LITERATURE REVIEWS

1. Evapotranspiration

1.1 Definitions

Evapotranspiration (ET) is the sum of evaporation and plant transpiration. Evaporation accounts for the movement of water to the air from sources such as the soil, canopy interception, and waterbodies. Transpiration accounts for the movement of water within a plant and the subsequent loss of water as vapour through stomata in its leaves. Evapotranspiration plays an important role in the water cycle. Evapotranspiration is controlled by three conditions : firstly, the capacity of the air to take up water vapor. Secondary, the amount of energy available for use in the evaporation and transpiration processes. Thirdly, the degree of turbulence of the lower layers of the atmosphere, necessary to replace the saturated air layers near the earth's surface by unsaturated air from higher levels (McGregor and Nieuwolt, 1998).

1.1.1 Evaporation

Burman and Pochop (1994) defined that evaporation (E) as the process whereby a liquid is transformed entirely into a gas. In this volume water is liquid, and water vapor is the gas. Andy and William (1995) stated that evaporation occurs when water is changed from a liquid to a vapor. Increase in air and temperature, wind and solar radiation on increase evaporation rates while a high water vapor percentage in the air (high relative humidity) decreases the potential for evaporation. Through the process of evaporation, water moves back to the atmosphere in the term of vapor.

1.1.2 Transpiration

Transpiration (T) is the process where liquid water within a plant is transformed into water vapor in the air surrounding the plant (Burman and Pochop, 1994). Andy and William (1995) described that water can take several paths after it enters the soil. Some water becomes part of the soil storage. While in storage near the surface, some of this water is used by plants and is eventually returned to atmosphere as water vapor. The process by which plants release water vapor to the atmosphere is called transpiration. This water vapor is a natural by-product of photosynthesis.

1.1.3 Potential Evapotranspiration

Potential evapotranspiration (PET) is a representation of the environmental demand for evapotranspiration. It is a reflection of the energy available to evaporate water, and of the wind available to transport the water vapour from the ground up into the lower atmosphere. Evapotranspiration is said to equal potential evapotranspiration when there is ample water.

Potential evapotranspiration (ET_p or PET) is the rate at which water removed from the soil surface of profile if available. ET_p has been used in a variety of ways. It always has referred to plants adequately supplied with water and usually not limited by disease or fertility (Burman and Pochop, 1994). Many authors treat potential evapotranspiration and potential evaporation as synonymous but the original intent was the potential evapotranspiration involved and actively growing crop and potential evaporation did not (Sue, 1995).

1.1.4 Actual Evapotranspiration

Actual evapotranspiration (ET_a or AET) or actual evaporation (E) is the amount or rate of ET occurring from a place of interest and it is the value we want to estimate. In practice, ET_a is obtained by first calculating ET_p and then multiplying by suitable crop coefficients to estimate the actual crop evapotranspiration. The methods for determining ET_a are variable and confusing. Scientists have attempted to remedy this problem by introducing reference crop evapotranspiration (Sue, 1995).

1.1.5 Reference Crop Evapotranspiration

Reference crop evapotranspiration (ET_r) is the rate at which water, if available, would remove from the soil and plant surface of a specific crop, arbitrarily called a ET_r. Typical reference crops are grasses and alfalfa. The crop is assumed to be well-watered with a full canopy cover (foliage completely shading the ground). The major advantage of relating ET_r to a specific crop is that it is easier to select consistent crop coefficients and to calibrate reference equations in new areas. (Sue, 1995)

1.2 Measurement of Evapotranspiration

Singh (1989) explained that various methods were used for measuring ET such as tanks and lysimeters, field plots and inflow-outflow measurement, Each of them has the following detail :

1.2.1 Tanks and Lysimeters

Lysimeters are tanks filled with soil in which crops are grown under natural conditions and the amount of water lost by evaporation and transpiration is measured. Lysimeters and tanks are equipped to record precipitation, runoff, gravity flow, and changes in soil moisture and provide information from which ET can be directly evaluated. This is the only method providing direct measurement of ET and is frequently used to investigate climatic effects and evaluate other methods of ET estimation. However, there exists some controversy over the differences between the lysimeter and natural conditions in soil profile, soil moisture regime, plant root characteristics, methods of water application, and net energy exchange, Needless to say, lysimeters provide reasonably reliable measurements of ET for short time periods if their installations satisfy certain minimum standards. The water balance for a lysimeter is :

$$ET = PT + \Delta SM + \text{water added} - \text{water removed} \qquad (1)$$

Where ET = evapotranspiration

PT = total precipitation

ΔSM = change in soil moisture

1.2.2 Field Plots

Field plots must be selected in places where the water table is deep so that plants do not extract water therefrom. Crops are grown under natural conditions. An inventory of water added by precipitation and irrigation, surface runoff, and changes in soil moisture is prepared. Since water is added to plots in small quantities, deep percolation is minimized and not measured. ET is the residual in the water balance. The primary advantage of using field plots is the ET is measured under field conditions.

1.2.3 Inflow-Outflow Measurements

Evapotranspiration can be determined on large watersheds under natural conditions by observing differences between inflow and outflow and adjusting for changes in ground water storage. These differences are considered as ET values. This method provides only gross estimates of ET for large time intervals such as a season or a year and should not be used for short-term rates within the season. A mass balance equation is used on a watershed, field or other area.

$$ET = INFLOW - OUTFLOW + P_e + \Delta SM + WT - DR \qquad (2)$$

Where INFLOW = measured surface inflow to a regime

OUTFLOW = measured outflow from an regime

WT = water contributed from a water table

DR = water loss by deep percolation

P_e = effective precipitation

1.2.4 Soil Moisture Depletion

Evapotranspiration can be established under natural conditions by observing changes in soil moisture over a period of time. The soil is usually sampled at several representative sites, where the water table is deeper than the plant root zone, during the periods of light rainfall to minimize drainage. The same sites are measured each time to minimize error due to soil variability. The major source of error with this method is drainage from the zone sampled or upward movement from a saturated zone.

1.3 Methods for Estimating Evapotranspiration

Singh (1989), Chow *et al*.(1988) and Burman and Pochop (1994) are drawn the several methods are available for estimating evapotranspiration which can be summarized as following :

1.3.1 Temperature Methods

Temperature methods are some of the earliest methods of estimating ET. Three methods are presented here.

1.3.2 Blaney-Criddle method

Singh (1989) described that Blaney-Criddle method was an empirical relation between ET, mean air temperature, and mean percentage of daytime hours. This is well known in the western United State of America and has been used extensively elsewhere also. The underlying assumption is that the heating of air and evaporation share the heat budget in a fixed proportion. Consequently, ET varies directly with the sum of the products of mean monthly air temperature and monthly percentage of daytime hours with an actively growing crop with sufficient soil moisture.

$$ET = KF = \sum kf, \text{and } f = Tp/100 \qquad (3)$$

Where ET = evapotranspiration or consumptive use in inches of water during the growing season,

K = an empirically derived seasonal consumptive coefficient applicable to a particular crop

F = the sum of monthly consumptive use factors

T = mean monthly air temperature in degrees Fahrenheit

p = mean monthly percentage of annual daytime hours

k = monthly consumptive use coefficient

Equation (3) is for seasonal estimates but can be written for monthly estimated of ET in inches as :

$$ET = kTp/100 \qquad (4)$$

Both meteorological and crop effects are included in this equation. Several workers have attempted to improve short-period estimates of ET by expressing k as a function of T and a daily percent of annual daytime hours. For Ladino clover, which is similar to alfalfa, it's used k = .04 + 5.52 (Tp/100) for conditions near Prosser, Washington.

1.3.3 Thornthwaite method

Singh (1989) described that Thormthwaite method was derived as equation under unlimited water conditions for monthly estimates for ET for east-central USA, wherein assumptions are similar to those of the Blaney-Criddle method: in which ET is in centimeters, T is mean monthly temperature in degrees celsius, and a is an exponent. The term a can be estimated as :

$$a = 67.5\text{x}10^{-8} I^{3} - 77.1\text{x}10^{-6} I^{2} + 0.0179 I + 0.492 \qquad (5)$$

Where I is a heat index expressed as:

$$I = \Sigma(T/10)^{1.51} \qquad (6)$$

I is an integral element of Thornthwaite classification of climates, so

$$ET = 1.62(10T/I)^{a} \qquad (7)$$

$$\ln ET = \ln 1.62 + a(\ln 10 + \ln T - \ln I) \qquad (8)$$

Obviously, ET is equal to 1.62 when I equal to 10T or ln 10T equal to ln I. It has been shown by Thornthwaite that all lines obeying this equation have a common point of convergence at T equal to 26.5 ^{o}C and E equal to 1.35 cm. The Thornthwaite method is one of popular methods which have prepared tables for use of this method.

1.3.4 Lowry-Johnson method

Singh (1989) described that Lowry-Johnson method was an empirical linear relation between ET and effective heat as:

$$ET = 0.8+0.156*I_e \qquad (9)$$

In which ET is annual in acre-feet per acre and I is effective heat in thousands of day-degrees. Effective heat is defined as the accumulation in day-degrees of maximum daily growing season temperature above 32 ^{o}F. It has been used successfully by the U.S. Bureau of Reclamation in the western United States and applies to a valley, not an individual farm. Equation (34) is a special case of the general form indicated by Tanner (1967) (Singh, 1989).

$$ET = c_1 d_L T(c_2-c_3 h) \qquad (10)$$

Where c_1, c_2 and c_3 are constants, d_L is day-length, T is mean air temperature, and h is a humidity term.

1.3.5 Humidity Methods

Singh (1989) described that some of the oldest equations for estimating ET are based on saturation deficit. As early equation proposed that :

$$ET = 0.5(e^{o}_{z}-e_z) \qquad (11)$$

Where ET = evapotranspiration (mm day^{-1})

e^{o}_{z} = saturation pressure (mb) at mean air temperature

e_z = saturation vapor pressure (mb)

Beside, Singh (1989) described that Papadakis method is one of humidity methods which can be presented as :

$$ET = 0.5625(e^{o}_{max} - e_{z}) \qquad (12)$$

Where ET = monthly evapotranspiration (cm)

e^{o}_{max} = saturation vapor pressure (mb) corresponding to average daily maximum temperature

e_z = average vapor pressure for the month

Similar equations have been developed as :

$$ET = Be_{d} \qquad (13)$$

$$ET = cd_{L}(q_{max} - q_{min}) \qquad (14)$$

Where ET = evapotranspiration (mm day^{-1})

e_d = average daily vapor pressure deficit (mb)

B = hydrometric coefficient (B = 0.56 for clover)

q_{max}, q_{min} = saturation absolute humidities corresponding to maximum and minimum air temperatures

d_L = fraction of annual daylight hours

c = 1 when ET is in millimeters per month.

1.3.6 Regression Methods

Singh (1989) described that Turc method was used for analyze data collected from 254 watersheds located in all parts of the world and related evaporation to rainfall and temperature as :

$$E = P/[0.9+(P/I_T)^2]^{0.5} \tag{15}$$

Where E = annual evaporation or ET (mm)

P = annual precipitation (mm)

$I_T = 300+25T+0.5T^3$

T = mean temperature (^oC)

The other equation incorporating soil moisture variability on ET as :

$$E = (P+E_{10}+K)/[1+\{(P+E)/I_T+(K/2I_T)\}^{0.5}] \tag{16}$$

Where E = evaporation or ET in a 10-day period (mm)

P = precipitation in a 10-day period (mm)

E_{10} = estimated evaporation (in a 10-day period) from bare soil assuming no precipitation and is not greater than 10 mm, and

$K = 25(cM\ G^{-1})^{0.5}$

In equation (16), 100M is final yield of dry matter in kilograms per hactare, 10G is the length of the growing season in days, and c is crop coefficient. I_T is evaporative capacity of the air,

$$I_T = (T+2)R_s^{0.5}/16 \tag{17}$$

Where T = mean air temperature over the 10-day period (^oC)

R_s = incoming radian energy (cal cm^{-2} day^{-1})

When soil moisture is not limiting and $I_T>10$, so that the equation expressed as:

$$E = (P+E_{10}+70)/[1+\{(P+E)/I_T+(70/2I_T)\}^{0.5}] \tag{18}$$

Where E = evaporation or ET in a 10-day period (mm)

1.3.7 Radiation Methods

Several methods of ET estimation depend largely on solar radiation. Three of the more frequently used methods are presented here.

1.3.7.1 Makkink method

Makkink method was used in estimating ET in millimeters per day over 10-day periods for grass under the cool climatic conditions of the Netherlands (Singh, 1989). The equation is :

$$ET = 0.61(\Delta/\Delta+\gamma)(R_s/58.5)-0.12 \quad (19)$$

Where net radiation is R_n, equal to $0.6*R_s$. This equation was applied to determine potential and agricultural evapotranspiration from agricultural crops in Denmark.

1.3.7.2 Turc method

Turc method was used in computing ET in millimeters per day for 10-day periods under general climatic conditions of western Europe (Singh, 1989). The eqution is :

$$ET = 0.013[T/(T+15)](R_s+50) \quad (20)$$

Where ET = evaporation or ET (mm day^{-1} in a 10-day period) and for relative humidity > 50 percent, and

$$ET = 0.013[T/(T+15)](R_s+50)[1+(50-h_r)/70] \quad (21)$$

For relative humidity < 50 percent, where T is the average temperature in degrees celcuis and R_s is in langleys per day.

1.3.7.3 Jensen-Haise method

Singh (1989) described that Jensen-Haise method was expressed as :

$$ET = C_T(T-T_x)R_s \quad (22)$$

Where C_T = temperature constant = 0.014,

T_x= intercept of the temperature axis = 26.4 if T in ^{o}F

C_T = 0.025 and T_x = -3 if T in ^{o}C

These coefficients are constant for an area. So, it was defined for C_T as :

$$C_T = 1/(C_1+C_2C_H),\ C_H = 50mb/(e_2-e_1) \quad (23)$$

Where e_2, e_1 = saturation vapor pressure at the mean maximum and mean minimum temperatures for the warmest month of the year respectively

C_2 = 13 ^{o}F, or 7.6 ^{o}C

C_1 = 68 ^{o}F-(3.6 ^{o}F x elevation in ft/1000 ft)

T_x = 27.5 ^{o}F-0.25(e_2-e_1) ^{o}F/mb - (elevation in ft/1000) (^{o}F)

For temperature in degrees celsius,

$C_1 = 38 - (2\ ^{o}C \text{ x elevation in m}/305)$

$T_x = -2.5 - 0.14(e_2 - e_1)\ ^{o}C/mb$ – (elevation in m/550).

1.3.8 Penman method

Burman and Pochop (1994) stated that Penman method is a result of a combination of a theoretical energy balance with an empirical wind function. Penman method is known as combination methods. Singh (1989) described that the generalized Penman method may be shown as follow :

$$ET = (\Delta/\Delta+\gamma)[-(R_n+G_s)-(\rho C_p k/\Delta)\{(e_a^{o}-e_a)-(e_0^{o}-e_0)] \quad (24)$$

Where ET = Potential evapotranspiration (mm day^{-1})

R_n = net radiation (MJ d^{-1} m^{-2})

G_s = soil heat flux (MJ d^{-1} m^{-2})

Δ = slope of saturated vapor pressure is temperature relationship (kPa $^{o}C^{-1}$)

$= (e_a^{o}-e_a)/(T_a-T_0)$

γ = psychometer coefficient (kPa $^{o}C^{-1}$)

$e_a^{o}-e_a$ = vapor pressure deficits (kPa)

$e_0^{o}-e_0$ = water-vapor saturation deficit (kPa)

a, o = at height a in the air and at the evaporating surface

For a free water surface, $e_0^{o}-e_0$ equal to zero, Therefore,

$$ET = (\Delta/\Delta+\gamma)[-(R_n+G_s)-(\rho C_p k/\Delta)(e_a^{o}-e_a)] \quad (25)$$

In practice, T_0 is seldom know and is replaced by T_a for computing Δ

1.3.9 Modified Penman method

Singh (1989) described that the Penman's equation has given a good account of the wind function for well-watered short grass, so, the modified Penman method was presented as :

$$ET = [(\Delta/\Delta+\gamma)(Rn+G)]+[15.36(\Delta/\Delta+\gamma)(1.0+0.0062U_2)(e_a^o-e_a)] \quad (26)$$

Where U_2 = wind speed at 2 m height (km day^{-1}) and e in mb.

Wind speed at 2 m height can be approximated from measurements at another height a using power log law,

$$U_2 = U_a(2/a)^{0.2} \quad (27)$$

1.3.10 Penman-Monteith method

Burman and Pochop (1994) described that the Penman-Monteith method is the Penman method involves the use of a plant resistance parameter and a more general use of an aerodynamic resistance parameter. The original equation was intended for short term calculation such as for an hourly period. Special adaptations are required for daily period :

$$\lambda ET = \frac{\Delta(R_n+G_s) + \rho C_p Q}{\Delta + \gamma^o} \quad (28)$$

$$Q = (e^o_{Tz} - e_z)/r_a \quad (29)$$

$$\gamma^o = \gamma[1+(r_c/r_a)] \quad (30)$$

Where Δ = slope of vapor pressure and temperature relationship.

ρ = Atmospheric density

γ = Psychometric coefficient

e = vapor pressure

C_p = coefficient of specific heat at constant pressure (kJ kg^{-1} K^{-1})

r_a = aerodynamic resistance (sec m^{-1}) to diffusion of water vapor from the evaporating surface

z = elevation in measurement (m)

r_c = canopy resistance (sec m^{-1})

h = 1/ r_a = convertive heat transport coefficient (W m^{-2} K^{-1})

1.3.11 Pan evaporation

Evaporation from an open container such as a class A evaporation pan provides a simple method of estimating ET. Pan evaporation is a very common measurement and provides excellent supporting data useful for correlation and prediction. In addition, pan evaporating is commonly measured at reservoir and lakes sites to be used in estimating water surface evaporation losses (Burman and Pochop, 1994).

Burman and Pochop (1994) stated that the both pan evaporation and ET are involve the same basic process, it is easy to assume that a reasonable estimate of reference ET might be found by multiplying measured pan evaporation by a factor usually less than unity. The general relationship is :

$$ET_r = K_p E_{pan} \tag{31}$$

Where E_{pan} = measured pan evaporation in any desired units

K_p = pan coefficient

1.3.12 Energy Balance and Bowen Ratio

1.3.12.1 Energy balance

Geiger (1971) described all parameters in equation (57) that the radiation (S) is the major factor of heat exchange, heat arrives at the earth's surface from the sun, the sky, and the atmosphere (insolation). Heat is sent back into space (outgoing or terrestrial radiation). The add heat to the surface of the ground is considered positive, the heat from it is negative. The sum of insolation and outgoing radiation, that is the balance, decides in individual cases whether S is positive or negative (unit : cal cm^{-2} min^{-1}, also called langley per minute, abbreviated ly min^{-1}).

The second factor B is determined by the flow of heat from the ground to the surface or in the reverse direction during the winter night heat flow upward through the ground are B is therefore positive, on summer afternoon, B is negative because heat transported downward from the surface. The third factor L may be positive or negative. Transport of heat to or from the ground depends not only on physical heat conduction, but also on mass exchange (eddy diffusion) because of the great mobility of the air. Finally there is the effect of evaporation (V). This is measured, like all the other heat-economy factors (unit : cal cm^{-2} min^{-1}). The quantity of heat in calories required to evaporate 1 gm of water is called the latent heat vaporization (r_w) and varies with temperature. Normally V is negative, but positive values are possible, as when dew form the surface and heat of condensation or sublimation is released. Since the earth's surface as a boundary surface can absorb no heat, according to the law of conservation of energy the following equation must be satisfied for all units of time.

$$S + B + L + V = 0 \qquad (32)$$

This is the fundamental equation that governs heat exchange at the earth's surface. Over oceans, lakes, and rivers the factor W, for the exchange of heat between water and its surface, is used instead of B, Q is extra factor and N is precipitation. So,

$$S + B(\text{or } W) + L + V + Q + N = 0 \tag{33}$$

Billing (1972) presented that W can be discarded and Q will become 0 and made N equal to zero. The equation expresses the various components of energy as they stream to and from the surface of the earth in the boundary layer in which most terrestrial organisms spend their lines.

$$S + R + LE + G + C + s = 0 \tag{34}$$

Where S = solar radiation flux
R = thermal radiation flux from the ground
L = latent heat of evaporation
E = rate of evaporation
G = sensible heat flux by conduction from the ground
C = sensible heat flux by convection in the air
s = a storage term for storing energy or giving up energy over short period of time

McGregor and Nieuwolt (1998) described that the total radiation inputs and outputs from the earth-atmosphere system are referred to as radiation balance. For the condition that radiation inputs to the earth's surface or atmosphere are greater (less) than the outputs, then the net radiation balance is positive (negative). The balance of net radiation surplus at the earth's surface and the upward fluxes of sensible and latent heat are referred to as energy balance. The main components of the energy balance are the fluxes of the net radiation, sensible heat, latent heat and surface heat. The symbols Q, H, E and G respectively are often used to the present these. Sensible and latent heat are convective fluxes, surface heat is a conductive fluxes. The energy balance may be written as :

$$Q = H+E+G \qquad (35)$$

The seasonal variation and the magnitude of the various energy balance components varies in nature according to the climate type was shown in Figure 1 For humid equatorial climates where there is plentiful rainfall in all month (a), wet and dry monsoon climates (b), dry grassland and steppe climates (c) and tropical deserts (d).

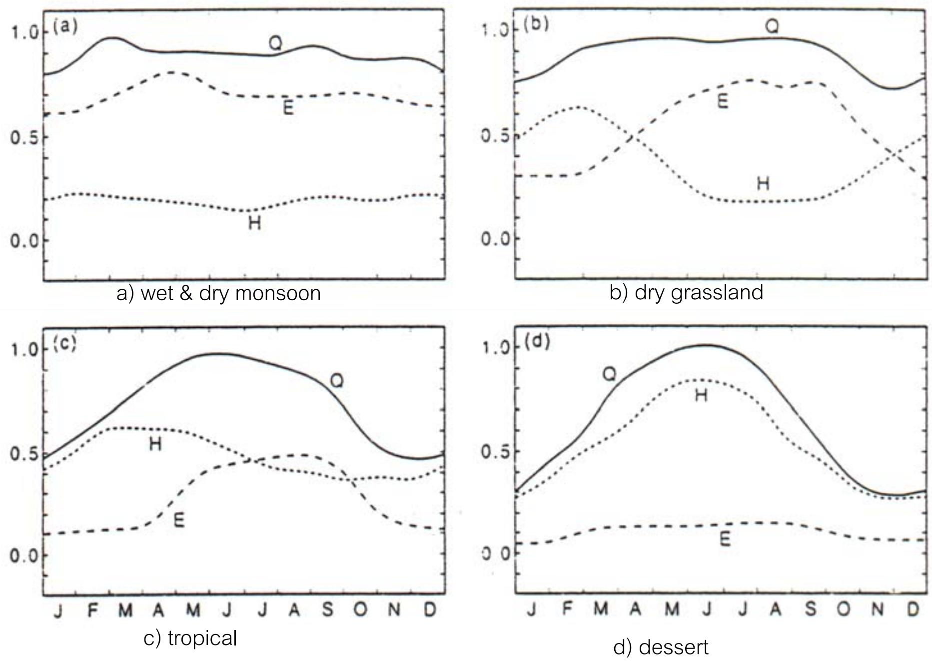

Figure 1 The energy balance for various low latitude climate types.

Source : McGregor and Nieuwolt (1998)

1.3.12.2 Bowen ratio

The simplified energy budget equation can be used to determine the sum of sensible and latent heat fluxes (H and LE) from measurement of net radiation and ground heat fluxes (R_n and G_s in the Bowen ratio energy balance method) (Tanner, 1988). McGregor and Nieuwolt (1998) stated that the ratio between sensible and latent is a useful tool for climate analysis. This is called the Bowen ratio and is expressed as :

$$\text{Bowen ratio } (\beta) = \text{sensible heat flux (H)/ latent heat flux (LE)} \quad (36)$$

From energy balance equation, the equation can be simplified as :

$$R_n - G_s - H - LE = 0 \quad (37)$$

So, the Bowen ratio and equation can be presented as :

$$\beta = H/LE = \frac{(\rho C_p P K_h \partial T/\partial z)}{(0.622 \rho L K_w \partial e/\partial z)} \quad (38)$$

Where ρ = the air density

C_p = specific heat of constant pressure

K_h, K_w = turbulent exchange coefficients for heat and water vapor transport

T, e = temperature and vapor pressure

P = air pressure

z = the height

Assuming that K_h equal to K_w and integrating over the same interval, so,

$$\beta = H/LE = \gamma(\Delta T/\Delta e),\ \gamma = (C_p P)/(0.622L) \quad (39)$$

Inserting equation (64) in equation (62), the evapotranspiration can be estimated by equations followed :

$$LE = (R_n-G_s)/[1+\gamma(\Delta T/\Delta e)] \tag{40}$$

$$ET = (R_n-G_s)/L(1+\beta) \tag{41}$$

1.3.12.3 Soil heat flux (G_s)

Soil heat flux (G_s) is small compared to R_s, R_n, ET and other term of the energy balance and therefore, is often neglected in making estimates of ET. Because of the relatively small magnitude of G_s, it's influence on final estimates is small. The principal reasons for including G_s and its inclusion recognizes a realistic phenomenon, and any simple model of G_s is easy to include in making computerized estimates of ET. Many differential equations defining soil heat flux. Because air temperature is a trigonometric function, the more theoretically based models involve periodic functions. Such as daily model and ten day to month model. Daily model is based on broad assumptions of the root zone depth, the soil heat capacity, and the notion that soil temperature lags air temperature.

$$G_s = 0.3768[T_D - (T_{D-1} + T_{D-2} + T_{D-3})/3] \tag{42}$$

Where G_s = daily soil heat flux (MJ $m^{-2}d^{-1}$)

D = calendar day number under consideration

T = daily average temperatures (^oC)

0.3768 = units of G_s and air temperatures (T)

The sign convection results in G_s being negative in the spring when the soil is warming. At that time G_s will slightly decrease estimated ET. The ten day to monthly model is a simplified model for G_s for periods of from 10 to 30 days. The formula is based on a soil with 50% solids, negligible organic material, and a volumetric soil water capacity of 27%. A 2 m effective soil depth is assumed.

$$G_s = 4.19[T_{J+1} - T_{J-1}]/\Delta time \qquad (43)$$

Where G_s = soil heat flux (MJ m^{-2} day^{-1})

T = average air temperature (^{o}C)

J = periods selected (days)

The rate at which temperature increases with depth in the ground is, on the average, 1 ^{o}C every 33 m (a unit of geothermal depth), the upward flow of heat can be only about 0.0001 cal m^{-2} min^{-1}, which can be neglected (Geiger, 1971).

Nakamura *et al*. (1996) found that the daytime soil heat flux can be estimated using the following equation which includes daytime net radiation (R_n) and vegetation coverage (VC).

$$G_s = (0.174-0.00086VC)*R_n \qquad (44)$$

The equation was obtained from the data regarding beets, potatoes and wheat. The standard error of estimation in this equation is 0.46 MJ m^{-2}.

1.3.12.4 Latent heat of vaporization (L)

Latent heat of vaporization (L) is the heat required to change a unit mass of water from a liquid to a water vapor in a constant pressure and constant temperature process. It may be estimated using the following linear regression equation.

$$L= 2501 - 2.3601*T \quad (45)$$

Where L = latent heat of vaporization (kJ kg^{-1})

T = average daily air temperature (^{o}C)

1.3.13 Aerodynamic Method

Chow *et al*. (1988) and Singh (1989) presented aerodynamic method which is the well-known Thornthwaite-Holzman aerodynamic equation. It was obtained by

assuming a flat uniform surface,. The equation can be expressed as :

$$ET = \frac{\rho k^2(u_2-u_1)(q_2-q_1)}{[\ln(z_2/z_1)]^2} \quad (46)$$

Where ρ = air density (g cm^{-3})

k_w, k_m = turbulent transfer coefficients or eddy diffusivities for water vapor and momentum respectively

u_2-u_1 = the wind shear between the same two heights

q_2-q_1 = the difference in specific humidities at heights z_2 and z_1

k = von Karman's constant (0.41)

Assuming adiabatic atmospheric condition and logarithmic distribution of wind speed and moisture in the vertical, the equation was derived as :

$$ET = \frac{0.623\rho k^2(u_8-u_2)(e_2-e_8)}{P[\ln(800/200)]^2} \quad (47)$$

Where u_8-u_2 = wind velocity (cm/s) at the heights of 8 and 2 m respectively

e_2-e_8 = vapor pressure (mb) at the heights of 2 and 8 m respectively

P = atmospheric pressure (mb)

1.3.14 Combination Method

Chow *et al*. (1988) and Singh (1989) stated that evaporation may be computed by the aerodynamic method when energy supply is not limiting and by energy balance method when vapor transport is not limiting. But normally, both of these factors are limiting, so a combination of the two methods is needed. If the two levels are taken at evaporative surface and in the overlying air stream, respectively, it can be shown that the evaporation rate computed from the rate of net radiation and the evaporation rate computed from aerodynamic methods can be combined to yield a weighted estimate of evaporation, by

$$ET = [E_r(\Delta/(\Delta+\gamma)]+[E_a(\gamma/(\Delta+\gamma)] \qquad (48)$$

Where Δ = the slope of the saturated vapor pressure curve at air temperature (kPa $^{o}C^{-1}$)

γ = the psychometric constant (kPa$^{o}C^{-1}$)

E_r = evaporation rate computed from the rate of net radiation

E_a = evaporation rate computed from aerodynamic

The combination method of calculating evaporation from meteorological data is the most accurate method when all the required data are available and the assumptions are satisfied. The chief assumption of the energy balance are that steady state energy flow prevails and that changes in heat storage over time in the water body are not significant. This assumption limits the application of the method to daily time intervals or larger, and to situations not involving large heat storage capacity, such as a large lake possesses. The chief assumption of the aerodynamic method is associated with the form of the vapor transfer coefficient. The combination is well suited for application to small areas with detailed climatologically data. The required data include net radiation, air temperature, humidity, wind speed and air pressure. When some of these data are unavailable, simpler evaporation equations requiring fewer variables must be used. For evaporation over very large areas, energy balance considerations largely govern the evaporation rate (Chow *et al*., 1988).

2. Remote sensing

2.1 Principle of Remote sensing

Remote sensing has been defined in many ways. It can be thought of as including traditional aerial photography, geophysical measurements such as surveys of the Earth's gravity and magnetic fields, and even seismic and sonar surveys. However, in a modern context, the term remote sensing usually implies digital measurements of electromagnetic energy often for wavelengths that are not visible to the human eye. A big advantage in remote sensing is that the final product is usually an image ("picture") of the Earth's surface which we can visualize and interpret as if it were as picture. Thus, many of the terms and concepts (e.g., brightness, contrast, color, intensity) are familiar, and we have a physical intuition for their meaning. Frequently, the ultimate goal of a remote sensing study is simply to be able to "see" some feature or changes in it well. Thus, the definition of a good result can be as subjective as deciding which of a series of photographs is best. However, one must be careful to remember that remote sensing images are more than just digital pictures, and we need to have a good understanding of their physical meaning (Chada, 2000).

Remote sensing techniques measure the interaction of the Earth's surface (or at most the upper few meters) and electromagnetic energy from the sun and therefore are inherently a form of geographic information. Thus, the use of geographic information systems (GIS) to store and display remote sensing information is so common that the terms remote sensing and GIS are almost synonymous. The use and generation of digital elevation models is an example of how these two fields are merging. When properly geographically referenced (ie., the location of each measurement is carefully determined), images ("pictures") created from remote sensing measurements become maps of the Earth's response to various wavelengths of electromagnetic energy (Chada, 2000)

2.2 Basic Principals of Electromagnetic Wave Propagation

Sharma *et al*. (1997) provided most of the key physical principals, we need to formulate a basic understanding of electromagnetic energy are familiar to us. In remote sensing, electromagnetic energy is classified by its wavelength. Visible light is the type of electromagnetic energy with which are most familiar, but there is much to be learned from waves which wavelengths are either longer or shorter that those of visible light. The basic theory needed to understand electromagnetic energy well enough to use remote sensing techniques intelligently is surprisingly simple. The mathematics needed is not difficult, and the physical principles are straightforward. The trick is to link your training in mathematics and physics with your practical knowledge of physical phenomena (like light, x-rays, radar, and radio waves) to develop an intuitive understanding of the propagation of electromagnetic waves through the atmosphere and their interaction with the Earth's surface. One advantage is that many terms used in remote sensing are familiar and have the same meaning as in every day life (i.e., bright, dark, high and low contrast, and intensity). In physics, one learns that light (and electromagnetic energy of similar wavelengths) can be thought of as either a propagating wave or a stream of particles. In remote sensing, we usually think in terms of waves, and sensors are designed to detect waves which wavelengths lie in specific bands (ranges of wavelengths).

2.3 Basic Concepts, Equations and Terms

There are a number of basic equations (and the terms involved in them) which form the basis for an understanding of electromagnetic waves and how to use them in practical applications (Arlene, 1999) .

2.3.1 Electromagnetic energy has been classified by wavelength and arranged to form the electromagnetic spectrum (Figure 2).

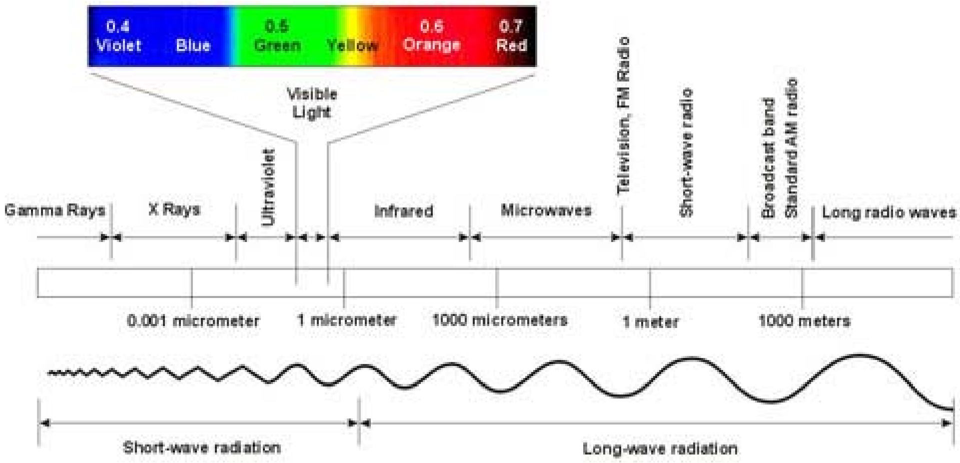

Figure 2 Spectrum of electromagnetic energy.

Source : http://www.crisp.nus.edu.sg/~research/tutorial/optical.html

Spanner *et al*. (1990) explanation this spectrum that visible light occupies only a very narrow band of wavelengths. Gamma rays, x-rays, and most ultra-violet energy do not penetrate the atmosphere so they are not used in remote sensing. However, infrared and microwave (radar) energy is measured regularly and is very useful in addition to energy at visible wavelengths. Thus, we must remember that almost images made from remote sensing data are false color in nature and are not pictures in the sense that humans would never see the Earth in this particular fashion. The images in fact detect objects and phenomena which could never be "seen" by the humen eye.

Primarily as the result of ozone (O3), CO2, and water vapor, the atmosphere absorbs energy in several discrete bands so sensors are designed to avoid these bands since little energy survives transmission through the atmosphere. Most remote sensing devices are passive in that they merely measure the intensity of the Sun's electromagnetic energy which is reflected off the surface of the Earth or is emitted (radiated) as heat. However, some systems (radar in particular) are active in that the energy that is measured is generated by the measuring system not the Sun.

2.3.2 As electromagnetic energy interacts with the atmosphere and the surface of the Earth, the most important concept to remember is the conservation of energy (i.e., the total energy is constant). (Chada, 2000)

Our main concern is the fate of the energy that makes it through the atmosphere to hit the surface of the Earth. The three main processes are reflection (scattering is a form of reflection), absorption (this energy is converted to heat and some is emitted with a change in wavelength and a time delay) , and transmission.

Each of these phenomena are a function of wavelength (*l*), and conservation of energy leads to the following equation:

$$Ei(l) = Er(l) + Ea(l) + Et(l),$$

where, $Ei(l)$ = the incident energy,

$Er(l)$ = the reflected energy ,

$Ea(l)$ = the absorbed energy, and

$Et(l)$ = the transmitted energy.

2.3.2.1 In the case of a lens, almost all of the energy is transmitted. However, the case of the Earth's surface most energy is reflected or absorbed. At a particular location, the ratio of the reflected energy to the incident energy as a function of wavelength (expressed as a %) is called the **reflectance spectra**, and is expressed as: *R(l) = Er(l) / Ei(l)* (Figure 3)

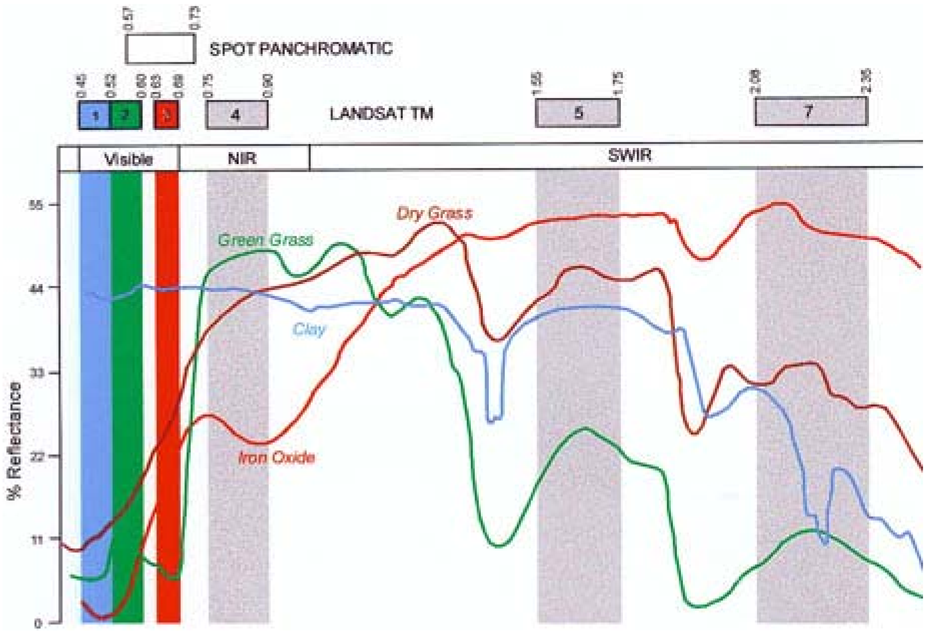

Figure 3 Percentage of Absorption and reflection in each types of satellites.

2.3.2.2 The interaction of the incident energy with the atomic structure of soil, rocks, plants, bodies of water, man-made objects, etc. governs how much energy is absorbed and thus how much is reflected. Materials such as minerals and leaves have a (at least somewhat) distinctive reflectance spectra which can be measured by a laboratory or field-portable spectrometer. These values can then be compared to remote sensing data in order to identify which materials are present in the area of an image.

2.3.3 As electromagnetic waves travel , they encounter objects (discontinuities in velocity) that reflect some energy like a mirror and transmit some energy after changing the travel path (reflect - like a lens).

2.3.4 The distance (*d*) a electromagnetic wave travels in a certain time (*t*) depends on the velocity of the material (*v*) through which the wave is traveling;

$$d = v\ t.$$

Since all electromagnetic energy travels at the speed of light (usually denoted by c) this is equation is very simple to apply. For example, we can generate radar waves and time their return after reflection from the Earth's surface. The distance to the reflecting object can be obtained from this simple equation.

2.3.5 The velocity (*c*), frequency (*f*), and wavelength(*l*) of a electromagnetic wave are related by the equation: $c = f\ l.$

Strictly speaking, this equation applies to simple harmonic motion in which we are dealing with a wave that consists of motion with a single frequency (i.e. a sine wave), but in remote sensing, one usually treats measurements that are made for a narrow band of frequencies as representing a single frequency at the middle of the band. (Spanner, 1990)

Figure 4 to the right, the wave is depicted as what happens to a particular point as the wave passes so the horizontal axis in this diagram is in units of time and the distance between two peaks is the period. We can also think of depicting the wave as if we could take a snapshot of it at one instant of time. In this case, we are looking at a spatial picture in which the horizontal axis is in units of distance and the distance between two peaks on the wave form is the wavelength.

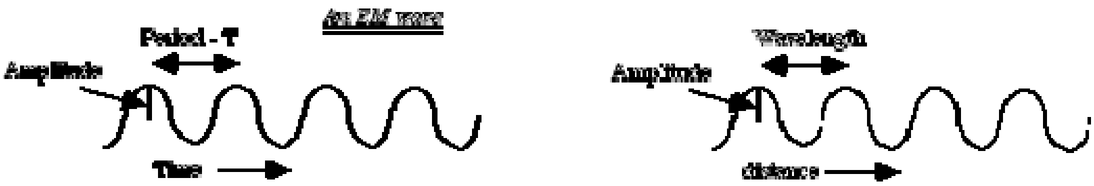

Figure 4 Normal wavelength and an existing wavelength

We usually think of electromagnetic waves as a constant stream of energy from the Sun for a particular frequency. Thus, the depictions in the figure of an EM Wave apply. In radar applications, we generate our own energy, and in this case, it is appropriate to think of these waves as a pulse (wavelet) of energy that is as short as possible in time. Strictly speaking, it is no longer simple to think of the wavelength or frequency of this wave although such terms are still commonly used in radar studies. One reason for this usage is the utility of wavelength as a measure of the spatial dimension of a electromagnetic wave relative to some object whose interaction with the wave is of interest. (Chada, 2000)

2.3.6 It is quite appropriate to look at the *amplitude* of a electromagnetic wave and think of it as a measure of the energy in that wave.

Obviously, the larger the amplitude the more energetic the wave. However, technically we must remember that introductory physical treatments of simple harmonic motion show that energy is proportional to the square of the amplitude.

2.3.7 Electromagnetic waves lose energy (amplitude) as they travel because of several phenomena: (Danie, 1990)

2.3.7.1 Spherical spreading phenomena is simply the fact that the amplitude dies off with distance traveled (i.e., the further you are from the source, the weaker the signal). This obvious physical effect is no different than observing that as one gets farther and farther from a source of sound, a weaker and weaker sound is heard. Technically, as the wave travels, the wavefront is a sphere whose radius gets larger and larger. Since the area of a sphere is 4¶r^2, at some radius (distance traveled), the energy per unit area on the wavefront is equal to the original energy divided by 4¶r^2. Since the energy is proportional to the square of the amplitude, the amplitude is proportional to 1/R. Strictly speaking, the energy is not diminished by this effect, it is just spread over the surface of larger and larger spheres.

2.3.7.2 Absorption It is appropriate to think of absorption as the tendency for materials to simply soak up electromagnetic energy and convert it to heat. We can measure some of this energy as radiated (emitted) heat that is at a longer wavelength than the original energy.

The Scattering interaction of electromagnetic waves with objects that are small relative to a wavelength (i.e., molecules in the atmosphere) leads to energy being reflected (scattered) in an unorganized fashion. This term is very intuitive, and we can just think of the energy being randomly dispersed in three dimensions. The primary scattering process at work is called Rayleigh scattering which proportional to 1 over the wavelength to the 4^{th} power. Thus in the atmosphere, short wavelengths are subject to selective scattering which is basically due to the fact that many particles and molecules are too small to effect the longer wavelengths significantly. Short wavelength energy like UV and blue light are in fact scattered about twice as strongly as red light. For wavelengths in the visible light range, selective scattering causes us to see that color. Blue sky, which is the effect of selective scattering of blue light, is the classic example of this effect. This effect (along with absorption) is a fundamental restriction in remote sensing and limits the range of wavelengths that can be employed. When large particles like water droplets are encountered, the scattering effects a broad range of wavelengths (non-selective), and then we simply see unorganized white light. Classic examples of this effect are clouds and fog. (Spanner, 1990)

2.3.8 Electromagnetic waves lose energy (amplitude) as they travel because of several phenomena:

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2.3.9 Resolution

Yang (1997) found the question which constantly arises in electromagnetic studies is whether the target of a remote sensing study can actually be "seen" (resolved) in the data. When one looks at a printed copy of a particular image or picture, resolution can be quantified to some extent but is really no more complex than determining if the interpreter can see the object, pattern, signature, texture, etc. that is of interest. In the case of digital images, resolution is a technical issue in two ways. First of all by its very nature, a digital image is composed of pixels (ground resolution cells) of finite size. The measurement for each cell is the intensity of the reflected energy for a particular band of wavelengths, and the size of the pixel is a measure of spatial resolution. The simple fact is the smaller the pixels, the better the resolution

Good resolution is of course desirable, and as new satellites and airborne systems become available, the tendency is for the pixel size to decrease (30 m for Landsat Thematic Mapper, 10 m for SPOT, 5 m for IRS). There is also the question of spectral resolution. In this case, the question is how well have we determined the variation of reflectance with wavelength (the reflectance spectra). As a practical matter, this simply a question of how many bands a particular device is capable of measuring and what is their width (i.e., what is the range of wavelengths). A large number of narrow bands is desirable and again technology is providing data with better resolution. For example, Landsat Thematic Mapper makes measurements for 7 bands while the AVIRIS airborne system makes measurements for 224 bands. Thus in digital terms, the denser the measurement, the better the resolution. However, the data volume also increases greatly as the sampling rate (spatially or in wavelength) increases, and this will affect the speed and ease of the data processing. (Yang, 1997).

3. Related studies

3.1 Evapotranspiration

Cargnel *et al*. (1996) estimated the evapotranspiration (ET) over a soybean (*Glycin max* Linn.) field during growing season in Argentina using the energy balance technique and based on the Bowen ratio method. The result reports that ET were approximately 90% of the available energy (R_n-G_s). The observed Bowen ratio value were zero during and a day after rain and increased to approximately 0.45 several days after the rain. Besides, they stated that one of the well know micrometeorological methods is based upon the Bowen ratio. This method has the advantage that it is not necessary to measured turbulence or wind speed and is independent of atmospheric stability. The soil evaporation contribution to ET diminishes when plant soil cover increased to at least 45% of the ground surface and plant evaporation was not limited by soil water. Under these circumstances, daily net radiation expressed as ET, was approximately equivalent to daily ET.

Aoki *et al*. (1996) have tried to find out the empirical formula by measurement data of actual evapotranspiration in different crop fields in Hokkaido, Japan. The empirical equation for estimating the Bowen ratio using four climatic factors, that are mean air temperature (T_a), mean relative humidity (RH), means soil water tension (pF), mean wind speed(WS), and the vegetation coverage (VC). The equation allowed to estimate evapotranspiration in the daytime by the error of less than 0.5 mm., for the crop fields. They reported that the equation was derived from the data in three beet fields in August 1992 as :

$$\beta = B - 0.30 \qquad (49)$$

$$\text{Where } B = -0.0279T_a + 0.00566RH + 0.0290pF + 0.00744WS \qquad (50)$$

(multiple correlation coefficient is = 0.89)

Otherwise, they found that this behavior was related to vegetation coverage(VC) or leaf area index (LAI). The better correlation was obtained when VC or LAI was added in the equation. Finally, the following equations was derived by adding VC :

$$\beta = 0.498(1 + 0.01VC)B + 0.299(1 - 0.01VC) + 0.24 \quad (51)$$

The equation of β was also derived by adding LAI :

$$\beta = 0.479(1 + 0.082\ LAI)Bi - 0.0526\ LAI + 0.34 \quad (52)$$

$$Bi = -0.0199T_a + 0.00569RH + 0.0518pF + 0.0434WS \quad (53)$$

The comparison between observed Bowen ratio and estimated Bowen ratio indicates rather good agreement between estimated and observed Bowen ratio. The estimated evapotranspiration agrees well with observed evapotranspiration.

In theory, the Bowen ratio method needs the measurement of all parameters between the top of the crop at height h and the top of boundary layer at height z. This height, in practice, cannot be determined precisely but the measurement on wind profiles suggested that z-d should not exceed 0.5% of the fetch over a uniform surface with a zero plane displacement of d in practice, measurements are often taken up to a height where z-d is 1% or more of the fetch and observations from the top of the profiles are discarded if they are obviously anomalous (e.g. if the potentials are not linearly related to each (Monteith, 1973).

Chunkao (1971) estimated the evapotranspiration (ET) above the dry evergreen forest and adjacent clearing area in the Sakaerat Experimental Research Station (SERS), Changwat Nahkhon Ratchasima, Thailand, during June 3 to September 3, 1971 by Bowen ratio, Aerodynamic and Penman method. He found that the actual evapotranspiration (ET_a) by Bowen ratio method in dry evergreen forest ranged between 0.8-5.8 mm day^{-1} with mean value of 3.7 mm day^{-1} and in adjacent clearing area, the ET_a was between 0.0-9.3 mm day^{-1} with mean value of 3.2 mm day^{-1}. The ET_a by Aerodynamic method in dry evergreen forest ranged between 0.2-7.3 mm day^{-1} with mean value of 2.0 mm day^{-1}. The potential evapotranspiration (ET_p) by Penman method in dry evergreen forest ranged between 4.9-17.1 mm day^{-1} with mean value of 10.5 mm day^{-1} and in adjacent clearing area, the ET_p was between 2.4-7.7 mm day^{-1} with mean value of 4.7 mm day^{-1}. While the pan evaporation ranged 1.3-6.1 mm day^{-1} with average 4.0 mm day^{-1}. Yabuki *et al*. (1983) studied the evapotranspiration in sugarcane area, dry evergreen forest and oil palm area in Thailand in dry period by Bowen ratio method, and found that the ET_a values in dry season were 3.3, 1.9 and 1.3 mm day^{-1} respectively. Tangtham *et al*. (1995) studied the water balance in northern river basin found that the annual water loss from Chao Phraya River Basin and Ping River Basin by evapotranspiration process was around 83-90 % and 74-85 % of annual rainfall respectively. Koichiro (1996) found that the mean annual evapotranspiration for natural forests in the humid tropics ranged about 1,450-1,750 mm and about 1,150-1,400 mm in the subhumid and wet-dry excluding tropical mountain forest area. Suksawang and Tangtham (1996) found that the annual pan evaporation (E_{pan}) at Linthin subwatershed of Mae Klong Watershed Research Station is about 1,340 mm. If ET_a was assumed to be about 0.7 E_{pan}, ET_a was estimated at 938 mm/annum. Also, Vudhivanich (2001) studied the monthly ET_p by using Penman-Montheith, modified Penman and pan evaporation methods with a long term climatologically data. He found that mean monthly ET_p of Thailand varied between 101 mm in December to 148 mm in April. The average annual ET_p was about 1,434 mm.

3.2 Remote sensing

Danie (1990) This approach proposed a methodology, which exploits the strong, approximately linear relationship between the amount of solar irradiant absorbed by plant pigments and short-wave vegetation indices calculated from red and near – infrared reflectance. A supervised binary decision tree classification of phytophenological variables derived from MultiMate normalized difference vegetation index (NDVI) imagery. Although interpretation of the various classes is limited considerably by the quality of global vegetation index imagery, the data show clearly the marked temporal asymmetry of terrestrial photosynthetic activity.

Spanner *et al*. (1990) calculated the monthly NDVI for 19 coniferous forest stands in Oregon, Washington, Montana, and California. The seasonal variation of NDVI was related to phenological changes in LAI, as well as the proportion of surface cover types contributing to the overall reflectance. It was concluded the AVHRR NDVI data of July were related to the seasonal maximum. LAI of coniferous forests of the western United States, and that seasonal differences in the AVHRR NDVI were related to : a) phonological changes in LAI caused by climate, b) the proportions of surface cover types contributing to the overall reflectance, and c) large variations in the solar zenith angle.

Eck (1990) The near-infrared channel of the NOAA advanced very high resolution radiometer (AVHRR) contains a water vapor absorption band that affects the determination of the normalized difference vegetation index (NDVI). This water vapor effect is quantified for the Sahel by radioactive transfer modeling and empirically using observations. In extreme cases, changes in water vapor are shown to result in reduction of the NDVI by 0.1. Variations of the NDVI of 0.01 would result from typical low atmospheric water vapor days within the wet season.

Groten (1993) derived ARTEMIS-NDVI data for 1984-1989 from the NOAA-AVHRR sensor for crop monitoring and early crop yield forcasting. In order to remove residual effect of clouds and other atmospheric influences on 10-day maximum, NDVI images. Various NDVI regression parameters were compared.

Foody *et al*. (1996) The spectral separability of thirteen topical vegetation classes, including twelve forest types, was assessed for three spectrally separable groups. On the basis of the class separability analyses the three spectrally separable groups were mapped, with an accuracy of 94.84 percent, from Landsat TM data by a maximum likelihood classification.

Kaneko *et al*. (1996) Taking forest areas specified according to information on the National digital land data, the vapor pressure deficit at leaf temperature and the degree of stomatal opening were presumed using the vegetation index of forest areas derived from Landsat TM data. The relation of the increase of latent heat flux with thevgetation index was obtained, and the plane distribution of latent and sensible heat fluxes in large areas is calculated.

Liu *et al*. (1996) NDVI (Normalized Difference Vegetation Index) images generated from NOAA AVHRR GVI data were recently used to monitor large scale drought patterns and their climatic impact on vegetation. The purpose of this study was use the Vegetation Condition Index (VCI) to further separate regional NDVI variation from geographical contributions in order to assess regional drought impacts. NDVI value reflected the different geographical conditions quite well. Seasonal and interannual comparisons of drought areas delineated by the VCI provided a useful tool to analyze temporal and spatial evolution of regional drought as well as to estimate crop production qualitatively.

Yang *et al*. (1997) Research designed to better define relations between 1-km. multi-temporal AVHRR – derived NDVI data and selected climatological parameters, soil hydrological properties and land cover characteristics was summarized. Temporal change in NDVI found is closely linked with the temperature regime. NDVI-precipitatin and NDVI-potential evapotranspiration relations exhibited time lags, although the length of lag varied with land cover type, precipitation, and soil hydrologic properties. This may reflect both the relatively homogeneous land cover characteristics of the study area and the effect of off-nadir viewing geometry on AVHRR data acquisition

3.3 Evapotranspiration and remote sensing

Evapotranspiration is one of important parameter in many scientific fields. Reliable evaporation data is required for planning, designing and operating reservoirs, shipping canals and drainage systems. The major work of engineering is searching the answer of actual demand. The techniques for the measurement of evaporation and evapotranspiration under natural condition are developing in many countries. Because of complicate step to calculate evapotranspiration to calculate the evapotranspiration day by day.

Now a days remote sensing is a new technology, which is using satellite to collects the information. It usually refers to the use of electromagnetic radiation sensors to record images of the environment, which can be interpreted to yield useful information. We can use satellite data to evaluate regional evapotranspiration literature related to this study are briefly summarize below.

Lillesand *et al*. (1987) defined remote sensing as a science and art of obtain information about an object, area or phenomenon through the analysis of data acquired by a device that is not in contact with the object, area or phenomenon under study. Information form remote sensed satellite data can be very useful in studying the land use/cover at any particular time and also the temporal change inland cover over time. Experience has shown that many earth features of interest can be identified, mapped and studied on the basis of their spectral characteristics. If we compare the spectral reflectance curves of vegetation, soil and water, we can observe their distinctive pattern.

Raymond *et al*. (1989) used remote sensing data to measure the areas covered by type of vegetation. Multiply each area by predetermined water use rate, the total evapotranspiration was estimated. The acreage correctly averaged 84+11% of actual acreage for the major crop and 64+27% for minor crop. The acreage correctly classified averaged 86+10% of the acreage classified for the major crop and 59+28% for the minor crops. All the evapotranspiration estimates.

Sucksdorff *et al*. (1990) found from their research result that if atmospheric and surface parameters are known, the energy and hydraulic budgets at the soil/vegetation/atmosphere interface can be simulated. Then, soil temperature estimated by thermal infrared remote sensing can be used to derive the energy fluxes as far as they are the equilibrium term of the energy budget. The NDVI and the surface temperature estimated from NOAA/AVHRR data have been used to calibrate the leaf area index and the minimum resistance to evapotranspiration.

Maidment *et al*. (1992) Remote sensing have two potentially role in estimating evapotranspiration. First, offer methods for extending empirical evapotranspiration relationship to area where gauged temperature data may be spare. Secondly, may be use to measure variables in the energy and moisture balance models of evapotranspiration. Using the temperature sounders on the meteorological satellite in a linear regression model. Price used thermal data from the Heat Capacity Mapping Mission (HCMM) to estimate region scale evapotranspiration rates which were found to be comparable to pan evaporation data.

Tada (1996) used remote sensing data for estimation of the evapotranspiration distribution in Tohoku region of Japan. All data set were overlaid onto NOAA-AVHRR satellite data and classified into three land use; forest, paddy field and urban area. The evapotranspiration is estimated by different method for each land use.

They compared the evapotranspiration estimation from NDVI with Thornwaite equation, bulk transfer simulation and Sakuartani and Horie equation.

Sharma *et al*. (1997) Monthly actual evapotranspiration (AET) for four humid catchmant in Kenya were compared with those based on a water balance analysis. The results of 34 years daily rainfall data indicated that both models tended to overestimate. The Grindley Morton model (AET/G) greater than water balance 32% but either equal or close to the Penman. The Morton model (AET/M) performed better were only 8% closer to the water balance-based estimates.

Yang *et al*. (1997) approach for estimate of local evapotranspiration (ET) on the concept of a vegetation Index/Temperature Trapezoid (VITT). The evapotranspiration rate was computed using surface temperature (T/s) and Normalized Difference Vegetation Index (NDVI) derived from the TM data. The remote sensing estimates were compared with results from a water balance model. This study shows that Landsat TM data can provide a practical means for the ET at a local scale.