Chapter 3

THEORETICAL BACKGROUND

3.1 Ground water hydrology

Ground water hydrology may be defined as the science of the occurrence, distribution and movement of water below the surface of the earth.

3.1.1 Ground water occurrence

Usable ground water occurs in permeable geologic formations known as "aquifers" these are the formations having structures that permit appreciable water to move through them under ordinary field conditions. An ‘aquifuge’ is an impermeable formation neither containing nor transmitting water.

3.1.2 Soil water distribution

Soil water was classified by Briggs (1897) into three categories depend upon its concentration in the soil zone. Hygroscopic water absorbed from the air forms thin films of moisture on soil particle surfaces. The adhesive forces are very large, so that this water is not available to plants. Capillary water exists as continuous films around the soil particles. It is held by surface tension, is moved by capillary action and is available to plants. Gravitational water is excess soil water, which drains through the soil under the influence of gravity.

Moisture content

<table>
<thead>
<tr>
<th>Gravitational water</th>
<th>Maximum water capacity</th>
</tr>
</thead>
<tbody>
<tr>
<td>Field capacity</td>
<td>Moisture Equivalent</td>
</tr>
<tr>
<td>Wilting point</td>
<td>Hygroscopic coefficient</td>
</tr>
<tr>
<td>Hygroscopic water</td>
<td>Zero vapor pressure</td>
</tr>
</tbody>
</table>
3.1.3 Field capacity
Field capacity is defined as the amount of water held in the soil after the excess gravitational water has drained away and after the rate of downward movement of water has materially decreased.

3.1.4 Wilting point
The “wilting point” is the moisture content at which permanent wilting of plant occurs.

3.1.5 Hygroscopic coefficient
The maximum moisture which an initially dry soil will absorb in contact with an atmosphere of 50% relative humidity at 25°C.

3.1.6 Moisture equivalent
The amount of water, which a saturated soil will retain after being centrifuged at a centrifugal force 1000 times that of gravity.

3.1.7 Divisions of sub surface water

<table>
<thead>
<tr>
<th>Suspended water (Vadose water)</th>
<th>Soil water</th>
<th>Soil water zone</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pellicular and Gravitational water</td>
<td>Intermediate zone</td>
<td>Zone of aeration</td>
</tr>
<tr>
<td>Capillary water</td>
<td>Capillary zone</td>
<td>Water table</td>
</tr>
<tr>
<td>Ground water</td>
<td></td>
<td>Zone of saturation</td>
</tr>
<tr>
<td>Bedrock</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Ground surface
### 3.1.8 Root constant

The degree to which potential and actual evaporation rates diverge is a function of soil and vegetation characteristics and is explained by Penman (1949) in terms of a “root constant”; This is a measure of the amount of water readily available within the root range, expressed as an equivalent rainfall. (Rushton and Ward, 1979)

### 3.1.9 A conceptual soil moisture budgeting procedure

![Soil moisture budget diagram](image)

(2000)

### 3.1.10 Ground water movement

In order to make the optimum use of groundwater resources, it is necessary to be able to predict its movement.

### 3.1.11 Darcy’s law

Darcy’s law, an equation that describes laminar groundwater movement in aquifers, contains three variables: (1) groundwater velocity, (2) hydraulic gradient, and (3) hydraulic conductivity. (Hall, Lutterll and Cronin, 1991)
In 1895 Henry Darcy, in a treatise on water supply, reported on experiments of the flow of water through sands. He expressed,

\[ v = K \frac{dh}{dL} \]  \hspace{1cm} (18)

where, \( v \) is the velocity of flow, \( \frac{dh}{dL} \) is the hydraulic gradient, \( K \) is the hydraulic conductivity.

Darcy's law is applicable only within the laminar range of flow where resistive forces govern flow. Fortunately for most natural ground water motion, Darcy’s law can be applied. (Varshney, 1986)

### 3.1.12 General form of the ground water flow equation

The flow of groundwater can be described on the basis of techniques developed from fundamental principles governing the movement of water in the saturated zone.

If the velocity components in \( x, y, z \) directions are.

\[ V_x = K \frac{\partial h}{\partial x} \] \hspace{1cm} (19)

\[ V_y = K \frac{\partial h}{\partial y} \] \hspace{1cm} (20)

\[ V_z = K \frac{\partial h}{\partial z} \] \hspace{1cm} (21)

Then the general form of the ground water flow equation in three-dimensional Cartesian coordinates is,

\[ \frac{\partial}{\partial x} \left( K \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left( K \frac{\partial h}{\partial y} \right) + \frac{\partial}{\partial z} \left( K \frac{\partial h}{\partial z} \right) = S \frac{dh}{dt} \] \hspace{1cm} (22)

\[ \frac{\partial}{\partial x} \left( T \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left( T \frac{\partial h}{\partial y} \right) + \frac{\partial}{\partial z} \left( T \frac{\partial h}{\partial z} \right) = S \frac{dh}{dt} \] \hspace{1cm} (23)

where, \( S_s \) is the Specific storage coefficient, \( S \) - Storage coefficient, \( K \) is the hydraulic conductivity and \( T \) is the transmissibility.
3.1.13 Steady flow equation

For steady flow \( \frac{dh}{dt} = 0 \)  
\[ \text{----------------- (24)} \]

In a homogeneous and isotropic medium,  
\[ (K_x) = (K_y) = (K_z) \]  
\[ \text{----------------- (25)} \]

In this case the general equation for steady flow of water in homogeneous and isotropic media is,
\[ \frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = 0 \]  
\[ \text{----------------- (26)} \]

3.1.14 Unsteady flow equation

In a homogeneous and isotropic medium,  
\[ (K_x) = (K_y) = (K_z) = (K) \]  
\[ \text{----------------- (27)} \]

In this case the general equation for steady flow of water in homogeneous and isotropic media is given by,
\[ \frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} + \frac{\partial^2 h}{\partial z^2} = \frac{S}{K_x} \frac{\partial h}{\partial t} \]  
\[ \text{----------------- (28)} \]

3.2 Groundwater Modeling

A groundwater model is a system, which represents the flow of groundwater in a given aquifer.

Application of a simulation numerical model, which enabled evaluation of water balance, water table maps, hydrograph in various cells, and analysis of errors involved the following steps.

(a) Definition of boundaries and boundary conditions (flow, water table)
(b) Division into cells by intersecting lines and columns
(c) Definition of hydraulic coefficients \( T \), \( S \) and thickness
(d) Estimation of recharge-discharge for each cell (inflow-outflow)
(e) Preliminary water table map
(f) Calibration between measured water tables at observation points and calculated water table, by changing parameters in a trial and error method. (Issar and Passchier, 1990)
3.2.1 Objectives of Groundwater Modeling

In general, there are two idealized uses of simulation in hydrology. The first use is the prediction (or forecasting) of future events based upon a calibrated and validated model (Loague and Freeze, 1985). The second use is in the development of concepts for the design of future experiments to improve the understanding of processes (Loague, 1988; Loague et al., 1995).

3.2.2 Data required for developing a groundwater model

All groundwater resource studies are iterative because perfect data are not available and circumstances change in time. The improved assessment that becomes possible once more data are available. (Issar and Passchier, 1990)

The first phase of a groundwater model study consists of collecting all existing geological and hydrological data on the groundwater basin in question. This will include information on surface and subsurface geology, water tables, precipitation, evapotranspiration, pumped abstractions, stream flows, soils, land use, vegetation, irrigation, aquifer characteristics and boundaries. Developing and testing the numerical model requires a set of quantitative hydro geological data that fall into two categories.

<table>
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<th>Physical framework</th>
<th>Hydrological stress</th>
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<td>1 Topography</td>
<td>1 Water-table elevation</td>
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<td>2 Geology</td>
<td>2 Type and extent of recharge areas</td>
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<tr>
<td>3 Types of aquifers</td>
<td>3 Rate of recharge</td>
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<tr>
<td>4 Aquifer thickness and lateral extent</td>
<td>4 Type and extent of discharge areas</td>
</tr>
<tr>
<td>5 Aquifer boundaries</td>
<td>5 Rate of discharge</td>
</tr>
<tr>
<td>6 Lithological variations within the aquifer</td>
<td></td>
</tr>
<tr>
<td>7 Aquifer characteristics</td>
<td></td>
</tr>
</tbody>
</table>
3.2.3 Topography

An accurate topographical map of the groundwater basin to be modeled is a basic requirement. If the basin is small or if a more detailed study of a local problem is to be made, the scale should be 1:50,000, 1:25,000, or even 1:10,000. Whatever the size of the basin or the purpose of the study, the topographical map should show all surface water bodies, streams, and other natural or man-made water course. It should also show contour lines of the land surface elevation. (Boonstra and Ridder, 1981)

3.2.4 Geology

In any groundwater study, the geological history of the basin must be known, as the resulting geological structure largely controls the occurrence and movement of groundwater. The number and type of water-bearing formations, their depth, interconnections, hydraulic properties, and outcrop patterns are all the result of the basin's geological history.

Intensive geomorphological and geological studies of the groundwater basin will be required to delineate its geomorphological features or land forms and to evaluate the manner and the degree in which they contribute to the basin's hydrology. Of special importance are the areas open to deep percolation, the subsurface areas where inflow or outflow to or from the aquifer occurs, the type of material forming the aquifer system, including its permeable and less permeable confining formations, the location and nature of the aquifer's impermeable base, the hydraulic characteristics of the aquifer, and the location of any structures affecting groundwater movement.

Driller's logs can reveal key beds, also called marker beds, which possess a recognizable lithology or fossil content that differs from the beds above and below them. (Boonstra and Ridder, 1981)
3.2.5 Type of aquifers

An aquifer can be defined as a formation; group of formations that contains sufficient saturated permeable material to yield significant quantities of water to a well or spring. The most common and most productive aquifers are unconsolidated sand and gravel.

Groundwater basins are usually defined as “hydro-geological units containing one large aquifer or several connected and interrelated aquifers” (Todd 1980).

Some basins do indeed contain one large aquifer formed in a single depositional environment. Others, especially deeply down warped ones, show several aquifers formed in different environments. Consecutive geological formations, different in age or origin but similar in water transmitting properties, should be grouped into single aquifer system. (Boonstra and Ridder, 1981)

3.2.6 Unconfined aquifer

Unconfined aquifer is one in which a water table serves as the upper surface of the zone of saturation. There is complete freedom for groundwater table to fluctuate. It is also known as a free, phreatic, or non-artesian aquifer. The water table varies in undulating form and in slope, depending upon areas of recharge and discharge, pump-age from wells, and permeability. (Varshney, 1986)

3.2.7 Confined aquifer

When an aquifer is confined on its upper and under surface, by impervious rock formations (i.e. aquicludes), and is also broadly inclined so as to expose the aquifer somewhere to the catchment area at a higher level for the creation of sufficient hydraulic head, it is called a confined aquifer. (Santosh Kumar Garg, 1989)
3.2.8 Aquifer thickness and lateral extent

Besides an isopach map of the aquifer, the numerical model requires a structural map of the aquifer's impermeable base. Such a map shows configuration and elevation of the surface of the base. To construct it, one uses the logs of all wells and bore holes that struck the impermeable base. (Boonstra and Ridder, 1981)

3.2.9 Aquifer boundaries

The conditions at the boundaries of the aquifer must be properly defined. Different types of boundaries exist, which may or may not be a function of time. (Boonstra and Ridder, 1981)

They are,

1) Zero - flow boundaries
2) Head controlled boundaries
3) Flow controlled boundaries

Mapping the aquifers, non-aquifers, and interconnections and establishing the hydraulic nature of the boundaries is essentially a geological exercise with a hydro geological interpretation superimposed. Establishing boundary conditions is essential for modeling, but does not come easily to those who with (hydro-) geological training. (Issar and Passchier, 1990)

If artificial boundary conditions are imposed in the aquifers, to minimize the effect of the artificial boundary conditions, the simulation domain should be extended well beyond the study area. (Mukhopadhyay, Al-Sulaimi and Barret, 1994)

The outer model boundaries should be set far enough from the area of interest to minimize their influence on results over the area of interest. (Mergia and Kelly, 1994)
3.2.10 A zero - flow boundary

A zero - flow boundary is a boundary through which no flow occurs. Examples of zero - flow boundaries are thick tight compacted clay layers, un-weathered massive rock, and a fault that isolates the aquifer from other permeable strata or a groundwater divides.

In practice, zero-flow boundaries can be defined as those places where flows are insignificant compared with the flows in the main aquifers. A groundwater divide, by definition, is a zero flow boundary as no flow occurs across the streamline running over the top of the divide.

In mathematical terms, the condition at a zero flow boundary is \( \frac{\partial h}{\partial n} = 0 \),

where \( h \) is the groundwater potential and \( n \) is the direction normal to the boundary. In the groundwater model, zero flow is simulated by setting the hydraulic conductivity at the boundary equal to zero (K=0). (Boonstra and Ridder, 1981)

3.2.11 A head controlled boundary

A head controlled boundary is a boundary with a known potential or hydraulic head, which may or may not be a function of time. Examples are large water bodies like lakes and oceans whose water levels are not affected by events within the groundwater basin.

Mathematically, a head controlled boundary that changes with time is expressed as \( h = f(x, y) \), i.e. the head is a function of place only. (Boonstra and Ridder, 1981)
3.2.12 A flow-controlled boundary

A flow-controlled boundary, also called recharged boundary, is a boundary through which a certain volume of groundwater enters the aquifer per unit of time from adjacent strata whose hydraulic head and/or transmissivity are not known. The quantity of water transferred in this way usually has to be estimated from rainfall and runoff data.

Flow-controlled boundaries are simulated by setting the hydraulic conductivity at the boundary equal to zero \( K=0 \), and entering the underflow into the model as a recharge term.

Mathematically the flow is represented, for steady state, by the normal gradient \( \frac{\partial h}{\partial n} \), taking a specified value \( \frac{\partial h}{\partial n} = - (\text{Velocity normal to boundary} / \text{permeability normal to boundary}) \).

When modeling a groundwater basin, it is adviceable to let the external boundaries of the model coincide with head-controlled and/or zero flow boundaries. (Boonstra and Ridder, 1981)

3.2.13 Aquifer characteristics

Without any doubt, aquifer tests are the most reliable methods of determining aquifer characteristics. The disadvantage of aquifer test is their high cost. (Boonstra and Ridder, 1981)

Theoretical and empirical investigations of the relation between particle size and inter-granular permeability have resulted in the well-known formula for intrinsic permeability \( k \)

\[ k = cd^2 \]

where, \( d \) is particle diameter, and \( c \) is a dimensionless constant. (Shepherd, 1989)
Data from classical analytical methods involving pump tests as well as statistical analysis using grain size characteristics of the aquifer seldom match the standard procedures. Hence, such procedures might be subject to difficulties in obtaining accurate results due to constructive assumptions and limitations. Nevertheless, aquifer test procedures may give reasonable results but the field data collection for such techniques requires costly boreholes. (Alyamani and Sen, 1993)

The slug test is one of the most common techniques for the in situ estimation of hydraulic conductivity in unconfined flow systems. (Hyder and Butler, 1994)

The use of specific capacity of a well to estimate the transmissivity of an aquifer is widespread because of the availability of specific capacity data from well driller’s logs and the relative expense of obtaining transmissivity through aquifer testing. The most common approach assumes that transmissivity is linearly proportional to the specific capacity of a well and that the constant of proportionality can be obtained by the Dupuit – Thiem equation. (Huntley, Nommensen and Steffey, 1992)

For areas with no existing field data, Transmissivity and storativity values were given to the aquifer according to hydrological test results of geologically similar areas. The computer simulation model supported the main water flow assumptions with only slight alterations in the transmissivity values. (Issar and Passchier, 1990)
3.2.14 Type and extent of recharge areas

An aquifer recharge area is defined as that area of the water table through which all water that flows in an aquifer enters the ground-water system. (Buxton and Modica, 1992)

Recharge varies across catchments because the controlling factors vary, both in their nature and size. These factors include;
- precipitation and other water supplies
- geology and soil
- vegetation and land use
- topography and landform
- groundwater condition

Because recharge is a non-linear process, it is not possible to use average values of each controlling factor to derive an average recharge. Recharge should be estimated separately for each homogeneous zone; the spatially varying values are of course essential for groundwater modeling studies. (Lerner, 1990)

3.2.15 Rate of recharge

Groundwater recharge may be defined in a general sense as the downward flow of water reaching the water table, forming an addition to the groundwater reservoir.

Recharge of groundwater may occur naturally from precipitation, rivers, canals and lakes and as a man-induced phenomenon via such activities as irrigation and urbanization. (Simmers, 1990)

A useful conceptual distinction is between actual recharge, which is the water that reaches the water table and potential recharge which is available, but which may go to another destination. (Lerner, 1990)
3.2.16 Main recharge sources

There can be several sources of recharge to a groundwater system. (Lerner, 1990)

The main recharge sources are,

1. Precipitation or direct recharge
2. River recharge,
3. Inter-aquifer flows
4. Irrigation losses, both from canals and fields
5. Urban recharge

3.2.17 Methods for recharge estimation

The total recharge to an aquifer is often estimated from the volume of water stored as the water table rises during the wet season. (Lerner, 1990)

Total recharge = change in saturated volume x specific yield + outflows

Each type of recharge can be quantified by several methods. There are similarities between methods for different recharges. The methods have been grouped into:

1. Direct measurement
2. Darcian approaches
3. Other, mainly empirical, methods.
4. Water balance methods
5. Tracer techniques

Direct measurement requires construction and is expensive, particularly as it only provides a point measurement.

Water balance methods estimates recharge as the residual of all the other fluxes. The principle is that other fluxes can be measured or estimated more easily than recharge. In soil moisture budgets, rainfall and potential evapotranspiration are inputs to soil moisture accounting procedure, with actual evapotranspiration and recharge as the outputs.
The advantage of water balance methods are that they use readily available data (rainfall, runoff, water levels), are rapid to apply, and they account for all water entering the system. Methods are available for all recharge sources; often they are the only feasible type of method. The major disadvantage is that errors can be high.

Darcian method's principal advantage of is that they attempt to represent the flow of water in the actual physical processes that we are interested in. However data requirements and computing load are high. Tracers do not measure water flow directly. Therefore a number of problems can arise, leading to over or under estimate of recharge.

Other methods are mainly empirical, in is which recharge is correlated with other variables (precipitation, elevation, canal flow). The relationship then used (a) to extend the recharge record in time, or (b) is transposed to other catchments of similar characteristics. (Lerner, 1990)

3.2.18 Direct recharge

Direct recharge is defined as water added to the groundwater reservoir in excess of soil moisture deficits and evapotranspiration, by direct vertical percolation of precipitation through the unsaturated zone. (Simmers, 1990)

Direct recharge is the recharge below the point of impact of the precipitation including moisture storage above the water table (Lerner, 1990).

Recharge = precipitation − runoff − actual evapotranspiration + storage change

The outcome of an evaluation of different rainfall recharge equations proved the following simple relationship:

\[ RE(I) = A(RF(I) - B) - SMD \]

Where, \( RE(I) \) denotes the recharge for year \( I \); \( RF(I) \) is the rainfall for year \( I \); \( A,B \) are lumped parameters optimized and \( SMD \) represents the accumulated soil moisture deficit which is also an integrated parameter. (Bredenkamp, 1990)
3.2.19 Methods for estimating direct recharge

(1) Direct measurement over areas up to 100 m²
(2) Empirical methods, usually simplify eqn to: Recharge = f (precipitation)
(3) Water budget method (usually soil moisture budgeting methods)
(4) Darcian approaches, that is making use of the flow equation.
(5) Environmental or applied tracers

3.2.20 Evapotranspiration

Good data on actual evapotranspiration is equally important as good precipitation data. Unfortunately actual evapotranspiration is rarely measured except in research projects.

A number of formulae have been devised for evapotranspiration, of which the Penman-Monteith (Monteith, 1965, 1981) is generally considered the best for actual evapotranspiration in humid climates. It is written

\[ e_t = \frac{sH + dc(e_a - e_s)/r_a}{(s + g(f + r_a/r_s))} \]  \hspace{1cm} (29)

- \( e_t \) - Actual evapotranspiration (kg/m²/s)
- \( s \) - Slope of saturated vapour pressure curve (mbar/K)
- \( H \) - Available energy = net radiation - soil heat flux - heat storage in vegetation (W/m²)
- \( d \) - Density of air (kg/m³)
- \( c \) - Specific heat of air (J/kg/K)
- \( e_s \) - Saturated vapour pressure at air temperature (mbar)
- \( e_a \) - Vapour pressure at screen height (mbar)
- \( l \) - Latent heat of vaporization (J/kg)
- \( r_a \) - Aerodynamic resistance (s/m)
- \( r_s \) - Stomatal resistance (s/m)
- \( g \) - Psychrometric constant (mbar/K)

Comparison of the recharge based on daily and average monthly potential evaporation values suggests that the use of monthly evaporation data lead to an underestimate of about 3%. (Rushton and Ward, 1979)
3.2.21 Indirect recharge
Indirect recharge results from percolation to the water table following runoff and localization in joints, as ponding in low-lying areas and lakes, or through the beds of surface watercourses. (Simmers, 1990)

3.2.22 Recharge from rivers
Recharge from rivers is probably the most difficult type of natural recharge to estimate. The river flow, the riverbed, and the aquifer control the amount of recharge. Controlling factors may include the following.

River flow: flow rate, depth, flow volume, peak flow, velocity, duration of flow, frequency of flows, temperature (affects hydraulic conductivity), silt content;

Riverbed: antecedent conditions, width, hydraulic conductivity;

Aquifer: boundaries, depth of water table, hydraulic conductivity, moisture retention in unsaturated zone.

General procedure for estimating river recharge:
(1) Consider how much water can be accepted by the aquifer
(2) Estimate the transmission capacity of the unsaturated zone, if there is one.
(3) Finally consider the river flow and riverbed process, estimating possible deep percolation and all other items of a water balance such as evaporation from the bed.

Direct measurement of river recharge is not possible. Empirical methods have been devised and are widely used. Water balance methods are possible for all three items, transmission loss, deep percolation and recharge. Darcian approaches that are based on the groundwater flow equation are possible in theory but are difficult to use in practice with important simplifications. Tracer techniques have very limited use for quantifying river recharge. (Lerner, 1990)
3.2.23 Irrigation losses, both from canals and fields

Losses from irrigation occur in two overlapping areas (a) canals (b) fields.

3.2.24 Water losses from irrigation canals.

Total loss from a canal is the sum of evaporative and seepage losses. Losses due to evaporation can be calculated by the Penman equation. Seepage losses may either flow to shallow water table and thence to drains and evaporation, or recharge deep groundwater systems.

Figures taken from the Periyar Vaigi Project in south India (Rushton, 1986) is reproduced in table below.

<table>
<thead>
<tr>
<th>Type of canal</th>
<th>Seepage losses (m/d)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Unlined</td>
</tr>
<tr>
<td>Main canal</td>
<td>0.37</td>
</tr>
<tr>
<td>Large distributary</td>
<td>0.18</td>
</tr>
<tr>
<td>Medium distributary</td>
<td>0.09</td>
</tr>
<tr>
<td>Small distributary</td>
<td>0.06</td>
</tr>
</tbody>
</table>

According to Lerner (1990), several methods are available to estimate recharge from irrigation canals. A direct measurement has been used for seepage rates through soil beds. Water balances are commonly and successfully used to estimate canal losses. Tracer techniques have not been widely used. Darcian approaches include flow nets, analytical solutions to the flow equations, and numerical models. Empirical formulas are perhaps more widely used for this type of recharge than any other. But these formulas are based on prolonged observation of canals within a region, and as such will not be valid for canal in the other regions with different field conditions (Krishnamurthy & Rao, 1969).
3.2.25 Recharge from irrigated fields

Recharge by deep percolation from irrigated fields has many similarities with recharge by precipitation, and the same methods can often be used to estimate it.

For direct measurement, Lysimeters can be used in principle. There are obvious difficulties, which may make their use impractical. Tracer methods are commonly used for estimating recharge from irrigated fields. Darcian approaches are equally applicable to irrigation fields. Water balance methods are more valid when there is a regular supply of irrigation water.

Inflow – outflow balances of irrigated areas have sometimes proved successful in estimating net recharge. These will take account of canal inflows, precipitation and other water sources, while outflows to be measured include drainage and crop water use. (Lerner & Gray, 1990)

Detailed field and model studies (Walker and Rushton, 1984) have shown that substantial quantities of water can pass through the bunds of rise fields to the underlying aquifers. This occurs because the puddle layer does not continue under the bund. (Rushton, 1990)

3.2.26 Recharge from Irrigation storages

Seepage from lakes could be a major source of inflow.

As a conclusion, it is obvious that the most reliable results are obtained in estimating natural recharge, if all the available data are used and a variety of methods can be applied. (Romijn, 1990)
3.2.27 Type and extent of discharge areas

An estimate of discharge over the aquifer boundaries provides an estimate of net recharge.

For an unexploited aquifer,
Average discharge = average net recharge.

For an exploited aquifer, which in general will not be in steady state,
Average discharge = average net recharge + rate of storage depletion

The possible routes of from a groundwater basin include springs, rivers, lakes and seas, evapotranspiration, abstraction by man and storage changes. Springs are discrete discharge points, usually at changes in lithology or in fracture zones. Rivers receive groundwater either at discrete points or as diffuse seepage. Lakes and seas also receive groundwater either at discrete points or as diffuse seepage. Evapotranspiration by vegetation may be a significant discharge of groundwater in arid and semi - arid areas. Abstractions by man are through wells, boreholes or drainage.

A portion of the water taken for irrigation or domestic use may well return to the aquifer locally. Storage changes would not normally be considered as a discharge. (Lerner, 1990)

3.2.28 Rate of discharge

Discharge from the shallow wells can be estimated from the number of wells and the method used for withdrawing the water. (Rushton, 1990)
This will include Q of abstraction within an aquifer.
3.2.29 Calibration

Calibration is generally defined as the adjustment of parameter values within known ranges to simulate the measured state of the flow system. (Bair, 1994)

However, because of its complexity, most practitioners still rely on “trial and error” methods throughout the world. (Olsthoorn, 1995)

Obviously, the longer the period used for calibration, the better the results will be. This is particularly so for unconfined aquifers, which have a long natural response time (Rushton and Redshaw 1979). As long-term records are seldom available, however, the model usually has to be calibrated with data covering only relatively short period, which, if possible, should be selected that extremes of water table behavior have occurred within it. The absolute minimum period, however, is two full years of data; the first year is being used to adjust the input data and the second year serving as a check to see whether the adjustments were adequate. If not, the process is repeated. (Boonstra and Ridder, 1981)

3.2.30 Validation

Validation is the simulation of a different measured state of the flow system using the final parameter values from model calibration. (Bair, 1994)

There have been a number of definitions of model validation. The international atomic energy agency (IAEA, 1982) defines validation as follows: “A conceptual model and the computer code derived from it are validated when it is confirmed that the conceptual model and the computer code provide a good representation of the actual processes occurring in the real system”.

Up to now, most validation efforts are simply a comparison of modeling results against field data. Since the goal of model validation is to ensure the modeling results provide a good representation of the actual processes in the real system (IAEA, 1982), validation should be applied to every step of the modeling process.
Steps of the Modeling Process and Their Validation (Tsang, 1991)

<table>
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<tr>
<th>The Modeling Process</th>
<th>Examples of Issues Requiring Validation</th>
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<tbody>
<tr>
<td>1. Data Review and Evaluation</td>
<td>Spatial correlation and parameter correlation.</td>
</tr>
<tr>
<td>3. Performance Criteria</td>
<td>Appropriate choice of quantities of interest. Are the criteria unnecessarily demanding?</td>
</tr>
<tr>
<td>4. Calculation models and Lumped Parameters for all “reasonable” alternative conceptual models and scenarios.</td>
<td>Simplification procedures and determination of lumped parameter from data.</td>
</tr>
</tbody>
</table>

6. Results Evaluation:
   (a) Uncertainty too large:
       Define new data needs;
       Design new site
       Characterization activities:
       • Feasible to perform further field studies, update data.
       • Not feasible within reasonable time and cost
   (b) Results with estimated uncertainty good enough

   ![Diagram](#)

   - GO TO STEP 1
   - STOP
   - INPUT TO DECISION MAKING