CHAPTER 3
METHODS FOR ESTIMATING GROUNDWATER RECHARGE

Groundwater is considered to be one of the most important water resources. Accurate of groundwater recharge estimation is important for proper management of groundwater systems. Many different approaches are available to quantify recharge. The present study presents a review of various methods that have been used to quantify recharge. The theories underlying the methods are explained. Various methods used in previous studies on Chiang Mai basin are also reviewed.

3.1 State of the art

The various methods for estimate groundwater recharge can be divided into three groups based on data from the different hydrologic zones (Figure 3.1). Within each zone, the methods for recharge estimate can be classified as hydrologic budget, physical, tracer, or numerical-modeling approaches. These are

3.1.1 Methods based on surface water studies

1) Channel water budget approach
2) Seepage meters
3) Stream hydrograph analysis
4) Heat tracer
5) Isotopic tracers
6) Watershed modeling

3.1.2 Methods based on unsaturated zone studies

1) Lysimeters
2) Zero flux plane
3) Darcy's law
4) Tracer approaches
5) Numerical modeling

3.1.3 Methods based on saturated zone or groundwater zone studies

1) Groundwater budget approach
2) Water table fluctuation approach
3) Darcy's law
4) Tracers approaches
5) Numerical modeling

In each approach, recharge estimates, attractive, limitation and reliability of the recharge estimation are described (Healy and Cook, 2002; Scanlon et al., 2002; Otto, 2001; Weight and Sonderegger, 2000; Fetter, 1994; Anderson and Woesser, 1992; Downing and Wilkinson, 1991; Viessman et al., 1977 and Walton, 1970). Choosing appropriate methods for quantifying groundwater recharge in varying space and time scale are of importance and depend on various hydrogeologic characteristics and conditions.

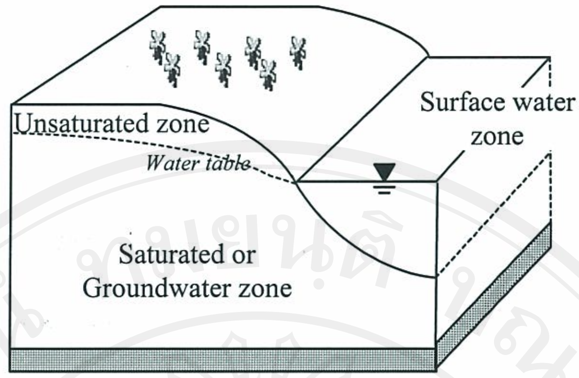

Figure 3.1 Hydrologic zones.

3.1.1 Methods based on surface water studies

Surface water commonly is hydraulically connected to groundwater. Stream interacts with groundwater in all types of landscapes. The interaction take place in three basic ways: streams gain water from inflow of groundwater through the streambed (gaining stream, Figure 3.2), they lose water to groundwater by outflow through the streambed (losing stream, Figure 3.3), or they do both, gaining in some reaches and losing in other reaches (Winter et al., 1998).

The water fluxes related to the degree of connection between surface water and groundwater system are described by Scanlon et al. (2002). The aquifer discharges to the river when the groundwater head is greater than the river stage, whereas the river recharges the aquifer when the river stage is greater than the groundwater head (Figure 3.4). Recharge values generally reach a constant rate when the water table depth is greater than twice the river width. Recharge can be estimated using surface water data in gaining and losing surface water bodies.

Recharge related to surface water bodies, stream or canal, is difficult to estimate due to the large variability in flows. Physical methods are practically impossible and tracer approaches are of little practical use. The most commonly used methods are channel water budget, seepage meters, stream hydrograph analysis and numerical modeling.

1) Channel water budget approach

Water budget or hydrologic budget for a basin is based on a simple statement of the law of mass conservation. It may be expressed as:

$$\text{Inflow} = \text{Outflow} \pm \text{Changes in storage} \tag{3.1}$$

The water budget considers all waters (surface, subsurface and groundwater) flow into the basin, leaving or stored within the basin and flow out of the basin. Volume of storage water is difference between rate of inflow and outflow in a water system. Groundwater recharge estimation using the water budget approach is the indirect approach. Groundwater recharge can be written as:

$$\text{Groundwater recharge} = \text{Inflow} - (\text{Outflow} \pm \text{Changes in storage}) \tag{3.2}$$

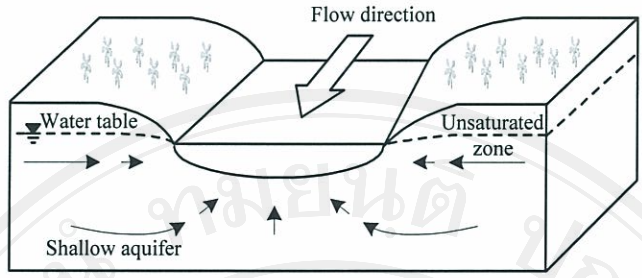

Figure 3.2 Gaining streams receive water from the groundwater system (Winter et al., 1998).

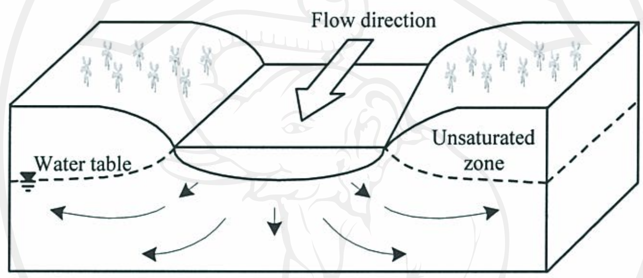

Figure 3.3 Losing streams lose water to the groundwater system (Winter et al., 1998).

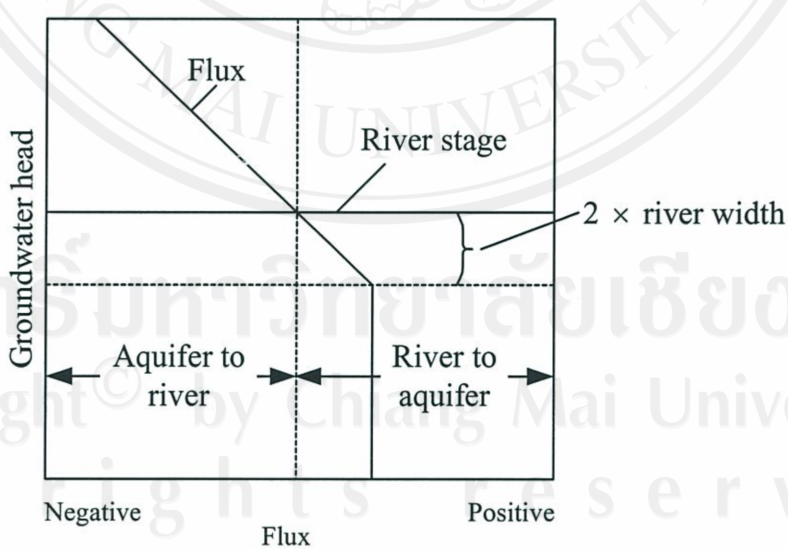

Figure 3.4 Water fluxes related to the degree of connection between rivers and aquifer (Scanlon et al., 2002).

All components in the equation except groundwater recharge are measured or estimated by separate methods and the groundwater recharge is set equal to the residual. The major components of inflow, outflow and changes in storage are listed in Table 3.1. All of the components are given as rates (e.g. millimeter per day; mm/d or millimeter per year; mm/y).

Table 3.1 Major inflow, outflow and changes in storage components (Fetter, 1994).

Inflow	1) precipitation 2) surface water inflow into the basin including streamflow and overland flow 3) groundwater inflow from outside the area 4) artificial import of water into the basin through pipes and canals
Outflow	1) evapotranspiration from land areas 2) evaporation of surface water 3) runoff of surface water 4) groundwater outflow 5) artificial export of water through pipes and canals
Storage	1) surface water in streams, rivers, lakes, and ponds 2) soil moisture in the vadose zone 3) ice and snow at the surface 4) temporary depression storage 5) intercepted water on plant surfaces 6) groundwater below the water table

However, for many basins, several items of the water budget can be eliminated because they do not measurably affect the balance between water gains and losses. The water budget approach for the surface water zone considers surface water gain or losses. The channel water budget approach is based on stream gauging data (Scanlon et al., 2002). The loss in streamflow between upstream (Q_{up}) and downstream (Q_{down}) gauging stations, transmission loss, can be measured. The simple budget can be considered from the transmission losses, the loss in streamflow between upstream and downstream gauging stations. This loss reflects to the potential recharge. The difference in flow between two gauged stations plus any inputs from tributaries minus evaporation from the river surface minus any change storage over change in time is attributed to recharge to surrounding groundwater. The recharge rate, R, can be written as:

$$R = Q_{up} - Q_{down} + \sum Q_{in} - \sum Q_{out} - E_{sw} - \frac{\Delta S}{\Delta t} \qquad (3.3)$$

where

Q_{up} is flow rate at the upstream,

Q_{down} is flow rate at the downstream,
Q_{in} is tributary inflows along the reach,
Q_{out} is tributary outflows along the reach,
E_{sw} is evaporation from surface water or streambed,
ΔS is change in channel and unsaturated zone storage, and
Δt is change in time.

The range of recharge rates depends on the magnitude of the transmission losses. The losses are relative to the uncertainties in the gauging data and tributary flows. The difficult to determine accuracy, as main problem, when the recharge component is often a small difference of a number of large term. The recharge values range from event scale, minutes to hours, to much longer time scales that are estimated by summation of individual events.

2) Seepage meters

The seepage flux between groundwater and surface water can be measured directly using a seepage meter. The seepage meter is a constructed by inserting an open-bottomed container into the stream sediments and then measuring the time it takes for a volume of water to flow into or out of a bag connected to the container (Weight and Sonderegger, 2000). Figure 3.5 shows schematic of installed seepage meter. The measured volume of water represents the seepage velocity. This is related to groundwater recharge rate. Seepage velocity or average linear velocity (v_s) can be calculated from Equation 3.4.

$$v_s = \frac{Q}{n_e A} \tag{3.4}$$

where

v_s is the average linear seepage velocity (m/s),
Q is the rate of water flow (m^3/s),
n_e is the effective porosity of unconsolidated sediments (% or fraction), and
A is the cross-sectional area of the container (m^2).

This method is rapid, direct and cheap to apply. Uncertainties depend on factor measurements. The seepage meters may be required at many locations to obtain a representative value. Time scales range from individual events to day and longer times estimated from a summation of shorter times.

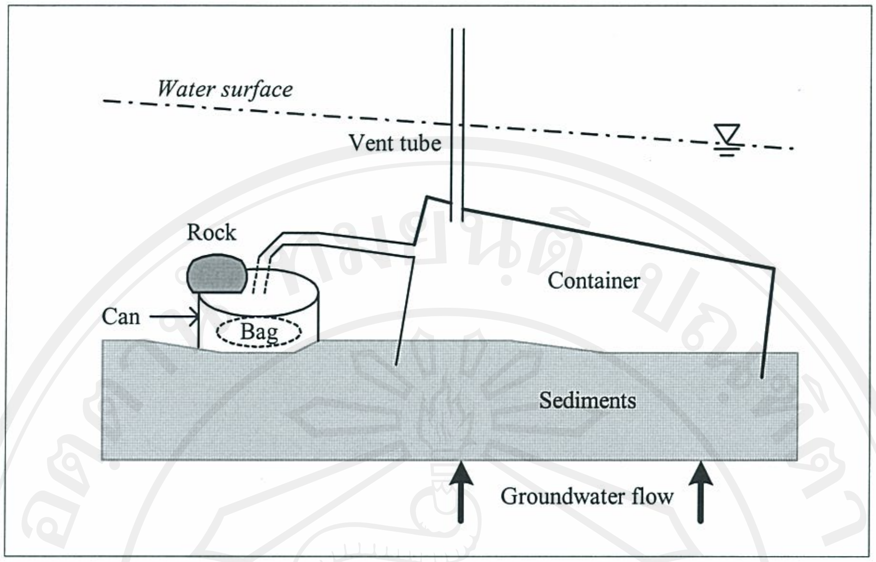

Figure 3.5 Schematic of installed seepage meter (modified from Weight and Sonderegger, 2000).

3) Stream hydrograph analysis

A stream hydrograph show the discharge of a river at a single location as a function of time. The total streamflow is possible to break down the hydrograph into components such as surface runoff and groundwater runoff. In watershed with gaining streams, groundwater recharge can be estimated from (i) baseflow separation and (ii) baseflow recession.

(i) Baseflow separation

Viessman et al. (1977) described three methods for baseflow separation as follows (Figure 3.6); (a) straight line method (b) fixed base length method and (c) arbitrary method.

(a) Straight line method The simplest separation is to draw a line between A and B, horizontal line from the point at with surface runoff begins to an intersection with the hydrograph recession. Addition, the widely used separation is to draw a line between A and F. F is the end of direct runoff.

(b) Fixed base length method This method projects the initial recession curve downward from A to C, which lies directly below the peak rate of flow. Then, point D on the hydrograph, representing N days after the peak, is connected to point C by a straight line defining the groundwater component. The N days can be calculated by

$$N = A_d^{0.2} \tag{3.5}$$

where

N is the time in days, and
A_d is the drainage area in square miles.

(c) Arbitrary method The procedure employs a base flow recession curve that is fitted to the hydrograph and then computed in a time decreasing direction. Point F, where the computed curve begins to deviate from the actual hydrograph, marks the end of direct runoff. The curve is projected backward arbitrarily to some point E below the inflection point and its shape from A to E is arbitrarily assigned.

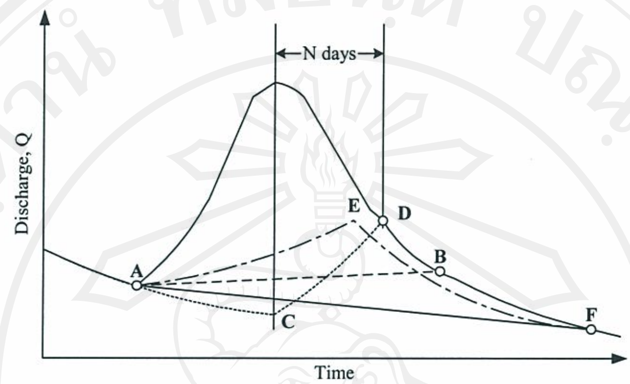

Figure 3.6 Hydrograph separation methods (Viessman, et al., 1977).

(ii) Baseflow recession

The stream hydrograph during a dry period will decay, following an exponential curve, because the stream drains water from the groundwater reservoir (gaining stream), leaving less and less groundwater to feed the stream (Figure 3.7). The part that groundwater seepage into the stream, baseflow period of baseflow recession, can be used to estimate groundwater recharge. The baseflow recession method is based the water budget approach, in which recharge is equated to discharge. The baseflow recession equation (Fetter, 1994) is

$$Q = Q_0 e^{-ct} \tag{3.6}$$

where

- Q is the flow at some time t after the recession started (m^3/s),
- Q_0 is the flow at the start of the recession (m^3/s),
- c is the recession constant for the basin, calculated from the rewritten equation; $c = -(1/t)\ln\left(Q/Q_0\right)$ (d^{-1}), and
- t is the time since the recession began (d).

Installation and maintenance of stream gauging stations are expensive and difficult. The accuracy of the groundwater recharge depends on the validity of the assumption. The minimum time scale is a few months. Recharge estimation for the longer period can be obtained from the summation of shorter period estimation.

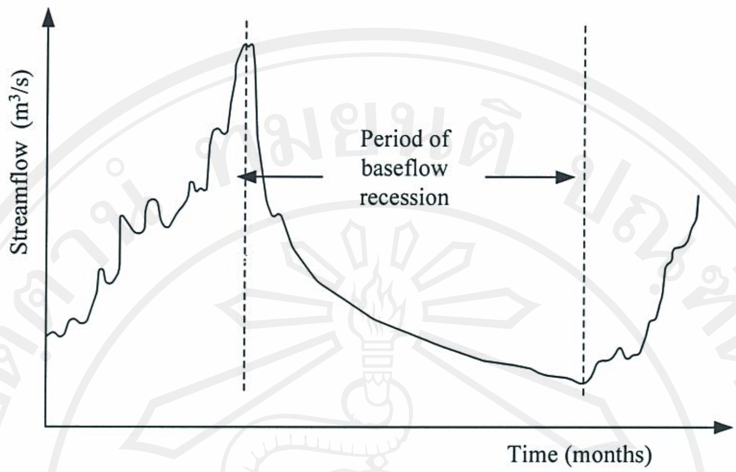

Figure 3.7 Annual hydrograph for a river with a long dry summer season (Fetter, 1994).

4) Heat tracer

Scanlon et al. (2002) described the groundwater recharge estimation from ephemeral streams in semiarid regions using the heat tracer. Heat can be used as a tracer to estimate infiltration. It is used with inverse modeling (mathematics modeling) using a nonisothermal variably saturated flow code, such as VS2DH, to estimate hydraulic conductivity of sediments and then estimated percolation rates. Temperature can be monitored accurately and inexpensively using thermistors or thermocouples at various depths. The monitoring depths depend on time scale, sediment types and anticipate water fluxes beneath the stream. Monitored temperature at depth of about 0.05 to 1 meter for fine grained material and 0.3 to 3 meters for coarse grained material, for example. The minimum net infiltration rate depends on surface water temperature and the time scale considered. Recharge can be estimated for time periods ranging from hours to years.

5) Isotopic tracers

Natural environmental tracers of oxygen and hydrogen are used to identify groundwater recharge. Due to rivers often retain the depleted isotopic signature of the headwater, the different in the isotope signatures of surface water and local precipitation can be used to determine the relative contribution of these two sources of groundwater recharge. However, the method that using tracer is generally difficult to quantify recharge rate. The time scales range from seasonal in areas of high flux to hundreds of years in areas of low flux.

6) Watershed modeling

Numerical modeling is generally used as a tool to evaluate flow processes and to assess sensitivity of model output to various parameters. The numerical modeling can generally be used to estimate any range in recharge rates. The reliability of recharge estimate using numerical modeling depends on uncertainties in the model parameters. There are seven main steps to work on the model; purpose determining, construction of the conceptual model, code selection, model design, model calibration, model verification and predication (Anderson and Woesser, 1992). Flow chart of groundwater modeling is shown in Figure 3.8.

Numerical modeling for surface zone, watershed or rainfall and runoff modeling, the recharge process and its relationship with rainfall have been made to estimate the amount and process of recharge by infiltration using precipitation data. Most of watershed modeling is based on the water budget equation. The minimum recharge rate depends on the accuracy of the water budget parameter and the time scale considered. The watershed models are applied at a variety of scales over large areas.

3.1.2 Methods based on unsaturated zone studies

The unsaturated zone is defined here as the zone of an unconfined aquifer above the water table. It comprises the materials from the land surface down to the water table; the soil, the unsaturated material, including the capillary fringe or zone. It has vertical water movement both downward and upward where lateral potential gradients are only significant in the rooting zone of the soil. Unsaturated zone methods for estimating groundwater recharge provide potential recharge based on drainage rates below the root zone. Term of drainage is used to describe downward water movement below the root zone and it is often equated to recharge. The unsaturated zone methods are applied mostly in semiarid and arid regions that are thick unsaturated zones. The recharge estimated from the unsaturated zone methods are generally apply to smaller spatial scales than the others. Most of unsaturated zone methods provide point estimates of recharge. The approaches for recharge estimation are lysimeters, zero flux plane (ZFP), Darcy's law, tracer approaches and numerical modeling.

1) Lysimeters

Lysimeters are essentially containers of rock and soil, with or without vegetable that are through which precipitation infiltrates to allow collection and measurement of its quantity. Surface areas of lysimeter range should not be small (range from 1-300 m^2)(Scanlon et al., 2002) and depth should be deeper than the root zone to protect the drainage overestimate. The lysimeter that have been used to measure the precipitation and water storage is called pan or non-weighing lysimeter. While weighing lysimeter is constructed on delicate balance capable of measuring slight changes in weight.

Lysimeters have often been used in agricultural work to study soil moisture and evapotranspiration measurement. They have not generally been used to measure direct recharge to aquifer because they are expensive and difficult to construct and have high maintenance requirements. Lysimeters are generally unsuitable for areas with deep-rooted vegetation. Recharge rates can be estimated at time scales from minutes to years. The minimum water flux that can be measured using a lysimeter depends on the accuracy of the drainage measurements and the surface area of the lysimeter.

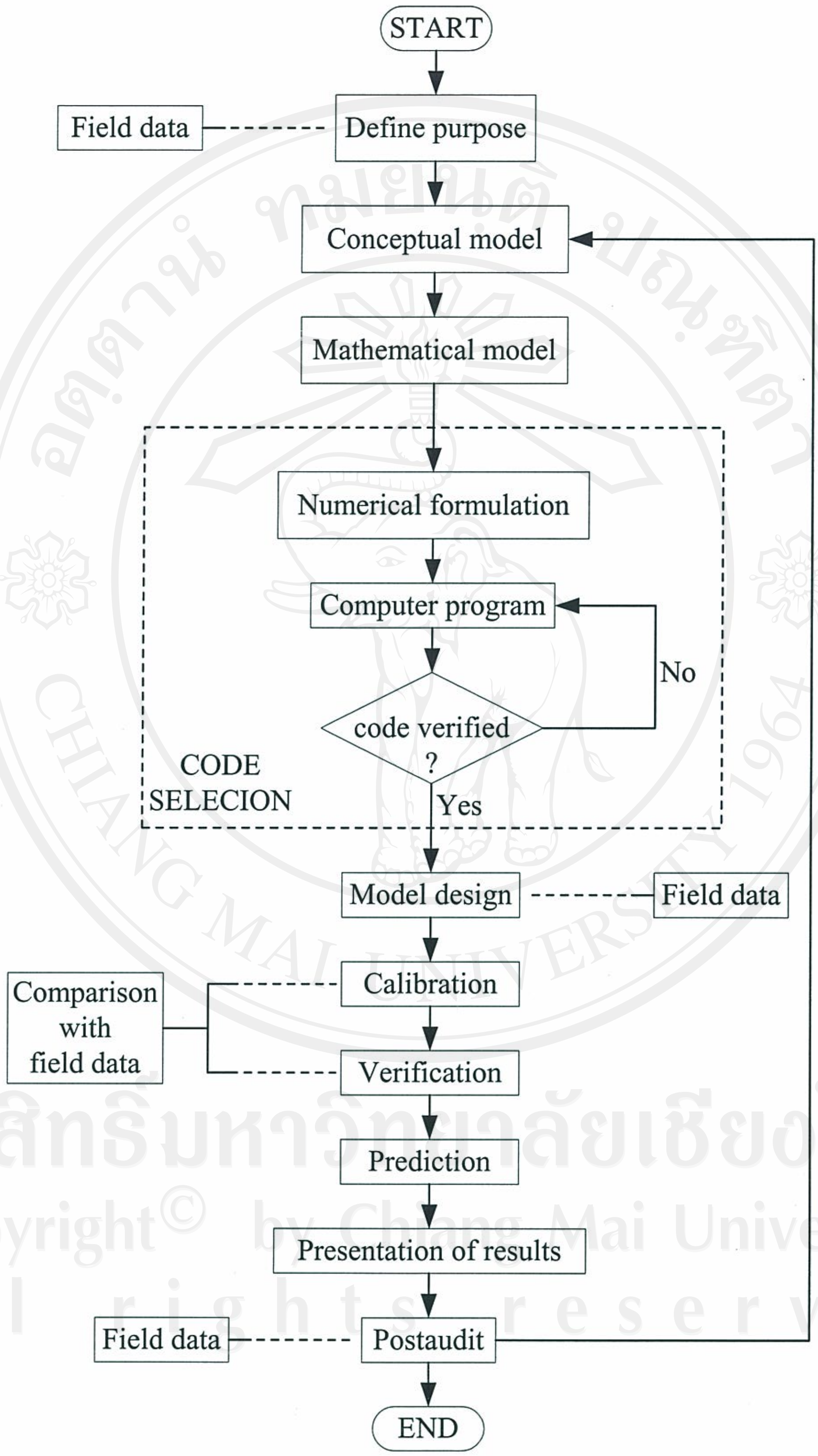

Figure 3.8 Flow chart of groundwater modeling
(modified from Anderson and Woesser, 1992).

For example, Otto (2001) studied the groundwater recharge in the southeastern Holstein region (1,392 km^2), northern Germany by lysimeter technique. The technique of lysimeter equations is to calculate the groundwater recharge by considering the relation between amount of percolation water and precipitation. Quantities of precipitation-water are determined on the basis of drainage-outflow amount at the base of lysimeter. A linear correlation between these parameters can be shown in the lysimeter equation:

$$Pw_{pot} = a \times (P - b) \tag{3.7}$$

where

Pw_{pot} is the amount of percolation water (mm/y),
P is precipitation (mm/y),
a is gradient of straight line, and
b is value of ordinate if $P=0$, respectively.

There are many different lysimeter equations due to the different of soil characteristics, vegetation and climate. The parameters a and b can be determined and vary considerably with many lysimeter data and lysimeter statistics. The equations that Otto (2001) used to assess groundwater recharge rate southeastern Holstein, Germany, are:

$Pw_{pot} = 1.1 \times P - 433$ for the area with sandy soils and field/grassland,
$Pw_{pot} = 1.1 \times P - 474$ for the area with sandy soils and forest,
$Pw_{pot} = 1.1 \times P - 558$ for the area with clayey/silty soil and field/grassland, and
$Pw_{pot} = 1.1 \times P - 578$ for the area with clayey/silty soils and forest.

Lysimeter equations are based on the assumption that surface runoff equal zero. So, the actual quantities of percolation water was determined after subtraction of the surface runoff:

$$Pw_{real} = Pw_{pot} - R_o \tag{3.8}$$

where

Pw_{real} is real percolation water quantity,
Pw_{pot} is potential percolation water quantity, and
R_o is surface runoff including interflow.

To calculate the groundwater recharge rate, the percolation water quantity must subtract various factors of water lost such as evapotranspiration and surface runoff. The potential amount of percolation water is calculated by subtraction the evapotranspiration from the summation of lysimeter equations in different conditions. So, the potential amount of percolation water is given by Otto (2001):

$$
\begin{aligned}
Pw_{pot} = {} & A_{sand/fields,grassland} \times Pw_{pot\ sand/fields,\ grassland} + A_{sand/forest} \times Pw_{pot\ sand/forest} \\
& + Pw_{clay/field,grassland} \times Pw_{pot\ clay/fields,grassland} + Pw_{clay/forest} \times Pw_{pot/forest} \\
& - A_{ld} \times ET_a
\end{aligned} \tag{3.9}
$$

where

ET_a is additional evapotranspiration, and
A_{ld} is area that consider evapotranspiration above shallow water table.

Total runoff comprises the surface runoff (R_o) and groundwater runoff (R_g), recorded by hydrological water gauges. With the help of R_o / R_g separation method, the groundwater runoff can be determined. The surface runoff (R_o) is the difference between the total runoff (R_{tot}) and groundwater runoff (R_g): $R_o = R_{tot} - R_g$. The groundwater recharge rate (GWR) is then determined after subtraction of the surface runoff: $GWR = Pw_{pot} - R_o$ (mm/y). For a unit cell of the area, the equation of groundwater recharge calculating is $GWR_{cell} = Pw_{pot} - R_{o\ cell}$ (mm/y).

Otto (2001) had used the spreadsheet program LOTUS 1-2-3 for estimate groundwater recharge rates in the southeastern Holstein region, northern Germany. All regional data required for the determination of groundwater recharge were stored in grid cells in the worksheets of a spreadsheet file (Figure 3.9). The calculation process is outlined in Table 3.2. The average groundwater recharge rate of the area from 1980 to 1999 is 165 mm/y.

2) Zero flux plane

The zero flux plane (ZFP) approach is based on soil water budget to estimate groundwater recharge. The soil water budget can be simplified by equating recharge to changes in soil water storage below the ZFP. The ZFP is the plane which the vertical hydraulic gradient is zero. The ZFP separates water movement into upward (evapotranspiration) and downward (drainage). Above the ZFP a potential gradient is upward, moving water towards the rooting zone. Below the ZFP the gradient is downward, inducing drainage. The soil water budget equation (Downing and Wilkinson, 1991) is as follows:

$$\Delta S_z = P - R - ET_A - D_z \tag{3.10}$$

where

ΔS_z is water content change above depth z,

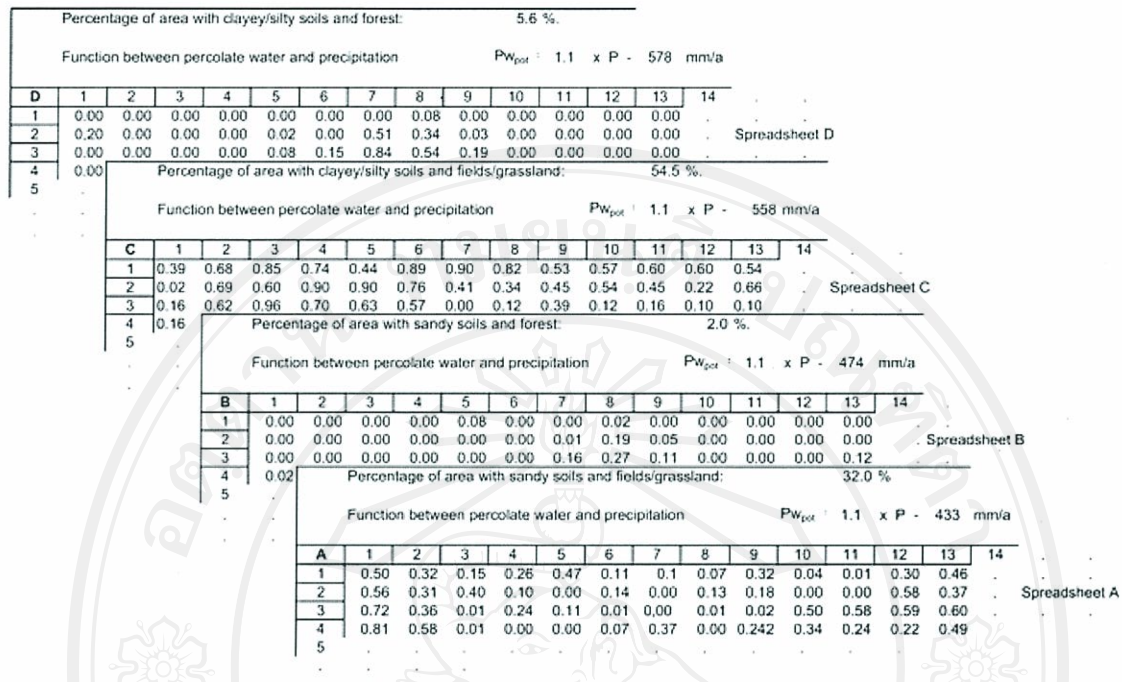

Figure 3.9 Partial worksheets for the groundwater recharge model of the Grosshansdorf area illustrating the percentages of types of soils and vegetation cover for each grid cell, spreadsheets E-I are shown in Table 3.2 (Otto, 2001).

Table 3.2 Calculation of groundwater recharge rates for each grid cell of the area studied using spreadsheet-calculation tools (Otto, 2001).

Spreadsheet no.	Content of grid cell	Formula
A	Percentage of area with sandy soils and fields/grassland	$A_{sand/fields,\ grassland}$ Lysimeter equation
B	Percentage of area with sandy soils and forest	$A_{sand/forest}$ Lysimeter equation
C	Percentage of area with clayey/silty soils and fields/grassland	$A_{clay/fields,\ grassland}$ Lysimeter equation
D	Percentage of area with clayey/silty soils and forest	$A_{clay/forest}$ Lysimeter equation
E	Additional evapotranspiration ET_a above shallow water table (90 mm/a), percentage A_{ld} of areas with shallow water table	$A_{ld} \times ET_a$
F	Precipitation	P
G[a]	Potential amount of percolation water of a cell	$Pw_{pot\ cell} = A_{sand/fields,\ grassland} \times Pw_{pot\ sand/fields,\ grassland} + A_{sand/forest} \times Pw_{pot\ sand/forest} + A_{clay/fields,\ grassland} \times Pw_{pot\ clay/fields,\ grassland} + A_{clay/forest} \times Pw_{clay/forest} - A_{ld} \times E_a$
H	Surface runoff including interflow	R_o
I	Groundwater recharge rate	$GWR_{cell} = Pw_{pot\ cell} - R_{o\ cell}$

[a]Grid formula in spreadsheet G: +A:C12*(A:$D$8*F:C12-A:$F$8) + B:C12*(B:$D$8*F:C12-B:$F$8) + C:C12*(C:$D$8*F:C12-C:$F$8) + D:C12*(D:$D$8*F:C12-D:$F$8) - E:C32

P is precipitation,
R is surface runoff,
ET_A is actual evapotranspiration, and
D_z is drainage below depth z.

The quantities of precipitation (P) are readily measurable, as are soil water content changes, but partitioning the residual water content change into evapotranspiration and drainage requires information about the direction of water movement (upward or downward) within the soil. Water content changes due to evapotranspiration and drainage can be separated when the ZFP is present. The depth of the ZFP can be identified by tensiometers. At the plane, water cannot flow through the plane. The soil water content can be measured by neutron probe or capacitance probe. Changes of water content in the soil profile above the ZFP will be due to exchange of water at the surface, either rainfall infiltration or evapotranspiration. Changes of water content below the ZFP will be due to drainage out of the base of the profile. The drainage will appear eventually as groundwater recharge.

The ZFP is usually present at times when, on average, evapotranspiration exceeds rainfall. If no ZFP can be identified it is most often because the profile is wet, water fluxes are downward throughout the entire profile. It is reasonable to assume that evapotranspiration losses will be close to the potential rate as calculated from meteorological data. Drainage from the profile, groundwater recharge, can be calculated from a simple hydrologic balance of the soil profile, knowing rainfall, evapotranspiration, and measured soil water content changes. The ZFP approach is best for the regions where large fluctuations exist in soil water content throughout the year and the water table is always deeper than the ZFP. The minimum estimated recharge depends on the accuracy of the water content measurements. Instruments and data collections for the ZFP approach are somewhat expensive.

3) Darcy's law

An approach to use Darcy's law to calculate recharge (R) requires the vertical total head gradient and the unsaturated hydraulic conductivity. The Darcy's law is described as (Scanlon et al., 2002):

$$R = -K(\theta)\frac{dH_{tot}}{dz} = -K(\theta)\frac{d}{dz}(h_p + z) = -K(\theta)(\frac{dh}{dz} + 1) \tag{3.11}$$

where

$K(\theta)$ is the hydraulic conductivity at the ambient water content (θ),
H_{tot} is the total head,
h_p is the matric pressure head, and
z is elevation.

The matric potential can be measured by the tensiometers. However, the assumption of a unit gradient (total head gradient is equal to 1) helps to remove the need to measure. This is for the thick unsaturated zones in thickly layered porous media, the matric pressure potentient is often nearly zero and moving water in the essentially gravity driven. The recharge within the unit gradient assumption is equal to the hydraulic conductivity at the ambient water content. The hydraulic conductivity can be obtained by using the steady state centrifuge (SSC) method (Scanlon et al., 2002). However, there is limitation due to the hydraulic conductivity varies by several orders of magnitude over the seasonal range of water content and is also spatially variable.

The method using Darcy's law provides a point recharge estimate throughout the entire year and may represent a larger area when it is applied at significant depths in thick vadose zone. The method accuracy is depending on the hydraulic conductivity and head gradient measurement if the latter is not assumed equal to one.

4) Tracer approaches

Tracer approaches can be used to give more basic information about water movement in the unsatured zone and rates of groundwater recharge. Various chemical or isotope, radioactive and environmental tracers can be used to estimate natural groundwater recharge in arid and semi-arid area. Tritium, H-3, Chlorine-36 and chloride are most common tracer used. Groundwater recharge can be (i) estimated from the vertical distribution of tracers (^3H and ^{36}Cl) and (ii) estimated from mass balance (chloride mass balance).

(i) Tritium-3 and Chlorine-36

Most of ^3H and ^{36}Cl are in hydrosphere and originates from cosmic radiation interacting with atmospheric gases. Large amount was introduced into the atmosphere through the nuclear testing during the [illegible] 3.10). The event marker can be used to estimate recharge rates during the past 50 years. A half-life of ^3H is about 12.4 years. Before 1953, tritium [illegible] precipitation were less than 10 tritium units (T.U.) and exceeded [illegible] T.U. in the period 1963-1964. The tritium concentration decreased during the subsequent decade to concentrations less than 100 T.U. (Charbeneau, 2000). Chlorine-36 (^{36}Cl) has a half-life of about 300,000 years. Chlorine is an excellent tracer for liquid phase water movement due to it is very strong, soluble and nonvolatile. Rainfall infiltration transports the tracers (^3H and ^{36}Cl), as a part of water vapor, to the soil profile. The net infiltration as recharge may be estimated from the depth of the peak tritium or chlorine-36 concentration in the soil profile. The recharge is given by (Scanlon et al., 2002).

$$R = \frac{\Delta z}{\Delta t}\theta \qquad (3.12)$$

where

Δz is the depth of the tracer peak,

Δt is the time between the tracer peak and sampling collection, and
θ is the average water content through the depth.

Theoretically, the maximum recharge that can be estimated is limited by depth to groundwater. If the water table were deeper, the tracer method could be used for higher recharge rates. However, soil sampling and locating the tracer peak may be more difficult and very expensive at the deeper depth. Tracer is applied for point areas over the last 50 years.

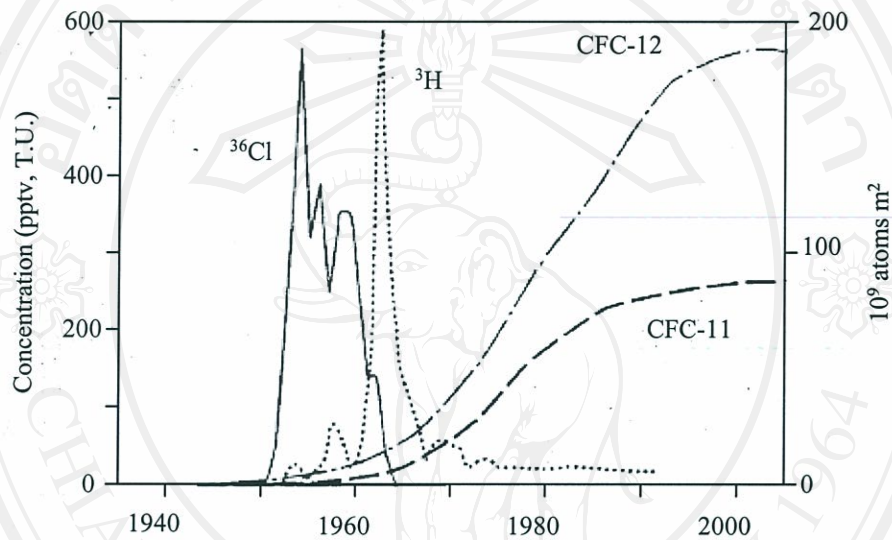

Figure 3.10 Input functions for historical tracers, including ^{3}H, ^{36}Cl, CFC-11, and CFC-12 (Scanlon et al., 2002).

(ii) Chloride mass balance (CMB)

Chloride (Cl) is produced naturally in the atmosphere. Natural chloride dissolves in precipitation. Infiltration of precipitation transports the chloride to depth. It is slowly accumulates in the soil profile and when water is evapotranspired back to the air, the dissolved chloride remains in the soil profile. The balance between the mass of Cl into and out of a system is used to estimate recharge rates. If C_p is the chloride concentration in precipitation and C_{uz} is the chloride concentration in the soil profile, then the drainage (D) is given by (Scanlon et al., 2002):

$$PC_p = DC_{uz} \quad \text{or} \quad D = \frac{PC_p}{C_{uz}} \tag{3.13}$$

where

P is the average precipitation rate.

The equation is valid where (1) all chloride in the system derives exclusively from precipitation, (2) chloride is not removed from the system, except transported by groundwater flow and chloride is not kept in storage in the soil. The chloride mass balance (CMB) approach provides point recharge estimation for decades to thousands of years. The accuracy of the estimation depends on the Cl concentration measurement.

5) Numerical modeling

Unsaturated zone modeling is used to estimate deep drainage below the root zone based on a variety approaches; soil-water storage, quasi-analytical and Richards equation (Scanlon et al., 2002). The reliability of recharge estimate should be checked against field information such as lysimeter data, tracers, water content and temperature.

3.1.3 Methods based on saturated zone or groundwater zone studies

Saturated zone or groundwater zone methods for recharge estimation provides evidence of actual recharge because water reaches the water table that is the top of the saturated zone in an unconfined aquifer. While surface and unsaturated zone methods provide estimates of drainage or potential recharge. The saturated zone methods commonly provide estimated recharge over larger areas than the other methods. The approaches of groundwater budget, water table fluctuation, Darcy's law, tracer approaches and numerical modeling are described.

1) Groundwater budget approach

The amount of precipitation percolates down through the soil and reaches the water table, becomes groundwater, can be estimated using the groundwater budget approach. The groundwater recharge is given by (Walton, 1970):

$$P_g = R_g + ET_g + U \pm \Delta S_g \tag{3.14}$$

where

P_g is groundwater recharge,
R_g is groundwater runoff,
ET_g is groundwater evapotranspiration,
U is subsurface underflow, and
ΔS_g is change in groundwater storage.

Groundwater runoff is groundwater recharged to stream. Periods are selected assuming that streamflow consists entirely of groundwater runoff. So, groundwater runoff is assumed here as streamflow. Groundwater evapotranspiration is groundwater discharged into the atmosphere by the process of evapotranspiration. It is largely a function of season and the mean groundwater stage. In summer months, groundwater evapotranspiration is great. It is very effective in reducing groundwater runoff. It is large at places where depth from surface to water table (mean groundwater stage) is small and it generally decreases as the mean groundwater stage declines. Groundwater evapotranspiration can be estimated from rating curves of mean groundwater stage versus streamflow on corresponding dates. The difference in groundwater on corresponding dates. The difference in groundwater runoff between the curve for the summer months (April through October) and the curve for the other period (November through March) is approximate groundwater evapotranspiration. Subsurface underflow can be estimated from Darcy's equation. Change in groundwater storage can be computed by substitute an appropriate values of gravity yield (Y_g) into the equation 3.10.

$$\Delta S_g = \Delta H_s (Y_g) \tag{3.15}$$

where

ΔH_s is the change in mean groundwater during an inventory period.

The groundwater budget approach can be applied ranging from small scale (centimeters, seconds) to large scale (kilometers, centuries). The accuracy of the recharge estimate depends on the accuracy of the measurement or estimation the components.

2) Water table fluctuation approach

Groundwater fluctuation follows an annual cycle as a result of seasonal variations in the quantity of effective rain. Groundwater levels rise in the recharge area can be used to estimate the annual volume of recharge to an unconfined aquifer. The annual volume of recharge (R) is given by

$$R = S_y \frac{\Delta h}{\Delta t} \tag{3.16}$$

where

S_y is specific yield (dimensionless),
Δh is water table height (m), and
Δt is an inventory period (y).

The water table fluctuation (WTF) approach provides a very simple, quick estimate of groundwater recharge.

3) Darcy's law

Darcy's law is used to estimate flow through a cross section. The Darcy's law approach assumes steady conditions and no water extraction. The approach requires estimates of hydraulic conductivity (K), hydraulic gradient and the cross sectional area. Estimates of the flow in the aquifer, both confined and unconfined aquifer commences with the construction of a flow net.

A flow net consists of two sets of lines; flow lines which represent the paths of water particles as they move through aquifer, and equipotential lines, which are lines of equal piezometric head or water level. The two sets of lines are arranged to form an orthogonal pattern of small squares (Figure 3.11). The flow through the orthogonal, Q, can be written in the form of

$$Q = A_{rec} K \frac{\Delta h}{\Delta l} \tag{3.17}$$

where

A_{rec} is the cross section area of a rectangles,
K is hydraulic conductivity, and
$\frac{\Delta h}{\Delta l}$ is the hydraulic gradient across the rectangle.

The hydraulic gradient can be estimated along a flow path at right angles to potentiometric contours. The advantage of the method based on Darcy's law is easy to apply, relatively quick and inexpensive. The method provides a point recharge estimate throughout the entire year and may represent a larger region when it is applied at significant depths in thick vadose zone. The region is about 1 to ≥ 10,000 km^2 with the time range from years to hundreds of year (Scanlon et. al., 2002). The flow net is subjective and difficult to construct when the flow pattern is complex. The highly uncertain of recharge estimates come from the high variable of hydraulic conductivity that are several orders of magnitude and head gradient measurement if the latter is not assumed equal to one.

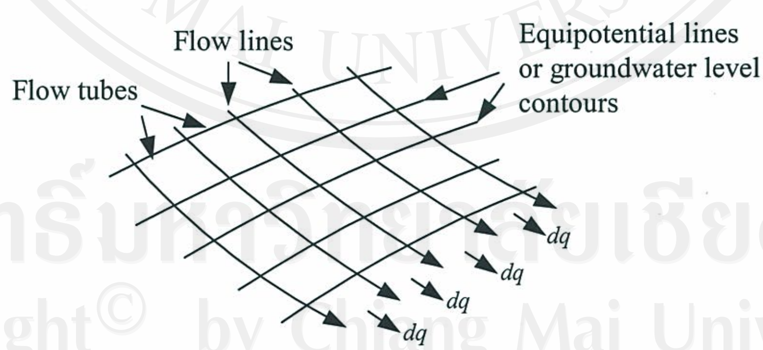

Figure 3.11 Flow net in two dimensional approximation (Hamill and Bell, 1986).

4) Tracers approaches

Tracers studies are used in both unsaturated and saturated zone to estimate recharge. Recharge rates can be determined from the two different ways: (i) estimating ages of groundwater (groundwater dating) and (ii) estimating groundwater chloride concentrations (Chloride Mass Balance approach).

(i) Groundwater dating

The tracers used to date groundwater are radioactive isotope (tritium, ^{3}H, and carbon-14, ^{14}C) and gasses such as chlorofluorocarbons (CFCs) and tritium/helium-3 ($^{3}H/^{3}He$) that are based only in saturated zone. The magnitude of ^{3}H concentrations in precipitation in the Southern Hemisphere is lower than in the Northern Hemisphere and more difficult to distinguish. In the Southern Hemisphere, the used of ^{3}H to date groundwater is replaced by the used of CFCs and $^{3}H/^{3}He$. Recharge rates can be determined by estimating groundwater age. Groundwater age is defined as the time since water entered the groundwater zone. Figure 3.10 shows groundwater ages that is readily estimated from CFCs by comparing CFCs concentration in groundwater with those in precipitation. The age of groundwater is calculated from $^{3}H/^{3}He$ data using the following equation (Scanlon et al., 2002).

$$t_{gw} = -\frac{1}{\lambda}\ln\left[1+\frac{^{3}He_{trit}}{^{3}H}\right] \tag{3.18}$$

where

t_{gw} is the age of the groundwater,
λ is the decay constant, and
$^{3}He_{trit}$ is tritiogenic ^{3}He.

Groundwater dating approach begins with dating groundwater at several points in a vertical profile and then inverting the age gradient to groundwater velocity. In unconfined aquifer, the vertical groundwater velocity decreases with depth and groundwater ages increase with depth. The recharge rate can be calculated by the groundwater velocity multiplied by the porosity of aquifer for the same depth interval. In a confined aquifer, the data of horizontal flow velocities that estimate from ^{14}C or ^{36}Cl can be used to estimate recharge rates (R) (Scanlon et al., 2002):

$$R = \frac{vnA}{S} \tag{3.19}$$

where

v is velocity,
n is porosity,
A is the cross section area, and
S is the surface area of the recharge zone.

The range of recharge rate depends on the ranges of groundwater ages that are determined by the tracer. CFCs and $^{3}H/^{3}He$ can be used to estimate groundwater ages up to 50 years while the radioactive decay of ^{14}C can date the groundwater ages of 200 to 20,000 years. There are various issues that need to be considered when using the tracers to date groundwater. CFCs can not used in contaminated area that affected by contamination, the effect of excess air for both CFCs and $^{3}H/^{3}He$, the effect of recharge temperature, sorption and degradation on CFCs. The sampling of CFCs and $^{3}H/^{3}He$ is complex and high cost to analysis.

(ii) The chloride mass balance (CMB)

The CMB approach can be used in unsaturated and saturated zone. Recharge estimated using Cl in groundwater is higher than in soil water because the chloride concentration in saturated zone is smaller. Chloride concentrations generally increase through the root zone as a result of evapotranspiration and then constant below this depth. The concentrate is diluted when the mass of chloride reaches groundwater. The CMB approach integrates recharge over areas upgradient from the estimated point. Scanlon et al. (2002) describes the range of the scale that is about 200 m. to several kilometers. The time scales range from years to thousands of years.

5) Numerical modeling

Initial groundwater recharge estimation was made by flow net analysis; however, numerical models or groundwater flow models have replaced the graphical analysis. Numerical models start with the basic equation of groundwater flow. This is solved for the head distribution in the aquifer. Various parameters that used for groundwater model calibration or inversion such as hydraulic heads and hydraulic conductivity are of importances. They are used to predict recharge rates. Because recharge and hydraulic conductivity are often highly correlated, the reliability of the recharge estimates depends on the accuracy of the hydraulic-conductivity data.

3.1.4 General considerations to choosing methods for recharge estimation

Choosing an appropriate technique to estimate groundwater recharge for a particular site is often difficult. General considerations to choosing a technique for recharge estimation are (1) climate of the study area, (2) space and time scale, (3) range of recharge rate, (4) reliability and (5) cost. These factors are based on different methods. The different methods may also be combined. The use of many different methods is recommended to constrain the recharge estimates because each approach has uncertainties (Scanlon et al, 2002).

(1) Influence of climatic region to the choice of method used

Regions with arid and humid climates have differences in recharge. Different climatic regions may require different approaches. Appropriate methods for recharge estimation in regions with arid, semiarid, and humid climates are shown in Table 3.3. Methods based on surface water and saturated zone are widely used in humid regions, whereas methods based on unsaturated zone are widely used in arid and semiarid regions. Some methods can be used to estimate recharge in various climatic regions, but their accuracy is difference. For example, watershed modeling approach can be used both in arid and humid regions; however, the modeling approaches may be more accurate in humid regions because of there are perennial streams that the stream flow can be used for model calibration. Some methods can be used in various hydrologic zone over the same climatic region; however, only one zone that more suitable. For example, historical tracer (CFCs, $^3H/^3He$) is generally used in the saturated zone in humid regions. These tracers can be used in the unsaturated zone in humid regions too; however, due to generally thin unsaturated zones in the regions, their technique used is limited. Some methods are used only in humid regions such as WTF and Darcy's law approaches. These approaches could also be used in arid and semiarid regions where water table is shallow.

Table 3.3 Appropriate methods for recharge estimation in regions with Arid-semiarid and humid climates (Scanlon et al., 2002).

Hydrologic zone	Method	
	Arid and semiarid climates	Humid climate
Surface water	Channel water budget Seepage meters Heat tracers Isotopic tracers Watershed modeling	Channel water budget Seepage meters Baseflow discharge Isotopic tracers Watershed modeling
Unsaturated zone	Lysimeters Zero-flux plane Darcy's law Tracers [historical (^{36}Cl, ^3H), environmental (Cl)] Numerical modeling	Lysimeters Zero-flux plane Darcy's law Tracers (applied) Numerical modeling
Saturated zone	- - Tracers [historical (CFCs, $^3H/^3He$), environmental (Cl, ^{14}C)] Numerical modeling	Water table fluctuation Darcy's law Tracers [historical (CFCs, $^3H/^3He$)] Numerical modeling

(2) Influence of space and time scale of the various methods to the choice of method used

Methods based on surface water and saturated studies provide regional recharge estimation, whereas unsaturated zone methods generally provide estimates at small scale. Surface water methods provide recharge estimates on the event times or longer by summation of individual events, whereas unsaturated zone and saturated zone tracer provide recharge estimates over long period (≤ thousands of years). Long-term recharge estimates provided by tracer approaches might be advantage for the studies related to radioactive waste disposal. However, they are not good for groundwater resource evaluation because they do not provide detailed information on variations in recharge rate.

(3) Influence of range of recharge estimates to the choice of method used

The various methods for estimating recharge differ in range of recharge rate (Figure 3.12). Numerical modeling approaches can be used to estimate recharge rate in any range; however, uncertainties of the model parameters affect to the reliability. Environment tracer-Cl can estimate very low recharge rates within accuracy. Some tracer approaches require a minimum recharge rate for their estimate because recharge help to transport or confine the tracers. Historical tracers-$^{3}H/^{3}He$ in the groundwater zone require a minimum recharge rate of about 30 mm/y to transport and confine the ^{3}He (Scanlon et al., 2002).

(4) Influence of reliability to choice of method used

The reliability of the technique used depends on the accuracy of technique and researcher's operation. The reliability and accuracy of recharge estimates is also quite variable. Groundwater zone methods generally provide recharge estimates that are more reliable because they estimate actual recharge, water has reached the water table, whereas surface water and unsaturated zone methods usually provide estimate of potential recharge. Methods that require hydraulic conductivity data, such as Darcy's law approach and modeling of unsaturated and groundwater zone, are generally high uncertainties because hydraulic conductivity can vary over several orders of magnitude. Water budget methods have errors from their terms that accumulate in the recharge rate. Using small time step can be minimized these errors. Uncertainties of tracer approaches are assumption of tracer transport processes, measurement of tracer concentration, and estimated inputs of tracers; however, tracer approaches may be more accurate than other approaches. Uncertainties associated with researcher are skill of data measurement and data applicable at scale of field or laboratory to the scale of recharge calculation.

(5) Influence of cost to choice of method used

Cost of the various methods including time and expense need to consideration. Tracer approaches are considered to be expensive because the cost of sampling, chemical analysis and isotopic tracers. However, they represent long time periods in only one-time sampling. The costs may be less than the approaches based on long-term monitoring such as water budget approaches. These require monitoring equipment, continual collection and data analysis.

3.2 Methods used to estimate groundwater recharge of Chiang Mai basin in the previous study

Study on groundwater recharge estimates of Chiang Mai basin is very scarce. Most of the methods are based on data from surface and groundwater zones e.g. meteorological data, stream stage and groundwater level.

Ramingwong (1976) assessed the groundwater recharge pattern of Chiang Mai basin using average monthly rainfall and estimated evapotranspiration data. The study area is about 3,000 sq. km. and covered the period of 1969-1975. Average annual rainfall is about 1,300 mm. Average annual evapotranspiration, estimated by

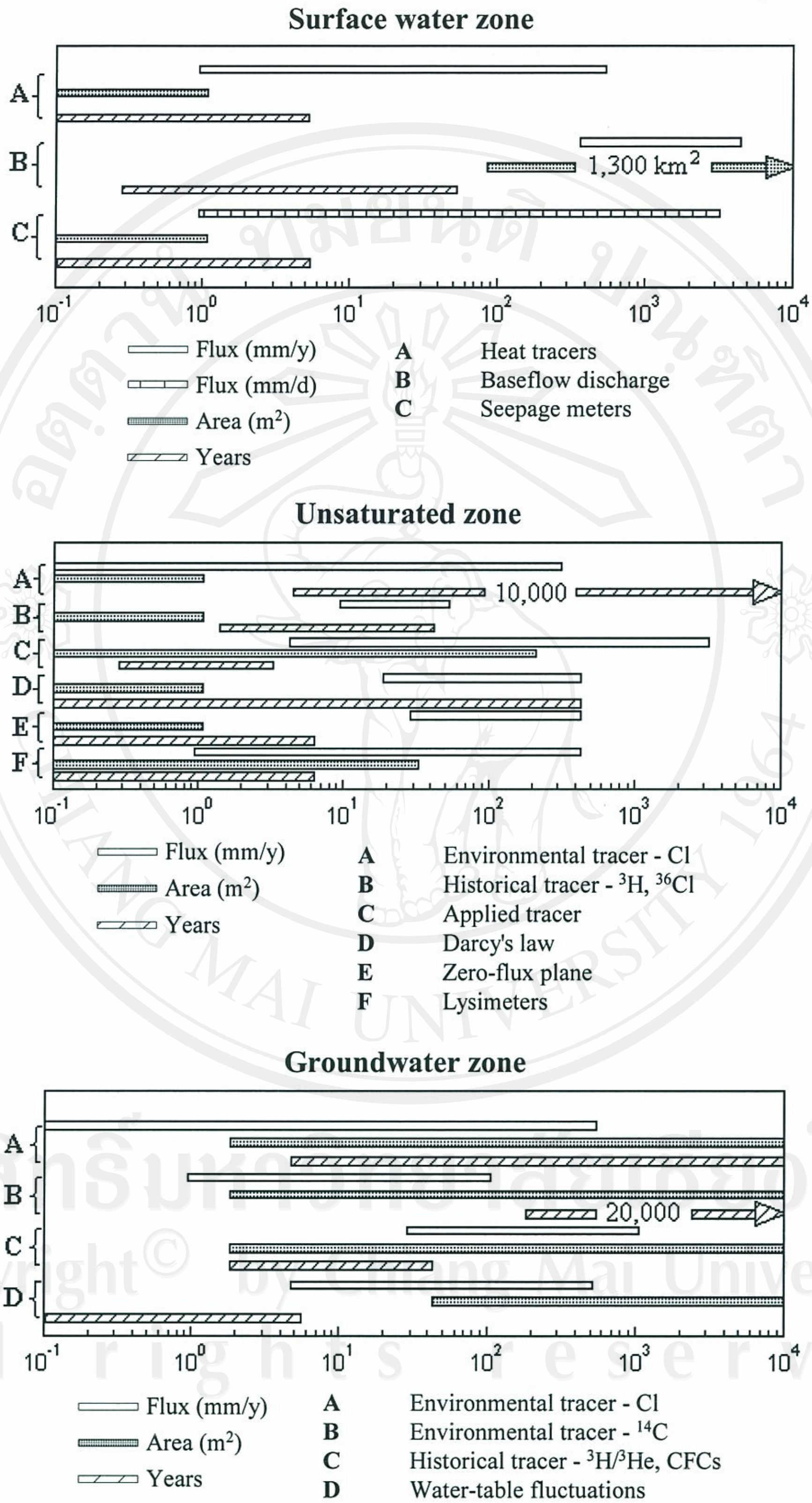

Figure 3.12 Comparison of range of recharge rates and spatial and temporal scales of the various methods (modified from Scanlon et al., 2002).

Penman's method is about 1,200 mm. This inventory period, groundwater is recharged mainly during June to September for each year. Approximately 63-65 percent of annual rainfall occurred during this period.

Suvagondha (1979) studied the groundwater recharge pattern and estimated groundwater recharge of the eastern part of Chiang Mai basin (the area of Amphoe Muang, Lamphun). The study area is approximately 200 sq. km. and covered the period of 1969-1976. Groundwater recharge pattern was controlled by effective rainfall (eventually groundwater recharge) from July to October. Average annual effective rainfall is 390.30 mm. However, not all of the effective rainfall will infiltrate downward because part of it will have to form runoff. From the stream hydrograph analysis, it is suggested that 30 percent of the effective rainfall form surface runoff, thus 70 percent is left over for groundwater recharge (273.20 mm/y).

Suvagondha and Jitapunkul (1982) estimated groundwater recharge of the eastern part of Ching Mai basin by the chloride mass balance approach. Net recharge can be calculated by substitute the 1,300 mm. average annual rainfall (1969-1976), chloride concentration of groundwater 2.5-4.0 ppm, and chloride concentration of rainwater 0.36 ppm into the equation of chloride mass balance. The minimum and maximum net recharge to the aquifer is 0.9 and 14.4 percent of rainfall. These values are equivalent to 11.7 and 187.2 mm/y.

Intrasuta (1983) estimated the recharge rates in unconsolidated sediments of Chiang Mai basin using flow net analysis combined with Darcy's law. Piezometric water level is observed at different period of 1981-1982. Rate of groundwater flow depends upon three components i.e. transmissibility, hydraulic gradient, and distance perpendicular to the flow path. Flow rate per area is recharge rate. The minimum recharge rate at 17 mm/y and the maximum at 143 mm/y are calculated. The groundwater in this region derived mainly from percolation of the rain, with the calculated rate about 10 % of the total precipitation. The annual total precipitation is approximately 1,200 mm/y.

Wongpornchai (1990) calculated groundwater recharge for the western part of the Chiang Mai basin (480 sq. km.) from streamflow measurements combined with meteorological data (the same approach of Suvagondha, 1979). During 1958-1984, effective rainfall occurred during 2 months periods from August to September with average annual effective rainfall value 91.14 mm. From hydrograph analysis, it can be observed that 71 % of effective rainfall form infiltration to water table. So, the groundwater recharge is about 65 mm/y.

Tatong (2000) estimated the recharge rates of Chiang Mai basin using mathematical modeling. The Modflow program is selected as a code for the model. The direct recharge estimated from water balance and one dimensional model is 25-293 mm/y. The groundwater is replenished in the rainy season between July and October.

Summary of the estimated groundwater recharge of Chiang Mai basin by different authors is presented in Table 3.4. There is wide range of recharge rate (11.7-293 mm/y). It is difficult to decide what approach is more accurate because each approach has uncertainties. Although the mathematical modeling approach, seems to be more reliable using computer program, the reliability of the recharge estimates should be evaluated in terms of the uncertainties in the model parameters.

Table 3.4 Summary of the estimated recharge of Chiang Mai basin in the previous study.

Authors	Study area	Methods	Months of recharge	Groundwater recharge (mm/y)
Suvagondha, 1979	Eastern part of Chiang Mai basin	Stream hydrograph analysis combined with meteorological data	Jul-Oct	273.2
Suvagondha and Jitapunkul, 1982	Eastern part of Chiang Mai basin	Chloride mass balance	Jul-Oct	11.7 – 187.2
Intrasuta, 1983	Chiang Mai basin	Flow net analysis and Darcy's law	No study	17 – 143
Wongpornchai, 1990	Western part of Chiang Mai basin	Stream hydrograph analysis combined with meteorological data	Aug-Sep	65
Tatong, 2000	Chiang Mai basin	Mathematical modeling	Jul-Oct	25 – 293

3.3 Methods used to estimate groundwater recharge of Chiang Mai basin in the present study

In the present study, the water table fluctuation (WTF) method has been adopted for estimating groundwater recharge because of (1) the method provides actual recharge (2) those parameters required for recharge calculation are existing data and (3) its simplicity and easy of use. Parameters required to recharge calculation are change in groundwater level and specific yield (Equation 3.16). These parameters can be analyzed from groundwater monitoring level and meteorological existing data from database of several Thai departments (Department of Groundwater Resources, Thai Meteorological Department and Royal Irrigation Department).

The WTF method is based on the premise that rises in groundwater levels are due to recharge water arriving at the water table. No assumptions are made on the mechanisms by which water travels through the unsaturated zone. For the WTF method to produce a value for total or net recharge requires application of Equation 3.16 for each water table rise. Equation 3.12 can also be applied over longer time intervals (seasonal or annual) to produce an estimate of change in subsurface storage.

Determination of water table height (Δh) can be done by well hydrograph analysis. It is measured by two different ways in order to compare the result of recharge estimations (Figure 3.13).

(a) The recession curve method The rise of water table (Δh_1) is set equal to the difference between the peak of the rise (point D) and low point (point B) of the extrapolated antecedent recession curve (AB or dashed line) at the time of the peak. Healy and Cook (2002) described the curve as the trace that the well hydrograph would have followed in the absence of the rise-producing precipitation.

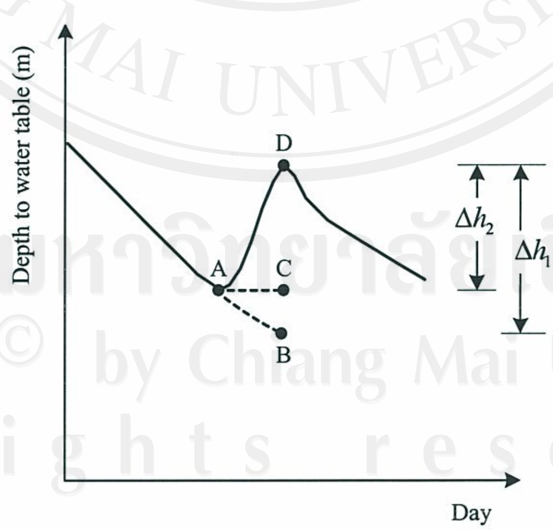

Figure 3.13 Hypothesis of water level rise in well in response to rainfall (modified from Healy and Cook, 2002). Δh_1 is difference between point D and B. Δh_2 is difference between point D and C.

(b) The horizontal line method The Δh_2 is measured by a horizontal line extrapolation from the lowest groundwater level. Figure 3.13, point A is the lowest water level at a given period. The horizontal line is extrapolated from point A. Point D is the peak (highest rising point) immediately followed the lowest groundwater level (point A). The Δh_2 can be measured directly from the graph (CD as shown).

Determination of specific yield can be done by laboratory and field approaches as described in Healy and Cook (2002). For the present study, two methods will be used (a) water budget approach and (b) aquifer test approach.

(a) Water budget approach The simple water budget methods for a basin as described by Walton (1970) can determine the specific yield. A simple water budget for a basin can be written as:

$$P_a = R_u + ET + U \pm \Delta S_s \pm \Delta S_g \tag{3.20}$$

where

P_a is precipitation,
R_u is stream flow,
ET is evapotranspiration,
U is subsurface underflow,
ΔS_s is change in soil moisture, and
ΔS_g is change in groundwater storage.

These quantities are volumes during the period for which the balance is made. During the months that evapotranspiration and soil moisture changes are very small, a reasonable estimate of evapotranspiration for this duration can be made. Soil moisture can be eliminated and evapotranspiration estimated without introducing serious error in the water budget. For Chiang Mai basin, the period low evapotranspiration and small soil moisture changes are during the rainy months i.e. July-September. Subsurface underflow is considered to be negligible in the present study. It is reasonable to assume that there is very small or no subsurface underflow that discharge out off the study area. Equation 3.20 may be rewritten as the simple water budget equation, and change in groundwater storage is then given by:

$$\Delta S_g = P_a - R_u - ET \tag{3.21}$$

The change in groundwater storage is the change in mean groundwater level during an inventory period multiplied by the specific yield of the deposits within the zone of groundwater fluctuation (Walton, 1970). Stated as an equation

$$\Delta S_g = \Delta H(S_y) \tag{3.22}$$

where

ΔH is change in groundwater level during an inventory period (mm), and
S_y is specific yield (dimensionless).

Equation 3.21 and 3.22 may be rewritten for an inventory period as:

$$S_y = \frac{P_a - R_u - E_T}{\Delta H} \tag{3.23}$$

The value determined for S_y from Equation 3.23 can then be used with the water level rises throughout the year to generate monthly and annual estimate of recharge.

(b) Aquifer test approach Values of S_y and transmissivity, T, for unconfined aquifers are commonly obtained from the analysis of aquifer tests conducted over a period of hours or days. Drawdown-versus-time data from observation wells are matched against theoretical type curves (Figure 3.14). Ideally, an unconfined aquifer is pumped from a well screened through the entire thickness of the saturated zone. Observation wells are located at various distances from the pumping well. These may fully or partially penetrated the aquifer. When pumping begins, declining water levels (drawdown) are recorded over time at each observation well. Early time recording is especially important. Drawdown versus time is plotted on a log-log scale, and type curves at the same scale are then overlaid on the plot. The curves are shifted (keeping coordinate axes parallel) until most of the data points lie on a curve. An arbitrary match point is then selected and the values of drawdown and time at that point are used to calculate T and S_y (Figure 3.14). In the present study, however, there are no observation wells, all pumping test data are from measurement in the pumped well itself. Radius of the pumped well is used as distance from the pumped well to the observation well.

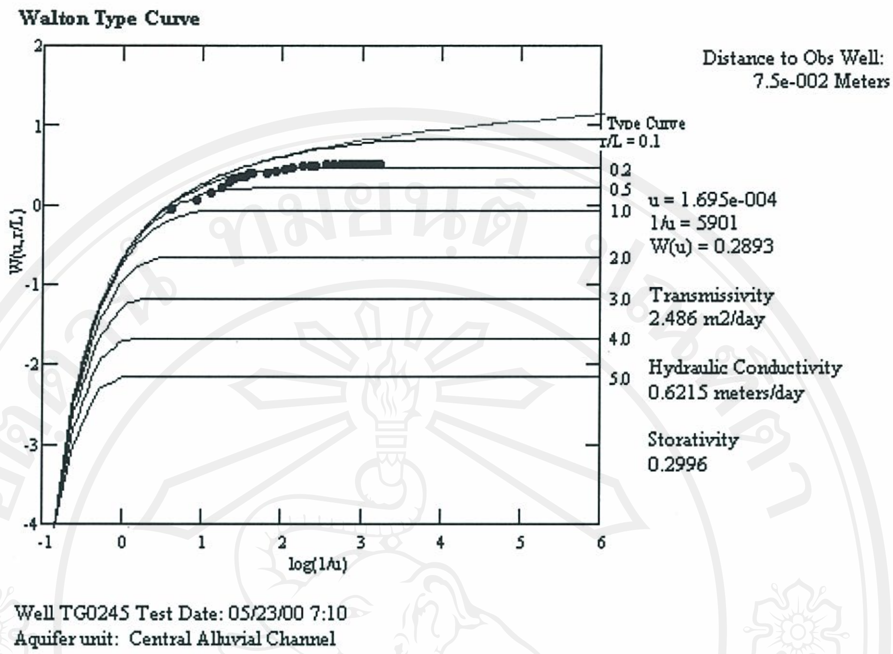

Figure 3.14 Time-drawdown data analysis using Walton graphical method (Starpoint Software, 2000).