeneous, isotropic.
3. The well has constant discharge rate.
4. At the beginning of pumping the the water table is horizontal.
5. The well fully penetrates the saturated thickness so that flow towards the well is horizontal.
6. Darcy Law is applicable.

For a steady sate well, the discharge flow across any cylindrical surface is constant [13], i.e. $\dfrac{\partial Q_w}{\partial r} = 0$ where $Q_w = 2\pi r h k \dfrac{\partial h}{\partial r}$

Thus, we can write

$$\frac{\partial}{\partial r}[2\pi k h \frac{\partial h}{\partial r}] = 0 \tag{4.1}$$

or a more simplified differential equation is

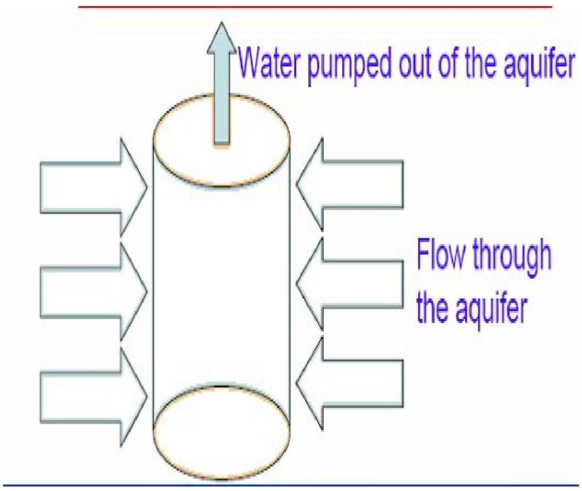



Figure 4.2: Cylindrical view of flow [9]

$$\frac{\partial^2 h^2}{\partial r^2} + \frac{1}{r}\frac{\partial^2 h^2}{\partial r} = 0 \tag{4.2}$$

Equation 4.1 is the governing equation for steady radial flow in homogeneous isotropic unconfined aquifer with no source or sink.

By integrating this equation within the limits from $r_e$ to $r_w$ where the head values are $h_0$ and $h_w$ respectively, gives

$$h_0^2 - h_w^2 = \frac{Q_w}{\pi k}[\ln \frac{r_e}{r_w}] \tag{4.3}$$

If $s_1$ and $s_2$ are the drawdown at two observation wells at a distance of $r_1$ and $r_2$ from well respectively, and $h_1$ and $h_2$ are the values of the groundwater head at these wells then Equation 4.1 can be written as

$$h_1^2 - h_2^2 = \frac{Q_w}{\pi k}[\ln \frac{r_2}{r_1}] \tag{4.4}$$

Further, drawdown can be written as $s_1 = h_0 - h_1$ and $s_2 = h0 - h2$.

Similarly, transient state equation for pumping well in unconfined aquifer is given as ,

$$\frac{\partial^2 h}{\partial r^2} + \frac{1}{r}\frac{\partial h^2}{\partial r} = \frac{S_y}{kh}\frac{\partial h^2}{\partial t} \tag{4.5}$$

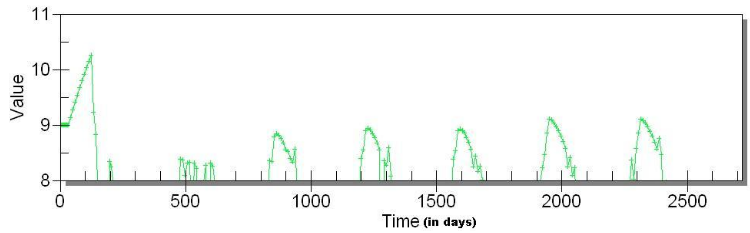
where $S_y$ is specific yield .

## 4.2 Well Package in MODFLOW and GMS

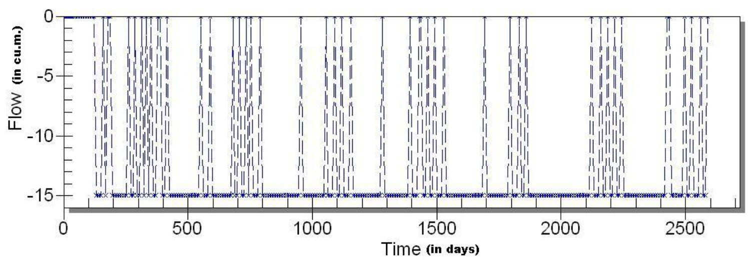
The Well (WEL) Package simulates features such as wells that withdraw water from or add water to the aquifer at a constant rate during a stress period, where the rate is independent of both the cell area and the head in the cell. The flow rate for a well is specified by the user as a fluid volume per unit time at which water is added to the aquifer. Negative values of $Q$ are used to indicate well discharge (pumping), whereas positive values of Q indicate a recharging well. The Package does not directly accommodate wells that are open to more than one layer of the model. Most well of this type and they can be represented as a group of single-cell wells, each open to one of the layers tapped by the multilayer well, and each having an individual Q term specified for each stress period. If this approach is used, the recharge or discharge of the multilayer well must be divided or apportioned in some way among the individual layers, externally to the model program. A common method of doing this is to divide the well discharge in proportion to the layer Transmissivity. When the cell having a well becomes dry then no discharge is possible. This type of well can be again active for discharge when this cell becomes wet again during the simulation.

**Results Plots:**

In addition to the earlier plots, here we can plot the water level of well. This plot (as shown in Figure 4.3a) is exactly the head plot of the cell where well is situated. When this cell (i.e. well) becomes dry then its head value hits its bottom elevation. We can plot a graph which shows amount of extracted water from well (as in Figure 4.3b). We can match the dry periods and zero water extraction in both plots. Countour surrounding wells becomes as Figure 4.4 due to extraction.




(a) Water level in a well at cell having bottom elevation 8m




(b) Water extaction plot of well

Figure 4.3: Plots in case of a well in given regime

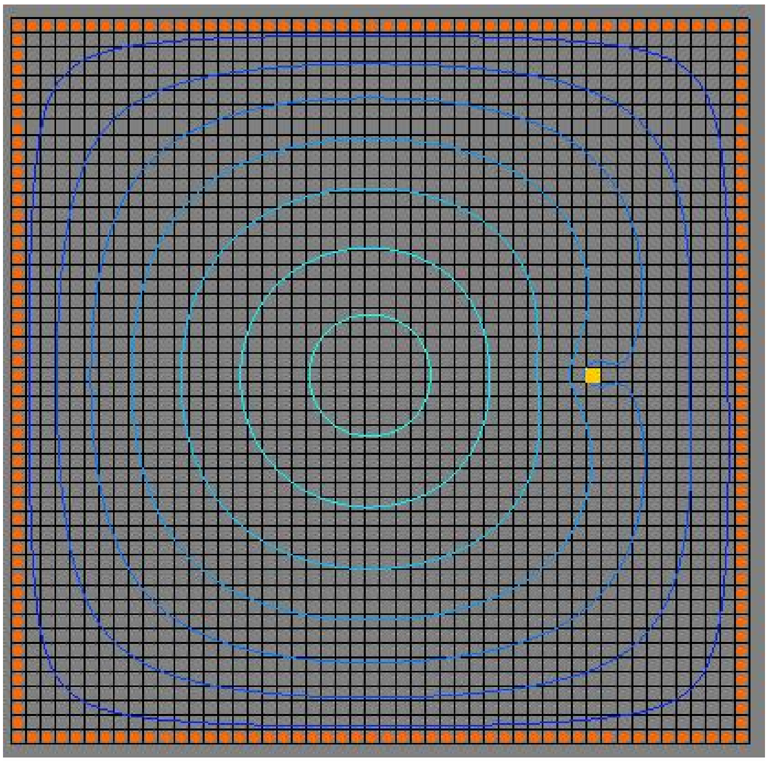


Figure 4.4: Contour in case of a well in regime

## 4.3 Simulation Example

Here, we have taken same regime as the Chapter 3 and added a well at position (row number = 25, column number = 25) in second layer (as Figure 4.5). Stress periods of this example is increased to 15 and pattern of cyclic rain every year as earlier. Pumping rate for different stress periods is the only input required for putting a well in a groundwater flow model. In this example, pumping rate is -15 cubic meter per day (-sign for extraction) in all stress periods excepts first two stresses.

We have studied this same example of well with lake (as simulation example of Chapter 3) in this regime. Figure 4.6 shows water level of wells with/without lake in the regime. In this figure whenever head plot hits 8m, the cell is considered dry (because cell bottom elevation is 8m) and extraction from well stops. Figure 4.7 shows amount of water extraction with respect to time. Using these plots, it is clear that we can extract more water from a well which is in a regime with lake (i.e. comparatively less dry periods of well). We have experimented this whole setup again with a changed pumping rate of -13 cubic meter per day. Figure 4.8 and 4.9 are same set of plots with pumping rate -13 cubic meter per day. In this case dry periods of wells are for lesser time compared to earlier set of plots of pumping rate -15 cubic meter per day.Figure 4.10 shows cummulative extractions of all the four cases.

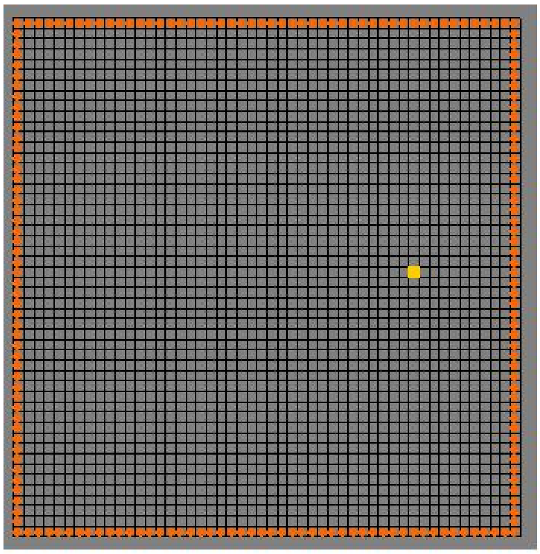
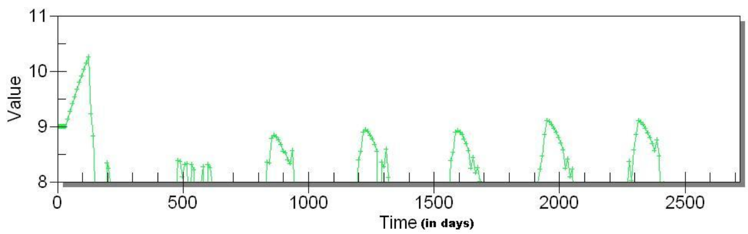
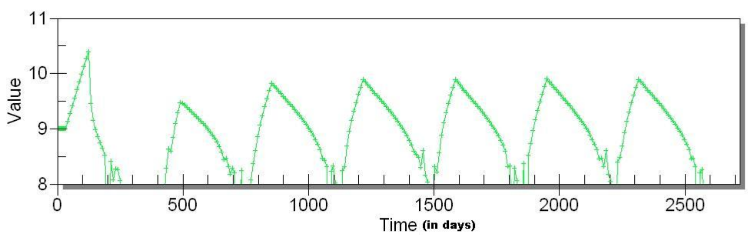


Figure 4.5: Well position in second layer

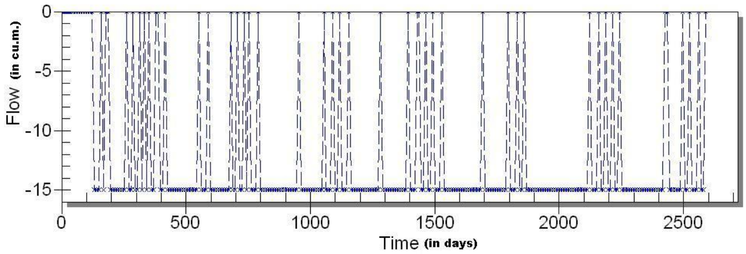



(a) Water level in well without lake in regime

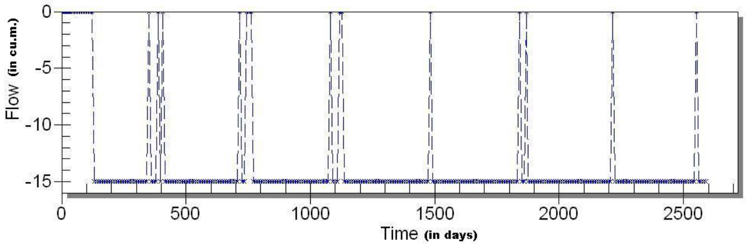



(b) Water level in well with lake in regime

Figure 4.6: Water level in well with pumping rate of 15 cubic meter per day

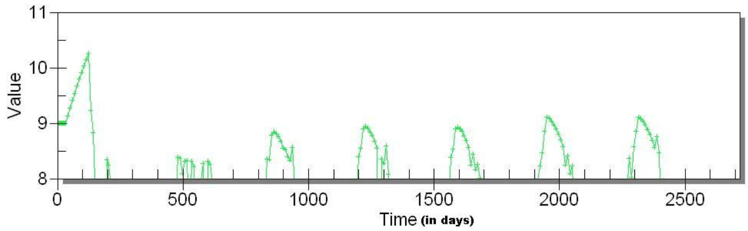


(a) Water extraction plot of well without lake in regime

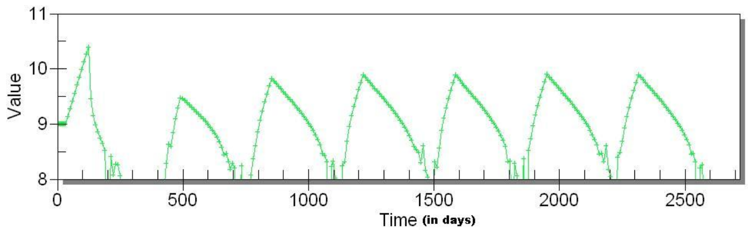


(b) Water extraction of well with lake in regime

Figure 4.7: Water extraction in well with pumping rate of 15 cubic meter per day

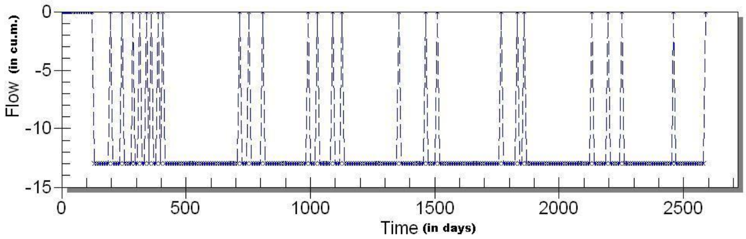


(a) Water level in well without lake in regime

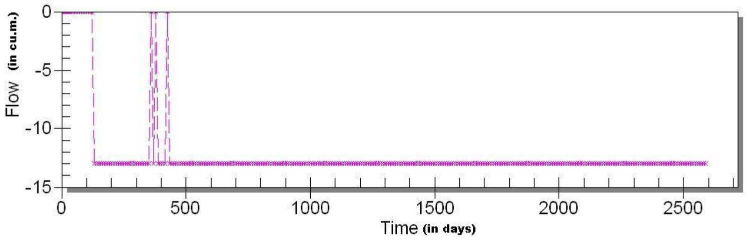


(b) Water level in well with lake in regime

Figure 4.8: Water level in well with pumping rate of 13 cubic meter per day)




(a) Water extraction plot of well without lake in regime




(b) Water extraction of well with lake in regime

Figure 4.9: Water extraction plot of well with pumping rate of 15 cubic meter per day)

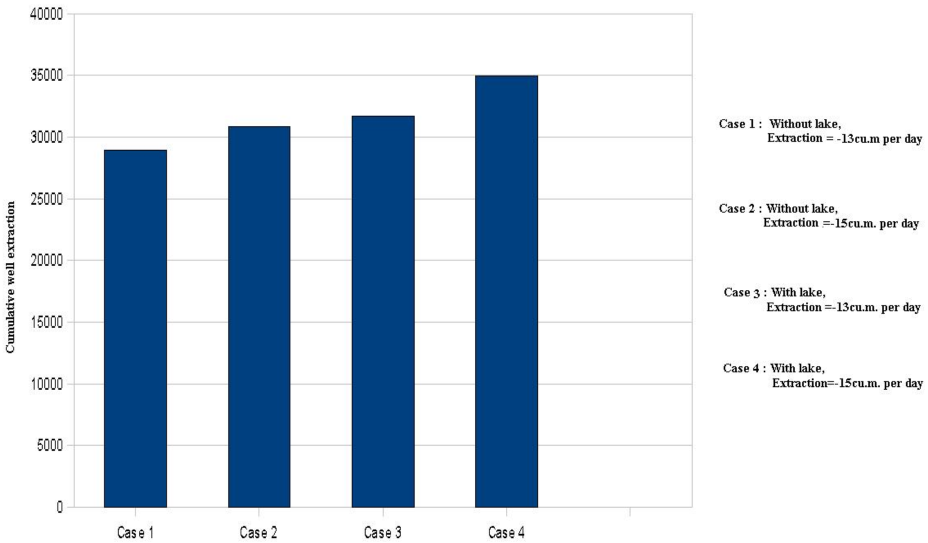



Figure 4.10: Cumulative well extractions (in cu.m.)

## 4.4 A Relatively Complex Regime to Simulate

In this section we describes a regime almost similar to Figure 4.11. Although this double dam system cannot be modelled as mentioned in Section 3.6. We have tried to simulate this regime with slopping topology and having one dam only.

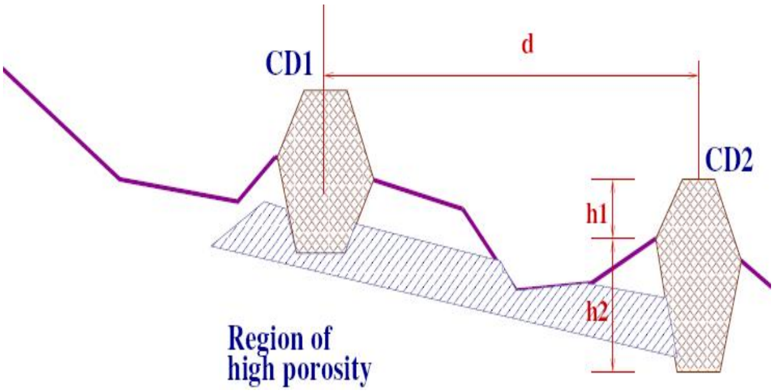



Figure 4.11: Double dam system [8]

### Regime Description

:

All input parameters are same as simulation example described in Section 2.1 except elevation of top layer and hydraulic conductivity zones. In addition, we have defined a reservoir in the middle of this regime using lake package. The parameters required to redefine are:

1. **Top elevation of top layer cells**: We have to design a slopping topology at top layer (as in Figure 4.12). Top elevation of left column's cell are set to 22m and as we go towards right from left, elevation of each columns cells are decreased by 0.25m. Thus we decrease our top elevation from 22m to 11m (from left to right), and then elevation of remaining cells are set to 11m.

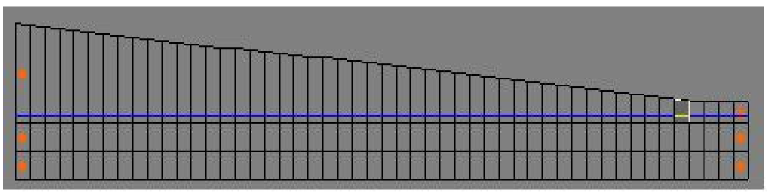


Figure 4.12: Sloping topology

2. **Conductivity**: Layer conductivity is same as table 2.1 except few changes as described in further inputs.

3. **Lake Parameters**: A lake is defined by using $7 \times 7$ cells in the middle of regime as in Figure 4.13. Lake bottom elevation is 8m and its initial water stage is 10.5m. All other parameters of lake except runoff (we set it as 200cu. m. per day) are same as lake parameters of example described in Section 3.2.

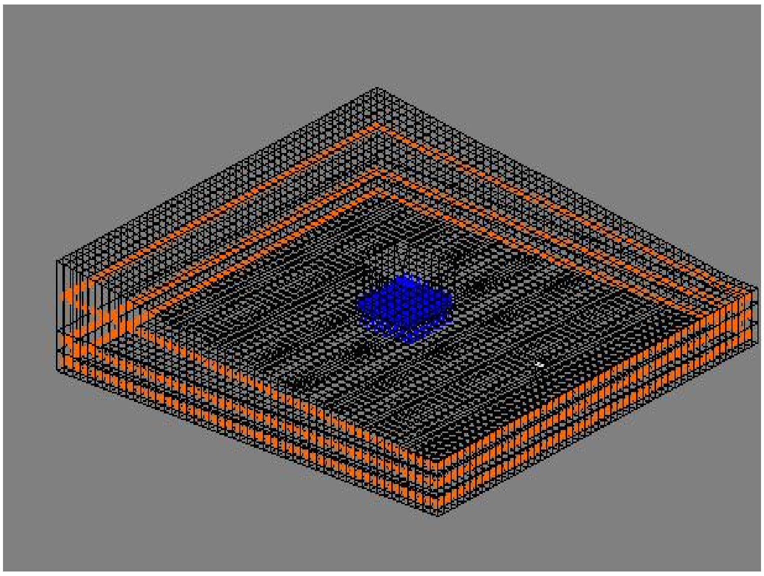


Figure 4.13: Three dimensional view of regime and a lake in the middle of regime

4. **High Conductivity Channel**: Here we are designing a regime similar to Figure 4.11. To model a high porous channel underneath of lake, we have selected some cells (as in Figure 4.14) from second layer and set their hydraulic conductivities as 10m/d.

5. **Creating a Barrier like construction in First layer**: We created a manmade barrier in first layer as positioned in Figure 4.15 with low Hydraulic conductivity values(i.e. 0.0000009m/d).

When we simulate this model, head values in a cell of low conductivity zone shoots up at some time steps as in Figure 4.16. Its head plot one of these cells is given as Figure 4.17a.

For experimentation we have extended our barrier of first layer up to second layer with same low conductivity. Now high conductivity zone of second layer is plugged in between. Now head plot is as Figure 4.17b. It is relatively lower this time. Clearly it also shoots up but not as much high as earlier.

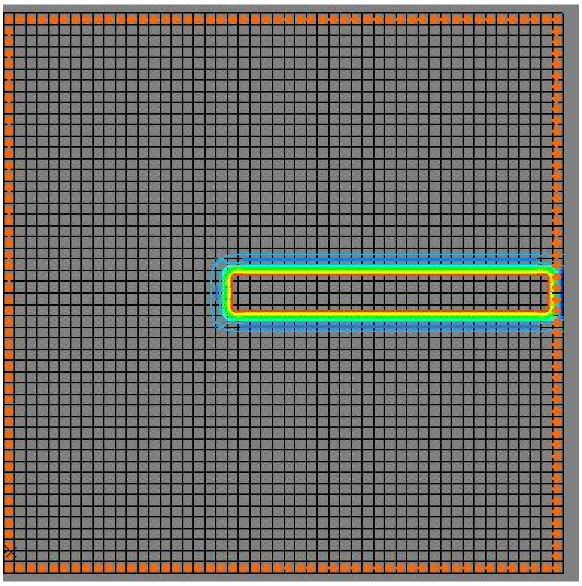


Figure 4.14: High conductivity channel of second layer

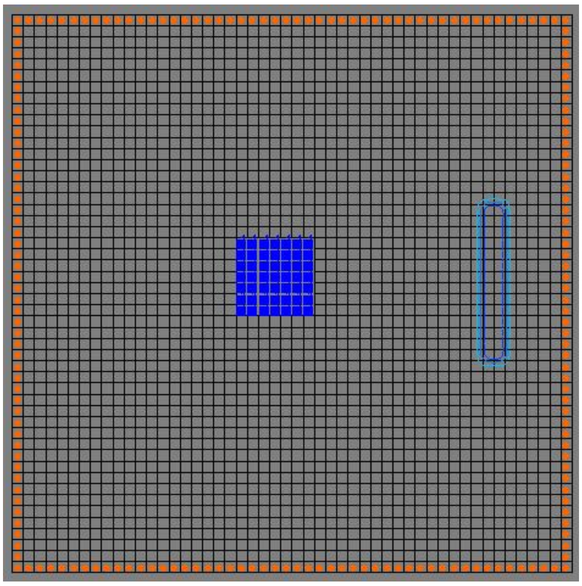


Figure 4.15: A man made barrier modeled using very low conductivity zone on first layer

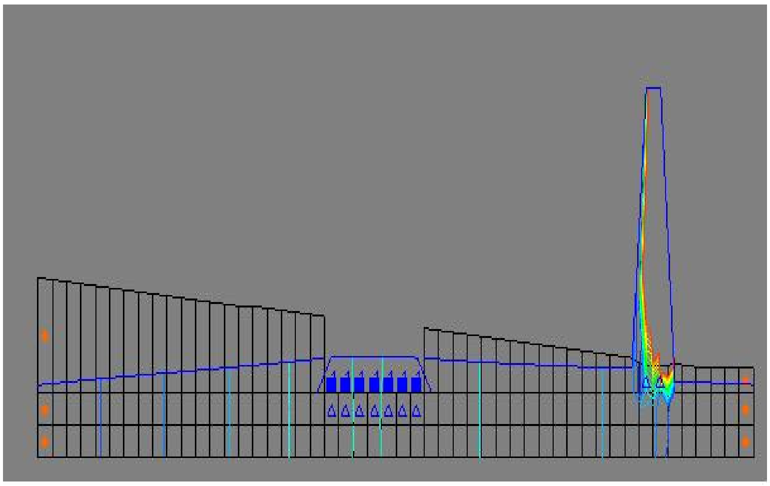
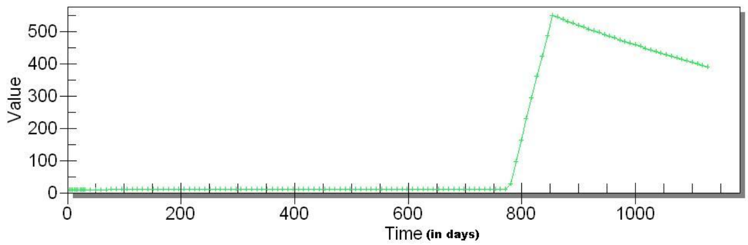
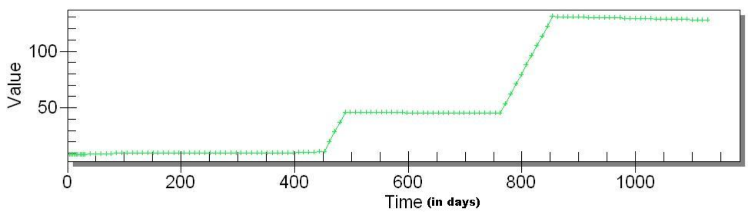


Figure 4.16: Water table at some time steps




(a) Head Plot of a cell from barrier of first layer




(b) Head Plot of same cell after extending barrier up to second layer

Figure 4.17: Head plots of a cell from low conductivity zone of first layer

## 4.5 Drain Specification

The Drain (DRN) Package is designed to simulate the effects of features such as agricultural drains, which remove water from the aquifer at a rate proportional to the difference between the head in the aquifer and some fixed head or elevation, called the drain elevation, so long as the head in the aquifer is above that elevation. If, however, the aquifer head falls below the drain elevation, then the drain has no effect on the aquifer. when $h_{i,j,k} > HD$

$$Q_{out} = CD(h_{i,j,k} - HD) \tag{4.6}$$

and when $h_{i,j,k} \leq HD$

$$Q_{out} = 0 \tag{4.7}$$

where $Q_{out}$ is the flow from the aquifer into the drain ;$CD$ is the drain conductance, $HD$ is the drain elevation; So drain is a type of IF..ELSE boundary conditions.

After extending our barrier up to second layer, we have experimented with drains on the same cells of barriers on first layer (As in Figure 4.18). In this experiment drain has elevation of 11m and conductance of 3. It means that whenever head in drain cells crosses beyond 11m, it removes water from the cell at a proportion of difference between head value and its elevation, 11m.

Now head plots of earlier cell is as Figure 4.19a. Amount of water coming out thorough drain is plotted in Figure 4.19b.

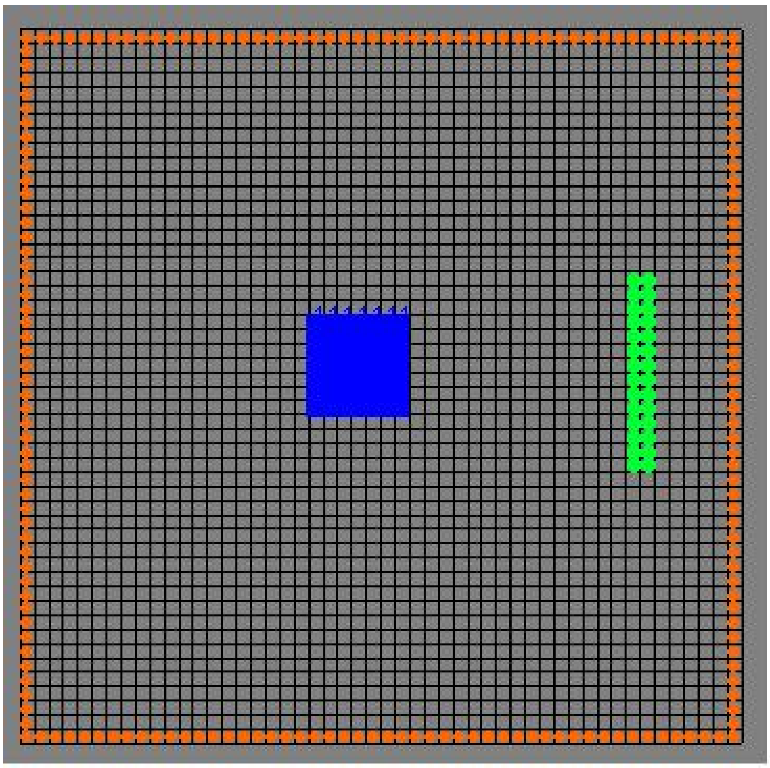
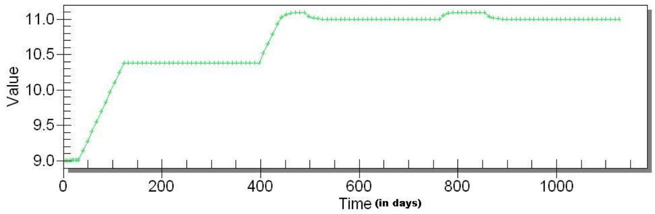
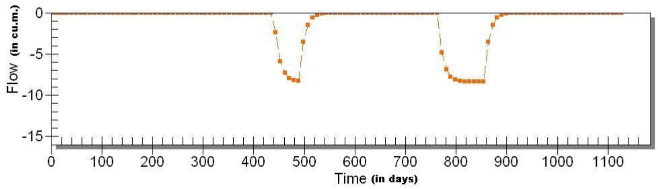


Figure 4.18: Drain cells on first layer



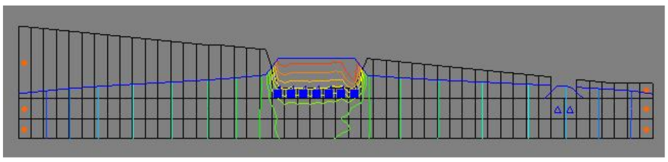

(a) Head Plot of a cell from barrier of first layer



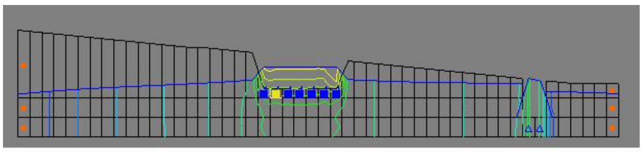

(b) Head Plot of the cell creating drains on fist layer

Figure 4.19: Head plots of a cell from low conductivity zone of first layer

When as an attempt to solve rising head value problem, we made the cells of barriers from first layer as inactive cell. Later we made barrier cells from second layer too. The resultant contour did not raise as high as earlier (as in Figure 4.20) but head values are almost same at other cells.



(a) Water table at some time steps when cells from first layer, forming barriers are made inactive

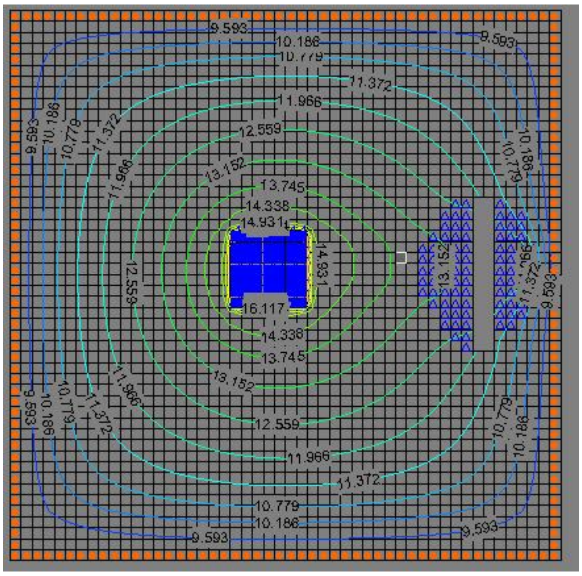


(b) Water table at some time steps when cells from first and second layers, forming barriers are made inactive

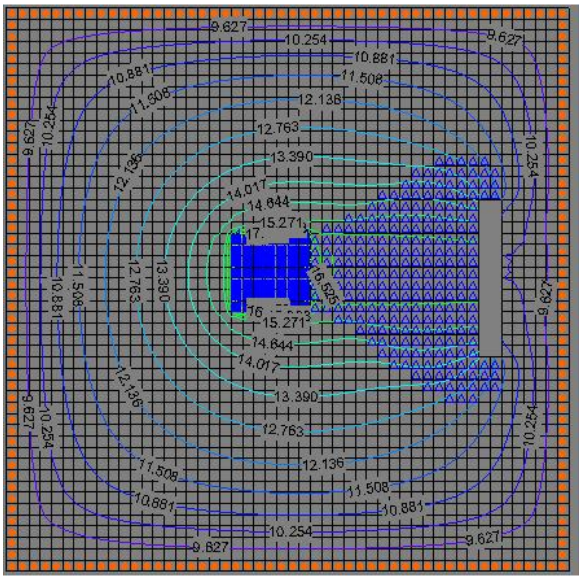
Figure 4.20: An another approach

In an another attempt of experimentation, we have made conductivity of first layer and second layer (except channel), 0.5m/d in place of 2m/d. In this case we get some difference in head values for both case (i.e. absence and presence of barrier in second layer). Figure 4.21 shows equipotential lines of both cases at time steps 22nd August, 2009. Till now, we can say that extending of barrier up to second layer will lead higher head values if surrounding domain is of low conductiviy.

Although we have attempted this model using different perspective but this simulation is a subject of future discussion and experimentation.




(a) When cells from first layer, forming barriers are made inactive




(b) When cells from first and second layer, forming barriers are made inactive

Figure 4.21: Equipotential lines at time step (22nd August, 2009) of first layer after simulation with low conductivity

# Chapter 5

# Conclusion and Future Work

In this final chapter, we summarize related issues, problems and possible directions for future work. In this report we presented our attempts of groundwater simulation examples and during the process we specified the way groundwater modelling is done. We used GMS interface to use MODFLOW for this purpose. We listed input requirements and obtained outputs for modelling of groundwater regime. We experimented with various parameters involved in groundwater flow. We presented a extracted specifications with the help of case examples and developed a test suite which will drive the development of an independent low-cost public domain simulator.

We explored limitations and issues during simulation of these examples. Related issues are already described in last section of each chapter. Overall, we conclude that lake modelling is not as realistic as we would like in GMS. Problems such as overflowing of lake, drying lake and sill elevation of lake are of serious concerns. We are not able to present more accurate and appropriate simulation due to these problems and issues. At first these problems need to be fixed for further development in this project domain. It requires some code writing in associated lake lackage of MODFLOW. All the examples and issues presented in this report should be tested for better modelling after such code manipulation.

We have also studied wells and drains and their simulation. Together with lake, These form a useful set of primitives by which a complex simulation may be done. This was attempted in Chapter 4. What remains to be done is a lot.

GMS has two modes of operations:

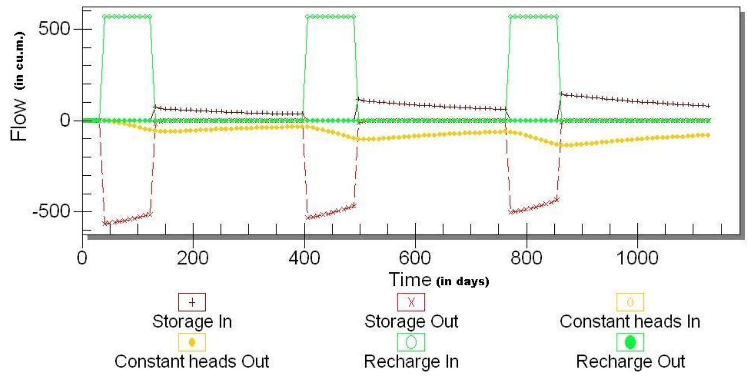
1. Grid approach, which is closer to MODFLOW.

2. Conceptual approach, which uses tools from GIS etc. as an interface. This approach produces a set up of grid approach as its output to run MODFLOW solver.

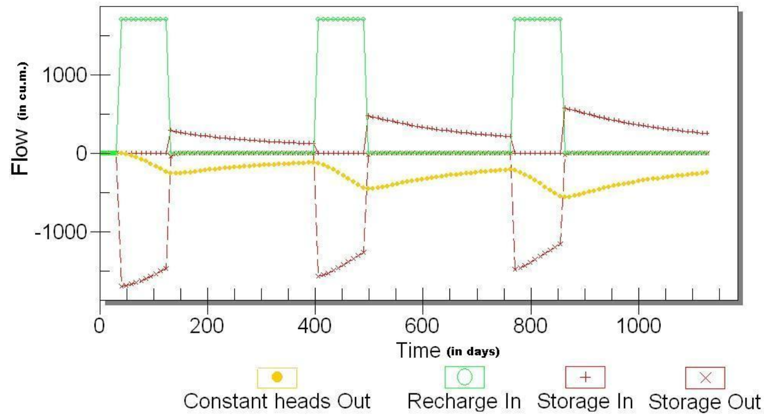
We have mainly concentrated on grid approach as this is more primitive core to a simulation.

# Appendix A

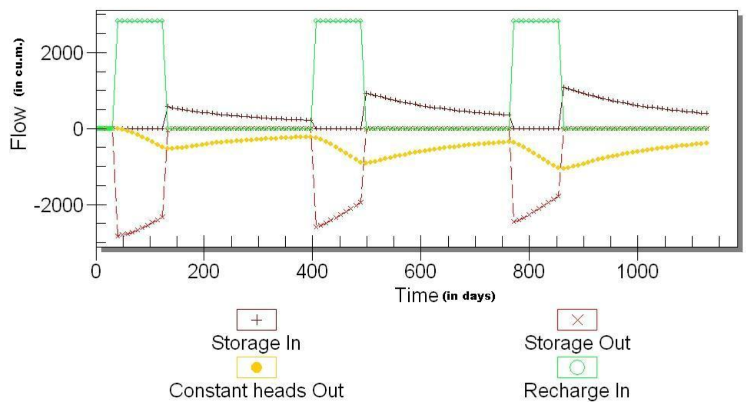
# Plots of Input Set 2




(a) Mass balance plot when recharge rate is 0.003m/d

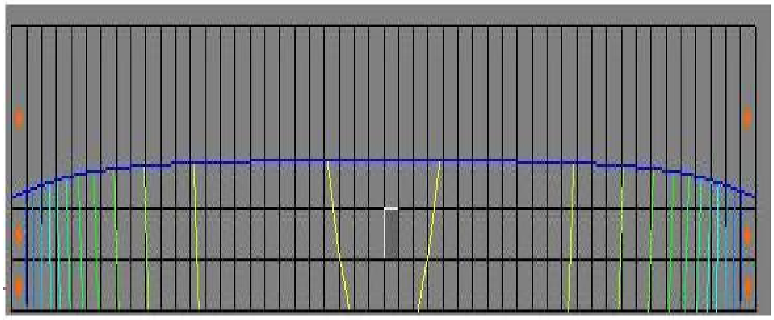



(b) Mass balance plot when recharge rate is 0.009m/d

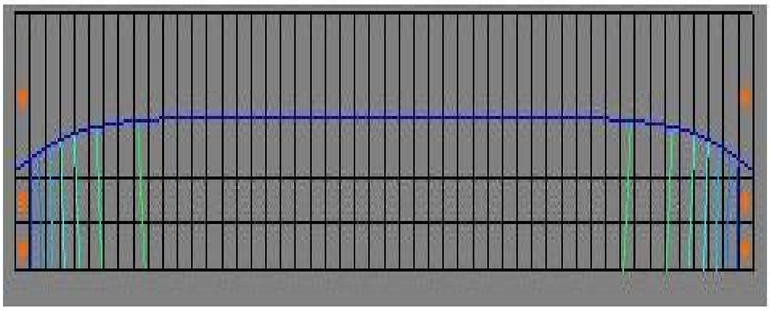



(c) Mass balance plot when recharge rate is 0.015m/d

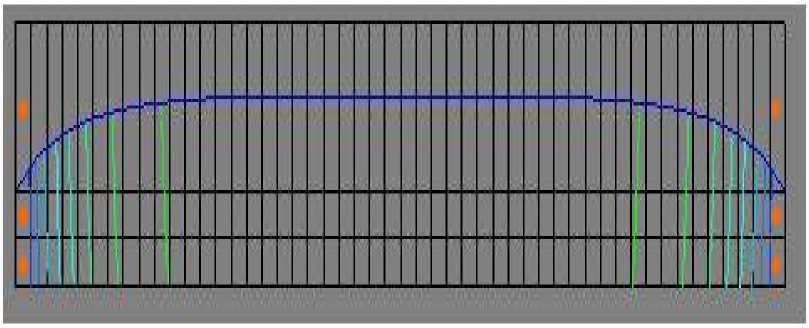
Figure A.1: Mass balance plots with different recharge rate



(a) Water table when recharge rate is 0.003m/d

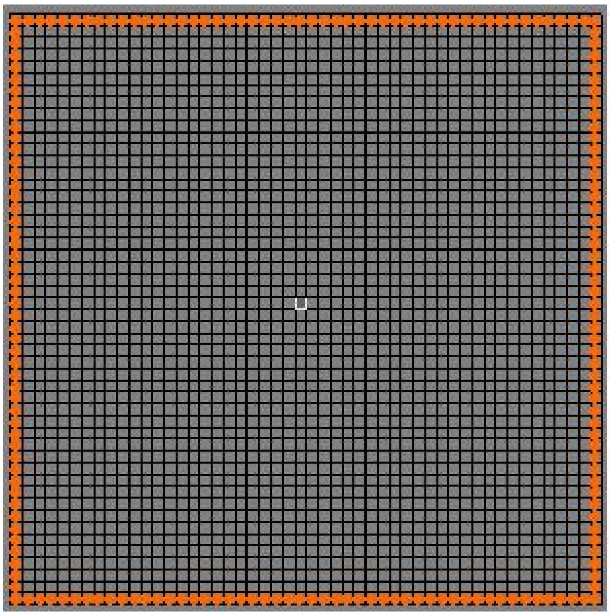


(b) Water table when recharge rate is 0.009m/d

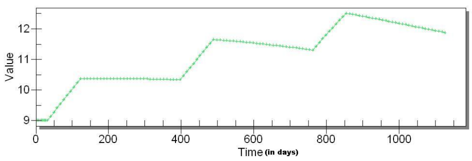


(c) Water table when recharge rate is 0.015m/d

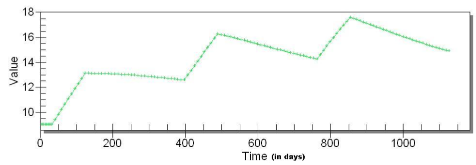
Figure A.2: Water table at the end of rain(i.e. 31st Aug, 2007) with different recharge rate



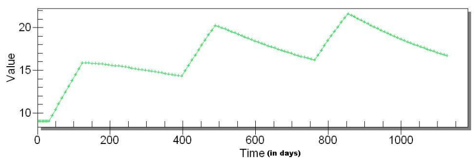
(a) Position of selected cell at $25 \times 25$, Id:= 1225




(b) Head time series of this cell when recharge rate is 0.003m/d

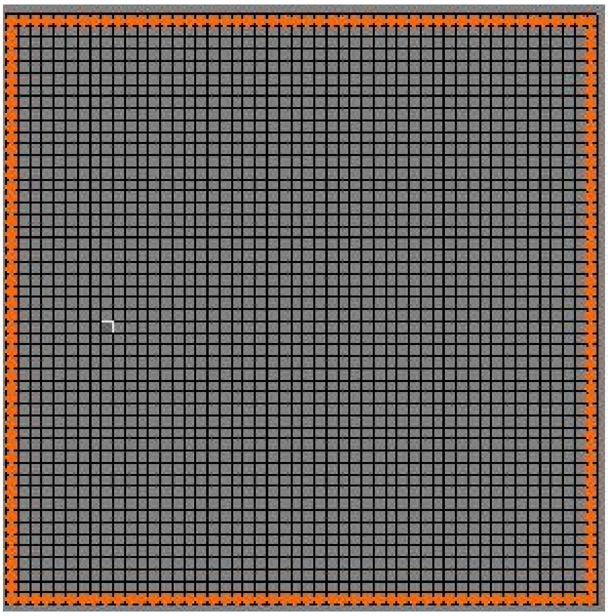



(c) Head time series of of this cell when recharge rate is 0.009m/d

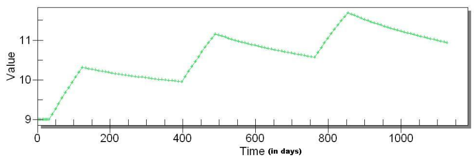



(d) Head time series of of this cell when recharge rate is 0.015m/d

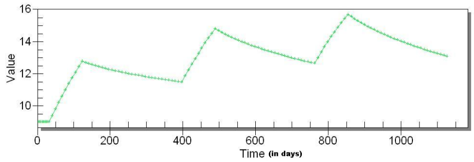
Figure A.3: Head vs Time plot of selected cell (cell id = 1225) at different recharge rate



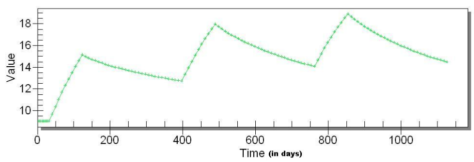
(a) Position of selected cell at $27 \times 9$, Id:= 1309




(b) Head time series of this cell when recharge rate is 0.003m/d

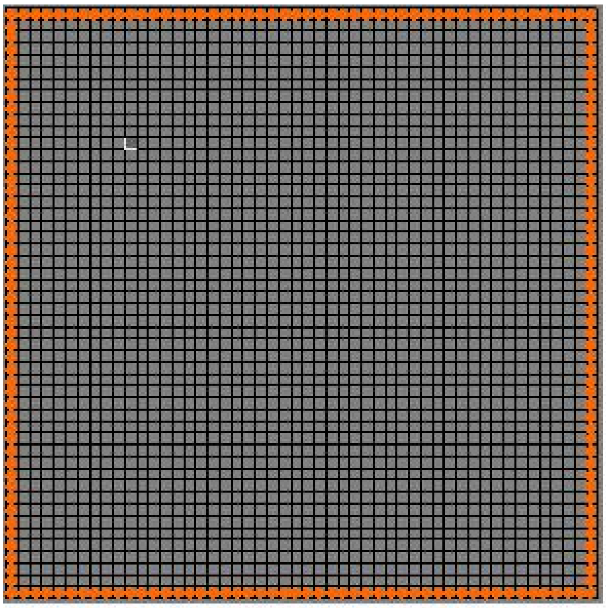



(c) Head time series of of this cell when recharge rate is 0.009m/d

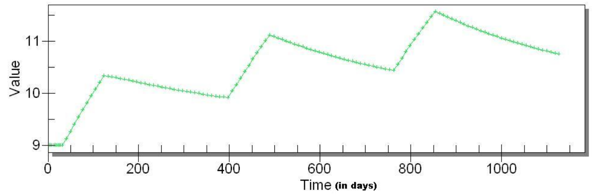



(d) Head time series of of this cell when recharge rate is 0.015m/d

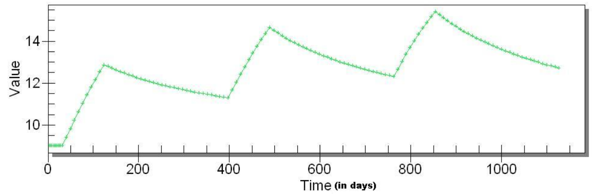
Figure A.4: Head vs Time plot of selected cell (cell id = 1309) at different recharge rate



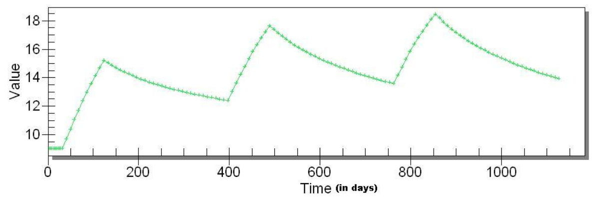
(a) Position of selected cell at 12 × 11, Id:= 561P




(b) Head time series of this cell when recharge rate is 0.003m/d

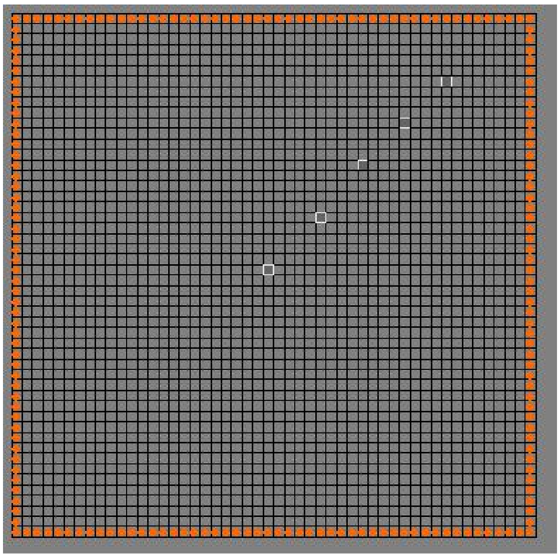



(c) Head time series of of this cell when recharge rate is 0.009m/d

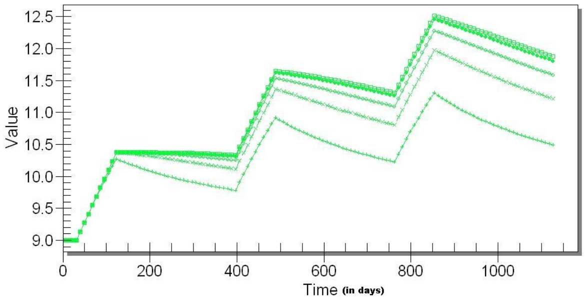



(d) Head time series of of this cell when recharge rate is 0.015m/d

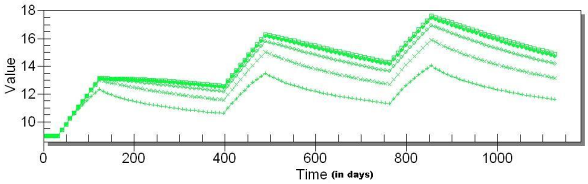
Figure A.5: Head vs Time plot of selected cell (cell id = 561) at different recharge rate



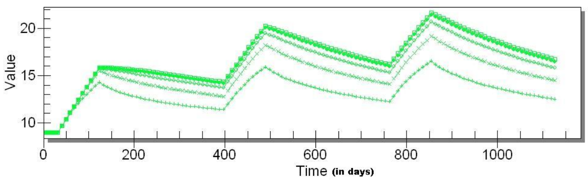
(a) A set of selected cell almost on diagonal




(b) Head time series of this cell when recharge rate is 0.003m/d

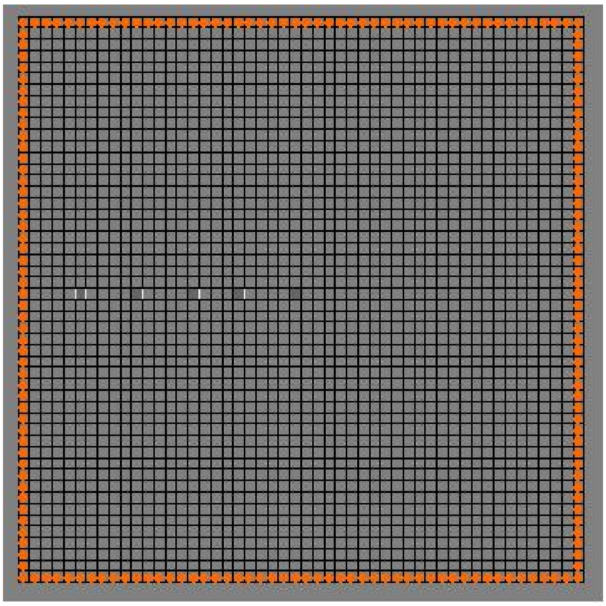



(c) Head time series of of this cell when recharge rate is 0.009m/d

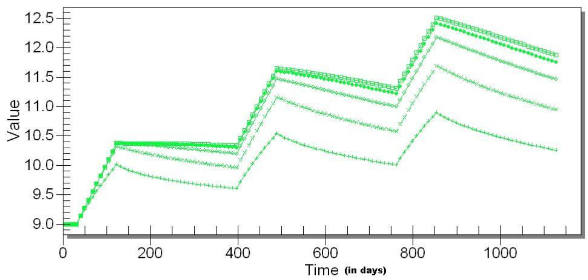



(d) Head time series of of this cell when recharge rate is 0.015m/d

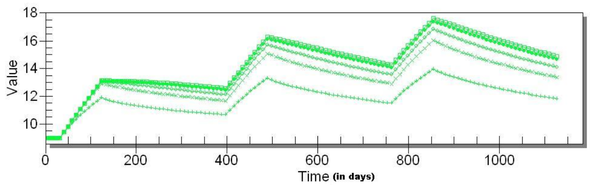
Figure A.6: Head vs Time plot of a group of selected cell at different recharge rate



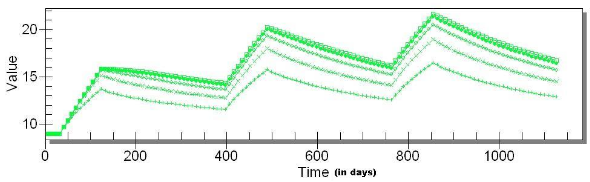
(a) A set of selected cell from middle cell towards west




(b) Head time series of this cell when recharge rate is 0.003m/d




(c) Head time series of of this cell when recharge rate is 0.009m/d

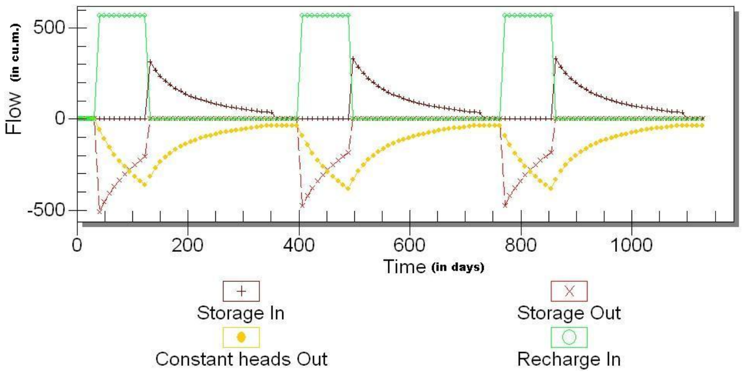



(d) Head time series of of this cell when recharge rate is 0.015m/d

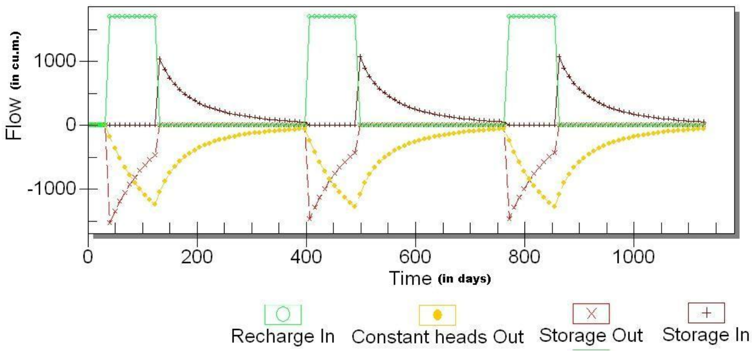
Figure A.7: Head vs Time plot of another group of selected cell at different recharge rate

# Appendix B

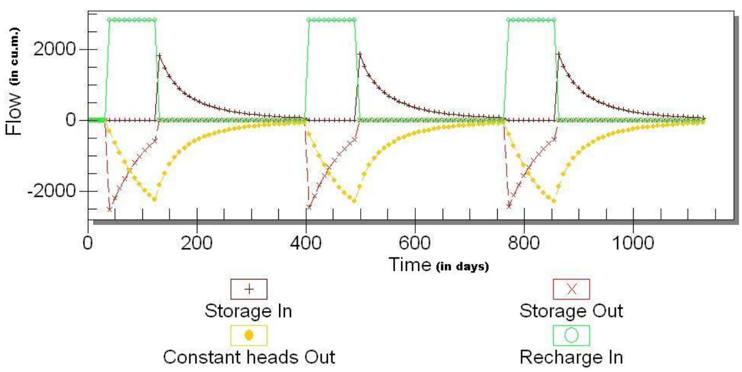
# Plots of Input Set 3




(a) Mass balance plot when recharge rate is 0.003m/d

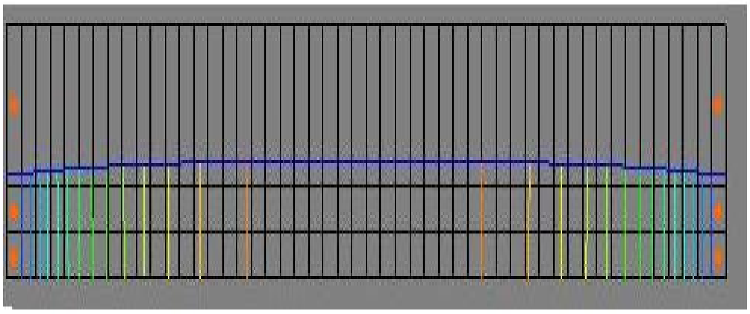



(b) Mass balance plot when recharge rate is 0.009m/d

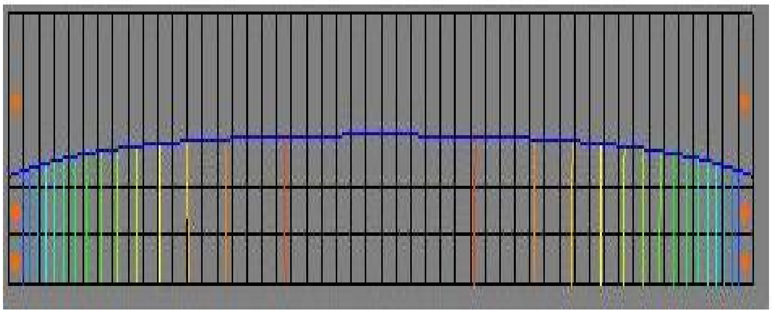



(c) Mass balance plot when recharge rate is 0.015m/d

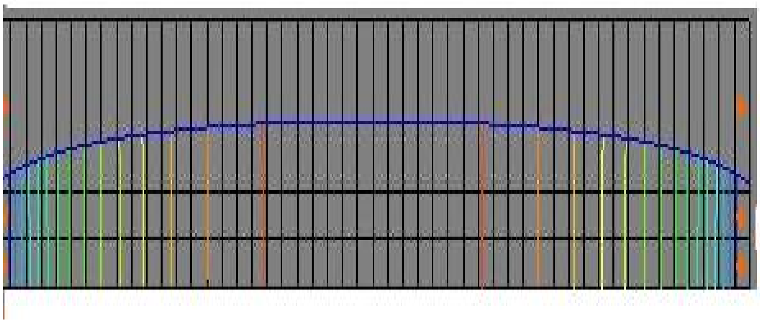
Figure B.1: Mass balance plots with different recharge rate



(a) Water table when recharge rate is 0.003m/d

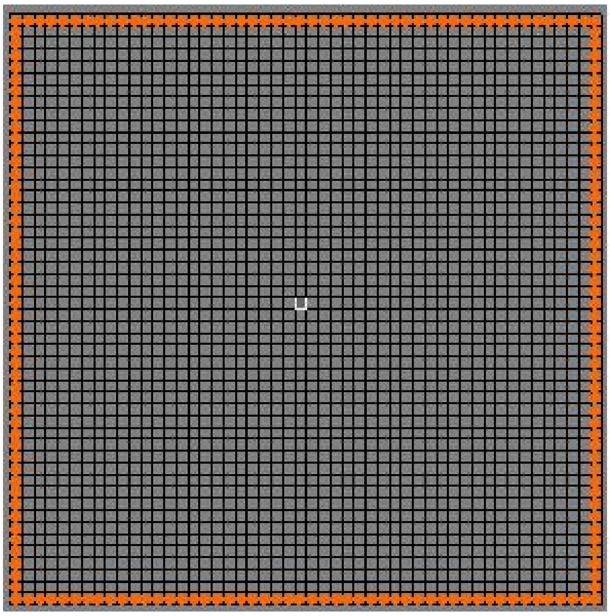


(b) Water table when recharge rate is 0.009m/d

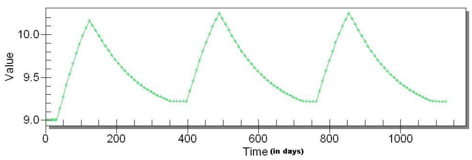


(c) Water table when recharge rate is 0.015m/d

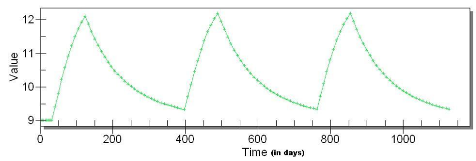
Figure B.2: Water table at the end of rain(i.e. 31st Aug, 2007) with different recharge rate



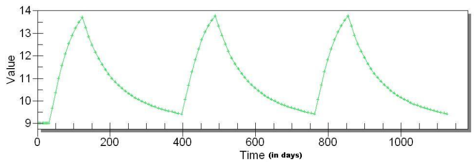
(a) Position of selected cell at 25 × 25, Id:= 1225




(b) Head time series of this cell when recharge rate is 0.003m/d

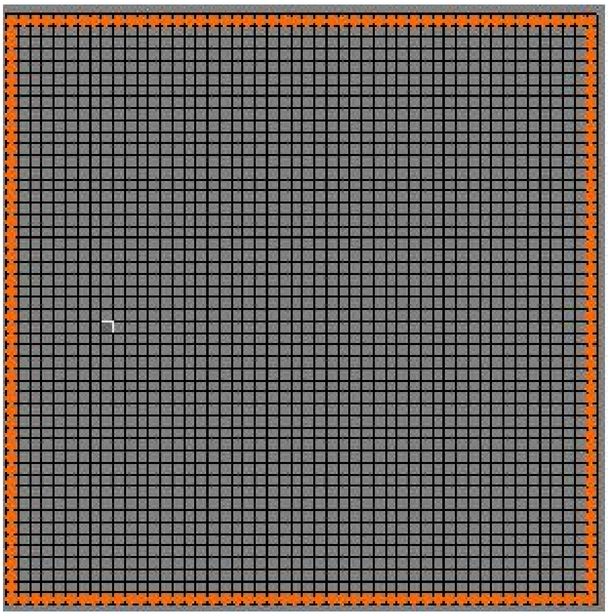



(c) Head time series of of this cell when recharge rate is 0.009m/d

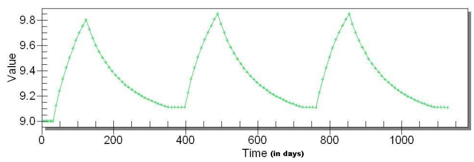



(d) Head time series of of this cell when recharge rate is 0.015m/d

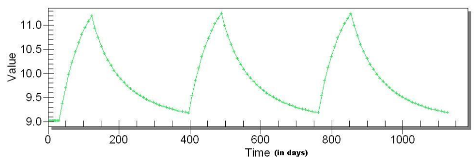
Figure B.3: Head vs Time plot of selected cell (cell id = 1225) at different recharge rate



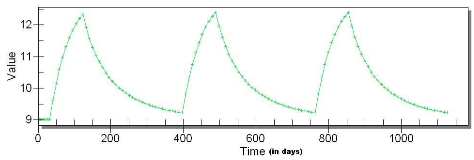
(a) Position of selected cell at $27 \times 9$, Id:= 1309




(b) Head time series of this cell when recharge rate is 0.003m/d

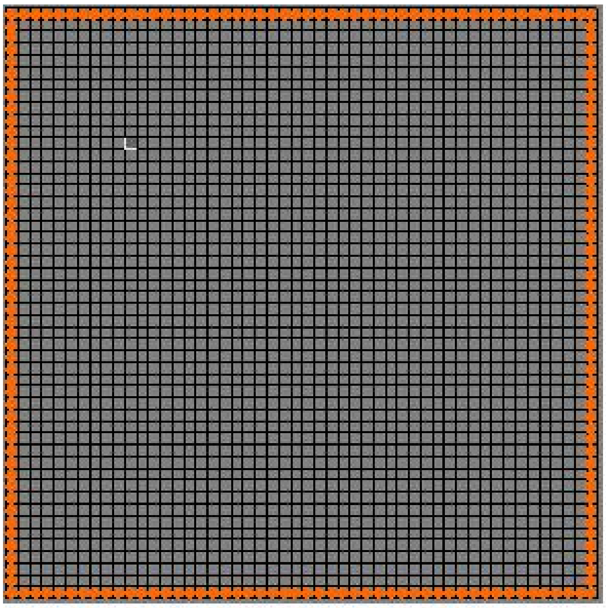



(c) Head time series of of this cell when recharge rate is 0.009m/d

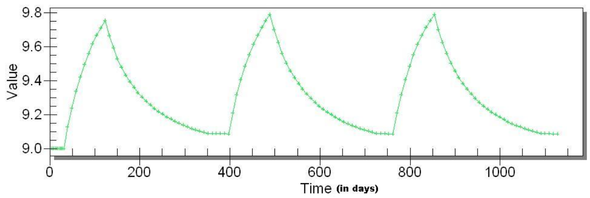



(d) Head time series of of this cell when recharge rate is 0.015m/d

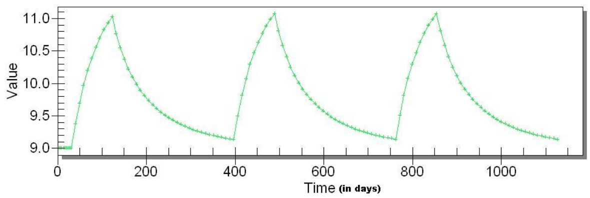
Figure B.4: Head vs Time plot of selected cell (cell id = 1309) at different recharge rate



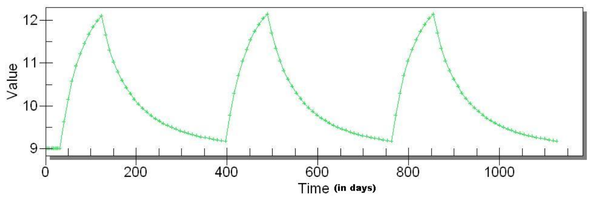
(a) Position of selected cell at $12 \times 11$, Id:= 561




(b) Head time series of this cell when recharge rate is 0.003m/d

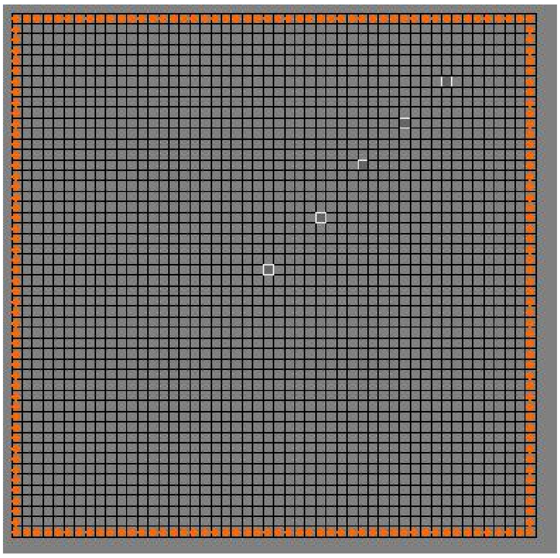



(c) Head time series of of this cell when recharge rate is 0.009m/d

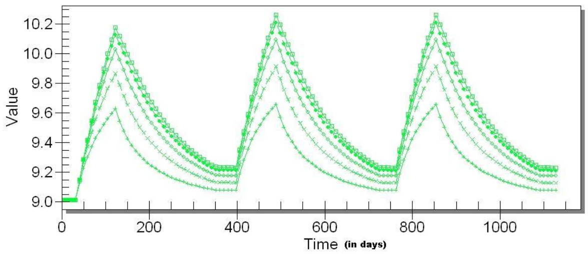



(d) Head time series of of this cell when recharge rate is 0.015m/d

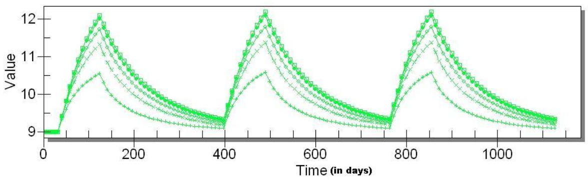
Figure B.5: Head vs Time plot of selected cell (cell id = 561) at different recharge rate



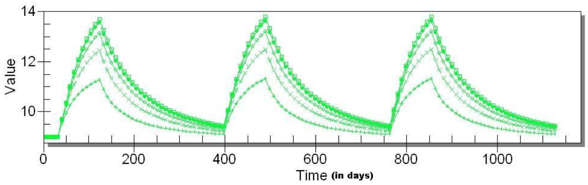
(a) A set of selected cell almost on diagonal




(b) Head time series of this cell when recharge rate is 0.003m/d

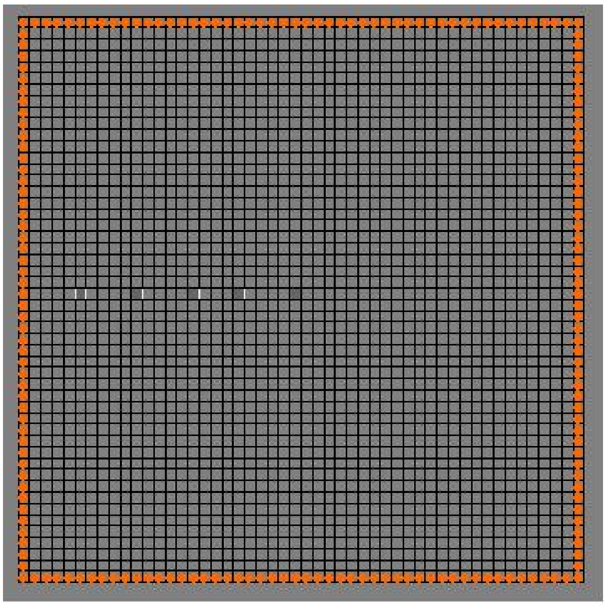



(c) Head time series of of this cell when recharge rate is 0.009m/d

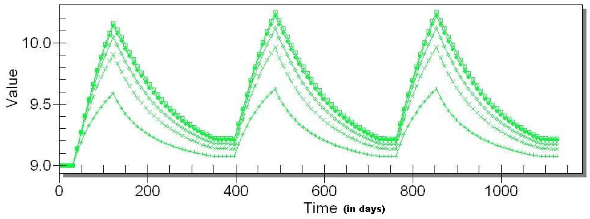



(d) Head time series of of this cell when recharge rate is 0.015m/d

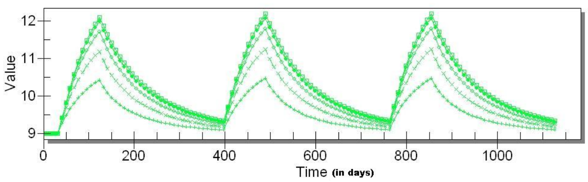
Figure B.6: Head vs Time plot of a group of selected cell (cell id = 561) at different recharge rate



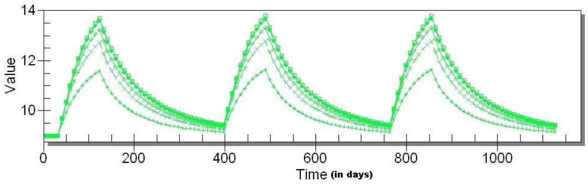
(a) A set of selected cell from middle cell towards west




(b) Head time series of this cell when recharge rate is 0.003m/d




(c) Head time series of of this cell when recharge rate is 0.009m/d

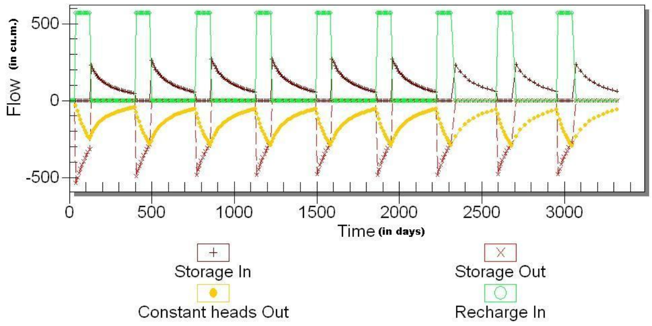



(d) Head time series of of this cell when recharge rate is 0.015m/d

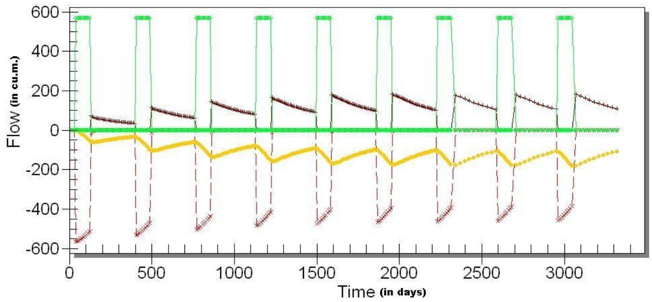
Figure B.7: Head vs Time plot of another group of selected cell at different recharge rate

# Appendix C

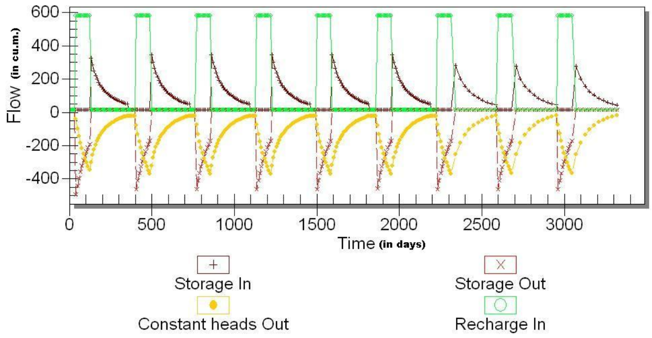
# Stability time and Conductivity Plot




(a) Mass balance plot for input set 1

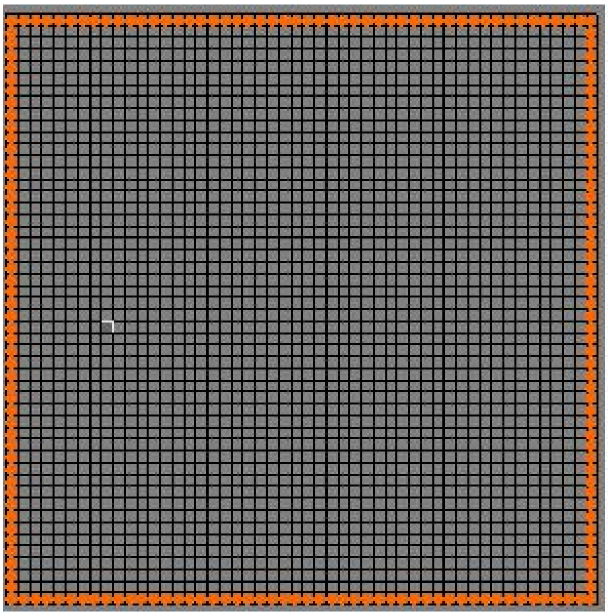



(b) Mass balance plot for input set 2

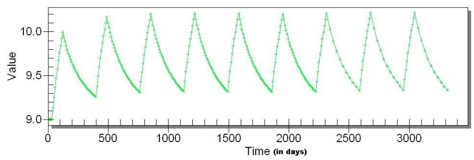



(c) Mass balance plot when recharge rate is 0.015m/d

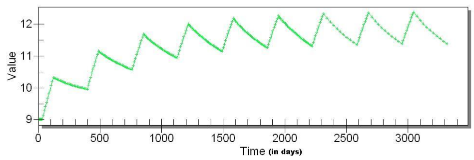
Figure C.1: Mass balance plots with different input set



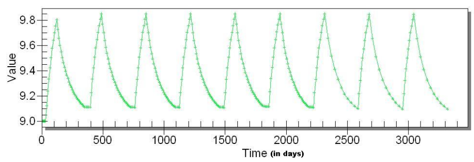
(a) Position of selected cell at $27 \times 9$, Id:= 1309




(b) Head time series of this cell when recharge rate is 0.003m/d

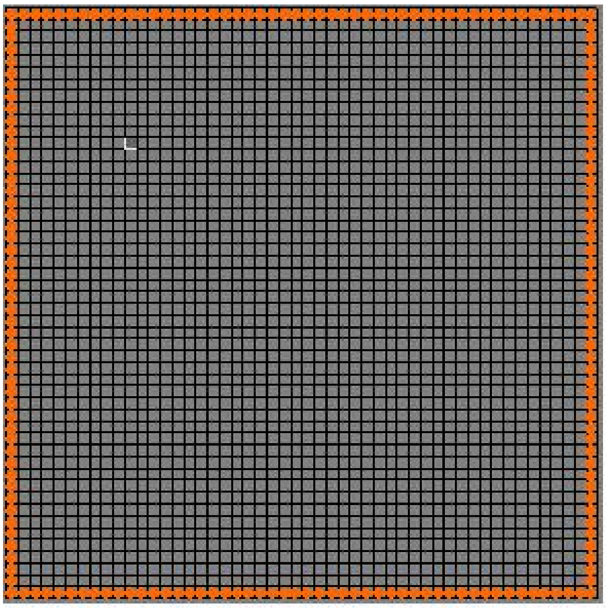



(c) Head time series of of this cell when recharge rate is 0.009m/d

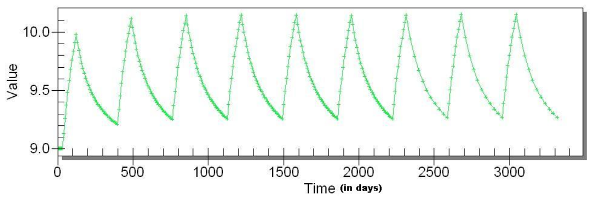



(d) Head time series of of this cell when recharge rate is 0.015m/d

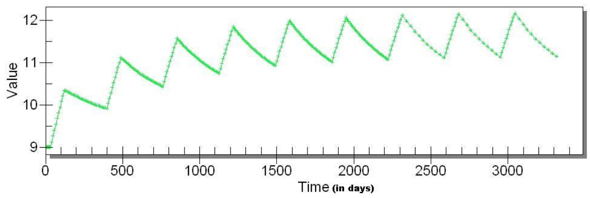
Figure C.2: Head vs Time plot of selected cell (cell id = 1309) at different recharge rate



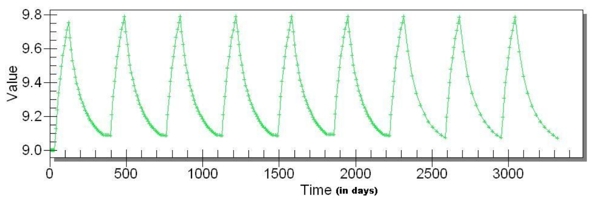
(a) Position of selected cell at $12 \times 11$, Id:= 561




(b) Head time series of this cell when recharge rate is 0.003m/d




(c) Head time series of of this cell when recharge rate is 0.009m/d




(d) Head time series of of this cell when recharge rate is 0.015m/d

Figure C.3: Head vs Time plot of selected cell (cell id = 561) at different recharge rate

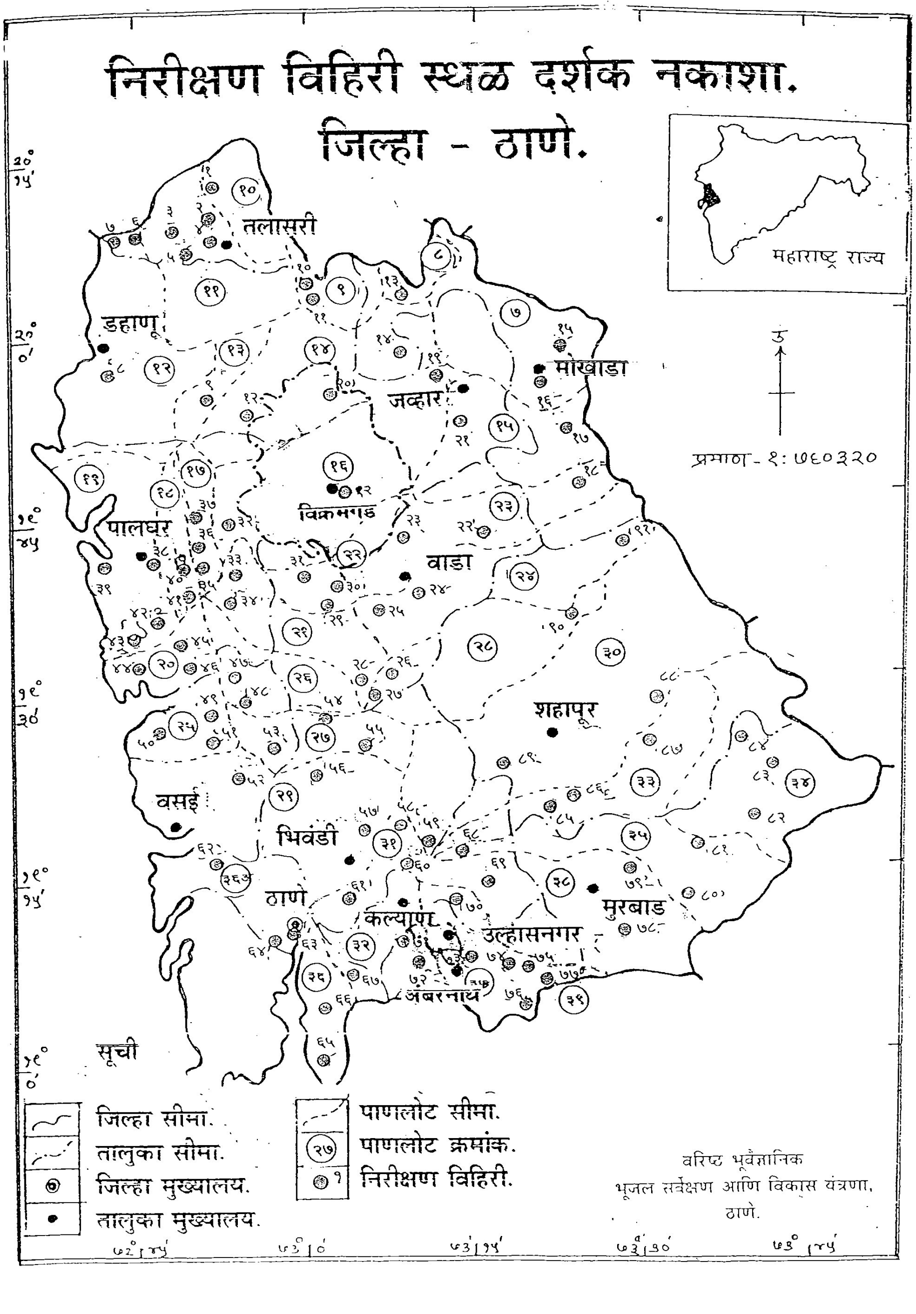
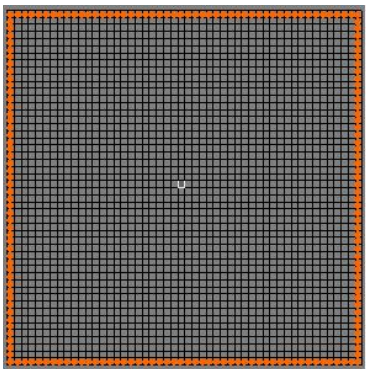


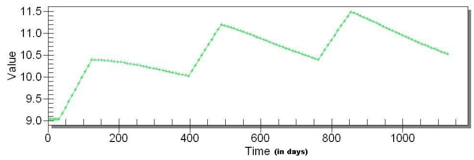

Figure C.4: Map of observation wells(numbered) in area

# Appendix D

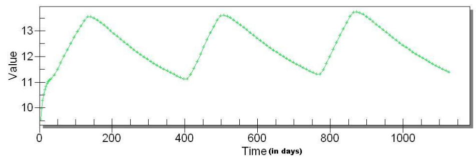
# Improvements in Groundwater due to Lake



(a) Position of selected cell at $25 \times 25$ on second layer

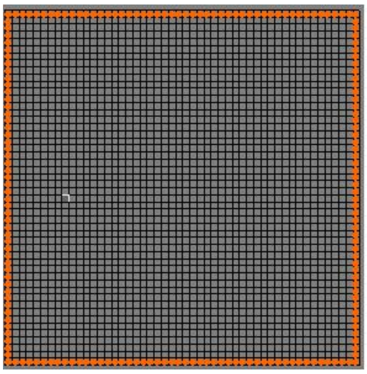



(b) Head time series of this cell without lake

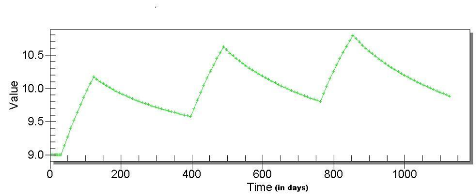



(c) Head time series of this cell with lake

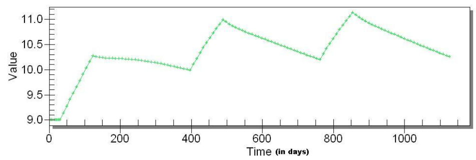
Figure D.1: Head vs. time plot for a selected cell of a system without lake and with lake



(a) Position of selected cell at $27 \times 9$ on second layer

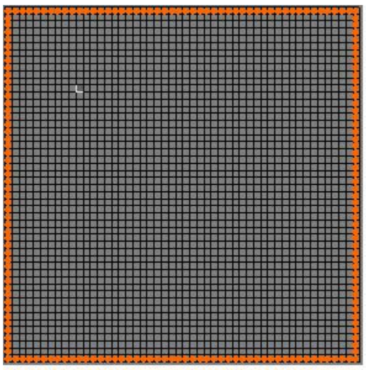



(b) Head time series of this cell without lake

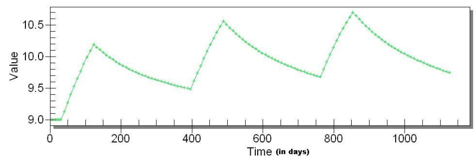



(c) Head time series of this cell with lake

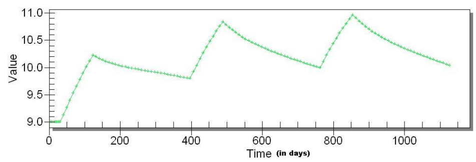
Figure D.2: Head vs. time plot for a selected cell of a system without lake and with lake



(a) Position of selected cell at $12 \times 11$ on second layer




(b) Head time series of this cell without lake

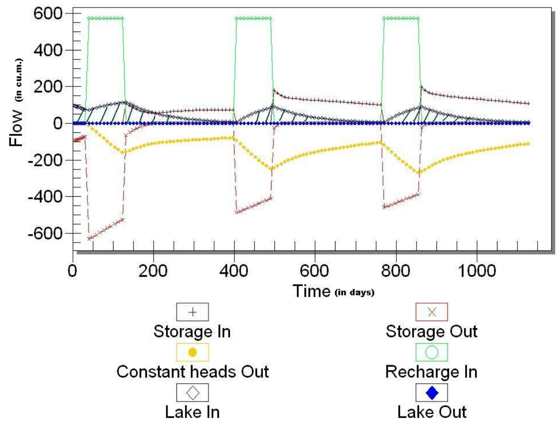



(c) Head time series of this cell with lake

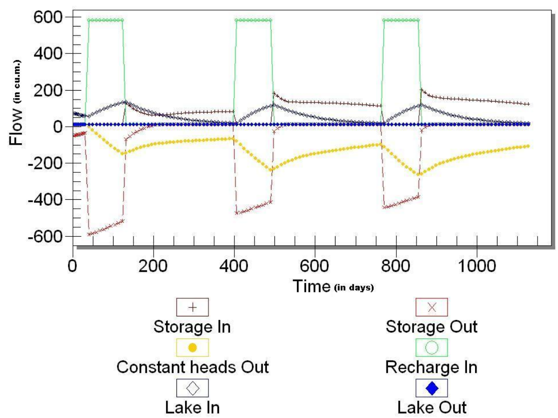
Figure D.3: Head vs. time plot for a selected cell of a system without lake and with lake

# Appendix E

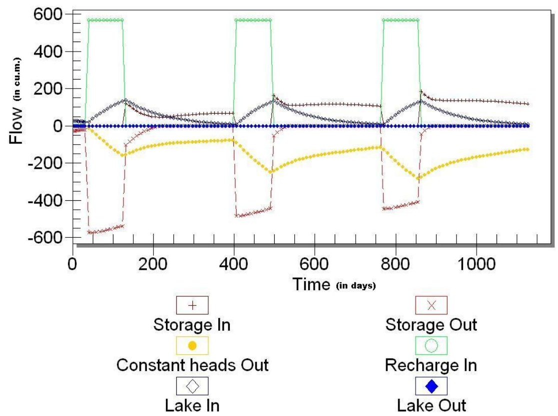
# Experiments with dimension of lake




(a) Mass balance plot with a lake of size $50m \times 50m \times 7m$

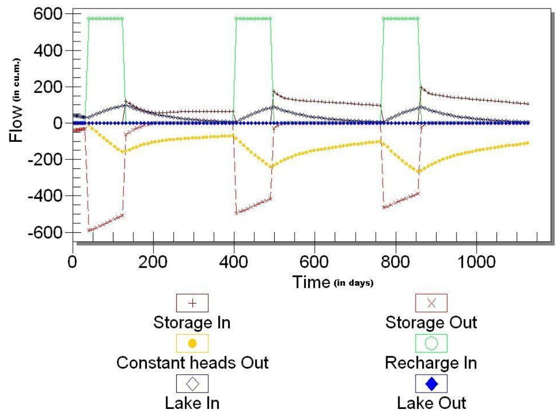



(b) Mass balance plot with a lake of size $60m \times 60m \times 4.8m$

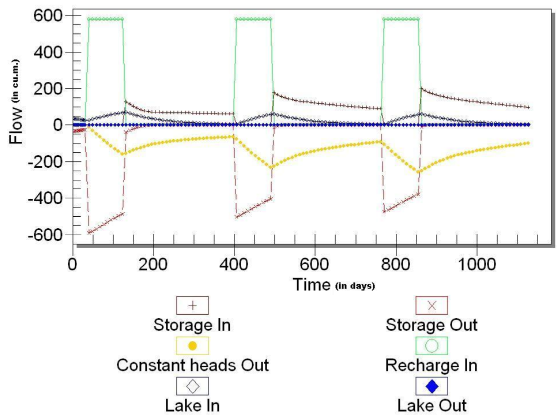



(c) Mass balance plot with a lake of size $70m \times 70m \times 3.6m$

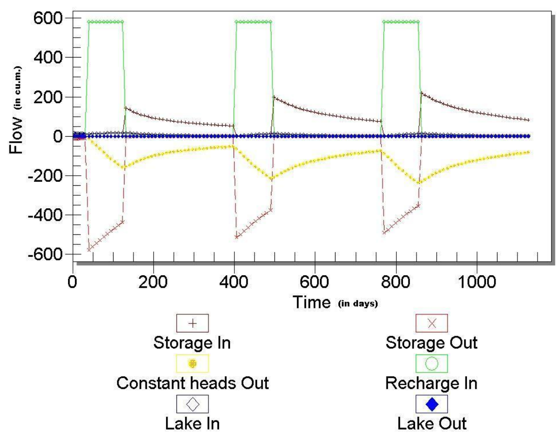
Figure E.1: Mass balance plots of systems of lake with different dimension




(a) Mass balance plot with a lake of size $50m \times 50m \times 4.8m$

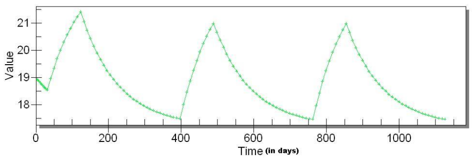



(b) Mass balance plot with a lake of size $30m \times 30m \times 7m$

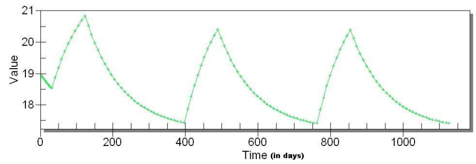



(c) Mass balance plot with a lake of size $20m \times 20m \times 9m$

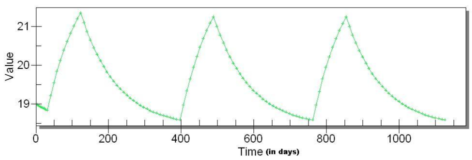
Figure E.2: Mass balance plots of systems of lake with different dimension



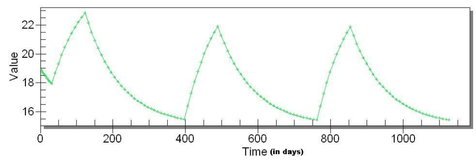

(a) A lake of dimension $50m \times 50m \times 4.8m$



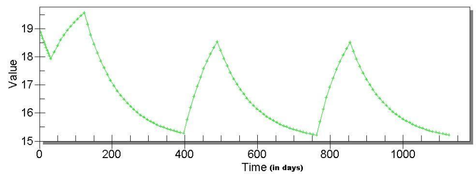

(b) A lake of dimension $60m \times 60m \times 4.8m$




(c) A lake of dimension $70m \times 70m \times 3.2m$

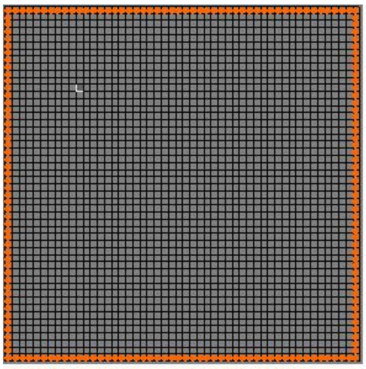



(d) A lake of dimension $30m \times 30m \times 7m$

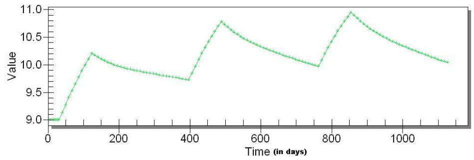



(e) A lake of dimension $20m \times 20m \times 9m$

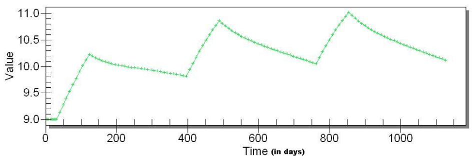
Figure E.3: Lake stages with respect to different dimensions



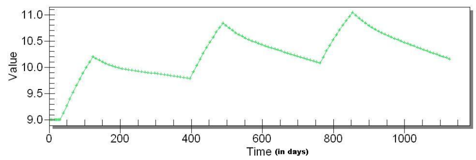
(a) Position of selected cell at $12 \times 11$ on second layer



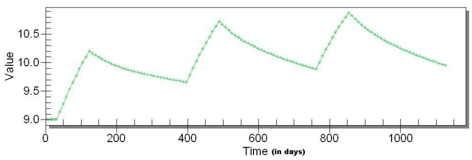

(b) Head time series of the cell with lake of $50m \times 50m \times 4.8m$



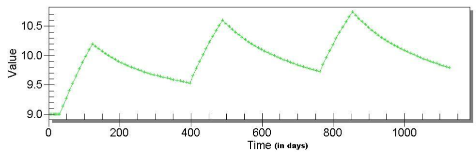

(c) Head time series of the cell with lake of $60m \times 60m \times 4.8m$




(d) Head time series of the cell with lake of $70m \times 70m \times 3.2m$

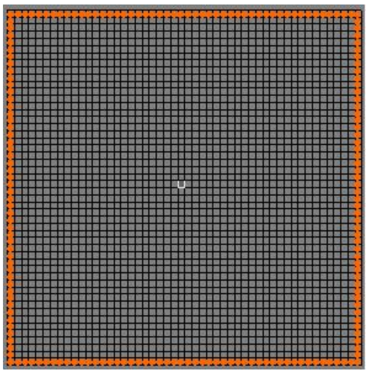



(e) Head time series of the cell with lake of $30m \times 30m \times 7m$

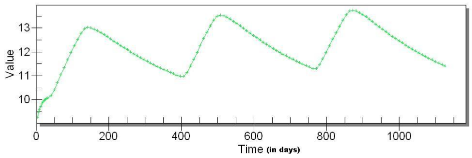



(f) Head time series of the cell with lake of $20m \times 20m \times 9m$

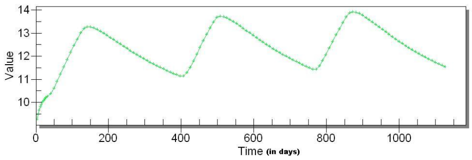
Figure E.4: Head vs. time plot for a selected cell of a system lake of different dimensions



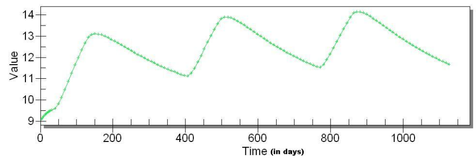
(a) Position of selected cell at $25 \times 25$ on second layer



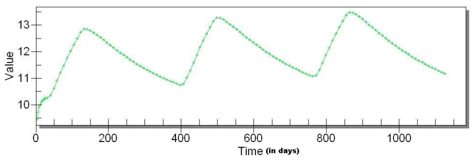

(b) Head time series of the cell with lake of $50m \times 50m \times 4.8m$



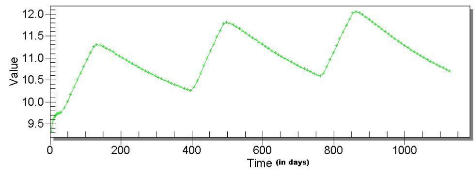

(c) Head time series of the cell with lake of $60m \times 60m \times 4.8m$




(d) Head time series of the cell with lake of $70m \times 70m \times 3.2m$

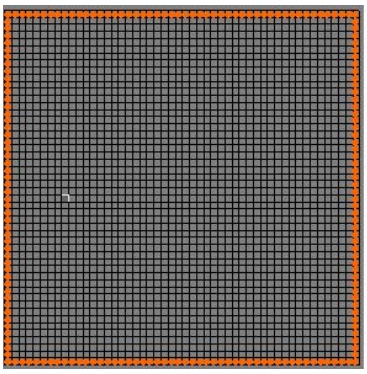



(e) Head time series of the cell with lake of $30m \times 30m \times 7m$

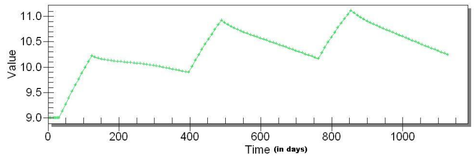



(f) Head time series of the cell with lake of $20m \times 20m \times 9m$

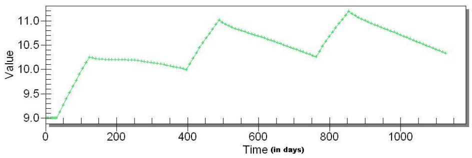
Figure E.5: Head vs. time plot for a selected cell of a system lake of different dimensions



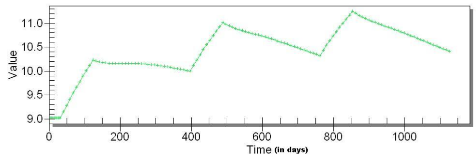
(a) Position of selected cell at $27 \times 9$ on second layer



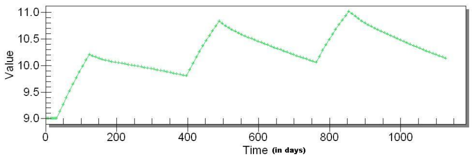

(b) Head time series of the cell with lake of $50m \times 50m \times 4.8m$



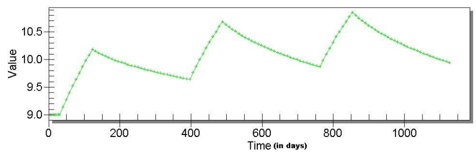

(c) Head time series of the cell with lake of $60m \times 60m \times 4.8m$




(d) Head time series of the cell with lake of $70m \times 70m \times 3.2m$

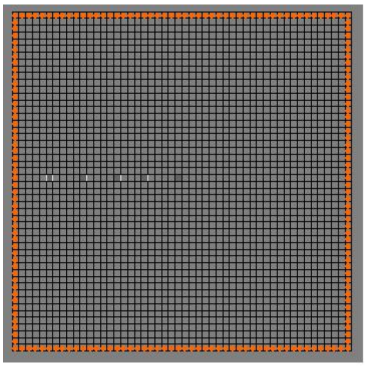



(e) Head time series of the cell with lake of $30m \times 30m \times 7m$

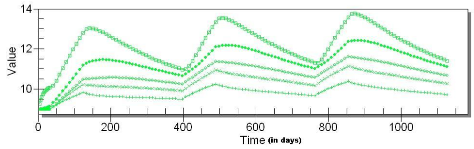



(f) Head time series of the cell with lake of $20m \times 20m \times 9m$

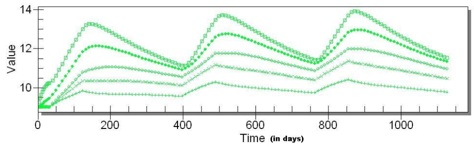
Figure E.6: Head vs. time plot for a selected cell of a system lake of different dimensions



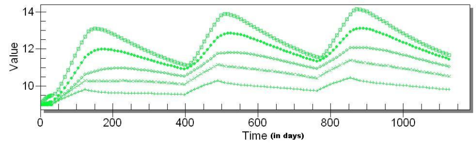
(a) Positions of selected cells in a group on second layer



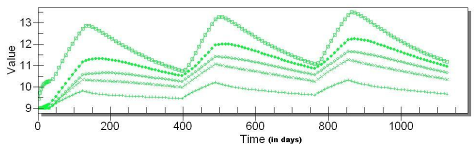

(b) Head time series of the cell with lake of $50m \times 50m \times 4.8m$



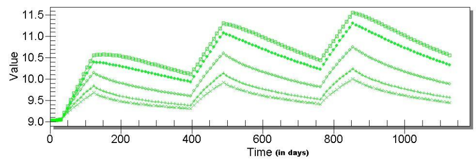

(c) Head time series of the cell with lake of $60m \times 60m \times 4.8m$




(d) Head time series of the cell with lake of $70m \times 70m \times 3.2m$

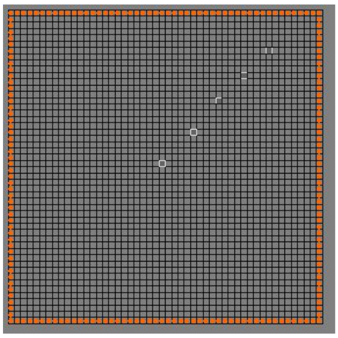



(e) Head time series of the cell with lake of $30m \times 30m \times 7m$

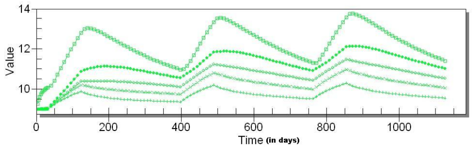



(f) Head time series of the cell with lake of $20m \times 20m \times 9m$

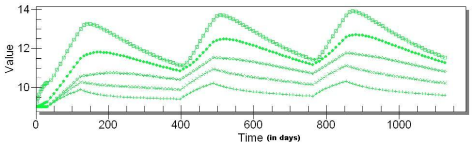
Figure E.7: Head vs. time plot for a set of selected cell of a system lake of different dimensions



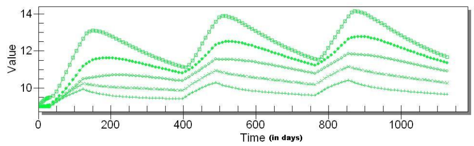
(a) Positions of diagonally selected cells in a group on second layer



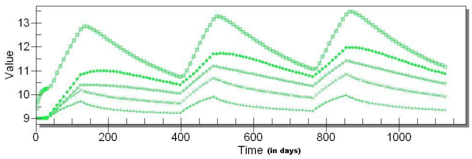

(b) Head time series of the cell with lake of $50m \times 50m \times 4.8m$



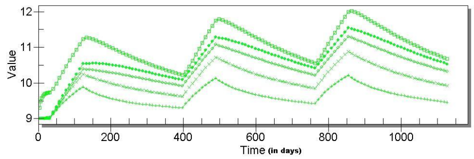

(c) Head time series of the cell with lake of $60m \times 60m \times 4.8m$




(d) Head time series of the cell with lake of $70m \times 70m \times 3.2m$




(e) Head time series of the cell with lake of $30m \times 30m \times 7m$

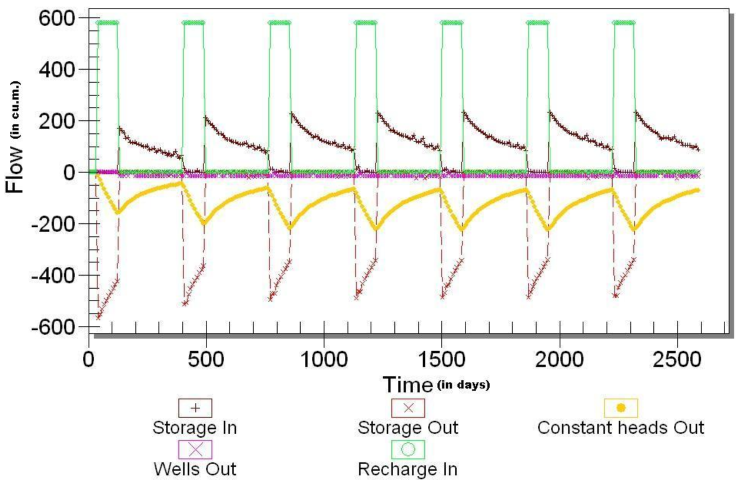



(f) Head time series of the cell with lake of $20m \times 20m \times 9m$

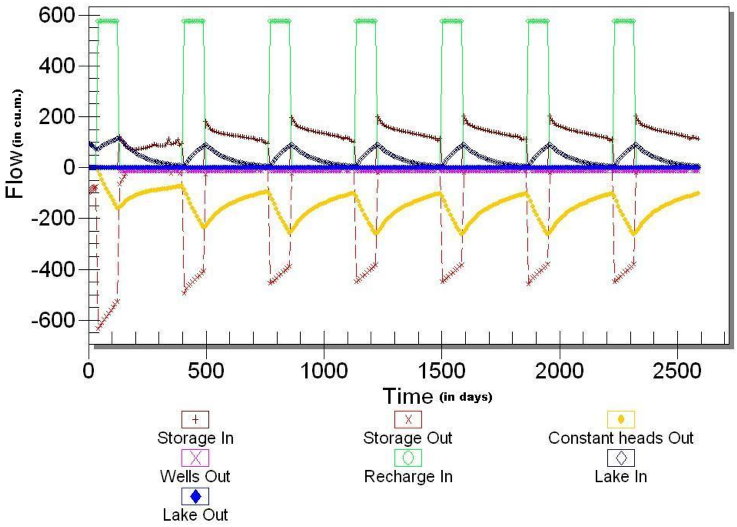
Figure E.8: Head vs. time plot for a set of selected cells of a system lake of different dimensions

# Appendix F

# Plots of Well

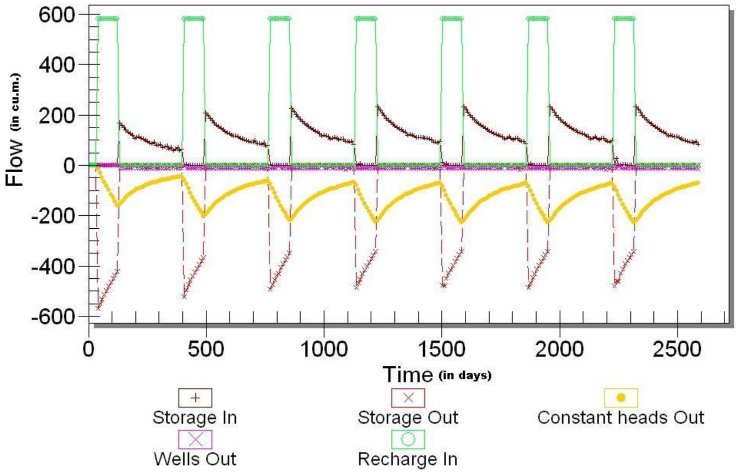



(a) Mass Balance plot of well without lake in regime

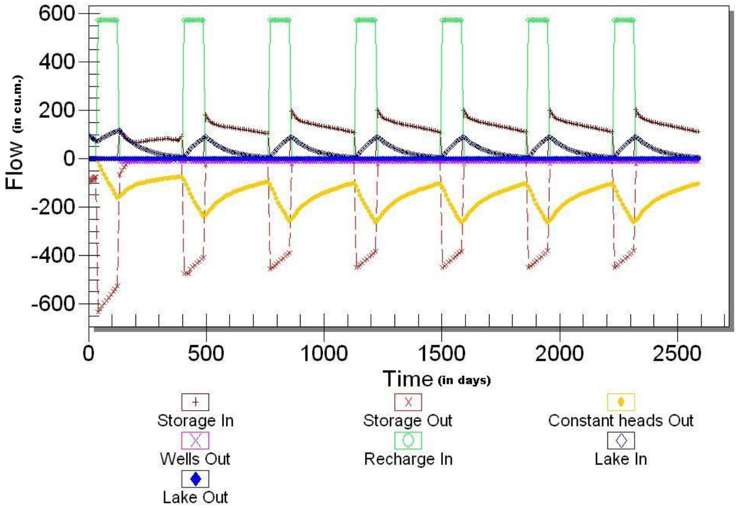



(b) Mass Balance of well with lake in regime

Figure F.1: Mass balance plot of regime with a well of pumping rate of 15 cubic meter per day




(a) Mass Balance plot of well without lake in regime




(b) Mass Balance of well with lake in regime

Figure F.2: Mass balance plot of regime with a well of pumping rate of 15 cubic meter per day

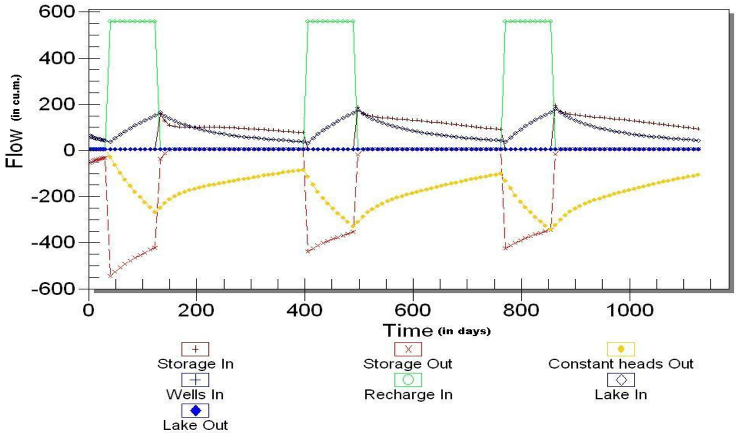



Figure F.3: Mass balance plot with barrier on first layer only

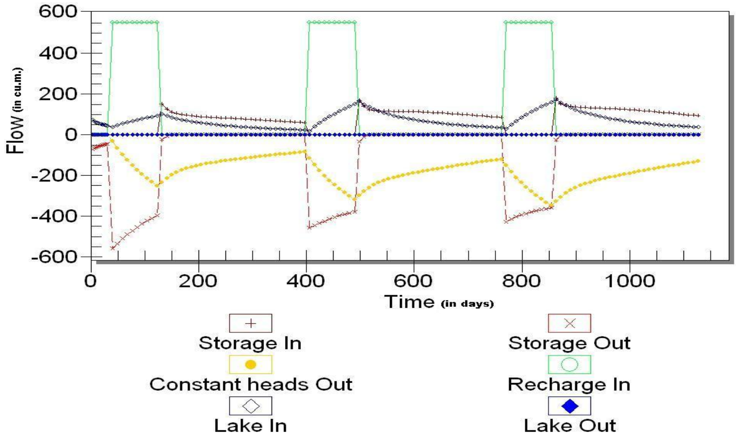



Figure F.4: Mass balance plot with barrier extended upto second layer

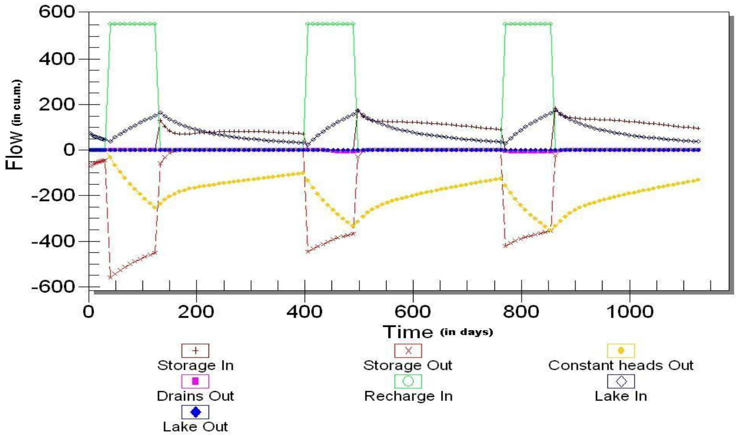



Figure F.5: Mass balance Plot with extended barrier and drain

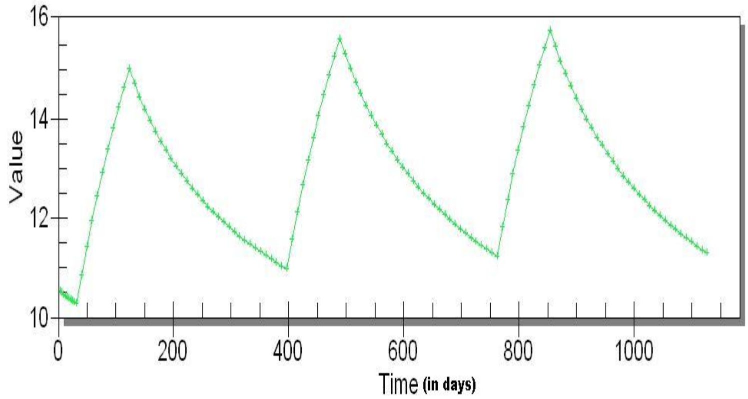



Figure F.6: Lake stage

# Bibliography


[1] [Online; accessed 10-jul-2008]. http://www.groundwater.buffalo.edu/software/VBB/images/Dam%20Flownet.JPG.

[2] [Online; accessed 10-jul-2008]. www.clas.ufl.edu/users/screaton/06fall/Unsaturated_Zone.doc.

[3] [Online; accessed 10-jul-2008]. www.northern.edu/natsource/images/sda3.jpg.

[4] [Online; accessed 15-Jul -2008]. http://uwp.edu/~li/geol460-00/chapter2.htm.

[5] [Online; accessed 15-Jul -2008]. http://www.connectedwater.gov.au/processes/hydrological.html.

[6] [Online; accessed 15-Jul-2008]. http://hercules.gcsu.edu/~sdatta/home/teaching/hydro/slides/darcy_law.gif.

[7] [Online; accessed 15-jun-2009]. http://www.if.uidaho.edu/~johnson/ifiwrri/sr3/swgw.html.

[8] [Online; accessed 15-jun-2009]. http://www.cse.iitb.ac.in/~sohoni/waterstory.pdf.

[9] [Online; accessed 15-jun-2009]. userwww.sfsu.edu/~enssani/Well%20Hydraulics.ppt.

[10] Gregory W. Council. A lake package for modflow (lak2), June 1999.

[11] R. A. Freeze and J. A. Cherry. *Groundwater*. 1979.

[12] Arlen W. Harbaugh. Modflow-2005, the u.s. geological survey modular ground-water model,the ground-water flow process, techniques and methods. 2005.

[13] A. K. Rustogi. *Numerical Groundwter Hydrology*. 2007.