CHAPTER 4

METHODOLOGY

4.1 General

The development of groundwater flow model is the first and essential step to know the river-aquifer interaction. In groundwater modeling, the groundwater flow equation has to be solved numerically to simulate the three-dimensional groundwater flow. Thus, it helps to understand the behavior of an aquifer system under changes in naturally and human induced hydrological stresses. Further, the groundwater model assist in organizing the large amount of regional topographical, geological, hydrological, hydro-geological etc., data. The complexity in the study of river-aquifer (R-A) interactions is due to temporal variations in surface water and groundwater diffusion phenomenon. Effective and efficient management of these water resources requires the ability to predict the movement of the groundwater system as well as predict the changes that may be applied to the groundwater system by nature and anthropogenic activities. For this, groundwater modeling was carried out to simulate the groundwater head in the study area using GMS 10.2 along with the river-aquifer interaction techniques are discussed in this chapter.

4.2 An approach to the Modeling

An accurate groundwater model presents the real condition of groundwater system under certain hydrological stresses. It can be easy, like one-dimensional analytical solution, or very complex three dimensional multi layered models (Olsthoorn, 1985). There are lots of complexities associated with the groundwater flow modeling; it is always suggested to start with an easy model. After successful conceptualization of the simple model, complexity in the model can be added step by step (Hill, 2006). Apart from the modeling approach, there are other factors like initial head value (condition), boundary conditions, time and space discritization, and quality of gathered/collected data can affect the modeling results.

4.2.1 Conceptual Model

The conceptual model can be defined as the thorough representation of the groundwater system, which includes the interpretation of the hydrological and the geological conditions of the aquifer as well as information about the water budget. After the objective identification, it is the most important part and next step of the modeling. Conceptualization of the model requires good knowledge about the geology, hydrology of the concern area, boundary conditions and hydraulic properties of the aquifer, soil characteristics. An excellent conceptual model must portray the actuality in an easy manner that meets the modeling goals and management requirements (Bear and Verruijt, 1987). The main issues involved in the conceptual model are:

1. Geometry (Horizontal & Vertical) of the Aquifer and model domain.
2. Initial conditions as well as Boundary conditions of the area.
3. Aquifer properties like hydraulic conductivity (K_x, K_y & K_z), porosity, storativity, etc.
4. Recharge into the Groundwater.
5. Sources and sinks identification for discharge information.
6. Water Budgeting.

After successful completion of the conceptual model, next step are setting up the numerical model. The numerical model presents the conceptual model in the form of mathematical concepts and language. Mathematical equations can be solved either analytically or numerically. Figure 4.1 shows the adopted methodology for the groundwater flow model development and conceptualization.

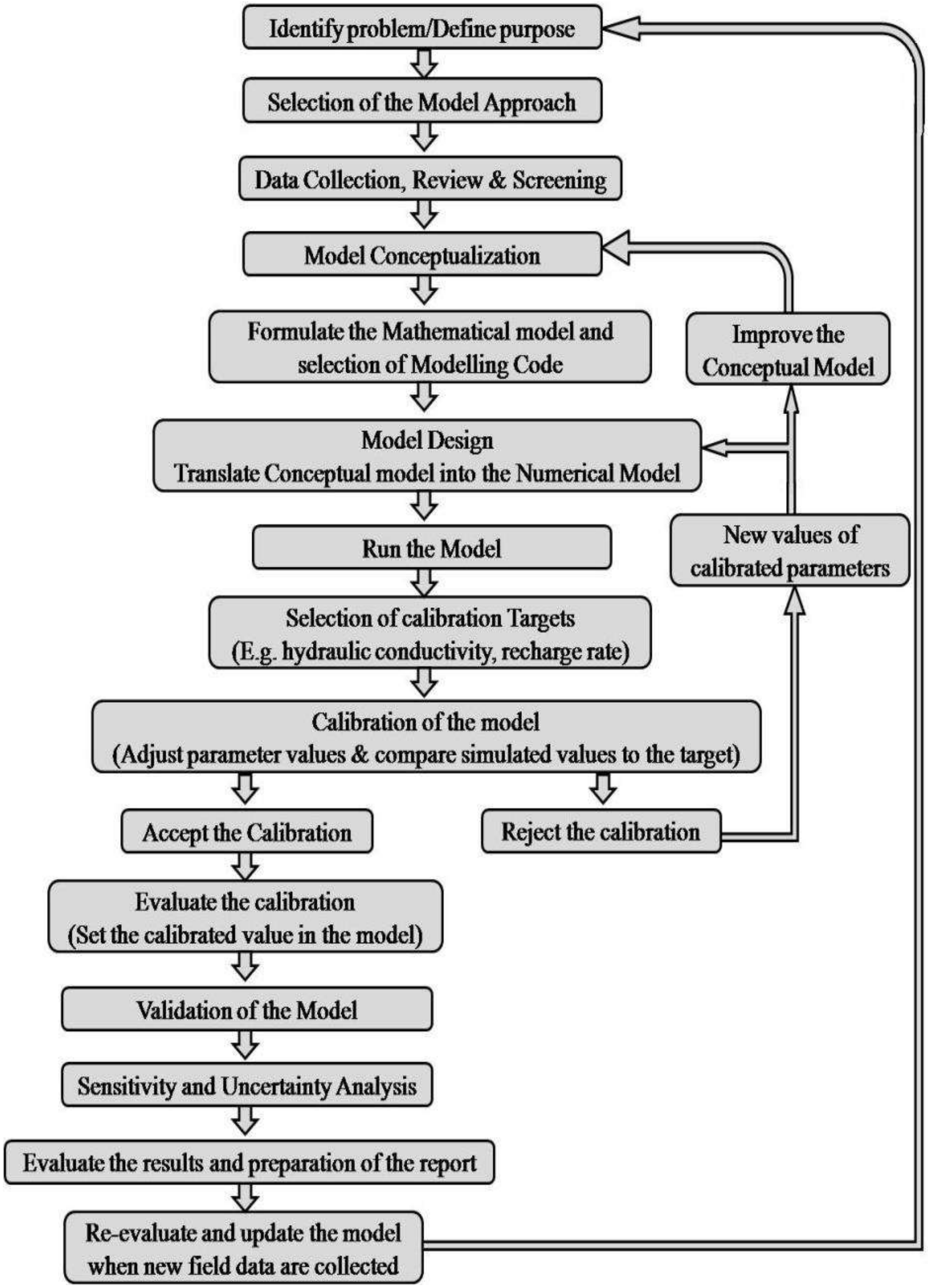

Figure 4.1 Flowchart of methodology for preparation of the Groundwater flow model

A mathematical model is a set of partial differential equations (PDE) that represent the physical processes of flow of water in the groundwater system. A discretized numerical method has been used to solve the governing equations of the groundwater flow in the study.

To derive mathematical equations describing the groundwater flow systems, the basic principles of conservation of mass of liquid are applied. The fundamental principle of conservation of mass (normally known as the continuity equation) is used in conjunction with the mathematical equations of related processes to derive the differential equations that describes the water flow.

The process of groundwater flow is controlled by Darcy's law and the law of mass conservation. In 1856, Henry Darcy, a French water works engineer, studied experimentally the flow of fluids through porous media and established a relationship that describes the flow of a fluid through a porous medium. According to Darcy's law, the flow occurring per unit time is directly proportional to the head causing flow and the area of cross-section of the soil sample but is inversely proportional to the length of the soil sample. However, Darcy's law have some limitations such as it is only valid for laminar and viscous flow. In Darcy's law, constant of proportionality i.e. hydraulic conductivity depends on the properties of flowing liquid and the properties of the permeable medium. By combining the Darcy's law with the continuity equation, a general form of equation describing the transient flow in the anisotropic, non-homogeneous aquifers can be derived. The governing partial differential equation for three-dimensional (3-D) groundwater flow through a porous medium for a confined aquifer is given by the equation below (Rushton and Redshaw, 1979):

$$\frac{\partial}{\partial x}\left[K_{xx}\frac{\partial h}{\partial x}\right]+\frac{\partial}{\partial y}\left[K_{yy}\frac{\partial h}{\partial y}\right]+\frac{\partial}{\partial z}\left[K_{zz}\frac{\partial h}{\partial z}\right]+W=S_S\frac{\partial h}{\partial t} \qquad 4.1$$

Where,

K_{xx}, K_{yy} & K_{zz} are the hydraulic conductivity (HC) of the aquifer along the x, y and z directions respectively (LT^{-1}).

W is the volumetric flux per unit volume representing sources/sinks (T^{-1})

S_s is the specific storage of the porous material of aquifer (L^{-1})

h is the potentiometric head (L).

t is the time (T)

4.3 Groundwater Flow simulation

Groundwater flow can be categorized into two states, one is steady state and another is transient state. The general equation of flow (Laplace equation) is developed on the assumption of the flow in the porous medium is transient. But when the storage term of the aquifer system becomes zero, the flow will change to steady state. In steady state, the depended variable (head h) is a function of space only while h is function of both space and time in unsteady/transient state.

In a steady-state flow condition, the water particle entering the flow field from the inflow boundary will continue to flow directly to the outflow boundary in a direction parallel to the path lines, which coincide with the flow lines regardless of time. While in a transient-state flow condition, the flow lines and path lines do not coincide with each other. Whereas, the flow lines keep varying with time. Actually, these flow lines represent the flow direction at a particular time and cannot be used in the estimation of the absolute path of the flow particle. Consequently, understanding of the groundwater flow simulation techniques is very essential for water resource engineers and groundwater hydrologists to accurately evaluate the steady state and transient flow conditions using suitable and accurate mathematical equations.

4.4 Boundary and initial conditions

To get a distinctive solution of the partial differential equation (PDE) corresponding to a given conceptual problem, further data about the physical state of the problem is requisite. This additional data can be supplied by initial conditions and boundary conditions of the area. Model boundaries for any groundwater flow models can be defined as the points at which the head (dependent variable) or the derivation of the head (flux) is known and location of this point can be underground or over the ground.

Apart from mathematical constraint, boundary conditions also represent the sources and sinks in the groundwater (Reilly and Harbaugh, 2004). For getting solution of transient type problems, initial and boundary conditions are required while for getting solution of steady state type of problems only boundary conditions must be specified. Improper boundary conditions identification may affect the solution and lead to inaccurate output of the model (Franke et al. 1987). Scientifically, the boundary conditions take account of the type (geometry) of the boundary and the values of the dependent variable or its derivative normal to the boundary. In the model, boundary conditions symbolize the locations, where water flows into or out of the model area due to any external factors. River, evapotranspiration, recharge, wells and lakes are some general examples of the boundary conditions. For groundwater flow model applications, the boundary conditions may be categorized into 03 types:

1. **Specified heads:-** The head remains constant in the specified head boundaries. Movement of the flow (in or out) in the model in such a way that specified head remains maintain. It is also called *Dirichlet* or *type I boundary*.
2. **Specified flux (flow):-** The rate of flow from the boundary in or out of the model is specified in the specified flux boundary and the head value may change at the specified flux cell according to the flux. It is also called a *Neumann* or *type II boundary*.
3. **Head dependent flux (flow):-** The rate of flow moving in or out of the model varies according to the changes in the head value in the boundary cell. It is also called a *Cauchy* or *type III boundary.*

In modeling, it is advisable to use physical boundaries (e.g., rivers, stream channels, impervious boundaries, lakes) when possible, as the model boundary conditions. The reason behind this is that it can be identified and conceptualized easily. More care should be taken to identify the natural boundaries. In addition to the above-mentioned types, there are further sub-types of boundaries. For example groundwater divides (Groundwater divide is the line connecting the high points on the water table or other potentiometric surface) or surface water divides can shift their position as the condition varies in the field. In a transient model, if contours of water table used to set the boundary condition it is recommended to identify the flux instead the head. In any case, the groundwater system is heavily stressed due to any hydro-geological condition, the boundary conditions may vary over the specific time. In that situation, during the simulation of the model, the boundary conditions should always be checked.

4.4.1 Examples of Different Boundaries

In 2004, Reilly and Harbaugh has been investigated the various types of physical framework and corresponding mathematical depiction. According to their investigation, different boundaries have been presented in the following sections:

***Constant head boundary*:** In constant head boundary, it is assume that the head remains constant in the cells irrespective of the time. It's a special case of specified head boundary (Franke et al. 1987).

***Specified head boundary*:** In specified head boundaries, head value can be specified as a function of time and place. Natural boundaries such as rivers, streams and lakes are the examples of this type of boundaries.

***No flow boundary*:** No flow boundary normally occurs at a place where impermeable structure exists and at a line normal to streamline (i.e. normal to flow direction). It's a special case of a *specified flux boundary*. An example of no flow boundary can be a water divide, line connecting the high points on the water table or other potentiometric surface.

***Specified flux boundary*:** Specified flux boundary arises when the flow across the boundary can be specified spatially and temporally. Recharge across the water table can be used as a specified flux boundary.

***Head-dependent flux boundary*:** This type of boundary takes place when flow rate across the boundary depends on the head value neighboring to that boundary. An example of such a boundary is a semi-confined aquifer, where the water head depends on the flow through the semi-confining layer.

***Free-surface boundary*:** Free-surface boundary normally occurs in the coastal aquifers. Interface of the water-table and the fresh-saline water are an example of free-surface boundaries. At free-surface boundary, pressure head must be zero and the total head equal to the elevation of the head.

4.5 Recharge

Recharge is one of the most important input parameters that affect the behavior and water level of the groundwater aquifer systems. Accurate estimation of the recharge rate is a difficult phenomenon due to various processes associated with this (Wood and Sanford, 1995). **Groundwater recharge** is a hydrologic process, in which water moves downward direction from surface water and in the form of leakage losses from rivers, water bodies, lakes, streams etc. to groundwater. It is a primary method through which water percolates to the aquifer.

In groundwater modeling, usually, the value of the recharge rate is taken positive to signify that there is an increment in the groundwater of the aquifer system. Generally, in the wet seasons groundwater receives recharge in large amount as compare to dry seasons.

Major groundwater recharge in the study area is through Rainfall, Canal seepage, infiltration of the river water, and irrigation return flow. The recharge enters the aquifer through infiltration in the overlying confining unit and seepage through riverbeds. Study area receives about 80% of its annual rainfall, i.e., 1020 mm from south-west monsoon during the period of July to August. Rainfall data were procured from Central Water Commission (CWC), Varanasi for the period of 16 years from 2000-2016, as shown in table B.1 in the Appendix B. Canal seepage was assumed as 10% of the water released for irrigation. The value of canal seepage depends on the type of canal, i.e., lined or unlined (Rao, 1993).

Land-use and land-cover (LULC) pattern was obtained through satellite image classification. Digital satellite imagery of Landsat-8 satellite, launched by US, has been used in the LULC classification. Landsat Satellite images were downloaded from the USGS website (Earth_Explorer). To bring all the satellite imagery under one geometric coordinate system, all the downloaded satellite imagery were geo referenced and projected to Transverse Mercator system. The coordinate system used in the study is WGS_84_UTM_Zone_45N and the Datum is WGS 84.

The satellite images were collected once in four years, which were then taken the same for the subsequent four years, starting from the year 2006. The image was classified for the year 2006 and then the same classification was used for the period of 2006-2008. Further satellite imagery for the year 2009 was classified, which was further used for the period of 2009-2012 and so on. After the classification of the images, different recharge zones were marked. Every recharge zones have different recharge coefficients depending on the type of soil available, and the amount of rainfall occurring in that recharge zone. There are empirical equations that can be used in the groundwater recharge estimation. These empirical equations relate the rainfall with the groundwater recharge. A modified version of Chaturvedi's (1973) equation was proposed by Irrigation Research Institute, Roorkee, which was used in this study.

$$R = 1.35\,(P - 14)^{0.4} \qquad 4.2$$

Where, R is the net recharge due to rainfall and P is the rainfall measured in inch. Recharge can be changed into millimetres (mm).

The recharge coefficient can be described as the ratio of recharge to the effective rainfall, expressed in % (Muteraja, K. N., 1986).

$$R_{coeff} = \left({}^{R}/_{P_{eff}}\right)\% \qquad 4.3$$

Where R is the net recharge and P_{eff} is the effective rainfall.

4.6 Hydraulic conductivity

Hydraulic conductivity (K) is the most significant and essential input parameters required for a groundwater flow modeling. Hydraulic conductivity can be define as a physical property of the soil or rocks which measures the ability of the medium (material) to transmit the water through pore spaces or fractures in the presence of an applied hydraulic gradient. This can be explained by Darcy's law as mentioned below.

$$v = k * i = k * \frac{dh}{dl} \qquad 4.4$$

Where, v is the discharge velocity (LT^{-1});

k is the soil hydraulic conductivity (LT^{-1}), which depends on the porous medium and flowing fluid;

h is the head of groundwater table (L) in unconfined flow and peizometric head in case of confined flow.

l is the length of flow (L).

According to the Darcy's laws, it is the ratio of the average velocity of the water flow through a cross-sectional area to the applied hydraulic gradient. Hydraulic conductivity depends on several factors such as the degree of saturation of the soil, density and viscosity of the water. In partly saturated soils, the entrapped air greatly reduces the hydraulic conductivity.

4.7 Well discharge (Pumping)

In the study area, there are many wells, hand pumps and tube-wells are available from which local people extract the groundwater from the aquifer to meet their daily demands. Water demand for the irrigation purpose is fulfilled by the tube-wells and in some places by canals as well. Whereas, domestic water demand is met by the wells and hand pumps. These wells play an important role in the movement of water from groundwater to the surface water. The effect of pumping out of the water from the wells would be called on the local scale if extraction of water was done by a well or a small group of wells. However, when the water extraction is done by many wells over the large region, it will have an impact on the regional scale.

4.8 Calibration

Calibration of the model is a process of adjustment of the model parameters, which are sensitive to the model results. Calibration is a very important component of any model development, especially for hydrological problems. It forces the model parameters to fit within the margins of the uncertainties and to obtain a model representation of the processes of interest that satisfies pre-agreed criteria i.e. Goodness of fit of the model. Model calibration is performed by adjusting the model input parameters and the value of these parameters is changed based on the collected field data. In the groundwater science, the most accurate definition of the calibration is given by Anderson et al. (1992). According to Anderson et al. (1992), calibration is the manipulation of the input data of groundwater model so that the output of the model, (water head or water flow) is close to the actual field condition. In other words, it can be said that calibration ensures that the developed model produces the most accurate results. This adjustment (change) in the model parameters can be made manually (trial and error) or automatically, and the effect of this adjustment is checked by statistical techniques. The acceptance of the result of the calibration process depends on the closeness between the simulated and observed data.

4.9 Validation

Although for any models, numerical simulation and calibration are valuable tools for analyzing any groundwater systems their prediction accuracy is limited without the model validation. Model validation is carried out after the model calibration process. Model validation suggests that the simulated model is consistent in providing the groundwater head data.

For the validation process in the groundwater studies, groundwater head data (sample) is collected from the field. After the model development and calibration, this collected field data is inserted into the calibrated model. The response of the model is checked by comparing its results with the field observations and this procedure is called validation of the model.

4.10 Model Development using Analytical Approach

The analytical element method (AEM) is a numerical method used to analytically solve the equation for each element (wells, rivers, etc.). It was initially developed by O. D. L. Strack (1989) at the University of Minnesota. An analytical solution is used for simplified groundwater problems, as it requires less field data than other available groundwater model methods. The advantages of analytical solutions are that they are easy to implement and give consistent and accurate results for *simple groundwater problems*. Unlike numerical solutions, analytical solutions give a continuous output at any point in problem domain as shown in Figure 4.2. However, in the conceptualization of the analytical solutions many assumptions, like homogeneity and isotropy of the groundwater aquifer, are to be made. Generally, these assumptions are not valid in the real field conditions. This is why analytical solutions are not being used for complex groundwater systems. In groundwater model development using AEM, “analytic elements” has been used to represent the line sources and sinks like rivers, streams, drains or specified head and specified flow boundaries. The pumping wells are represented by point sources; evapotranspiration, recharge rate and aquifer characteristics (hydraulic conductivity) are represented by polygons.

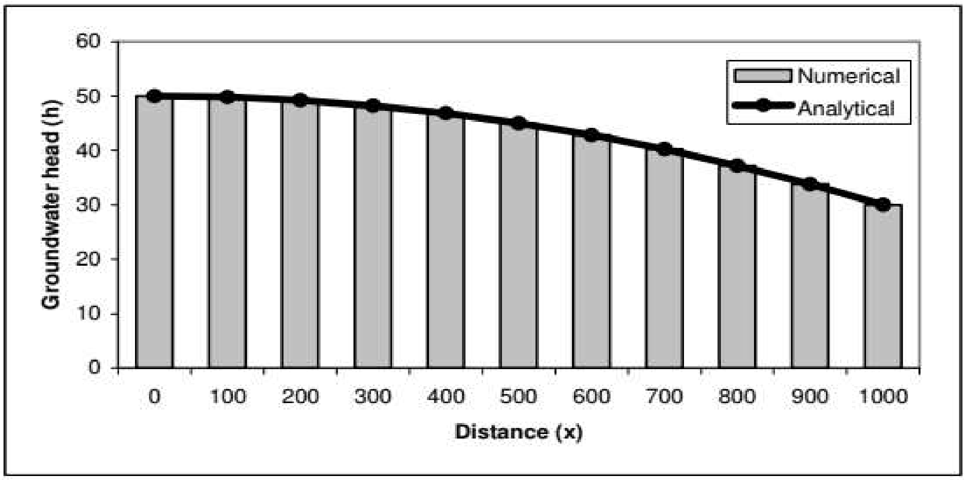

Figure 4.2 Numerical and Analytical solutions for one dimensional groundwater flow.

4.11 Groundwater-surface water (river) interaction

In water resource management, generally one focuses on groundwater or surface water as a separate entity. However, in general, all surface water resources like rivers, lakes, ponds, estuaries, streams and wetlands are interconnected to the groundwater resources. This type of interaction between these two water resources is dynamic in nature and leads to the exchange of the water and solutes between them. Water quantity and quality is affected due to this dynamic interaction. Therefore, for better understanding of this system study of surface water (river) and groundwater (aquifer) interaction is needed. This study will helps in the effective and sustainable land and water management (Winter et al. 1998).

Precipitation contributes to the surface water only and after the infiltration and percolation process, this surface water reaches the groundwater. A hyporheic zone (exchange zone) exists underneath the riverbed where the exchange of the groundwater and surface water takes place (Kazezyilmaz and Medina, 2006). The characteristics and behavior of the hyporheic zone is very complex due to the presence of hydrological and geological actions in the zone (Conant, 2004). Hyporeic zones have the potential to affect the quality and quantity of the groundwater source due to its connection to the surface water source. Any impact in this zone directly affects the groundwater and surface water (Kazezyilmaz and Medina, 2006). If any pumping activity takes place near to this hyporheic zone, impact on the interaction phenomenon will become more significant. Direction of the water flow movement between surface water and groundwater depends upon the hydraulic gradient. Water flux always moves from the higher gradient to the lower one (Townley, 1998).

Interaction between surface water and groundwater bodies can be categorized into three parts based on their spatial and temporal movement (Schaller and Fan, 2009). Local, intermediate and regional flow systems are the three patterns based on the spatial scales as shown in the Figure 4.3. This categorization of the river-aquifer connection suggests that the local flow systems are assumed to occur when the exchanges happen on a small scale, whereas the regional flow system have exchanges at a large scale. Intermediate flow systems occur when they are neither happening at a small scale nor very large scale (Toth, 1963).

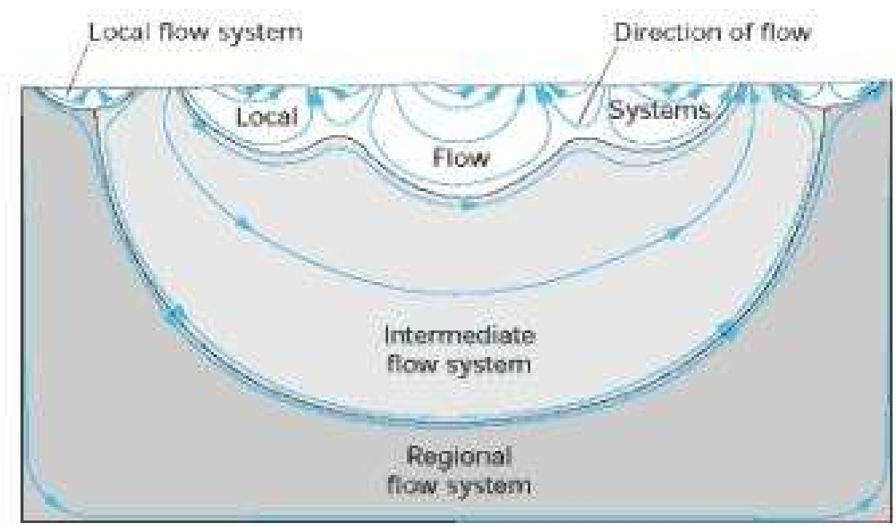

Figure 4.3 Scales of the groundwater system (Modified and adopted by Toth, 1963)

Generally, rivers lose the surface water to the groundwater in different climatic conditions. But is some seasons, the same river may receive the water inflow from the groundwater aquifer. The water quantity of this groundwater inflow varies across the river length (spatially) and in different climatic seasons (temporally). Winter et al. (1998) categorized River-Aquifer interaction into three types, discussed as below:

1. Gaining stream (water movement to the surface water): When the groundwater table adjacent to the riverbed is higher than the water level of the river, then the riverbed will allow the groundwater to seep through it and this will increase the water level of the river through the groundwater.
2. Losing stream (water movement to the groundwater): If the groundwater table, which is contiguous to the riverbed, is below the water level of the river, then it would cause the restoration of the groundwater from the river water by outflow from the river bed.
3. Stable stream flow (no exchange takes place): When the groundwater table and the river water level is at the same level, it means that there is no flow takes place between the surface water and groundwater. However, it is relatively rare to occur like this case for a long river stretch and a longer period. This type of system is termed as *stable stream flow.* Another case of no flow can occur when the groundwater table is so much lower than the water level of the river that an unsaturated zone has formed between them. This unsaturated zone separates the river water from the groundwater. In this case, river is known as a disconnected river from the groundwater system.

4.12 Summary

A study of the flow simulation process along with its methods applicable to the groundwater models is explained in the chapter. The process of developing a conceptual groundwater flow model for an area with boundary conditions and required necessary aquifer properties is investigated with a schematic pattern indicating the steps required to reach a groundwater model with fair accuracy. The basic governing groundwater flow equation is mathematically described. The critical techniques required to assess the accuracy of groundwater models to represent the actual field conditions, such as calibration, validation, are discussed. Another approach to model development, i.e., AEM, has been investigated and highlights the benefits of this method. Since river network exists in the study area, the relationship of the river with the groundwater is also discussed and described. The river aquifer interactions and their types are also discussed.