CHAPTER-IV
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ESTIMATION OF NATURAL RECHARGE

4.1 INTRODUCTION:

Natural recharge is a hydrogeologic parameter, governed primarily by the permeability of the soil and the subsurface. Hence textures of the soils and the subsurface grain size distribution below the soil zones determine the magnitude of natural recharge. Moreover, natural recharge is controlled by the terrain characters like geomorphology and the metrological parameters like rainfall and its intensity. As it is dependant on many variable parameters, its estimation will be very difficult. The thesis presents an interesting relation of this most important but complicated parameter with the geophysical parameter, which is obtained by carrying out simple geophysical investigations on the surface (vertical electrical sounding). The natural recharge, on the other hand, is obtained by carrying out cumbersome subsurface hydrogeological investigations. Geophysical methods are quicker and cost effective, unlike the hydrogeological investigations used to estimate the natural recharge, which are complicated and cumbersome and moreover time consuming. This is because the various hydrological and hydrogeological parameters required to estimate natural recharge are complicated for estimation and vary in space and time. Hence the financial expenditures involved in the estimation of natural recharge are many times higher than the geophysical investigations. The significant qualitative relation recognized between the simple geophysical parameter and the complex hydrogeological parameter of natural recharge could facilitate in optimizing the locations required for recharge estimations when recharge estimates are to be obtained for large regions. It is therefore essential to understand the processes and mechanisms of natural
recharge and the various established hydrogeological methods widely used to estimate natural recharge.

4.2. PROCESSES AND MECHANISM OF RECHARGE:

Some portion of the precipitation falling on the surface, flows into the open gullies and channels and is responsible for the drainage. The drainage in turn is the cause of surface runoff. Most of the precipitation is absorbed by the surface. Some part of this flow as interflow and the major part moves further downwards through the soil, subsoil and the rocky layers. The process of natural recharge through the unsaturated zone occurs in three steps. They are infiltration, storage in soil moisture and down movement and deep percolation to the water table.

Infiltration takes place namely due to rainfall or due to irrigation. During the process of infiltration, the near surface layers will be saturated immediately. However, comparatively the moisture content in the deeper layers is appreciably less. In such a subsurface environment, by virtue of suction gradient, gravitational gradient and adsorptive forces, the moisture moves down words in the soil profile. This is not always a continuous process because variation in rainfall intensity and infiltration causes variation in the depth of the wetted soil in the subsurface (Ward 1975). When the infiltration front reaches the capillary fringe, it displaces the pore spaces and hence the water table rises. Time taken for the infiltrating water to reach the water table is a function of thickness and the hydraulic conductivity of the unsaturated zone. Presence of low permeability layers like silt and clay can retard the recharge rate. The time lag for the infiltrating water to reach the water table may be few hours in humid regions for very coarse soils with shallow water table. In arid environment, the water may take years to pass through the unsaturated zone (Fetter 1988).
Datta et al. (1980b) has put forth the following mechanism, which causes the downward percolation of water and recharges the aquifer. When a storm of high intensity rainfall occurs, the rate of rainfall exceeds the infiltration capacity of the soil. A saturated water sheet is formed within the top surface layer, and the soil moisture movement is due to the movement and dissipation of this sheet. During the interim spells, in addition to downward movement and dissipation of the infiltration sheet formed during the pervious rainfall, evapotranspiration loss from above the sheet reduces its moisture content till it is below field capacity. Hence the next infiltration sheet must first satisfy the moisture deficit before proceeding further. This means that infiltration of rainwater and thereafter the recharge to the groundwater reservoir takes place in pulses.

4.3: FACTORS AFFECTING RECHARGE:

The factors affecting recharge are amount and intensity of rainfall, time gaps between rainfall events, infiltration capacity of surface soils, evapotranspiration, surface topography, compaction and structure of soil, hydrologic conditions etc. These conditions impose a limit on the minimum amount of threshold rainfall required to initiate and continue the process of natural recharge. The rate at which the recharge to the water table takes place depends on the thickness of the unsaturated zone. In saturated zones time required for the recharging water to reach the water table is less. In the upland regions the time required for the recharging water to reach the water table is relatively more. Due to the above-mentioned important hydrogeological factors, recharge varies across the catchment areas because the controlling factors vary in nature and size.
4.4: METHODS OF RECHARGE ESTIMATION:

The methods of estimation can be divided into main groups. A) Conventional or Hydrogeological Methods and B) Radioactive Tracer Methods. The important conventional methods are Empirical methods, Water Balance Methods, Groundwater Balance Method, Lysimeter Method, Storage Method and the use of Darcy’s Law.

Empirical method is the simplest one. A simple empirical formula considers recharge as a proportion of precipitation.

\[ R = f \times P \]

Where
- \( R \) = average recharge rate (mm/year)
- \( P \) = Annual precipitation (mm/year)
- \( f \) = Empirical constant

'\( f \)' varies with the terrain and climate.

A second level of the formula includes a threshold. For example the formula expressed by Chaturvedi for recharge in India is (Sinha and Sharma, 1988);

\[ R = 50.8 \times (P / 25.4 - 15)^{0.4} \quad (P \text{ is }> 380 \text{ mm/year}) \]

Where
- \( R \) = Average recharge (mm/year) and
- \( P \) = Annual precipitation (mm/year)

According to Allison (1988) the empirical methods are useful to estimate recharge where annual recharge is fairly high, say greater than 50 mm/year.

Soil water balance models are useful in estimating long term recharge to the groundwater system. The method is cumbersome, as it requires a large data obtained from measurements of daily rainfall, evapotranspiration (ETP), water holding capacities of soils, saturated field capacities, wilting point moisture values and runoff generated in the basin or any study area, made in the field.
In this method, it is presumed that rainfall in excess of ETP losses is utilized in bringing the soil moisture up to its field capacity and the rest is available as surplus water to be shared by recharge and runoff. Ignoring the runoff, the soil water balance can be represented by the equation,

\[ R = p - (E_a + \Delta s) \]

Where,

- \( R \) = Recharge
- \( P \) = Precipitation
- \( E_a \) = Actual evapotranspiration
- \( \Delta s \) = Change in soil moisture storage

The data required to estimate recharge by this method is large and cumbersome for acquisition. The parameters used to estimate recharge are complicated variables in space and time. Hence Howard and Lloyd (1979) suggested that frequency of data acquisition should be less than ten days. Rainfall and pan evaporation can be recorded regularly. A range of techniques for estimation of actual evapotranspiration based on Penman type equations are available. The method works well for seasonal patterns of recharge and in well developed soils which do not dry completely, when actual and potential evapotranspiration are similar and with wide spread and relatively uniform precipitation.

Groundwater balance method applies the hydrologic equation for groundwater regime to estimate recharge. The equation is expressed as;

\[ \text{[Surface inflow + Subsurface inflow + precipitation + imported water} + \]
Decrease in surface storage + Decrease in groundwater storage = Surface outflow + Subsurface outflow + consumptive use + exported water + Increase in surface storage + Increase in groundwater storage (Todd, 1959)

A few of these parameters can be directly measured. Few others can be determined by estimating the difference between measured volumes and some require indirect methods of estimation. The method uses easy to obtain parameters like rainfall, runoff, water level etc; and accounts for all the water entering the system. However a major drawback of the method is that the recharge is estimated as a residual product. Recharge is a small difference between large numbers. Hence errors can be high.

Lysimeter is a system filled with soil of the experimental area. This is a direct measurement of recharge and evapotranspiration. Lysimeters are weighed on weigh-bridges, sensitive enough to record rainfall as low as 1 mm. The density of the soil and the slope of the surface are maintained similar to natural conditions as far as possible. Recharge is measured by collecting the percolating water at certain depth, with the help of a funnel. Examples of use of construction and use of Lysimeters for recharge measurements are given in detail by Kitching et al. (1977, 80, 82), Kitching and Bridge (1974), Kitching and Shearer (1982).

Though the method is direct measurement of recharge, insitu conditions of the area or site are disturbed during the process of measurement. However, when the topography and the soil conditions are fairly uniform and show consistent behavior, the recharge estimations can be taken up for larger areas.

Storage method of estimating the recharge utilizes the knowledge of moisture regime of unsaturated and saturated zone. The change occurring in the
unsaturated zone (Δ μ) and the amount of water stored in the saturated zone (Δ ms) are directly related to recharge (Walton, 1970).

(ρ μ) is calculated by recording periodically the moisture distribution in soil profiles, by using neutron moisture gauges, at representative sites in a given area. Hydraulic potential is measured by using tensiometers. This data along with the rainfall, runoff and ETP (evapotranspiration) are used to calculate the soil moisture balance and the recharge, for the period of investigation over an area (Vinsen & Mahar, 1981).

(Δms) is the net change that occurs in the quantity of water stored in the saturated zone. It is calculated by continuously recording the water levels in the observation wells. The change in the water level along with the data on specific yield can be used to estimate the recharge. Specific yield is defined as the volume of water released by per unit surface area for per unit decline of water table of an unconfined aquifer (Krueseman & Rider 1990). The specific yield is expressed in percentage. Specific yield of the aquifer is determined through pumping tests. Specific yield of the geological formations, in the zone of water table fluctuations, computed from pumping tests are as follows. Sandy alluvial areas: 12-18 %, Valley fills: 10-14 %, Silty or clayey area : 5-12 %, Granites 2-4 %, Basalts (give the value) %, Laterite: 2-4 %, Weathered Phyllites, shale's, Schist's and other associated rocks, 1-3 %, Sand stones 1-8 %, Limestone's: 3%, Highly karstified Limestone's: 7 %. Recharge is obtained by using the following equation.

\[
Gr = h \times Sy
\]

Where Gr = Groundwater recharge in meters.

\[
h = \text{water level fluctuation in meters}
\]

\[
Sy = \text{Specific yield in percentage (\%)}
\]
Using the premonsoon and post monsoon water level data computes water level fluctuations. Recharge to the groundwater system can be computed for large areas, for a representative aquifer, by using area representative data of water level fluctuation and the specific yield data (CGWB 1979).

4.5: RADIOACTIVE TRACER METHODS OF RECHARGE ESTIMATION:

The application of radioactive (nuclear techniques) for the estimation of natural recharge is the most latest among all methods of recharge estimation. These are called as the tracer techniques.

Tracer is defined as a substance such as a labeled chemical or an atom, which is used to trace the course of any physico-chemical processes such as various stages of the hydrogeological cycle. For studying hydrogeological processes, the tracer should be such that it behaves like water and be synonymous, but at the same time should be distinguishable for purpose of detection. Tracer used in groundwater studies, must posses the following most essential and important properties (Athavale, 1980).

I) The tracer should follow all movements of water under investigation as closely as possible.

II) It should be easily detectable.

III) The tracer should not change the hydraulic characteristics Of the medium (unsaturated zone incase of natural recharge).

IV) It should not be adsorbed by the medium.

V) It should not be present in large concentrations in the formation under study.
VI) The tracer should possess low toxicity, to ensure safety of the user handling the radiotracer and to avoid environmental hazards in the study area.

VII) The most important character, which is a must, is its half-life. The tracer should possess an optimum half-life, long enough to sustain the duration of the experiment and short enough to minimize radioactive contamination.

Various types of tracers, both natural and artificial, have been used in studying hydrological problems in general and in the unsaturated zone in particular. The natural tracers used are the environmental Tritium \( (3\text{H}) \), Deuterium \( (\text{D or}2\text{H}) \), Oxygen-18 \( (18\text{O}) \), Chloride \( (\text{Cl}^-) \) etc. The thesis uses the data of injected tritium method to estimate natural recharge. Hence the same is dealt in detail, in the text to follow.

4.6: NATURAL RECHARGE BY INJECTED TRITIUM TECHNIQUE:

Tritium \( (3\text{H}) \), is a heavier isotope of hydrogen. It is used for estimating natural recharge at any site. Tritium is radioactive and emits beta \((\beta^-)\) radiation. A known quantity of Tritium is artificially injected into the subsurface at a particular depth, before monsoon. The same is again retrieved after the monsoon to know the vertical movement of the tracer. Tritium \( (3\text{H}) \) is artificially produced by irradiating Lithium \( (\text{Li}) \) with neutrons in a nuclear reactor. The reaction is as follows.

\[
\begin{align*}
3^6\text{Li} + 0^1_n & \rightarrow 1^3\text{H} + 2^4\text{He} \\
\end{align*}
\]

The radioactive material (tritium) decays spontaneously to a Helium isotope as follows.

\[
1^3\text{H} \rightarrow 2^3\text{He} + \beta^-
\]
Tritium is the most suitable radioactive tracer for soil moisture movement and recharge estimation. It can be incorporated into water molecule as tritiated water (HTO). It is a soft ($\beta$) emitter of low energy. (18 kilo electron volt maximum) and belongs to lowest radio toxicity class. It has a half-life of 12.26 years and can be measured with high detection sensitivity. The half-life has dual advantage. It is short enough to maintain an efficient soft ($\beta$) emission and facilitates reduction of radioactive contamination. The half-life is long enough to facilitate the detection of ($\beta$) activity during the period of the experiment. Being a soft ($\beta$) emitter, it cannot easily be measured in the field and hence the measurements are required to be carried out in the laboratory. The unit of measuring radioactivity in tritium is curie (Ci). Table 4.1 shows the various units in an ascending order.

<table>
<thead>
<tr>
<th>TABLE 4.1: Units of Measurement of Beeta ($\beta$) Radiation</th>
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<tbody>
<tr>
<td>1 Picocurie = 1 Ci x $10^{-12}$ Disintegrations per seconds</td>
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<tr>
<td>1 Nanocurie = 1 Ci x $10^{-9}$ Disintegrations per seconds</td>
</tr>
<tr>
<td>1 Micro curie = 1 Ci x $10^{-6}$ Disintegrations per seconds</td>
</tr>
<tr>
<td>1 Mill curie = 1 Ci x $10^{-3}$ Disintegrations per seconds</td>
</tr>
<tr>
<td>1 Curie (Ci) = 3.7 x $10^{10}$ Disintegrations per seconds (dps)</td>
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The thesis uses the recharge estimated by injected tracer technique for comparison with the geophysical parameters, the chapter to follow explains in detail, the method of estimation of recharge by using the injected tracer (Tritium) technique.