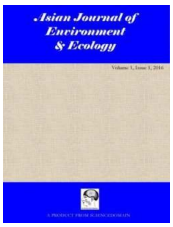

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Approaches and Methods of Quantifying Natural Groundwater Recharge – A Review

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Authors' contributions

This work was carried out in collaboration between both authors. Author MHA designed the study, wrote the protocol and wrote the first draft of the manuscript. Author SM managed the format of the study, performed literature review and formatted the tables. Both authors read and approved the final manuscript.

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ABSTRACT

Groundwater is the main source of water supply to both urban and rural populations as well as to industry and agriculture. Among various water cycle characteristics, groundwater recharge is the leading hydrologic parameter determining groundwater resources availability and sustainability. Accurate estimation of groundwater recharge is extremely important for proper development and management of the resource. Different approaches and methods are available to quantify groundwater recharge – from direct approaches, inferred from more easily measurable physical and chemical parameters, to simulation models with varying complexity. The methods have their own merits, demerits, and limitations. For proper selection of a method in a particular geo-hydrologic and climatic condition, detail knowledge of the methods along with their applicability/limitations, and the governing factors affecting recharge are essential. This paper presents an overview of the methods along with the theory underlying the methods (physical basis), assumptions, advantages, limitations, and selection procedure under the prevailing situation of technological, hydro-geological, and

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resource availability; with a view to help proper selection of a method. The overview synthesized and exemplified the above issues, and concludes with discussion of challenges and research needs in this evolving field.

Keywords: Groundwater; recharge; sustainability; hydro-geology; aquifer; isotope.

1. INTRODUCTION

Water is available on the planet both on the surface and under the surface of the earth. Surface water is renewable, usually within few months or a year, while groundwater is completely renewable, as it takes several months or a year. Many reasons make groundwater a good choice for a water supply. At present world, groundwater is the main source of water supply to both urban and rural populations as well as to industry and agriculture [1]. As demand for groundwater increases, groundwater managers are faced with the difficult task of ensuring the future viability of the resource [2]. With the rise in public environmental awareness, groundwater managers are also concerned with protecting natural environments that are dependent upon the ground water, such as stream base-flows, riparian vegetation, aquatic ecosystems, and wetlands. Sustainable use of groundwater must ensure not only that the future resource is not threatened by overuse and depletion, but also those natural environments that depend on the resource [3]. Trade-offs between groundwater use and potential environmental impacts always will exist, and therefore a balanced approach to water-use between development and environmental requirements needs to be adopted [4]. To properly manage groundwater resources, managers need accurate information about the inputs (i.e., recharge) and outputs (i.e., pumpage and natural discharge) within each groundwater basin, so that the long-term behavior of the aquifer and its sustainable yield can be estimated or reassessed [4,5]. Recharge is a major component of the groundwater system and has important implications for shallow groundwater quality.

Quantitative determination of the rate of natural groundwater recharge is a pre-requisite for efficient groundwater resource management. It is particularly important in regions with large demands for ground water supplies, where such resources are the keys to economic development. Recharge is critical in any analysis of groundwater systems and the impacts of withdrawing native water from them. It is also required for robust model predictions as groundwater recharge is one of the main drivers of the hydrological system [6]. In water-resource investigations, groundwater models are often used to simulate the flow of water in aquifers, and, when calibrated, may be used to predict long-term behavior of an aquifer under various management schemes. Without a good estimate of recharge and its spatio-temporal distribution, these models become unreliable. Accurate estimates of recharge and recharge mechanisms are also necessary to assess the risk of groundwater contamination, particularly diffuse agricultural contamination and concentrated *or* point-source of contamination (for example, landfill). Spatial and temporal variability of recharge, an inappropriate conceptual model, measurement or calculation errors can lead to inaccurate or non-representative recharge estimates [7].

Realizing the importance of estimation of groundwater recharge, numerous studies focused on various approaches and methods of recharge estimation. These range from simple seepage meter method to complex numerical modeling and isotropic tracer techniques - under different physiographic, climatic condition, technology level, and resource availability situations. Recharge has been estimated by lysimeter (Rushton et al. [8]; Xu and Chen [9]), seepage meter (Otto [10]), water table fluctuation method (Callahan et al. [11]; Ordens et al. [12]; Obuobie et al. [13]; Yin et al. [14]; Misstear et al. [15]; Sibanda et al. [16]), water balance method (Risser et al. [17]); base-flow /hydrograph-separation (Coes et al. [18]), soil moisture budget (Cuthbert et al. [19]; Bakundukize et al. [20]; Yin et al. [14]; Misstear et al. [15]), the zero flux plane method (Scanlon et al. [17]), the Darcy's method (Coes et al. [18]; Flint et al. [21]), inverse modelling (Kendy et al. [22]), hybrid water fluctuation method (Sophocleous [23]); chloride mass balance (Ordens et al. [12]; Obuobie et al. [13]; Huang and Peng [24]), Darcyan flow-net computation (Yuan et al. [25]; Sibanda et al. [16]), empirical methods (Saghravani et al. [26]), radioactive isotopic tracer (Jimenez-Martinez et al. [27]; Wang et al. [28]; Kaste et al. [29]; Rangarajan and Athavale [30]), stable isotopes (Yeh et al. [31]; Sukhija et al. [32]), modeling approach (Yidana et al. [33]; Githui et al. [34]), integrated surface-water – groundwater modeling

approach (Chung et al. [35]), GIS based approach and satellite imageries (Subramania et al. [36]). Approaches for specific local conditions have also been advocated, such as for arid and semi-arid regions (Wood and Sanford [37]), granitic terrain (Chand et al. [38]), land-use change (Walker et al. [39]), etc. Many researchers advocated for using multiple methods to increase reliability in recharge estimate. But it involves huge cost, manpower and instruments.

Although numerous methods have been suggested and used for evaluation of groundwater recharge, this parameter is still the most difficult to measure as far as the evaluation of groundwater resources is concerned. In arid and semi-arid environment, recharge is often heterogeneous. With the increasing aridity of the climate, the recharge flux becomes smaller and variable in space and time (Sibanda et al. [16]). The geological structure/formation play an important role in recharge rate (Marie et al. [40]). Groundwater recharge is indeed a complex function of several factors and mechanisms, including meteorological conditions, soil types, land use, physiographic characteristics, depth to the water-table, antecedent soil moisture, properties of the geological materials, interaction between surface and groundwater, available groundwater storage, etc. Hence, selection of an appropriate method for a particular climatic, geo-hydrologic, land-type and vegetation requires knowledge of system as well as thorough knowledge of the methods themselves such as principles, background, limitations, suitability, and capability/range. This paper highlighted the recharge mechanisms, factors affecting recharge, reviews the recharge estimation methods, and selection procedure of a method with a view to help proper selection of a method for a particular situation.

2. RELEVANT TERMINOLOGIES

Before going to discuss the mechanism of recharge and methods of estimation, some relevant terminologies are defined below.

Actual recharge

It is the recharge in an area under existing conditions of the topographic, hydrologic and hydro-geologic settings.

Potential recharge

The maximum possible recharge that can occur in a geological formation under non-limiting conditions of other factors (such as availability of water, infiltration barrier, aquifer fullness, etc.).

Induced recharge

Recharge to groundwater by infiltration, either natural or anthropogenic, from a body of surface water as a result of the lowering of the groundwater level below the surface-water level.

Direct /diffuse recharge

Direct or diffuse recharge refers to recharge derived from precipitation or irrigation that occurs fairly uniformly over large areas.

Indirect recharge

Recharge that 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-water courses.

Localized/focused recharge

It is the concentrated recharge from small water-bodies such as ponds, depressions, joints *or* rivulets.

Natural recharge

Naturally occurring water added to an aquifer. Natural recharge generally results from snowmelt and precipitation or storm runoff.

Artificial recharge

Artificial act of adding water to an aquifer by means of a recharge project, also the water so added. Artificial recharge can be accomplished via injection wells, spreading basins, or in-stream projects.

Preferential recharge

Recharge that takes place preferentially through macropores, as opposed to *diffuse recharge*, which takes place through the entire vadose porous medium.

Total recharge

Recharge that has occurred to the aquifer, measured before losses.

Net recharge

Total recharge minus losses.

Piston flow or plug flow

Purely advective flow without dispersion or diffusion of the dissolved components.

Recharge area

The area that contributes water to an aquifer. It is normally considered to be the natural area of recharge, as contrasted with a constructed recharge basin.

Recharge coefficient

It is the ratio of recharge to rainfall amount, normally expressed in percentage.

3. MECHANISMS AND PATHWAYS OF RECHARGE

Groundwater recharge estimation can be made with more accuracy if different types of recharge mechanism and their relative contribution/ importance can be assessed from the beginning. The relative importance of direct and indirect recharge mechanisms is one of the criterions for the selection of an appropriate estimation method. The method to be used for estimating recharge would have an impact upon the magnitude of the recharge estimates due to the spatial and temporal scales over which the different methods estimate recharge [41].

In humid and sub-humid regions, precipitation normally exceeds potential evapo-transpiration during most of the years, which leads to almost continuous recharge. In contrast, in arid- and semi-arid regions, no such precipitation surplus exists on an annual time scale. The large spatial and temporal variability of precipitation, and preferential flow can cause gross recharge. The following major processes contribute to recharge in semi-arid and arid regions [42]: recharge on hard-rock outcrops through fissures, cracks and large karst conduits (fracture recharge); colluvial infiltration at the bottom of hill slopes, where the upper part of the hill slope acts as a 'micro-catchment' to concentrate the water; streambed infiltration; and direct recharge by flow through the soil matrix.

Both soil-matrix recharge (piston-flow mechanism) and preferential flow recharge can be present in the same area based on the soil condition and geology (Demile et al. [43]). The regions having significant water-bodies/stream-channels (or wet-lands) during most periods of the year, recharge from the stream-bed contribute a major part of the total recharge. In arid and semiarid regions, recharge may be concentrated in the outcrop of permeable/coarse-grain limestone and fault zone.

4. FACTORS AFFECTING RECHARGE

Recharge can vary substantially both within and between basins because of variations in precipitation, geological and geomorphologic settings. In general, the following factors influence the recharge rate and total recharge: soil, climate, land use, land-surface cover, sub-surface geology, existence of water bodies, storage capacity of the aquifer, depth to aquifer, etc.

Soil factor

Soil type, bulk density, organic matter content, etc. influence the recharge rate. Coarse-grained soils generally result in higher recharge rates than do fine-grained soils. Cook et al. [44] reported negative correlation between clay content in the upper 2 m and the recharge rate. For an event basis consideration, the antecedent soil moisture and the thickness and nature of the unsaturated zone (clay or sand) play the important role in determining recharge.

Land-surface and vegetation

If the surface slope of the land is very steep, water moves quickly to downward, thus there is less time or opportunity to enter into the soil. The reverse is true for flat land. The presence and type of vegetation is of paramount important to the physical processes occurring at land surface (Ali [45]). In particular the runoff, heat and water vapor are controlled by soil cover and leaf area index. The surface energy and water balance varies with vegetation dynamics. Vegetation cover is important in assessing recharge potential at a site. Recharge is generally much greater in non-vegetated than in vegetated regions and greater in areas of annual crops and grasses than in areas of trees and shrubs. Allison et al. [46] reported the impact of vegetation in Australia, where replacement of deep-rooted native Eucalyptus trees with shallow-rooted crops resulted in recharge increases of about two orders of magnitude (<0.1 mm/year for native mallee vegetation to 5–30 mm/year for crop/pasture rotations).

Sub-surface geology

Recharge rate is greatly influenced by the ability of the subsurface formation (up to the aquifer) to transmit water. The thickness and nature of the

unsaturated zone play a major role in determining recharge (Ali [47]).

Climate

Recharge rates are affected by climatic factors, such as precipitation and evapotranspiration. The total rainfall amount and its distribution (i.e. intensity, duration, spatial variability) greatly influence the recharge. If other conditions remain constant, the more water is available at the soil surface, the more will be the recharge. For the same amount of rainfall, recharge is affected by the temporal distribution of rainfall. The evaporative demand of the atmosphere controls the recharge in the way that, if the evaporation or evapo-transpiration rate is high, the surface water ceases quickly facilitating less time for infiltration or recharge. The opposite is true for the location having less evaporative demand.

Existence of water bodies

Recharge rates and total recharge are largely limited by the availability of water at the soil surface. If the sources of water, such as rivers, lakes, streams, low-lands, depressions, irrigated fields, etc. exist and the area is large, a higher amount of recharge will take place.

Recharge varies with irrigation regimes. Recharge in flood-irrigated cropland is higher than that of sprinkler-irrigated cropland (Wang et al. [28]).

Capacity of the aquifer

If the aquifer has smaller storage capacity and shallow water-table, it is often full and water is usually discharged through evapo-transpiration and base-flow to streams.

Depth to aquifer

If the depth to aquifer is too high, there is little chance of recharge to take place. There are huge possibilities to divert water in other ways – such as evaporation, transpiration, base-flow to stream, etc. The reverse is true for shallow and large storage-capacity aquifers.

5. AVAILABLE APPROACHES AND METHODS FOR RECHARGE ESTIMATION

The recharge estimation techniques have been classified in different ways by different researchers based on the mode of classification. In a generous sense, the techniques can be categorized in the following four broad groups: Physical, chemical, indirect, and empirical. Under each group, sub-groups can be made. Physical methods include: Lysimeter, seepage meter, field-plot water balance, etc. which involves direct or physical measurement. Chemical group includes application of chemicals/tracers to estimate the recharge, such as application of dye, chemicals and isotopic tracers. Indirect group includes the methods which estimate recharge from other variables, such as general water balance (catchment or basin scale), water-table fluctuation, fallout of environmental tracers, groundwater aging, etc. Empirical group includes estimation of recharge from empirical relationship of recharge with other factors of recharge (having 'cause and effect relationship').

The methods are described below in detail along with their applicability and limitations.

5.1 Lysimeter Method

The simplified water balance at lysimeter scale can be written as:

$$P + I + SM_i = ET \text{ (or } E) + R + SM_f \quad (1)$$

Where, P is the precipitation, I is the irrigation (if water added), R is the recharge, ET is the evapo-transpiration, E is the evaporation (if the no crop/vegetation on the surface), SM_i and SM_f are the initial and final total soil moisture within the lysimeter soil, respectively. Thus, the recharge (R) is:

$$R = P + I - ET \text{ (or } E) + (SM_i - SM_f) \quad (2)$$

For recharge study, measurement should be started from the beginning of the rainy period, if rainfall occurs during a part of the year. For the regions having rainfall all over the periods of the year, measurement can be done for a certain period, or the whole year. Although daily observations of drainage, soil water content, etc. can be made; water balance and ET values are normally calculated on a weekly or 10-days basis.

Lysimeter method is a direct method. It has the potential to overcome problems of low flux, if lysimeters are large enough and monitoring period is long enough [48]. The problems associated are high cost, lower boundary not identical to natural condition, possibility of preferred flow through side-walls, possibility of

non-identical soil condition (profiling and density) compared to natural one [49]. Lysimeter soils may not represent spatial variability produced by natural and human-induced changes in surface- and subsurface-flow pathways, and hence may deviate the actual result [50].

5.2 Seepage Meter Method

Originally, seepage meter was developed to measure canal seepage. Downward seepage, that is, deep percolation or recharge from surface-water bodies can be measured by using seepage meters. Recharge rate is calculated as:
Recharge rate (depth/time) = (Volume of water infiltrates)/(Internal area of the meter × time)

That is,

$$R = V / (t \times A) \quad (3)$$

Where, R is the recharge rate ($m^3/m^2/hr$), V is the volume of water lost (m^3), t is the time period (hr), and A is the Area covered by the meter (m^2).

Multiple measurements and longer time period can give confidence to the recharge rate measured. Merits of this method include: it is a direct method, gives a rapid measurement, easy to calculate the recharge, relatively cheap, easy to transport and apply the instrument in field. Demerits include: it provides point estimates of water fluxes, measurements may be required at many locations for a representative value of recharge.

5.3 Water Budget (or Water Balance) Method

Water budget of a hydrological unit (or basin) is an account of all quantities of water added to, subtracted from, and stored (within a given volume of soil) during a given period of time:

Inflow – Outflow = Change in storage (4)

Water-budget (WB) methods are those which are based, in one form or another, on a water-budget equation. To use this approach in practice, two types of boundaries are required: (i) physical *or* spatial boundary, and (ii) temporal *or* time boundary. The spatial boundary may range from individual farm to regional catchment. In this approach, recharge is estimated as "residual" of water-budget equation.

Forms of WB that have been used and/or suggested ranged from simple to more detail budgeting Yin et al. [14]; Scanlon et al. [17]). Writing the water-budget equation to incorporate many of the subcomponents, and equating for recharge (R) [which is a component of "groundwater in"] results in:

$$R = (P + Q_{in\text{-}sw}) - R_0 - (ET_{sw} + ET_{pw}) - (\Delta S_{sw}+\Delta S_{pw}+\Delta S_{snow}) \quad (5)$$

Where, P is Precipitation /Rainfall; Q_{in} and Q_{out} are water flow into and out of the site, respectively; ET is evapotranspiration; E is the evaporation; and ΔS is change in water storage; subscripts *sw*, *pw, gw,* and *snow* represent surface water, profile water, groundwater, and snow, respectively. For a particular basin (or site), some of the terms of the above water budget equation may be negligible, and hence may be ignored. This approach requires detail understanding of regional hydrological processes.

Many hydrological models use the water budget equation. The merits of this approach include: simplicity in understanding, applicability for a wide range of space and time scales, and easy to apply if other components can be measured/estimated accurately.

Accuracy of recharge estimates using this approach depends on accuracy of other components. It is not easy to measure all the components with sufficient accurately. Water budget method yields reasonable estimates of recharge when the precipitation exceeds potential evapotranspiration (PET). However, if the PET is similar magnitude to precipitation (such as under semi-arid conditions), the recharge estimated by this approach must be treated with caution (Sharma [51]; Lerner et al. [52]). Although there is sufficient advancement of ET estimation methods, the spatial variation of ET under such environment is uncertain. In humid region (where rainfall is much higher than ET), the correct estimation of surface runoff component is the main source of uncertainty of WB approach (Ali [53]; Ali [54]). In addition, this approach requires topographically closed basin to accurately account for incoming and outgoing surface runoff. Inflow of stream-flow to groundwater during flash-flood and outflow of groundwater to river (as base-flow) during recession/dry-period are also of concern in WB method.

The combined error and uncertainty associated with ET and surface runoff measurements can be equal to greater than the recharge (Halford and Mayer [56]), or much as an order of magnitude (Gee and Hillel [57]). The environments where recharge is only a portion of precipitation, the uncertainty of recharge estimates by this approach is magnified, and can produce misleading value of recharge.

For regions where rainfall amount is not too much higher than the ET (say, $ET<R_{ain}<3ET$), if spatial variability of land-surface are taken into account, and short time steps are used, the WB approach may produce reasonable estimate of recharge.

5.4 Soil-moisture Balance (or Unsaturated Zone Water Balance)

In the absence of significant runoff, simple soil-water balance at plot level have been used to estimate recharge (Allison et al. [48]; Bakundukize et al. [19]):

$$R = P - ET + S \quad (6)$$

Where P is the precipitation, ET is the actual evapo-transpiration, and S is the change in storage (i.e. $S_i - S_f$).

The method is simple to understand and apply. This approach requires considerable field measurements of crop ET, soil-water holding capacity, and estimation of rainfall interception and runoff losses (if any). For humid region, this method is not suitable due to uncertainty in runoff estimation, and that the error in runoff estimation may be higher than the recharge. Delin et al. [58] observed inconsistent recharge estimates using this method compared to the WTF method. In arid region, where recharge takes place mainly through preferential pathway or from depressions, this method gives inaccurate estimates of recharge.

5.5 Darcy's Law Approach

5.5.1 Darcy's law method for unsaturated zone

If the flow under field conditions is steady and governed by gravity alone, then, according to Darcy's Law, downward flow rate (i.e. recharge, R) will be numerically equal to the hydraulic conductivity of the soil (at the measured in-situ water content) multiplied by hydraulic gradient, as:

$$R = K(\theta)dH/dz = K(\theta)\frac{d}{dz}(h+z) = K(\theta)\left(\frac{dh}{dz}+1\right) \quad (7)$$

Where, $K(\theta)$ is the hydraulic conductivity at the ambient water content, H is the total head, h is the matric pressure head, and z is the elevation.

The fundamental assumption behind this approach is that, the water draining below the root zone (or passing through unsaturated zone) is contributing to recharge. The Darcian method assumes one-dimensional flow, which may be a reasonable approximation in flat topography with vertical uniform profile (Sharma [51]). The situation where the above assumptions are violated, the recharge estimates become unreliable. The major pitfalls of this approach are: difficulties of measuring soil-water potential gradient at deeper layer/profile, variabilities of hydraulic properties of field soil, field measured data of hydraulic properties, etc.

For thick unsaturated zones, below the zone of fluctuations related to climate, in uniform or thickly layered porous media, the matric pressure gradient is often nearly zero, and water movement is essentially gravity driven. Under these conditions, little error results by assuming that the total head gradient is equal to 1 (unit-gradient assumption). The unit-gradient assumption removes the need to measure the matric pressure gradient and sets recharge equal to the hydraulic conductivity at the ambient water content.

The minimum recharge rate that can be estimated by using Darcy's law depends on the accuracy of the hydraulic conductivity and head-gradient measurement if the latter is not unity. If hydraulic conductivity is strongly dependent on water content, uncertainty increases. The demerits of this method include: it provides a point estimate of recharge over a wide range of time, accuracy of recharge estimates depends on the accuracy of hydraulic conductivity and matric pressure head (or moisture content, θ) value, does not indicate total recharge as it only accounts for diffuse or matrix flow. Recharge due to preferential flow is inherently non-Darcian and, if significant, must be determined separately.

5.5.2 Darcy's law method for saturated zone

Darcy's law can be used to estimate flow through a cross section of an unconfined or confined aquifer (assuming steady flow and no water

extraction). The sub-surface water flux (q) is calculated by multiplying the hydraulic conductivity by the hydraulic gradient. The hydraulic gradient should be estimated along a flow path at right angles to potentiometric contours. The volumetric flux through a vertical cross section of an aquifer (A) is equated to the recharge rate (R) times the surface area that contributes to flow (S):

$$qA = RS \quad (8)$$

or,

$$K(\theta) \times (d\theta/dz) \times A = R \times S \quad (9)$$

Or,

$$R = [K(\theta) \times (d\theta/dz) \times A\ ]/\ S \quad (10)$$

The cross section should be aligned with an equipotential line. This technique can be applied to large regions (~1 to ≥10,000 km^2). This method is not suitable for areas where hydraulic conductivity and hydraulic gradient vary significantly with space (Yin et al. [14]). Recharge value is highly uncertain because of the high variability of hydraulic conductivity (several orders of magnitude) over large area (even within a field). In addition, accurate estimation of saturated aquifer thickness and length of cross-section are matter of concern.

5.6 Zero-flux Plane Method

The zero-flux plane (ZFP) represents the plane where the vertical hydraulic gradient is zero. The ZFP method is based on the premise that soil-water moves upward in response to ET above the ZFP and below that level percolates downward to the water-table (Delin et al. [58]). The ZFP method requires soil matric-potential measurements to locate the position of ZFP and soil-water-content measurements to estimate storage changes.

This technique works best in regions where large fluctuations exist in soil-water content throughout the year and where the water-table is always deeper than the ZFP. Accuracy of recharge estimate is dependent on the accuracy of the water-content measurements. Demerits of the method include: the method yields a point estimate of recharge, requires multiple readings at multiple locations for representative estimate, data recording and calculation requires expertise, and relatively expensive in terms of the required instruments and amount of data collection.

5.7 Water-table Fluctuation (WTF) Method

The WTF method is most widely used to quantify recharge, and frequently used to compare/validate other methods and simulation models. This is an indirect method for recharge estimation, but the "fluctuation of water-table" is the only direct indication (observed phenomena) of recharge available. In essence, the WTF method is based on volume-balance principle - the 'recharging water' entering the aquifer system is contributing to rise in the water-table depth, assuming bounded/closed aquifer basin (that is, no inflow-outflow to/from the aquifer), and with same 'recharge-area' and 'aquifer areal extent'. It is also presumed that there is instantaneous response of the water level with recharge input to the aquifer (and vice versa). The method is suitable where a distinct rainy season(s) with the remainder of the year being relatively dry period exists, and only applicable to unconfined aquifer.

The refill depth of aquifer is multiplied by the specific yield of the aquifer (S_y) to obtain the recharge depth (i.e. total recharge, R_T). The recharge rate (R_r) is obtained by dividing the recharge depth by the time period, i.e.

$$R_r = \frac{S_y \times (h_2 - h_1)}{(t_2 - t_1)} = \frac{S_y \times \Delta h}{\Delta t} \quad (11)$$

Where, Δh represents change in (rise) water-table depth, and Δt represents the time difference between successive measurements of depth to water-table.

If there is substantial amount of discharge from the aquifer during the study period, this should be taken into account as:

$$R_T = \left(S_y \times \Delta h\right) + \left(Q_p + Qs\right) \quad (12)$$

Where, Q_p and Q_s are the pumpage and spring discharge (or aquifer leakage), respectively.

If only pumping (abstraction) is present (i.e. aquifer discharge can be neglected), recharge rate (R_r) is:

$$R_r = \frac{S_y[\Delta h + h_p]}{\Delta t} \quad (13)$$

Where, $h_p = Q_p/A$, ratio of the pumped volume (Q) during the recording interval to the areal extent of the aquifer (A) (or influenced area).

From the underlying principles and assumptions of the method, and the mathematical forms presented above, it is apparent that the following factors are the major issues (governing factors) for the correct estimation of recharge using this method (as well as weakness/uncertainties in estimates): aquifer boundary *or* areal extent of the aquifer/basin (A_{aqu}), inflow-outflow to/from the aquifer ($Q_{in\text{-}out}$), aquifer response type (that is aquifer type)(T_{aqu}), specific yield of the aquifer (S_y), measurement of depth to water-table (h, Δh), and the time interval of depth-measurement (Δt). They will be discussed under separate headings for clear interpretation and understanding.

Inflow-outflow to/from the aquifer ($Q_{in\text{-}out}$)

In this approach, it is assumed that the rise in water level (i.e. refill) is caused by recharge only and there is no inflow-outflow to/from other inter-connected aquifers. Groundwater outflow (to the river or sea) or any inflow through the aquifer boundary will affect the recharge estimate. Where there is withdrawal within the basin, although provision can be made to account for, the difficulties arise due to extraction rate, number of wells in operation, area of influence, variation of recharge responses with respect to both time and space etc.

In case of coastal aquifers, the tidal effect can influence the water-level; and hence should be taken into account. In addition, evapotranspiration by deep rooted plant in shallow aquifer can influence the water-table. Checking of water-level records with corresponding rainfall data is necessary in order to verify that WT rise is due to rainfall only (Nimmo et al. 2005). Water-table near the stream or river may also be influenced by river stage (Delin et al. [58]).

Although techniques are available to explore the aquifer boundaries (Morgen et al. [59]), in practice, it is very difficult to measure/estimate the inflow-outflow. Although analysis of hydrograph recession for such correction has been suggested (Healy and Cook [60]), this is a crude method and may not represent the actual effects. In addition, such data may not be available in many basins. However, if the objective is to calculate the net available recharge for groundwater development potential, these sorts of inflow-outflow may not be a problem.

Areal extent of the aquifer (A_{aqu})

In practice, it is difficult to ascertain the aquifer boundary (i.e. groundwater divides). Indeed, it requires huge number of pumping test to ascertain it. To account for the pumpage from the aquifer during the recorded period, the influenced area (A) has been equated/considered as the basin area, but this is not fair if the pumping is not uniform throughout the basin and the basin area is large.

Aquifer response type (i.e. aquifer type) (T_{aqu})

This approach is applicable for unconfined aquifer only, because of nearly instantaneous response of the water level with recharge input to the aquifer, and direct relationship with recharge volume. For confined aquifer, it needs complicated calculation. Without knowledge of the type of studied aquifer, it can give misleading recharge estimates. The method works best for shallow water-tables that display sharp water-level rises following rainfall events. Based on the depth to aquifer and type of overlying strata, the lag-time differs considerably (Lu et al. [61]). For deep aquifer, the lag-time for full response of the aquifer due to rainfall should be identified by frequent measurements; and subsequently should be considered when estimating recharge on event basis.

It is worthwhile to mention here that, amount of recharge also depends on the aquifer capacity (to accommodate potential recharge) under certain circumstances. If the aquifer is full or nearly full, but there is climatic potential (sufficient availability of water on the soil surface) and hydro-geological potential (the geological setting is such that it has the ability to transmit water downward), no more recharge (or negligible amount) to the aquifer will be taken place; rather the water will be lost as surface runoff or seepage to spring/river. Under such condition, the apparent recharge will represent the actual recharge, but not the potential recharge. Induced recharge can occur under such condition, and hence can stimulate the actual recharge (up to potential recharge level) (Shamsudduha et al. [62], Ali [47]). Water resources planning based on the apparent recharge will be misleading.

Specific yield of the aquifer (S_y)

In this approach, the specific yield of the aquifer (S_y) must be known. The approach presumed

that the specific yield is constant throughout the aquifer (both in vertical and horizontal direction). The accuracy of recharge estimates depends on the accuracy of the specific yield. For example, if the S_y value is deviated from 0.10 to 0.08 (i.e. 20% deviation), the resultant recharge estimate will also be deviated by 20% from the true value. Weighted average of specific values determined at different locations should be used to reduce the uncertainty of specific yield value. Long-term pumping test may be fair enough to obtain reasonable S_y. If stratified aquifer is found by bore-hole, calculation should be done layer-wise and then weighted average should be calculated.

Measurement of depth to water-table (h, Δh)

The WTF method requires monitoring wells to obtain time variant data of depth to water-table. The 'observation wells' / 'boreholes' in the basin should be at representative locations (and also up to the correct aquifer depth), covering the basin. Indeed, the accuracy of observed WT data depends on the diffusivity of the aquifer, and the distribution in time and space of the abstraction well (if any) (Ordens et al. [12]). Additionally, factors such as changes in barometric pressure, the presence of entrapped air, evapotranspiration by deep rooted plants (specially for shallow aquifer), etc. can influence WTF (Heely and Cook [60]). Although these factors are not common for all basins.

Time interval of depth measurement (Δt)

The WTF for determining recharge from rainfall and WT measurements was originally developed for an event basis (Crosbie et al. [63]). For thick unsaturated zone (i.e. for deep aquifers), the WTF is not suitable for an event basis, rather seasonal basis is appropriate. The WT quickly rises in response to recharge in shallow aquifers, while there is a delay where the groundwater level is at greater depth (Ali [47]). The time interval of WT recording/measurement (short or long) should be chosen depending on the length of wet/dry spell, depth to aquifer, and objective of the recharge estimation. The estimate is not accurate for very short period, as there is a time-lag between arrival of recharge component in the aquifer and the measurement of precipitation at the surface. If the objective is the estimation of sustainable withdrawal amount during dry period from a basin, long interval, say monthly data, is sufficient (specifically at the beginning of the wet and dry seasons are essential). If the objective is to estimate leaching amount (e.g. for contaminant transport), frequent measurements are needed; because the groundwater can flow due to hydraulic gradient, and also the WL may fluctuate due to withdrawal and time-lag effect. In this case, the annual recharge amount will be the summation of recharge amounts obtained during the year.

Water-level fluctuations occur in response to spatially averaged recharge, which is desirable, and advantageous than some other point estimation approaches. As the WTF is the combined effect of all possible sources of recharge (which is desirable for recharge estimation technique), the existence of preferential flow paths is not a restriction in its application. This method has the merit over some other methods in that it provides insight into transient recharge trends (Ordens et al. [12]). Long-term changes in recharge caused by climate or land-use change can be determined using this method.

The WTF method is attractive and widely used because of its simplicity, ease in use, and less data requirement. Other advantages of the method are: it is independence of the water displacement mechanism in the unsaturated zone, as well as areal integration of the recharge (Carretrro and Kruse [64]). The recharge estimates with this method should be compared with time series of precipitation to identify other processes/mechanisms and sources of error.

If the above discussed factors are considered during application and calculation, certainly the uncertainty in recharge estimates will be reduced (i.e. the accuracy will be increased).

Park [65] presented a physically based WTF method, in which the concept of transient fillable porosity has been proposed and computed with unsaturated hydraulics model. Lorenz and Delin [66] provided a method for regional recharge estimates by regressing the local estimated recharge based on rainfall, topography, and other factors, which they found causes variability in recharge within the particular region. The WTF method can be used in basin scale by weighted averaging the point/local data.

5.8 'Base-flow Discharge' / 'Hydrograph Separation' Method

Use of base-flow discharge to estimate recharge is based on the water-budget approach, in which

recharge is equated to discharge. That is, recharge equal to base-flow, assuming equilibrium between recharge and discharge (Sharma [51]). This approach is not applicable where the stream is losing. This method assumes that pumpage, ET and under-flow (from the aquifer) are negligible. Hydrograph analysis is an essential step of this method. Various approaches are used for hydrograph separation, including digital filtering (Nathan and McMahon [67]; Arnold et al. [68]) and recession-curve displacement methods (Rorabough [69]).

The critical assumptions of this method include that the hydraulic characteristics of the contributing aquifer can be reliably identified, and that stream-discharge peaks approximate the magnitude and timing of recharge events (Coes et al. [18]). These assumptions are violated for most basins, and therefore not reliable for determining recharge for all cases (Halford and Mayer [56]). The base-flow can come from various sources besides groundwater, and the basin response may not be linear because the response is a function of various geologic and hydrologic factors in addition to those considered in mathematical derivations of hydrograph analysis techniques (Hall [70]).

If the pumpage, evapotranspiration, and underflow to deep aquifers are significant, recharge (R) can be estimated as (assuming no change in groundwater level):

R = Pumpage + ET + Under-flow + base-flow(14)

The discharge components (right-hand side of the equation) should be estimated independently. The method is not suitable for basins having significant pumpage, evapotranspiration, and underflow to deep aquifers. Other demerits include: minimum time scale is a few months, problematic to apply this method for large basins because of difficulties in separating surface-water and groundwater flow, the accuracy of recharge rates depends on the accuracy of other elements, and may not represent potential recharge for all cases (i.e. where pumpage and ET are absent) but actual recharge.

The merits of this method include: simple, no sophisticated instrument is needed, recharge over longer times can be estimated by summation of estimates over shorter times. This approach is usually appropriate for a particular scale of catchment. It works best where the water-table is relatively shallow and streams are typically gaining (Delin et al. [58]).

5.9 Numerical Method

5.9.1 Numerical method for Watershed modeling

Rainfall-runoff modeling is used to estimate recharge rates over large areas. Watershed models generally provide recharge estimates as a residual term in the water-budget equation (as discussed earlier).

The minimum recharge rate that can be estimated is controlled by the accuracy with which the various parameters in the water budget can be measured and the time scale considered. The various watershed models differ in spatial resolution of the recharge estimates. Some models are termed lumped and provide a single recharge estimate for the entire catchment. Others are spatially disaggregated into hydrologic-response units (HRUs) or hydro-geomorphological units (HGUs).

This approach can be applied at a variety of scales (Flint et al. [21]). Small-scale applications allow more precise methods to be used to measure or estimate individual parameters of the water-budget equation. Time scales in models are daily, monthly, or yearly. Daily time steps are desirable for estimation of recharge because recharge is generally a larger component of the water budget at smaller time scales.

Demerits of this approach include: indirect method, recharge is estimated as residual term; accuracy of recharge estimates is dependent on the accuracy with which the various parameters in the water budget can be measured.

5.9.2 Numerical modeling for Unsaturated-zone studies

Recent advances in computer technology and in computer codes have made long-term simulations of recharge more feasible. Unsaturated-zone techniques provide estimates of potential recharge based on drainage rates below the root zone; however, in some cases, drainage is diverted laterally and does not reach the water table. In addition, drainage rates in thick unsaturated zones do not always reflect current recharge rates at the water table. Unsaturated-zone techniques for estimating recharge are applied mostly in semiarid and arid regions, where the unsaturated zone is generally thick. The recharge estimates generally apply to smaller spatial scales than those calculated from surface-water or groundwater approaches.

Theoretically, the range of recharge rates that can be estimated using numerical modeling is infinite; however, the reliability of these estimates should be checked against field information such as lysimeter data, tracers, water content, and temperature. A variety of approaches is used to simulate unsaturated flow, including soil-water storage-routing approaches, quasi-analytical approaches (Simmons and Meyer [71]), and numerical solutions of the Richards equation. Examples of codes that use the Richards equation include HYDRUS-1D, HYDRUS-2D (Simunek et al. [72]), SWIM (Ross [73]), VS2DT (Hsieh et al. [74]), and UNSATH (Fayer [75]). Bucket-type models can be used over large areas (Flint et al. [21]); however, models based on the Richards equation are often restricted to evaluating small areas (≤100 m^2) or to one-dimensional flow in the shallow subsurface (≤15 m depth). Many recharge modeling studies evaluate periods of 30 to 100 years because of availability of meteorological information.

Merits of this approach include: fast, with the aid of computer; time scales that can be evaluated range from hours to decades; numerical modeling can be used as a tool to evaluate flow processes and to assess sensitivity of model output to various parameters, and they allow predictions of future recharge regimes resulting from different land-uses and climatic changes.

Demerits include: models based on the Richards equation are often restricted to evaluate small areas or to one-dimensional flow in the shallow subsurface; because of uncertainties in hydraulic conductivity and non-linear relationships between hydraulic conductivity and matric potential or water content, recharge estimates based on unsaturated-zone modeling that use the Richards equation may be highly uncertain.

5.10 Use of GW Models

Groundwater models can be used to estimate recharge. If the other model parameters are known well enough, then the model could be used to constrain the recharge (Sanford [76]). The uncertainties are associated with the model parameter values, which rarely can be measured with sufficient accuracy. Models are verified/compared with hydraulic head, WT data (Lu et al. [61]), or infiltration equation (Krishnamurthi et al. [77]). Simulation models may show error from two directions – within the model itself (due to inherent assumption), and in the validation process.

5.11 Tracer Techniques

The tracer technique is based on the conservation of mass of the tracer, and the assumption that the tracers moves freely with water, no other sources of the tracer nor no absorption or uptake by the soil/rock or by vegetation. If these assumptions are violated, then the recharge estimates by this approach becomes uncertain. Tracer may be chemical or isotopic. The chemical tracer may be natural (or environmental) or artificially applied one. Applied tracers give more accurate recharge estimates because it is driven by recharge component (i.e. independent of runoff); while in CMB method, the Cl concentration is dependent upon runoff, which is a major source of error, especially in humid region. The main advantages of the applied tracers are that the investigators have control over the timing, placement, and amount of tracer (Wang et al. [28]). In addition, the complexities of the top meter of soil (action of roots, and cultural disturbances) can be avoided if tracers are injected below one meter depth.

5.11.1 Applied chemical tracer

Commonly used chemical tracers include bromide, chloride, rhodamine, and other visible dyes. Organic dyes are generally used to evaluate preferential flow. To estimate recharge, chemical tracers are applied as a pulse at the soil surface, or at some depth within the soil profile. Infiltration of precipitation or irrigation transports the tracer downward. The subsurface distribution of applied tracers is determined sometime after the application by digging a trench for visual inspection and sampling, or by drilling test. The vertical distribution of the tracer is used to estimate the velocity (v), and the recharge rate (R) is (Chand et al. [36]; Scanlon et al. [17]; Ali [54], Ali [55]):

$$R = v\theta = \frac{\Delta z}{\Delta t}\theta \quad (15)$$

Where, Δz is the depth of the tracer peak, Δt is the time between tracer application and sampling, and θ is the volumetric water content.

Chemical tracers are generally applied at a point, or over small areas. The calculated recharge rates represent the time between application and sampling, which is generally months to years. The minimum water flux that can be measured with applied tracers depends on the time between application and sampling, and in the

case of surface-applied tracers, the root-zone depth. Hydrological interpretation of tracer results depends, among others things, on the validity of the physical model of water-flow for the system in question (Sharma [51]). Thus, the accuracy of recharge estimates using a tracer technique would depend on how realistic was a particular model used for interpreting the results and how realistically the required assumptions for the model met for the system.

Merits of chemical tracers include: no environmental hazard, easy to apply and sampling, low cost (Ali [55]), and visual observation is possible for visible dye. Demerits include: the observed recharge rate will be higher than the actual if there are preferential pathways, sorption may be significant for organic dyes, uptake by plants is often significant for some chemicals, concentration towards greater depth becomes negligible if the initial concentration/amount is not high enough. As tracers do not measure water flow directly, a number of problems can arise, leading to over- or under-estimation of recharge. These problems include secondary (unknown) tracer inputs, mixing, and dual flow mechanisms. Such problems only arise if the sources, sinks, and pathways of tracer are not fully understood. As this technique yields point estimates of recharge (and through soil matrix only), the areas where preferential flow contributes major part of the total recharge, the tracer results in such regions should be interpreted with caution (Ali [54]).

Isotopic tracer

With conventional hydro-geochemical studies, in many cases, it is not sufficient to characterize groundwater dynamics or to detect recharge areas and source areas of recharged water. The application of isotope-based methods has become well established for water-resource assessment, development and management in the hydrological sciences. Isotopes used are stable and radio-active.

Stable isotope

Commonly used stable isotopes to identify groundwater recharge from rivers and lakes are oxygen (^{18}O) and hydrogen (^{1}H, ^{2}H). Since the isotopic composition of O and H in groundwater does not change as a result of rock–water interactions at low temperatures, it provides a useful means to detect source of recharge. Stable isotopic tracers provide information on recharge sources; however, it is generally difficult to quantify recharge rates. The time scales range from seasonal in areas of high flux to hundreds of years in areas of low flux.

Radioactive isotope

Commonly used radioactive tracers include tritium (^{3}H), Carbon-14 (^{14}C), Cesium-134 (^{134}Ce). Although ^{3}H is the most conservative of all tracers, its use is prohibited in many areas because of environmental-protection laws. Radioactive isotopes are applied as a pulse at the soil surface or at some depth within the soil profile to estimate recharge. Infiltration of precipitation or irrigation transports the tracer to depth. The subsurface distribution of applied tracers is determined sometime after the application (3 months to 1 year) by drilling test holes for sampling. The vertical distribution of tracers is used to estimate the recharge rate (R):

$$R = v\theta = \frac{\Delta z}{\Delta t}\theta \quad (16)$$

Where, Δz is the depth of the tracer peak, Δt is the time between tracer application and sampling, and

θ is volumetric water content.

Commonly used radioactive isotopes (also called radio-isotope) in hydrological studies and their half-life are given in Table 1. Half-life is the time period after which the radioactivity of an isotope is decayed to half of the original activity. Among the radio isotopes, Tritium is widely used in groundwater studies.

Application of tracer at multiple sites and appropriate averaging of the results (such as Kriging, Thiessen Polygon method) can give more realistic value of recharge (Chand et al. [38]). The minimum water flux that can be measured with applied tracers depends on the time between application and sampling and, in the case of surface-applied tracers, the root-zone depth.

Merits of radio isotope include: it is direct method, more accurate results can be obtained, no absorption or loss of tracer, does not require frequent visits to the field (generally require only one-time sampling and may represent long time periods), smaller fluxes can be estimated with this technique than with other methods.

Table 1. Commonly used radioisotopes in hydrological studies and their half-life

Sl	Isotope	Half-life (year)
1	^{134}Cs (Cesium-134)	2.06
2	^{222}Rn (Radon-228)	5.75
3	^{85}Kr (Krypton-85)	10.76
4	^{3}H (Tritium/Hydrogen-3)	12.43
5	^{39}Ar (Argon-39)	269
6	^{226}Ra (Radon-226)	1600
7	^{14}C (Carbon-14)	5730
8	^{81}Kr (Krypton -81)	2.1×10^{5}
9	^{36}Cl (Chlorine-36)	3.01×10^{5}
10	^{129}I (Iodine-129)	1.57×10^{7}

Demerits include: radioactive material - may not be permitted by environmental protection law in all areas, needs precaution, needs costly instrument for reading the sample and technical hand to operate the instrument, point estimates of recharge - therefore several measurements are needed to get the average/ representative value, difficulties of soil sampling at higher depths (specially in areas having higher recharge rate) and locating the tracer peak (specially for longer time period between injection and sampling), water in the root zone can be evapotranspired thus water content within the root zone can be underestimated.

5.11.2 Natural tracer

The natural or environmental tracers for recharge study include oxygen-18 (^{18}O), deuterium (^{2}H), chloride (Cl), nitrogen-15 (^{15}N), tritium (^{3}H), carbon-14 (^{14}C), and chloride-36 (^{36}Cl).

Recharge estimates based on environmental tracer gives areally integrated value, which is the advantage over soil physical methods. The most widely used natural tracer is Cl, which is well known in the literature as 'Chloride Mass Balance' (CMB) method, and due to its importance, it is described in details under a separate heading.

5.12 Chloride Mass Balance (CMB) or Chloride Budget Approach

Erikson and Khunakashem [78] first explored the possibility of estimating recharge from groundwater chloride concentration. After then, chloride mass balance approach has been emerged, suggested and used to estimate recharge; and different simplified forms of CMB have been used (Allison and Hughes [79]; Wood and Sanford [37]; Scanlon et al. [17]; Ordens et al. [12]; Jack and Traore [80], amongst others).

The approach is based on the assumptions that: (1) The environmental chloride (dry deposition of airborne Cl, and Cl from precipitation) is the only source of groundwater Cl, (2) No Cl loss during evaporation of precipitation water as vapor. (3) The geological formation (both unsaturated zone and aquifer) do not contain/absorb/release chloride, (4) No surface runoff of precipitation water (or, quantity and concentration of Cl in runoff water are measurable), (5) All the percolating water reaches to the groundwater (no subsurface drainage to streams, rivers, etc), (6) No Cl input from agricultural activities (e.g. fertilizers, insecticides, pesticides, weedicides), (7) No uptake of Cl by vegetation or plants, (8) No Cl input to land and water-bodies by human activities/endeavors such as urine, stool, industrial release (so that there is no possibilities of percolating), (9) The recharge takes place mainly through soil matrix (Erikson and Khunakashem [78]; Allison et al. [48]; Wood [89]; Scanlon et al. [17]).

Among the different simplified forms of CMB equation suggested and used, Allision and Hughes [79] used a simplified balance (assuming no runoff):

$$P\times C_p = R\times C_R \quad (17)$$

Where, P is the annual rainfall (mm), R is the drainage flux beneath the root zone, and C_P and C_R are the Cl concentration of rainfall and drainage flux, respectively. The weighted-average Cl concentration has been used for samples of dry and wet precipitation (Wood and Stanford [37]). Wood [81] suggested a modification of the above equation, replacing C_R by Cl_{uz}, where C_{uz} is the Cl concentration in drainage water in the unsaturated zone. That is,

$$R = \frac{P\times C_p}{C_{uz}} \quad (18)$$

Subyani and Sen [82] used arithmetic average of each term in equation (13) over the time and space scale of sampling. Jacks and Traore [ ] used a more detail budget of Cl:

$$P\times (Cl_p + Cl_{DP} + Cl_{NACl}) = R \times Cl_{GW} \quad (19)$$

Where P is the mean yearly precipitation, Cl_p is the Cl concentration in rainwater, Cl_{DP} is the Cl concentration as dry deposition mixed up in rainwater, Cl_{NACl} is the Cl from use of common

salt mixed up in rainwater, and Cl_{GW} is Cl concentration in groundwater.

The CMB approach has been most widely used for estimating low recharge rates, largely because of the lack of other suitable methods. The maximum water flux that can be estimated is based on uncertainties in measuring low Cl concentrations and potential problems with Cl contributions from other sources, and is generally considered to be about 300 mm/year.

The uncertainty of the recharge estimate using CMB method depends on the violation of the above assumptions for the particular area or basin of interest. For arid and semi-arid regions, where rainfall is scanty (generally 500 mm -700 mm per year), runoff can be neglected. But for humid and sub-humid regions, where seasonal rainfall is high (normally 1000 mm – 2500 mm per year) and concentrated over few months, a substantial amount of rainwater is drain out as surface runoff. In those regions, the exact runoff amount and Cl concentration of runoff water (time variant concentration) must be taken into account. Similarly, where lands are cultivated twice or thrice a year, Cl input from fertilizer and insecticide/pesticide can not be ignored. In industry-based area, Cl input from the industries (both as vapor and liquid-discharge) is substantial. The region where macro-pore recharge (or recharge through streams/wet-lands, depressions) constitute a major part, the CMB approach may yield unrealistic estimate if the macro-pore recharge and its Cl concentration is not taken into account. In practice, such recharge amount and its Cl concentration is difficult to measure exactly. The regions where the Cl concentration of precipitation varies with time and space due to the variation of Cl concentration in the atmosphere, weighted average Cl concentration of precipitation (or arithmetic average) should be determined. In irrigated farmlands, Cl input from irrigation and absorption by crops should be taken into account (Lin et al. [83]). This method is not suitable where the aquifer is exposed to salt intrusion and rock-water interactions (Abu-Jaber [84]). In addition, uncertainty in CMB estimates of recharge is related to the uncertainties of atmospheric Cl deposition (Scanlon et al. [85]).

The merits of CMB approach includes its simplicity in understanding and application, less cost and less time consuming, and requires les technical skill. It is typically suited in relatively dry areas. Demerits include: it provides point estimates of recharge, and the accuracy of this approach decreases as surface runoff and recharge rate increase. Care should be taken in the application of the method when the assumptions are not fully meet.

In many simplifications of CMB approach, actually the “mass balance” of Cl no more satisfied; rather a relationship between rainfall, Cl concentration of rainfall, recharge, and Cl concentration of groundwater or ‘pore-water in unsaturated zone’ has been established which was first proposed by Erikson and Khunakashem [78]. This is because, the Cl concentration of present groundwater is a result of hundreds or thousands of years, not resulted from the current year rainfall. Similarly, the Cl concentration of pore-water in unsaturated zone depends on the Cl concentration in the preceeding year and other source-sink factors. The relationship should be verified with observed data for the target basin, such as water-table data. After then, the relationship can be used to estimate recharge simply from the seasonal or yearly rainfall amount and its Cl concentration, and Cl concentration of groundwater.

5.13 Historical Tracer

The historical tracers result from human activities or events in the past, such as contaminant spills or atmospheric nuclear testing (^{3}H and ^{36}Cl). These historical tracers or event markers are used to estimate recharge rates. Industrial and agricultural sources produce contaminants such as bromide, nitrate, atrazine, and arsenic, and these can provide qualitative evidence of recent recharge; however, uncertainties with respect to source location, concentration, and timing of contamination, as well as possible non-conservative behavior of contaminants, make it difficult to quantify recharge. The presence of an event marker in water suggests that a component of that water recharged in a particular time period.

Merits of this approach include: no extra hazard, no extra cost of tracer. The demerits include: uncertainties of the tracer (location and concentration), difficulties of soil sampling at higher depths and locating the tracer peak in areas having higher recharge rate, water in the root zone can be evapotranspired and thus water fluxes estimated from tracers within the root zone can overestimate water fluxes below the root zone.

Gaye and Edmunds [86] observed deep penetration of thermonuclear tracers in sandy soils in arid settings (^{3}H, 22 to 26 mm/year). Similar observation was also reported by Dincer and Davis [87] (^{3}H, 23 mm/year). Theoretically, the technique could be used for higher recharge rates if the water-table were deeper; however, the difficulty of soil sampling at these depths and locating the tracer peak may be prohibitive.

Historical tracers or event markers such as bomb-pulse tritium (^{3}H) has been widely used in the past in both unsaturated and saturated zones to estimate recharge. However, bomb-pulse ^{3}H concentrations have been greatly reduced as a result of radioactive decay. In the southern hemisphere, ^{3}H concentrations in precipitation were an order of magnitude lower than in the northern hemisphere. Thus, it is now often difficult to distinguish bomb-pulse ^{3}H from current ^{3}H concentrations in precipitation.

5.14 Groundwater Aging

Recharge rates can be determined by estimating ages of groundwater. Age is defined as the time since water entered the saturated zone. The age of the groundwater, t, can be calculated from ^{3}H/^{3}He data (ratio of tritium to tritiogenic helium) using the following equation:

$$t = -\frac{1}{\lambda} ln\left[1 + \frac{^3He_{trit}}{^3H}\right] \quad (20)$$

Where, λ is the decay constant (ln 2/$t^{1/2}$), $t^{1/2}$ is the ^{3}He half life (12.43 years), and He_{trit} is the tritiogenic ^{3}He. Use of this equation assumes that the system is closed (does not allow ^{3}He to escape) and is characterized by piston flow (no hydrodynamic dispersion). Radioactive decay of ^{14}C can be used to estimate groundwater ages of 200 to 20,000 years. The estimated recharge rates are average rates over the time period represented by the groundwater age. The residence time of groundwater can be calculated through the following decay equation:

$$\ln(A/A_0) \times 8266.7 = \text{Age} \quad (21)$$

where A_0 is initial radiocarbon concentration of water. The recharge rate, *R*, can then be calculated as:

$$R = \frac{L \times \varphi_e}{T} \quad (22)$$

where *L* is the distance along the flow path, φ_e is the effective porosity, and *T* is the travel time or age of the groundwater at the distance L.

In unconfined porous-media aquifers, groundwater ages increase with depth, the rate of which depends on aquifer geometry, porosity, and recharge rate (Cook and Bohlke [88]). The vertical groundwater velocity decreases with depth to zero at the lower boundary of the aquifer. The age increases linearly with depth near the water table and nonlinearly at greater depths. Near the water table, the influence of the aquifer geometry is greatly reduced. The recharge rate can be determined by dating water at several points in a vertical profile, calculating the groundwater velocity by inverting the age gradient, extrapolating the velocity to the water table if it is not measured near the water table (Cook and Solomon [89]), and multiplying the velocity by the porosity for the depth interval.

That is,

$t = v \times \rho$,

where ρ is the porosity. CFCs and ^{3}H/^{3}He are used to determine groundwater ages up to approximately 50 years, with a precision of 2 to 3 years (Cook and Solomon [89]).

Merits of this approach include: the approach is easy to implement if the instrument for reading the sample is available, no additional field setting/experiment is needed. Demerits include: costly instrument is needed for reading the sample, variation of isotopic signature with depth may occur (due to various reasons), so multiple sampling (throughout the depth up to aquifer) is needed.

5.15 GIS and Remote Sensing Application

For regional recharge estimates, combination of local data, remote sensing, GIS, and geostatistical techniques has been advocated (Simmers [90]). Several different approaches have been used to estimate recharge at regional scale using remote sensing and GIS (Szilagyi et al. [91]; Cherkauer [92]).

6. SELECTING AN APPROPRIATE METHOD FOR RECHARGE ESTIMATION

6.1 Factors Affecting Selection of a Method

The soil, vegetation, topography, geology and climate (specially rainfall amount and distribution, and temperature) of a site control the recharge,

and therefore impact on the choice of a technique for estimating recharge. In addition, purpose or aim of recharge estimation - specially accuracy needed, influences on the selection of a method.

Geomorphology of the target area along with climate dictates the recharge rate, and hence the recharge estimation method. Sources and mechanisms of recharge may also dictate the techniques to be used to quantify recharge. Surface-water sources require techniques such as channel-water budgets, and water-table fluctuations. Climatic regions (e.g. arid vs. humid) have fundamental differences in recharge that may require different approaches. Surface-water and saturated-zone techniques are more widely used in humid regions, whereas unsaturated-zone techniques are widely used in arid and semiarid regions (Table 2). Watershed-modeling approaches may be more accurate in humid regions, where perennial surface-water flow can be used for model calibration. Although historical tracers can be used in the unsaturated zone in humid regions, their use is limited because of generally thin unsaturated zones and the ease of using such tracers in the saturated zone. Water-table fluctuations and Darcy's law could also be used in arid and semiarid regions where water-tables are shallow. The estimated recharge rate at a site may determine the most appropriate procedures for quantifying recharge because different techniques measure recharge over different ranges. The recharge values estimated by various researchers using different techniques are summarized in Table 2 to Table 4.

A key factor in deciding on a recharge-estimation methodology is related to the spatial and temporal scale of interest. The space and time scales of the various techniques also affect the choice of technique to be used (Table 5). Surface-water and groundwater approaches provide regional estimates of recharge, whereas unsaturated-zone techniques generally provide estimates at points or small scales. Lysimeters, zero-flux plane, applied tracers, and saturated-zone techniques, such as water-table fluctuations, provide recharge estimates on event time scales also. In arid and semiarid areas where deep drainage fluxes are low and water-tables are deep, interpreting groundwater hydrographs and water-table rises may be misleading for estimating rates of groundwater recharge; chemical and isotopic methods are likely to be more successful than physical methods in such cases.

Cost and time requirement for the various approaches generally vary considerably. If recharge estimates have to be developed within a short time (months), then techniques based on long-term monitoring (several years) cannot be used. In that case, tracer techniques may be more suitable. Although sampling and analysis of chemical and isotopic tracers are usually considered to be expensive, one-time sampling is generally sufficient; therefore, the costs may be less than those associated with long-term monitoring that require monitoring equipment and continual collection and analysis of data.

Techniques that require hydraulic-conductivity data, such as Darcy methods and unsaturated- and saturated-zone models, are inherently inaccurate because hydraulic conductivity can vary over several orders of magnitude. Various sources of uncertainty include those related to measurement of hydraulic conductivity, applicability of data at the measurement scale (laboratory vs. field scale) to the scale of recharge calculation, and spatial variability in hydraulic conductivity. Uncertainties in hydraulic conductivity are even greater in unsaturated systems than in saturated systems because of nonlinear relationships between hydraulic conductivity and water content. These uncertainties in hydraulic conductivity could readily result in order-of-magnitude uncertainties in recharge estimates. Uncertainties in recharge estimates based on tracer data include those associated with measurement of tracer concentrations, estimated inputs of tracers, and assumptions about tracer transport processes. These uncertainties are generally less than those associated with water-budget approaches or methods that use hydraulic-conductivity data.

The tracer techniques such as ^{36}Cl, ^{3}H, $^{3}H/^{3}He$, CFCs, ^{14}C, and Cl can provide integrated, long-term estimates of recharge are. Tracers are very useful for estimating net recharge over long time periods but generally do not provide detail time series information on variations in recharge.

7. CHALLENGES AND RESEARCH NEEDS

In spite of numerous research efforts throughout the globe, both accurate measurement and modeling of recharge is still a challenging task. Addressing variation of recharge response in simulation model with the variation of precipitation, land-use, soil structure, geology (both in space and time) remains a challenge.

Table 2. Recharge estimates throughout the globe using different methods

Sl	Country, region	Aridity type / Yearly rainfall	Method of R estimation	Recharge value (mm/yr)	Recharge coefficient	Reference
1.	Australia, Southern	Semi-arid	CMB	52-63	-	Ordens et al. [12]
			WTF	47-129	-	
2.	China Ordos Plateau	Arid to Semi-arid	WTF	46-109		Yin et al. [14]
			Sat. zone Darcy	17-54		
		(Rainfall: 189-342 mm)	WB	21-109		
			CMB	5-74		
3.	Ireland		WB (3 yrs av.)	284	81-85%	Misstear et al. [15]
	(Gravel aquifer)		WTF (S_y = 0.19)		70-100%	
			WTF (S_y = 0.13)		40 - 80%	
			SMB (P-G method)	334		
4.	Burundi, northeastern		WB model (Thornwaite & Mather)	235.11		Bakundu-kize et al. [19]
5.	USA, east-central Pennsylvania	Humid Continental	Zero-tension Lysimeter	311		Risser et al. [17]
			WB	308		
		(Mean rainfall 1069 mm)	WTF	252		
			Rorabangh eqn.	357		
6.	Zimbabwe, Nyamandh area	Semi-arid	CMB	19-62		Sibanada et al. [16]
			WTF	2-50		
		(Av. Rainfall 555 mm)	Darcy	16-28		
			^{14}C	22-25		
			GW modeling	11-26		
7.	USA, North Carolina (Coastal Plain)	Rainfall: 1170 mm (JO-035 site, record: 1987-2004)	WTF	140		Coes et al. [18]
			Darcy's law	110		
			Hydrograph separation	34		
8.	India, Andhra Pradesh	Rainfall: 570 mm	^{3}H tracer	25	5%	Chand et al. [38]
9.	Saudi Arabia (Western side)	Arid (R: 345mm)	CMB		11%	Subyani and Sen [82]
10.	USA, Texas and New Mexico	Semi-arid	CMB	11 ±2	2%	Wood and Sanford [37]
11	North-eastern Bangladesh	Humid sub-tropic	Chloride tracer	228.7	11.2 %	Ali [54], Ali [55]
		Rainfall: ~2000 mm	WB	141.6	7.16 %	

Table 3. Recharge estimates throughout the globe using different methods

Sl	Country, region	Aridity type / Yearly rainfall	Method of R estimation	Rechar-ge value (mm/yr)	Recharge coefficient	Reference
1.	Sweden, southeastern (moraine area)		WTF based	134-194		Johansson [93]
2.	India (For 35 study areas)		^{3}H injection	24-198	4.1-19.7%	Rangarajan and Athavale [30]
3.	Australia (western, deep coastal sand)	Rainfall 775 mm	Environ-mental Cl		15%	Sharma and Hughes [94]
4.	Senegal (coastal quarter nary aquifer)	Sahel (Rainfall 280 mm)	Environ-mental CMB	0.11- 1.3 %		Edmunds and Gaye [95]
5.	China, North China Plain (For model, 2001-2009)		Tracer Model INFIL3.0	108 102	16% 14%	Tan et al. [96]
6.	USA, Minnesota (Glacial deposit)	500-900 mm (1971-2000)	SMB WTF Age dating of GW RRR model		33-40% 16-26% 24 % 23%	Delin et al. [58]
7.	USA, Southeastern Wisconsin	Rainfall 750-900 mm	Model	110		Cherkauer [92]
8.	Australia, Southern	Semi-arid	Cl (natural/ Env.)	0 - 3 mm		Allison and Hughes [79]
9.	Mali, Timbuktu	225 mm	CMB		1 - 2%	Jacks and Traore [80]
10	Palestine West Bank		CMB	95.2-323.6	15-50%	Marei et al. [40]
11	Canada (Kenogami Upland.)		1-D Dupuit-Forchheimer model	3.5	0.4%	Chesnaux [97]

Table 4. Recharge estimates throughout the globe using different methods

Sl	Country, region/other characteristics	Aridity type /Rainfall	Method of estimation	Recharge value (mm/yr)	Recharge coefficient	Reference
1	Argentina (Pampa plain)	Rainfall: 1064 mm	WTF, S_y = 0.09	210	18%	Varni et al [98]
	Shallow aquifer	(Av. of 18 years)	WTF, S_y = 0.07	164	14%	
2	China (northwest, Luanjng)		CMB (at natural site)	0.1	0.06%	Liu et al [99]
			CMB (at irrigation site)	268		
3	California		Soil moisture monitoring	180		Houston [100]
			CMB	42-141		Shivanna et al. [101]
			SWB	75-164		Kendy et al. [22]
4	Chile (northern, Atacama desert)		WTF		6%	Subyani and Sen [82]
5	India (Karnataka)	Semi-arid (Rainfall: 550 mm)	^3H	33		Zagana et al. [42]
6	China (Hebei Province, North china Plain)		Soil-moisture balance model (1949-2000)	50-1090		Houston [102]
7	Saudi Arabia	Arid to Semi-arid	Modified CMB		11% of effective Rainfall	
8	Jordan	Arid to semi-arid	CMB (shallow soil)	14	2.8%	Githui et al. [34]
	(Thick desert soil)		CMB (desert soil)	3.7	3.7%	
9	UK		CMB			Touhami et al. [103]
			Model	28		Hagedorn et al. [104]
10	Australia (Southeast)	Semi-arid	SWAT model	147-289	40% of (P+Irri)	Wang et al. [28]
11	Spain (Southeastern)	Semi-arid	HYDRO-BAL	0-59	0-18%	
12	Korea	Humid to semi-arid	WTF	687		
	(Jeju, volcanic island)		CMB	429		
			CFC-12	423		
			^3H,	394		Grismer et al. (2000) [105]
			SWB	911		
13	China (Hebei plain)	Rainfall: 857 mm	^3H, Br	0-1.05 mm/d	0 - 42.5%	

Table 5. Suitability of the methods in terms of spatial and temporal scale, capability of recharge estimate (actual or potential), and range of estimates

Sl	Method	Spatial scale	Temporal scale	Capability of estimate recharge (Actual/Potential)	Capability of range
1.	Lysimeter	Local	Event to year	Potential	Not limited (any)
2.	Seepage meter	Local	Event	Actual	Not limited (any)
3.	Dracy flux	Local	Event to year (by sum)	Potential	Small to medium
4.	WTF	Local to catchment/regional		Actual	Not limited (any)
5.	CMB	Local to catchment/regional		Actual*[1]	Small to medium
6.	Applied tracer	Local to catchment/regional	Year	Actual*[1]	Small to medium
7.	Radioactive tracer	Local to catchment/regional	Year	Actual*[1]	Small to medium
8.	GW age dating	Local to regional	Years to long-term average	Actual	Not limited
9.	Hydrograph separation	Watershed/ catchment/regional	Months to years	Net (i.e Actual- loss)	Medium to large
10.	Numerical models	Catchment to regional	Months to year		Medium to large
11.	SMB	Local	Event to year	Potential	Small to medium
12.	WB	Catchment to regional	Months to years	Potential	Medium

**1: Gives potential estimates, if withdrawal > recharge*

Heterogeneity of surface and sub-surface geology, climate (mainly rainfall and ET), land-use change, and their interaction in recharge process makes the complexity in correct estimation of recharge. Physically based model aided by remote sensing (RS) and GIS data is the hope in this aspect. But the RS and GIS technologies (both software and hardware) are not reachable in many areas/countries. Therefore, it is a crucial need to standardize a method which is easily implementable with physical measurements (such as WTF method) comparing with the RS and GIS data; and then regression model should be developed with rainfall so that recharge estimate for a particular year can be made from the rainfall data.

Most studies (and methods) estimated "point recharge" and inferred for basin scale recharge. To address the spatial variability and point(site)-to-basin(or regionalization) inferation, there is a lacking of systematic approach (statistical or others). Researcher should come forward in this issue.

In addition, a little effort is observed in the literature for estimation of recharge in humid climate (compared to arid climate) where a large amount of rainfall is concentrated over 4 to 6 months. The WB and CMB methods are associated with uncertainties in such regions due to lack of accurate estimation of runoff amount. The WTF method seems reliable than other methods in those regions. Few studies are explicitly concerned with confined aquifer, which occupies a large share of underground reservoirs in many regions. It is an urged need to adapt the WTF method for confined aquifer. Theoretical and experimental efforts are needed in this aspect.

8. CONCLUSION

Groundwater is the main source of water supply to both urban and rural populations as well as to industry and agriculture. To properly manage groundwater resources, managers need accurate information about the inputs (i.e., recharge) and outputs (i.e., pumpage and natural discharge) within each groundwater basin, so that the long-term behavior of the aquifer and its sustainable yield can be estimated or reassessed. Recharge is a major component of the groundwater system and has important implications for shallow groundwater quality. Quantitative determination of the rate of natural groundwater recharge is a pre-requisite for efficient groundwater resource management.

The soil, vegetation, topography, geology and climate (specially rainfall amount and distribution, and temperature) of a site control the recharge, and therefore impact on the choice of a technique for estimating recharge. In addition, geomorphology, sources and mechanisms of recharge, spatial and temporal scale, estimated recharge rate at the site, accuracy needed, and availability of technology/facility at the site also influences on the selection of a method.

COMPETING INTERESTS

Authors have declared that no competing interests exist.

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