RN-13

COMPARATIVE STUDY OF COMPONENTS OF WATERSHED MODELS

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ABSTRACT

The rainfall runoff process in a catchment is a complex and complicated phenomenon governed by large number of known and unknown physiographic factors that vary both in space and time. The rain or snow falling on a catchment undergoes number of transformations and abstractions through various component processes such as interception, detention, evapotranspiration, overland flow, infiltration, interflow, percolation, sub-base flow, base flow etc. and emerges as runoff at catchment outlet. Application of mathematical modelling techniques to the constituent processes involved in the physical processes of runoff generation have led to better understanding of the processes and their interaction. Different types of watershed models have been developed depending on the purpose such as flood forecasting, simulation of hourly or daily runoff or estimation of water yield etc.

Each watershed comprises of different types of soil cover, vegetation, land use, topography, drainage pattern and density, slopes etc. The processes that take place are not uniform throughout the basin, moreover they are also not uniform in time, eg.interception loss depends on type of vegetation cover and its density and also on rainfall amount, its intensity and duration. Interception loss is high at the beginning of rainfall but reduces gradually to a constant value equal to potential evaporation rate till rainfall continues. Similarly infiltration rate vaires in space and time and also depends on initial soil moisture condition. As such exact analysis of these complex component processes is very difficult.

To simplify analysis of these complex processes different water-

shed models have adopted different laid out approaches, methods or approximations for each process and the developed model as a whole is capable to simulate observed runoff. A comparative study of model structures for various processes considered in different watershed models has been done to ascertain suitable model structure for each component process for typical physiographic and hydrometeorological conditions of river basins in India.

The models that have been included in this review are standard watershed model(SWM IV) developed by Grawford and Linsley of USA, UBC model(UNiversity of British Columbia model) developed at the Civil Engineering Department of the University, Vancouver, Canada, stream flow synthesis and Reservoir Regulation(SSARR) model of the U.S.Army Engineers, Sacramento model of the Sacramento river forecast Centre, California, USA, TVA daily streamflow model of Tennessee Valley Authority, USDAHL-74 hydrologic model of U.S.Department of Agriculture, HBV-model of the U.S.Department of Agriculture, HBV-model of the Swidish Meteorological and Hydrological Institute, USGS Peakflow synthesis model of U.S.Geological Survey, RORB(Version-3) model of Monash University, Australia and Leavesley model of George,H.Leavesley of Colorado State University, USA. The component processes that have been considered in the review are interception, evapotranspiration, overland flow, infiltration, percolation, interflow and base flow.

Different simplified techniques and model structures for component processes which have been identified in this review, may be considered for developing rainfall runoff models suited to Indian conditions.

1.0 INTRODUCTION

There are three possible approaches to a modelling application viz.(i) using an existing model,(ii) modifying an existing model, and (iii) developing a new model. Application of existing model is to be done after selecting the model that suits best to the particular application. Similarly an existing model may suitably be modified to suite the need, or developing a new model is recognised as a major necessity to suit the needs.

The decision about the best approach to a particular modelling application is to be made on the basis of knowledge, modelling principles, available watershed models and available data.

A deterministic watershed model usually includes the following elements:

(i) Input parameters representing physical characteristics of the watershed.

(ii) Input of precipitation and other meteorological data.

(iii) Calculation of water flows, both surface and sub-surface.

(iv) Calculation of water storages, both surface and subsurface.

(v) Calculation of water losses.

(vi) Watershed outflow and other outputs, if desired.

A deterministic watershed model consists of a series of submodels each representing a particular hydrologic process and usually is structured accordingly. Each submodel represents basically flow of water and usually includes a storage. The submodel outflow is either an outflow to the next sub-model or a water loss. Water storages regulate flow in the watershed itself. Most flows in a model are into or out of a storage. The flow is related to the amount of water in storage as well as other factors.

Model building is a process of choosing appropriate submodels linking them together to form a watershed model, and making the resulting watershed model. Selection of appropriate model depends on the purpose of the overall model. The questions to be considered in this connection are: (i) Is the model intended purely for predicting watershed outflow or for other purposes also (ii) Is the model intended for a particular type of watershed in terms of size, topography or land use and (iii) Is it intended for use on any type of watershed?

The overall model developed by assembling the submodels may be too cumbersome or too costly to operate. The next step is to reduce the detail wherever it can be done without serious loss of accuracy.

Watershed models can be characterized also as event models or continuous models. The accuracy of the model output may depend on the reliability of the input conditions. Continuous watershed model keeps a continuous account of the basin moisture condition and determines the initial conditions applicable to runoff events. Most continuous watershed models utilize three runoff components, direct runoff, interflow and ground water flow, while an event model may omit one or both the sub surface components, and also evapotranspiration. In terms of scope, there are complete models or partial models. It is useful also to characterise watershed models as fitted parameter models or measured parameter models.

A fitted parameter model is one which has one or more parameters that can be evaluated only by fitting computed hydrographs to the observed hydrographs. A measured parameter model on the other hand is one for which all the parameters can be determined satisfactorily from known watershed characteristrics, either by measurement or by estimation.

For example, watershed area and channel length can be determined from existing maps. Two examples of direct measurement would be field measurements of channel cross-sections and laboratory measurements of channel cross-sections and laboratory measurements of soil characteristics. Channel roughness is often estimated. A measured parameter model can be applied to totally ungauged watersheds and, therefore, is highly desired. The development of such a model, that is also continuous, acceptably accurate and generally applicable is, however, a very difficult task.

Watershed models can be classified also as general models and special purpose models. A general model is one that is acceptable to watersheds of various types and sizes. A special purpose watershed model is one that is applicable to a particular type of watershed in terms of topography, geology and land use.

Watershed models or submodels are also classified as distributed models or lumped models. A distributed model is one in which areal variations of watershed characteristics e.g.soil and land use can be utilized directly in applying the model. In a lumped model, this can not be done and therefore, representative or mean values of land slope, channel slope, length, soil characteristics etc. are usually used.

The Stanford watershed model(SWM) is the first comprehensive watershed model which was developed based on different component processes. The model has been widely used in water resource studies and has undergone numerous modifications, additions and revisions. The Kentucky version of the model is the best known among them. Stanford model is a continuous, complete and general watershed model and is applicable to watersheds of all types and sizes. In this model each flow is an outflow from a storage and can be expressed as a function of the current storage and physical characteristics of the subsystem. The overall model is physically based although many of the flows and storages are represented in a simplified or conceptual manner. The model employs various surface and sub surface water storages, which in most cases, are not defined explicitly. For example the lower zone which represents soil moisture storage, but neither the depth nor the soil moisture characteristics are specified and is utilized as an index value of water storage. One important feature of the model is that infiltration, interflow and evapotranspiration values are considered to vary over the watershed, somewhat arbitrarily. Infiltration capacity, for example, is assumed to be linearly distributed over the watershed area from zero to a maximum value. A portion of this is retained in the soil and the remainder becomes interflow. Similar relationships are used to calculate actual evapotranspiration from potential evapotranspiration. Channel routing is considered in two steps. First, a time-area histogram is used to represent the effect of translation time from various parts of the watershed and their relative areas. Secondly, a conceptual reservoir at the watershed outlet represents the effects of channel storage. In the routing portions of the model, fitted parameters are utilized for interflow, groundwater flow and the conceptual reservoir. Anderson and Crawford(1964) developed subroutine to consider snowmelt component utilizing daily temperature data and a number of parameters. Primary data inputs to the model are hourly and daily precipitation, daily maximum and minimum temperature data are also used if snowmelt component is to be calculated. Data of several rainfall stations may be used to improve the accuracy. Observed daily streamflow data are used to compare with the calculated values. The model IV version has sixteen land surface, channel system and ground water parameters. Out of these twelve are clearly fitted parameters. The model can be applied to large catchments by dividing the basin into number of sub-basins. The most common applications of the Standford watershed model are to obtain hydrographs for unusual rainfall events and to extend a short flow record sufficiently for hydrologic analysis.

2.0 REVIEW

Components of ten commonly used watershed models namely SWM IV, UBC, SSARR, Sacramento, TVA, USDAHL-74, HBV, USGS RORB(version-3) and Leavesley model have been reviewed. The components that have been reviewed are interception, evapotranspiration, infiltration, overland flow, percolation and base flow. The Stanford watershed model has extensively dealt with almost all the components whereas in some models some of the components have been taken into consideration and other components have been approximated. Review has been done here parameterwise and the models that has dealt the respective parameters extensively have been taken into consideration.

2.1 Interception

Precipitation that falls in a basin is intercepted by leaves and stems of vegetation and other forms of cover. Interception can be defined as that segment of gross precipitation input which wets and adheres to above ground objects until it is returned to the atmosphere through evaporation. When the leave or stem surfaces become wet additional precipitation falling on vegetation will start flowing down the stems of plants and become stemflow or fall off the leaves to become part of the through fall.

The amount of water intercepted is a function of:

(i) the storm characteristics.

(ii) the species, age and density of prevailing plants and trees and

(iii) the season of the year.

The interception loss is more at the beginning of rainfall and it gradually reduces to a constant value equal to evaporation loss during the storm period. Percentage of interception loss is more for smaller amount of rainfall. Oak or Aspen leaves may retain as much as 100 drops of water. In average for a well developed tree, retention may be of the order of 20 drops per leave. For light shower, where $P < 0.01$ inch, 100% interception may occur, whereas for showers where $P > 0.04$ inch, losses occur in the range of 10 to 40%. This may be represented graphically as shown in the figure 1.

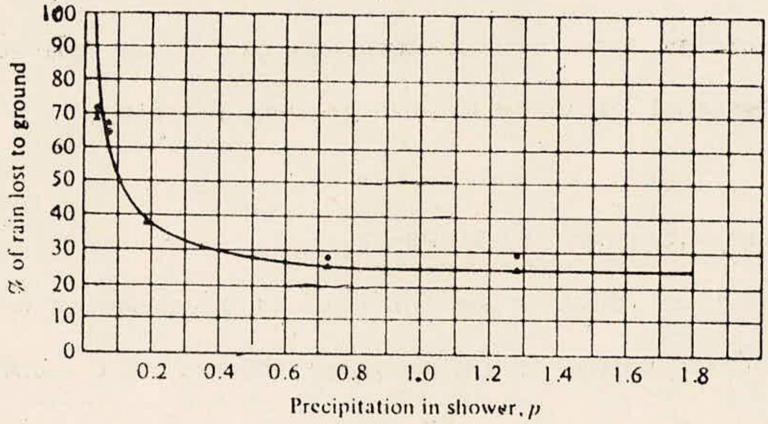

FIGURE 1 - MEAN CURVE SHOWING TOTAL PERCENT OF PRECIPITATION IN A SHOWER INTERCEPTED BY VARIOUS TREES

(ADOPTED FROM VIESSMAN ET AL'ENGG HYDROLOGY')

$$L_i = S_i + KEt \qquad \ldots(1)$$

where L_i = the volume of water intercepted(in).

S_i = the interception storage that will be retained on the foliage against the forces of wind and gravity(usually varies between 0.01 to 0.05 inches).

K = the ratio of surface area of intercepting leaves to horizontal projection of this area(value ranges from 1.7 to 9.2).

E = the amount of water evaporated per hour during precipitation period.

t = time in hour.

Total interception by an individual plant is directly related to the amount of foliage and its character and orientation. Interception loss function also varies with the storm characteristics. Using the above formulae only total amount of interception losses can be calculated but its distribution can not be known. Common practice is to deduct the estimated amount entirely from the initial period of the storm as initial abstraction.

A general equation for estimating such losses is not available since most studies have been related to particular species or experimental plots strongly associated with a given locality.

2.1.1 Stanford watershed model(SWM-IV)

Interception in any time interval is governed by watershed cover and by the current volume in interception storage. All incoming moisture enters interception storage until a pre-assigned volume EPXM is filled. This depends on types of cover and values considered are

Grassland	0.10 inch
Moderate Forest Cover	0.15 "
Heavy forest cover	0.20 "

Evaporation from interception storage is assumed to occur at a rate that corresponds to the current rate of potential evapaotranspiration. Therefore, interception will continue during a storm due to evaporation losses.

The flow chart of the model for interception component is given in figure 2.

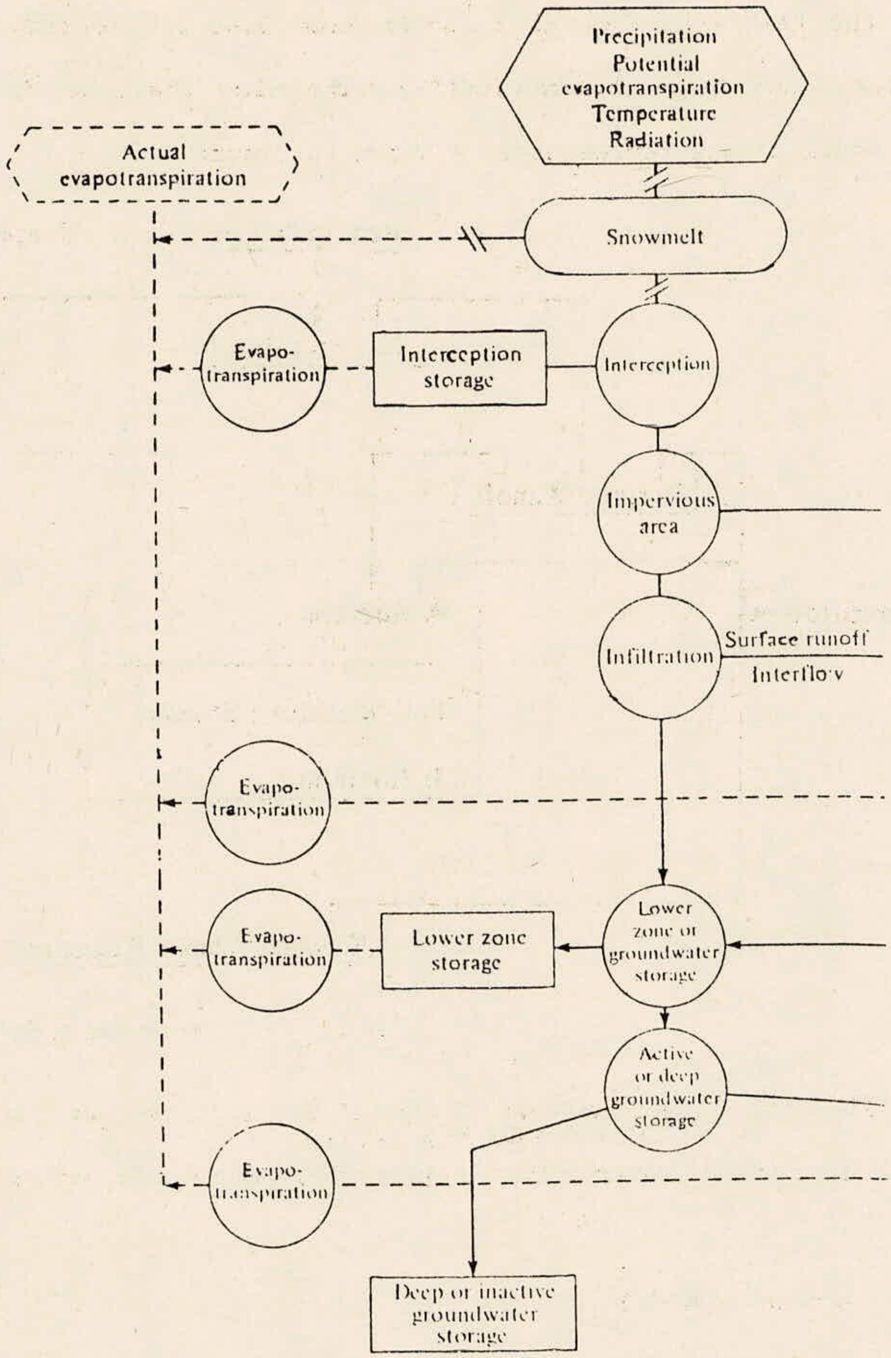

FIGURE 2- FLOW CHART OF SWM IV MODEL FOR INTERCEPTION COMPONENT
(Adopted from Viessman et al 'Engg.Hydrology')

2.1.2 Tennessee Valley Authority(TVA) model

In this model interception is considered to have non-parametric seasonal variations. The value is considered to be an initial abstraction from the rainfall. The value ranges from 0.05 inch during winter(WCEPT) to 0.25 inch during summer(SCEPT). The flow chart of interception in this model can be represented as shown in figure 3.

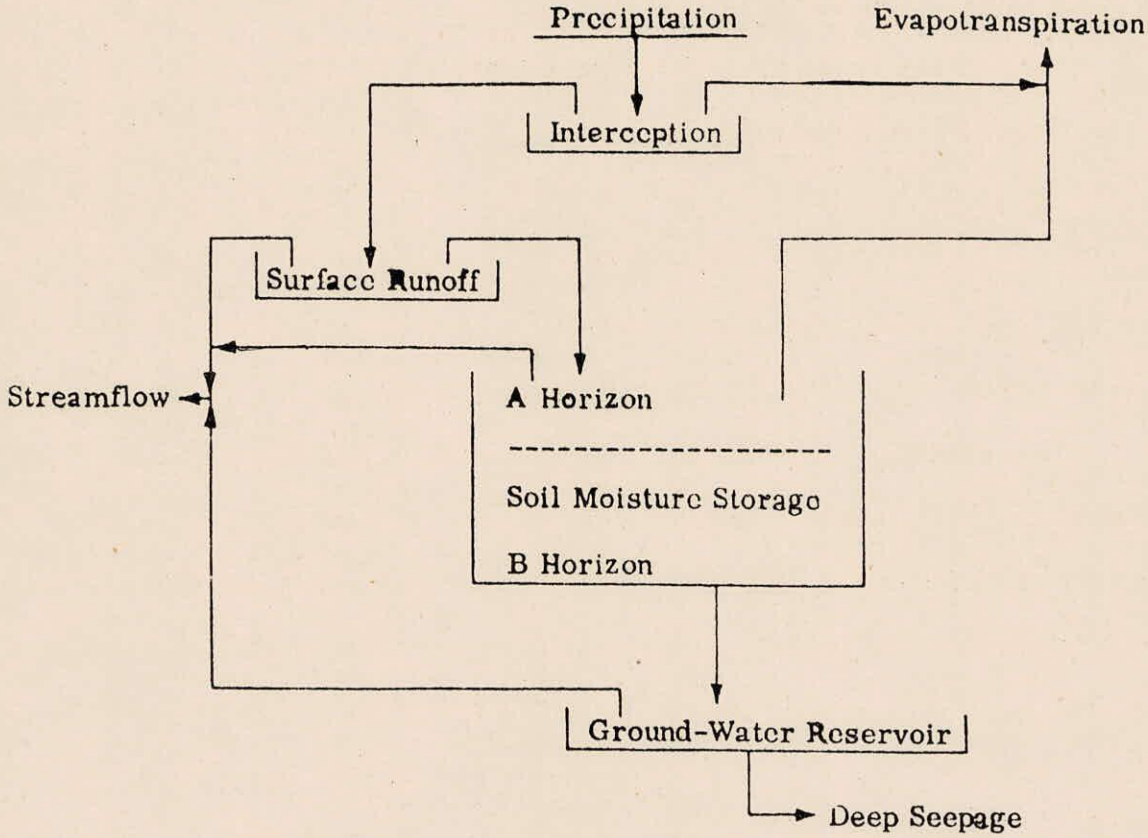

Figure 3 - Flow chart of TVA model for interception component. (ADOPTED FROM TVA 1972)
Storage in interception sets recharged through evapotranspiration.

2.1.3 Leavesley model

The factors like vegetation type, canopy density and precipitation type have been taken into consideration to compute interception(XIN) of each (HRU) Hydrologic Response unit of watershed (with homogeneous hydrologic response) cover density(COVDN) expressed as a percent the

HRU, surface covered by a horizontal projection of the vegetation canopy, a canopy storage for rain(RNST) in inches depth and a canopy storage for snow(SNST) in inches water equivalent depth are input for each HRU.

For the precipitation events occuring as all rain or all snow and where

(i) Precipitation depth PPT > RNST or SNST

$$XIN = (RNST \text{ or } SNST) * COVDN(inches) \qquad \ldots (2a)$$

(ii) Precipitation depth

PPT < RNST or SNST

$$XIN = PPT * COVDN(inches) \qquad \ldots (2b)$$

When the precipitation takes place as a mixture of rain and snow, it is assumed that rain occurs first followed by snow.

Interception losses through evaporation and sublimation are assumed to vary with precipitation form. For intercepted rainfall, all interception loss, XIN is considered lost through evaporation.

Interception losses of the intercepted snowfall is obtained from the energy available for sublimation on the deny following the storm.

2.2 Evapotranspiration

Evapotranspiration(ET) is the conversion of water to vapour and the transport of that vapour away from the watershed surface into the atmosphere. The ET varies both in space and time and mainly depends on available water and solar radiation. Water is available at plant surfaces, soil surfaces, streams and ponds or snowpacks.

Evapotranspiration flux moves large quantities of water from the soil back to the atmosphere. Lenpold and Langbein(1960) estimated that 70 percent of the precipitation falling on the United States is returned to the atmosphere through E.T.Accurate, spatial and temporal predictions of evapotranspiration are necessary for hydrologic models. Soil moisture storage can be determined from the difference of infiltration and evapotranspiration. But the parameters infiltration, percolation,evapotranspiration and other hydrologic variables are interdependent on the soil moisture both on quantity and its spatial distribution. The important influence of ET in hydrology has been shown and discussed by Woolhiser(1971, 1973), McGuinness and Harrold(1962), Knisel etal(1969) and Paramele(1972).

ET varies from place to place in a watershed and also varies throughout the day but spatially averaged daily E.T.values may be used for hydrologic analysis of watershed models.

The evapotranspiration phenomenon was observed by scientists since early recorded history(Biswas,1970). In 346 BC, Aristotle first wrote treatise on meteorology and evaporation. Fitzerald(1886) identified many of the important quantities and variables related to pan and lake evaporation. In the mid twentieth century Thornthwaite and Holzman(1942) described method of calculating evapotranspiration values. Penman(1948) ,in his model, described a method to calculate E.T.by combining the vertical energy budget with horizontal wind effects. Harrold and Dreibelbis(1958,1967) have done lysimeter studies and identified plant characteristics effects. Gates and Hanks(1967) have done extensive work on effects of plants on ET. Evapotranspiration from vegetated surfaces is the result of several processes like radiation exchanges, vapour transport and biological growth, operating within a system involving the atmosphere, plants and soil.

(a) Principles

Evaporation takes place from soil surface and water bodies. Evapotranspiration takes place from vegetated surfaces. The process requires solar energy as input, water availability and a transport process from the surface into the atmosphere. Researchers like Tanner (1957), Goodell(1966), Penman et al(1967), Gray (1970) and Campbell (1977) have provided good descriptions of these primary variables which determine E.T.rates.

Soil surface and water availability to the evaporating plant often limits ET . The rate of ET is limited to the diffusion rate of soil water to the soil surface and to the plant roots and through the plant system. Transport of water vapour upward from the evaporating surface for most vegetated situations does not often significantly limit the ET process. The horizontal advection of sensible heat from areas of excess energy to areas of limited energy is another important energy source for E.T. This is often called the clothes line or oasis effect.

Evapotranspiration varies spatially as a resualt of variations in climate, crops, or soils. Elevation, orographic effects and cropping patterns can cause large changes in E.T.Spatial averaging of E.T.values for a basin or sub-basins are generally done. The daily E.T.data indicate the annual distribution and daily variation of E.T.values. The considerable daily variation within each month demonstrates the dynamic behaviour of E.T.values.

Schematically soil-plant-atmosphere system may be represented as shown in figure 4.

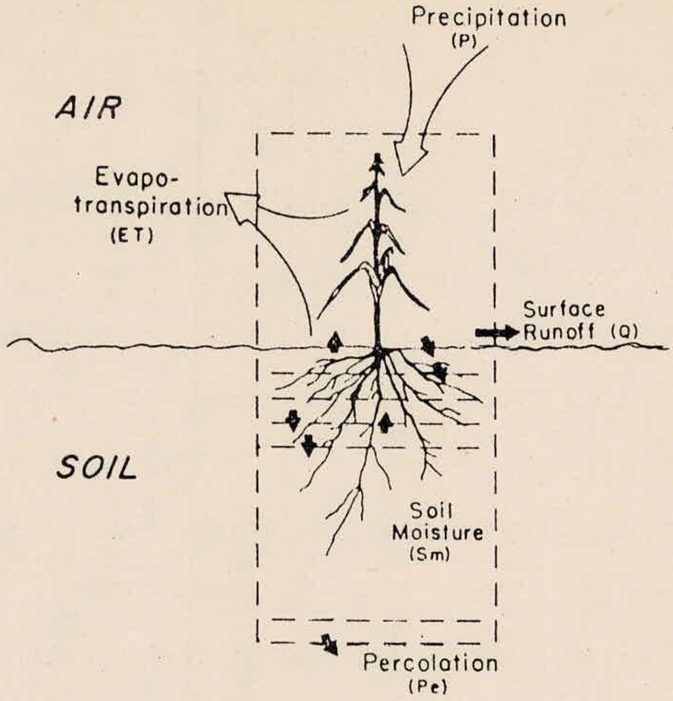

FIGURE 4- SCHEMATIC REPRESENTATION OF SOIL-PLANT-AIR SYSTEM WATER BUDGET(Adopted from Haan'Small Catchment Hydrology')

Estimation of E T follows a vertical water budget within a system. It requires to consider three sets of variables (i) determination of potential E T (ii) plant-water-related characteristics and (iii) soil-water related characteristics.

(b) Potential E T

The potential E T (or PET) is usually defined as an atemospheric determined quantity, which assumes that the E T flux will not exceed the available energy from both radiant and convection sources. Techniques for estimating potential E T are based on one or more atmospheric variables like solar or net radiation and air temperature and humidity or some measurement related to these variables, like pan evaporation. Measurement or prediction of some variables such as vapour or heat flux is difficult, only radiation is measured routinely.

(c) Pan Evaporation

Evaporation that takes place from shallow pan is called pan evaporation. This is one of the oldest and most common method of estimating potential E T which can be expressed as

$$PET = C_{ET} . E_p \qquad \ldots (3)$$

where C_{ET} = a pan to P E T Coefficient.

The coefficient C_{ET} is necessary because evaporation for a pan is generally more than that from a well-wetted vegetated surface. The value of C_{ET} generally varies from 0.5 to 0.8.

Methods to calculate pan evaporation from meteorological data are given by Penman(1948), Kohler et al(1955), Christiansen(1966,1968) and Kohler and Permele(1967)

(d) Energy Budget

In this method, calculation of potential E T is done by energy budget method. Energy limits evaporation where moisture is readily available and the necessary vapour transport occurs. Energy balance over a vegetated surface is shown schematically in figure 5.

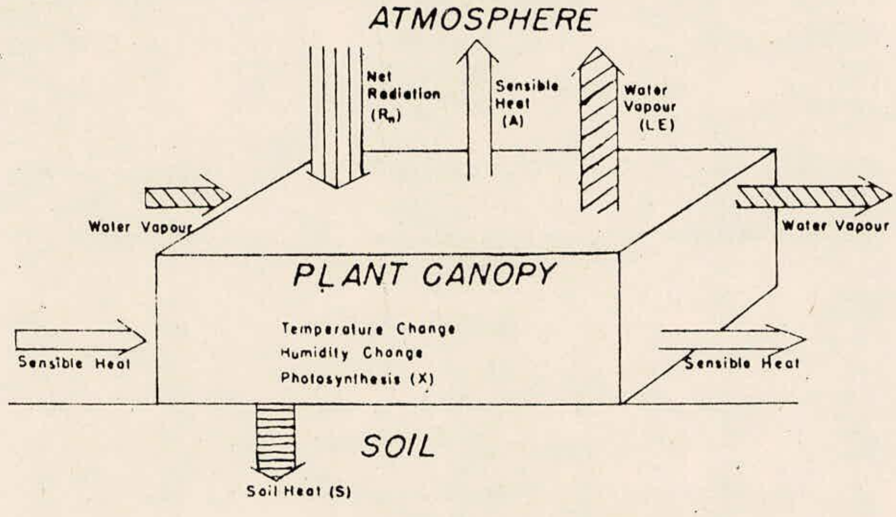

FIGURE - 5 ENERGY BALANCE OVER A VEGETATED SURFACE

In this method, vertical components may be expressed as

$$R_n = A + LE + S + X \qquad \ldots (4a)$$

and

$$R_n = R_s - a\,R_s + R_I - R_{Ir} \qquad \ldots(4b)$$

where

R_s = Incoming solar radiation(short wave)

aR_s = solar radiation reflected

R_I = Incoming radiation(Long wave)

R_{Ir} = Emitted long wave radiation.

R_n = Net radiation

A = Sensible heat of air

LE = Latent heat of water vapour.

S = Soil heat

X = Miscellaneous heat sinks, like plant and air heat storage and photosynthesis.

(e) Temperature Based Methods

Some correlation exists between the climatic variables causing potential E T and air temperature. Air temperature data are readily available. This is one of the most readily available climatic variables. There are several methods for predicting potential E T based on average air temperatures.

The Blaney-Criddle(1966) method is an extensively used method for irrigation design particularly in western US. The equation is

$$U = K_p \frac{45.7\,t + 813}{100} \qquad \ldots(5)$$

Where

U = estimated monthly evapotranspiration in mm.

K = an empirical consumptive use coefficient.

p = mean monthly percentage of annual day time hours of the year and

t = mean monthly air temperature in ^{o}C

Experience has shown the results of energy budgets are usually more **reliable than** temperature based method.

The other methods are (a) aerodynamic profile method as described by (i) Dyer(1961) for mass transfer eddy flux method or (ii) that of Parmele and Jocoby(1975) for the Bowen ratio measurements and (b) Combination method, Penman(1948,1956)

(f) Sensitivity Analysis

To assess the accuracy of prediction of potential evapotranspiration(PET), it is necessary to evaluate the relative effect of several variables that cause PET. Sensitivity analysis help to determine the required accuracy of instrumentation for measurements and calculations needed for estimating PET. Evaporation for each period is the result of a unique set of variable effects, so no single answer is possible. But average guidelines have been developed by McCuen(1974), Saxton (1975), Coleman and Decoursey(1976).

Among the energy related variables, the net radiation flux variable R_n is very important, Aerodynamic variables are usually less important except when there are very dry winds.

(g) Spatial Variation

Climatologic variables which determine PET tend to vary slowly with distance given that major land form features are reasonably similar. For some applications, when data are transferred from off-site, the effects of aspect and slope may be important. Foyster(1973) described a grid technique to determine regional PET and the method of computing actual ET in the stanford watershed model contains an empirical adjustment for spatial variation over large watersheds.

(h) Comparison of methods

The selection of a method for potential ET estimates depends on (i) data availability(ii) accuracy required(iii) time available to develop accurate estimates from available data sources. Studies comparing the results of several methods were reported by McGuinness and Bordne(1972), Bordne and McGuinness(1973) and Parmele and McGuinnes (1974). Doorenbos and Pruitt(1975) and Burman(1976) showed similar comparisons for a vareity of stations.

(i) Plant Transpiration

Plants control a large number of the processes that determine ET rates, such as(i) use of radiant energy (ii) stomatal control of leaf transpiration(iii) root interaction with available soil water etc. Federer(1975) showed the recent trend in research of ET from physically controlled process to a physiologically controlled process.

The effects of plants on ET can be divided into the main categories of(a) Canopy (b) phenology (c) root distribution and (d) water stress. There are many interactions among these categories. Many of the basic interactions of crops with the atmosphere and soil are provided by Monteith(1976), Kramer(1969) and Slatyer (1967).

The dynamic development, maturation and decay of crop canopies significantly influence plant transpiration effects. The canopy of any particular day largely determines the amount of intercepted solar radiation or absorbed advection, thus hydrologic models must provide a representation of this dynamic plant behaviour.

The phenological of plants often modifies plants ability to transpire. As crop matures,its need for water and ability to transpire diminishes.

The crop effects on ET have often been represented by crop coefficients, either as average seasonal values or as seasonal distributions. Most often the coefficients account for the combined effects of crop canopy, phenological development and soil evaporation.

Crop roots are also important in the process of connecting soil water with atmospheric energy and the resulting transpiration. However, root distribution and their effectiveness are difficult to study and quantify.

Transpiration process reduces at some level of deficiency of soil water and eventually ceases if water availability is severely limited.

(j) Soil Water Evaporation.

The process of evaporation from soil is similar to transpiration from a plant. Evaporation from soil takes place at three stages:

(i) In the first stage, the drying rate is limited by and equals the evaporative demand.

(ii) In the second stage, water availability becomes limiting and

(iii) In the third stage, it becomes limited to a more constant rate.

Gardner and Hillel (1962), Idso et al (1974) did some studies on this.

(k) Actual Methods

There are different methods to calculate ET from quite simple to very complex. A method should account for climatic, crop, and soil variables in some reasonable fashion under a range of moisture regimes. The methods are mentioned below in short:

(i) Based on daily water budget

$$ARI_i = (ARI_{i-1} + R_{i-1})K \qquad ...(6)$$

Where

ARI_i = Antecedent retention index for day i

R = Daily retention (infiltration)

K = A seasonally varied coefficient less than 1.0

(ii) Haan (1972) simulated daily ET in a model written to estimate monthly streamflow from daily precipitation by the relationship:

$$E = E_p(M/C) \qquad ...(7)$$

Where

E = Actual ET (mm/day)

E_p = Potential ET by the Thornthwaite method (mm/day)

M = Available soil moisture (mm) and

C = Maximum available soil moisture (mm)

(iii) Bair and Robertson (1966) gave a more complex soil moisture budgetting equation as:

$$AE_i = \sum_{j=1}^{n} K_j \left(\frac{S_i}{S_j}\right) Z_j \, PE_i \, e^{-W(PE_i - \overline{PE})} \qquad ...(8)$$

Where

AE = Actual ET(mm/day)

K = Coefficient of soil and plant characteristics

S_i = Available soil moisture(mm)

S_j = Capacity for available water(mm)

Z_j = Factor for different types of soil dryness curves.

W = Factor for effects of varying PE rates on AE/PE ratio

PE_i = Potential ET(mm/day) and

$\overline{PE}$ = Average for month or season(mm/day)

The equation(8) is summed for soil layers j for each day i.

(iv) Holtan et al (1975) gave a similar simple equation:

$$ET = (GI)\ K\ E_p [(S-SA)/S]^X \qquad ...(9)$$

where

ET = Actual ET(mm/day)

GI = Growth index of crop(Per cent)

K = Ratio of ET to pan evaporation for full canopy

E_p = Pan evaporation(mm/day)

S = Total soil porosity(Per cent)

SA = Available soil porosity(Per cent)

X = Exponent estimated to be 0.10

2.2.1 Stanford watershed model

In this model evapotranspiration is considered to take place from:

(i) Interception storage

(ii) Upper zone storage

(iii) Lower zone storage

(iv) Streams and lake surface and

(v) Groundwater storage

Evapotranspiration from interception and upper zone storage is considered to take place at potential rate, E_p, which is assumed to be the lake evaporation rate calculated as the product of a pan coefficient times the input values of the evaporation pan data.

Lake Evaporation Rate= Pan Coefficient x Input values of the evaporation pan data.

The evaporation of any intercepted water is assumed to occur at a rate equal to the potential evapotranspiration rate and ceases when the interception storage has been depleted.

Evaporation from stream and lake surfaces also occurs at the potential rate. The total volume is governed by the total surface area of streams and lakes(ETL).

$$K24EL = \frac{\text{Total streams and lakes area}}{\text{Total watershed area}}$$

Evapotranspiration from ground water storage also occurs at the potential rate and is calculated using a surface area equal to a factor(K24EL) multiplied by the watershed area.

$$K24EL = \frac{\text{Total area from which evapotranspiration from the ground water storage occurs}}{\text{Total watershed area}}$$

Generally value of this parameter is considered equal to the fraction of the watershed area covered by phreatophytes. Its value is normally small.

The upper zone simulates storage in depresssions and highly permeable surface soil while the lower zone is linkage to groundwater storage. If interception storage is depleted, the model will attempt to satisfy the potential for ET by drawing moisture from the upper zone storage at the potential rate. Once the upper zone storage is depleted, ET occurs from the lower zone but not at the potential rate; the ET rate from the lower zone is always less than E_p. Evapotranspiration opportunity is considered to control evapotranspiration from the lower zone. Evapotranspiration opportunity is defined as the maximum amount of water available for evapotranspiration at a particular location during a prescribed time interval.

Linear variation of potential evapotranspiration is assumed over a watershed as shown in figure 6.

The rate of evapotranspiration from the lower zone is determined

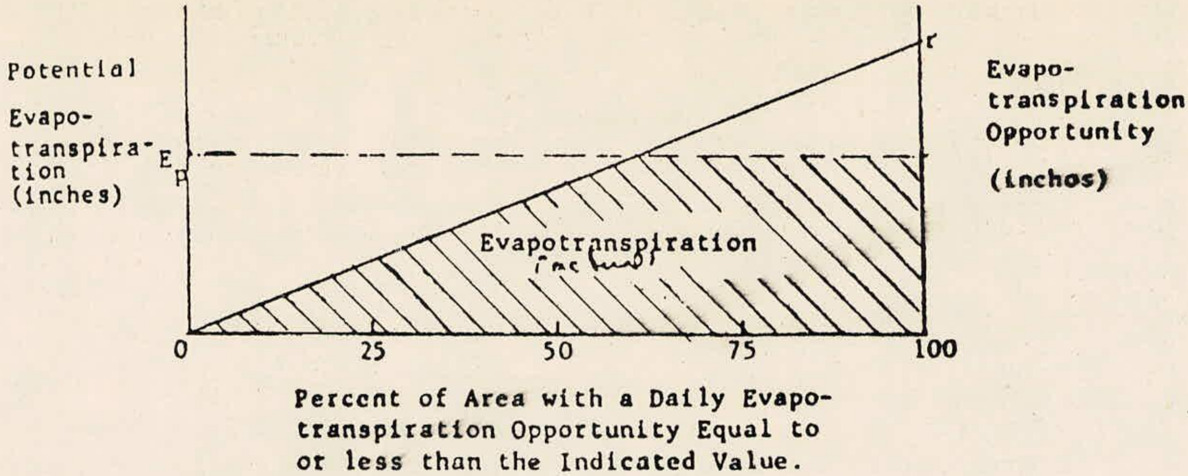

FIGURE-6 ASSUMED LINEAR AREAL VARIATION OF POTENTIAL EVAPOTRANSPIRATION

from the shaded area, or

$$E = E_p - \frac{E_p^2}{2r} \qquad \ldots(10)$$

The variable r is the evaporation opportunity. This factor varies from point to point over any watershed from zero to a maximum value of:

$$r = K3 \frac{LZS}{LZSN} \qquad \ldots(11)$$

where

LZS = The current soil moisture storage in the lower zone

LZSN = A nominal storage level, normally set equal to the median value of the lower zone storag

K 3 = An input parameter that is a function of watershed cover

Table:- Typical Lower Zone evapotranspiration Parameters

Watershed cover	K3
Open land	0.20
Grass land	0.23
Light forest	0.28
Heavy forest	0.30

Source:N.H.Crawford and R K Linsley, Jr.'Digital simulation in Hydrology:

Stanford Watershed Model IV', Deptt. of C.E.Stanford Univ.Tech. Report No.39,July 1966.

The ratio $(\frac{LZS}{LZSN})$ is known as the lower zone soil moisture into and is used to compare the actual lower zone storage with the nominal value at any time.

2.2.2 SSARR Model

Evapotranspiration loss is determined from potential evapotranspiration expressed either as watershed mean monthly values or determined from daily evaporation data. Flow chart of the evapotranspiration parameter of the model is shown in figure 7.

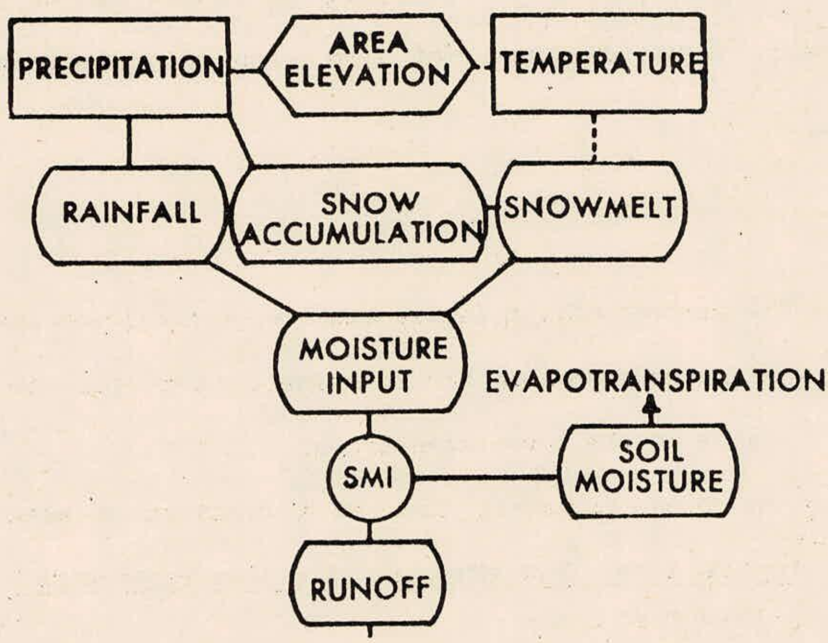

FIGURE-7 FLOW CHART OF SSARR MODEL SHOWING EVAPOTRANSPIRATION COMPONENT
(Adopted from U.S.Army Corps of Engineers 1972)

Evapotranspiration takes place from soil moisture. The soil moisture index(SMI) is an indicator of the relative soil wetness and is used to determine runoff. Soil moisture varies from minimum value equal to wilting point to maximum value equal to field capacity. The SMI is depleted only by the evapotranspiration index(ETI). The ETI can be specified either(i) in table form as month verses average daily potential evapotranspiration(Cm per day) or (ii) as weighted daily pan evaporation data(Cm per day) at one or more stations.

The tabler form of average monthly values of ETI is usually used when pan evaporation or other estimates on a day-to-day basis are not available, or when evapotranspiration amounts are not hydrologically significant. When monthly mean of daily ETI values are used the SMI is calculated at the end of each period as:

$$SMI_2 = SMI_1 + (WP-RGP) - [\frac{PH}{24} * KE * ETI] \qquad \ldots(12)$$

where

SMI_1 = Soil moisture index(in inches) at begining of period

SMI_2 = Soil moisture index(in inches) at end of period

PH = Period length in hours

ETI = Evapotranspiration index, in inches per day

WP = Weighted precipitation for the period

RGP = Generated runoff for the period

KE = A factor for reducing ETI on rainy days, specified(to the computer) in a table of KE versús rate of precipitation in inches per day.

For zero precipitation KE = 1.0 when rainfall occurs, the amount of soil water depletion by ETI diminishes and follows a relationship as shown in figure 8.

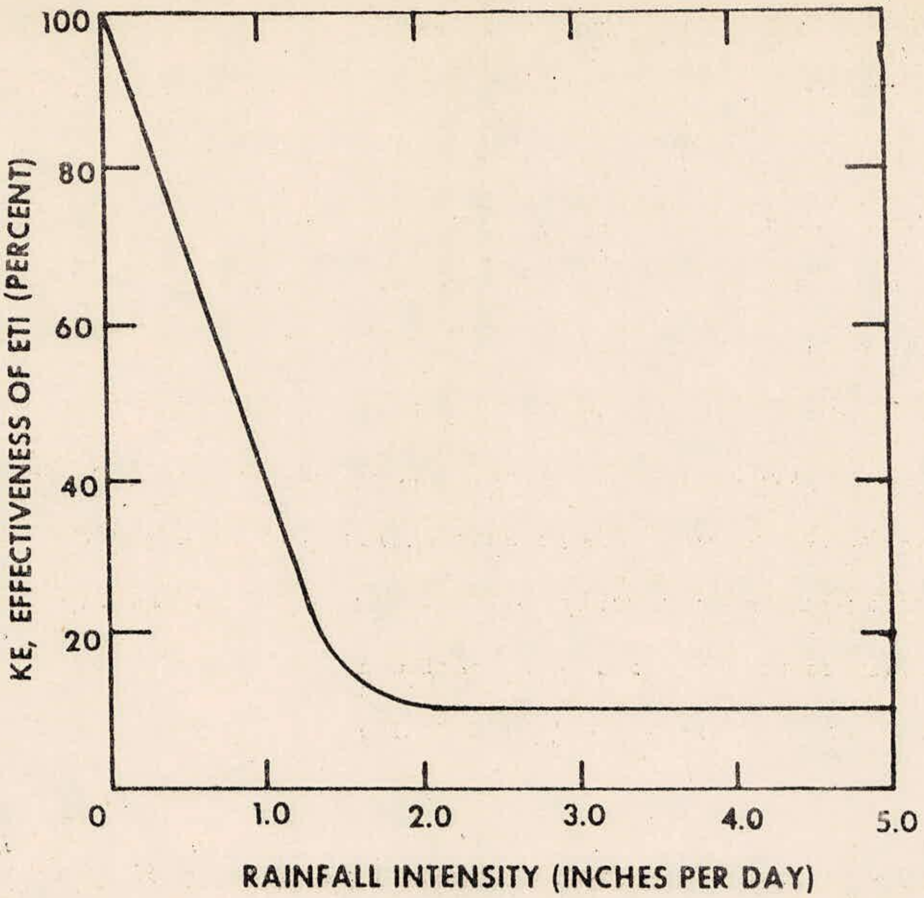

FIGURE 8 - SOIL WATER DEPLETION BY ETI WITH RESPECT TO RAINFALL INTENSITY (Adopted from U.S.Army Corps of Engineers 1972)

When accurate accounting of soil moisture change is required, daily ETI calculations can be made by entering daily evaporation data(such as pan evaporation). Daily estimates of ETI are desirable over monthly indices in arid and semi-arid basins where evapotranspiration losses are high in relation to precipitation input. Daily ETI values are calculated from pan evaporation data from one or more stations as:

$$ETI_d = \frac{(ETI_1 * Wt_1 + ETI_2 * Wt_2 + \ldots ETI_n * Wt_n)}{n} \quad \ldots(13)$$

where

ETI_d = Weighted daily evapotranspiration index(Inches)

$ETI_1 \ldots . ETI_n$ = Pan evaporation amounts (Inches per day) for each station.

$Wt_1 \ldots\ldots Wt_n$ = Weighting percentages for the respective pan evaporation stations to approximate actual evapotranspiration.

When daily ETI values are used SMI is calculated as:

$$SMI_2 = SMI_1 + (WP - RGP) - (DKE * ETI_d) \qquad \ldots(14)$$

where

DKE = A factor for reducing the daily ETI when soil moisture becomes depleted.

= 1.0 when soil moisture is adequate and then ETI approximates potential evapotranspiration.

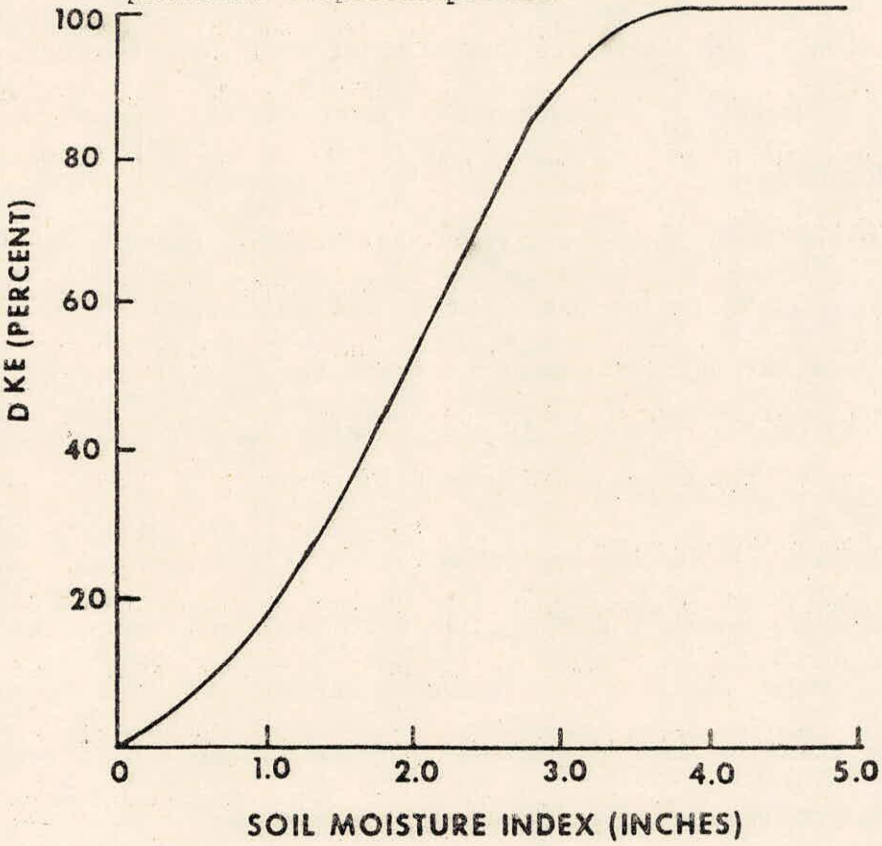

FIGURE - 9 RELATIONSHIP OF DKE VALUE IN PERCENTAGE WITH SMI (Adopted from U.S.Army Corps of Engineers, 1972)

When soil is dry DKE value in percentage diminishes with SMI values, this is shown in figure 9.

2.2.3 UBC Watershed Model

This is a daily rainfall runoff model which simulates daily runoff of a basin. The watershed model is designed primarily for mountaneous watersheds and calculates the total runoff contribution from both snowmelt and rainfall. The model is designed to operate on daily meteorological data inputs of maximum and minimum temperatures and precipitation. The basic structure of the model depends on a division of the watershed into a number of elevation bands. The elevation increment for each band is considered as same and an area for each band is specified. The model can be used for watersheds ranging from a few square miles to several thousand square miles.

In majority of the situations, most of the hydrometeorological stations are located in the valley. The model considers the following:

(i) Important aspect of the watershed model is the elevation distribution of data.

(ii) Temperature lapse rate is a key relationship because it influences the precipitation distribution and is significant in determining snowmelt rates at various elevations.

(iii) Precipitation inputs are made functionally dependent on elevation and on temperature regime.

The response of the watershed to snowmelt and rainfall is controlled by a soil moisture model. The soil moisture status of each area elevation band controls the subdivision of the total snowmelt and rain input into the various components of watershed runoff response. These components of runoff can be characterised as:

(i)	Fast	represents	Surface runoff
(ii)	Medium	.....	Interflow
(iii)	Slow	.....	Deep groundwater component

The total snowmelt and rain input to each watershed band is sub-divided on a priority basis. First priority is the satisfying of of any soil moisture deficit, a deficit which arises continuously because of evaporative demand.

Each component of runoff undergoes delay before reaching the outflow point of the watershed. These delays, or time distribution of runoff, are achieved by using unit hydrograph convolution. Various time distribution processes can be thought of in terms of cascades of linear reservoirs.

Estimation of evapotranspiration is divided into three processes:

(i) Potential evapotranspiration is estimated for the lowest meteorological station in the watershed (EVAP)

$$EVAP = K*MK*(10^{\frac{TX-14.5}{64}} - 10^{\frac{TN-14.5}{64}}) \qquad ...(15)$$

where

K = Evaporation constant

MK = Factor which is specified as a monthly factor to take into account seasonal variation of EVAP

$(10^{\frac{TX - 14.5}{64}} - 10^{\frac{TN-14.5}{64}})$ = Variation of the saturated vapour pressure curve as a function of maximum and minimnum temperature. Minimum temperature is a good approximation of the dew point temperature.

(ii) The EVAP value is then distributed to each elevation mid-band level and is designated by PET

$$PET(L) = \frac{TX(L) - 32}{TEX-32} * EVAP \qquad ...(16)$$

where

$TX(L)$ = Maximum temperature

TEX = Minimum temperature for meteorological station No.1.

(iii) PET values are then used in conjunction with the calculated soil moisture deficit to yield an actual evapotranspiration value for each band(AET)

Before any runoff can occur, other than fast runoff, the soil moisture deficits must be satisfied. While soil moisture deficits are being satisfied by incoming water, there is also an evaporative demand which is continually building up a deficit. On any given day, in any given elevation band, there will exist a specified potential evapotranspiration. The soil moisture deficit which exists in that band will represent the actual evapotranspiration capacity of that band.

$$AET = PET * 10^{(-BSD/AETGEN)} \qquad ...(17a)$$

AET = Actual evapotranspiration

PET = Potential evapotranspiration

BSD = Current value of the band soil moisture deficit

AETGEN = Specified constant which controls the rate at which BSD influences PET

For each day a new value of soil moisture deficit is computed. New value of BSD is:

$$BSD=BSD-PRN-BM=AET \qquad ...(17b)$$

where

PRN = Rain input

BM = Snowmelt input

This actual evapotranspiration demand will only influence the area of the watershed which is not impermeable.

2.2.4 Sacramento Model

In this model evapotranspiration is considered to take place from:

(i) Upper zone tension water(E_1)

(ii) Upper zone free water (E_2)

(iii) Balance of upper zone tension and free water storage

(iv) Portion of area which is pervious when the catchment is dry but becomes impervious as the tension water requirements are satisfied.

(v) Lower zone tension water

(vi) Balance of lower zone tension and free water storages

(vii) Portion of catchment area covered by riparion vegetation.

(viii) Portion of catchment area covered by streams and lakes

(i) Upper zone tension water(E_1)

The evapotranspiration from upper zone tension water takes place at the potential rate multiplied by the ratio of tension water contents to capacity:

$$E_1 = EDMD \frac{UZTWC}{UZTWM} \qquad \ldots (18)$$

where

EDMD=The evapotranspiration demand is computed from pan evaporation or meteorological variables. The potential evapotranspiration is multiplied by a factor representing the state of vegetation in the catchment, the catchments ability to transpire water from the ground to the atmosphere. It is also adjusted to correspond to the portion of the day comprising the computational time period.

(ii) Upper zone free water(E_2)

The evapotranspiration from upper zone free water is at the residual rate and is equal to RED or to UZFWC whichever is smaller, E_2 is substracted from UZFWC and RED is further reduced.

where

UZFWC = The quantity of water in storage at any time as upper zone free water.

RED = Remaining evapotranspiration demand

(iii) Balance of upper zone tension and free water storages:

The ratios:

$$\text{RATIO (T)} = \frac{\text{Tension Water Content}}{\text{Tension Water Capacity}} = \frac{UZTWC}{UZTWM} \quad \text{...(19)}$$

UZTWC = The quantity of water in storage at any time as upper zone tension water

UZTWM= The maximum amount of tension water which can be stored in the upper zone

and

$$\text{RATIO (F)} = \frac{UZFWC}{UZFWM} \quad \text{...(20)}$$

where

UZFWC = The quantity of water in storage at any time as upper zone free water

UZFWM = The maximum amount of free water which can be stored in the upper zone

If RATIO (F) >RATIO (T)

Free water is transferred to tension water in such quantity as to make the ratios equal.

But if RATIO<(F) RATIO (T)

No transfer of water takes place.

(iv) Evapotranspiration from portion of catchment area (ADIMP) which is pervious when the catchment is dry but becomes impervious as the tension water requirements are satisfied(E_6):

The evapotranspiration from ADIMP is equal to that taken from upper zone tension water, E_1, plus an additional increment. This increment is based on the amount by which the tension water in the area, ADIMP, exceeds the quantity which was in UZTW before the withdrawl E_1.

This excess=[ADIMC-(UZTWC+E_1)] ...(21a)

$$\text{or,} = (\text{ADIMC} - E_1 - \text{UZTWC}) \qquad \text{...(21b)}$$

This additional increment is equal to the ratio of this excess to the total tension water capacity multiplied by the remaining demand, RED.

Then,

$$E_6 = E_1 + \text{RED} \left[\frac{\text{ADIMC} - E_1 - \text{UZTWC}}{\text{UZTWM} + \text{LZTWM}} \right] \qquad \text{...(22)}$$

The quantity ADIMC is adjusted by subtracting E_6 and of course, E_6 may not exceed ADIMC. After the adjustment, E_6 is adjusted by multiplying it by the ratio of ADIMP to the total area. Mathematically this is equal to the parameter ADIMP.

(v) The evapotranspiration from lower zone tension water(E_3):

$$E_3 = \text{RED} \left[\frac{\text{LZTWC}}{\text{UZTWM} + \text{LZTWM}} \right] \qquad \text{...(23)}$$

where

LZTWC=Lower zone tension water content

(UZTWM+LZTWM) = Total tension water capacity

As in the upper zone, E_3 may not exceed LZTWC.

E_3 is subtracted from LZTWC and RED is further reduced.

(vi) Balance of lower zone tension and free water storage:

Ratios are computed in a manner similar to the upper zone:

$$RATIO\ [T] = \frac{LZTWC}{LZTWM} \qquad ...(24)$$

RATIO[F] = Involves both primary and supplemental free water and is the amount over and above that which is reserved

Reserved free water:

$$RFW = RSERV(LZFSM + LZFPM) \qquad ...(25)$$

where

RSERV=The portion of lower zone free water which is not available for transfer to lower zone tension water and subsequent evapotranspiration

LZFSM= The maximum amount of supplemental free water which can be stored in the lower zone

LZFPM= The maximum amount of primary free water which can be stored in lower zone

and $$RATIO\ [F] = \left[\frac{LZFPC + LZFSC - RFW}{LZFPM + LZFSM - RFW} \right] \qquad ...(26)$$

if RATIO[F]>RATIO[T]

a transfer is made such as to make the ratios equal.

The quantity to be transferred, DEL, may be computed as:

$$DEL = LZTWM \left[\left(\frac{LZFPC + LZFSC + LZTWC-RFW}{LZFPM + LZFSM + LZTWM-RFW} \right) \right]$$

$$- RATIO[T] \qquad ...(27)$$

if the quantity in lower zone supplemental free water storage, LZFSC is equal to or greater than DEL,the water is transferred from supplemental to tension. But if DEL> LZFSC, the remainder is taken from primary free water.

If RATIO[F] < RATIO[T], no transfer is made.

(vii) Evapotranspiration(E_4) from portion of catchment area(RIVA) covered by riparian vegetation.

Evaporation from RIVA is at the residual rate. It is equal to the remaining demand multiplied by RIVA.

$$E_4 = RED*RIVA \qquad ...(28)$$

(viii) Evaporating(E_5) from portion of catchment area(STLA) covered by streams and lakes.

Evaporation from STLA is at the potential rate and is equal to the demand multiplied by STLA:

$$E_5 = EDMD*STLA \qquad ...(29)$$

To determine the total evapotranspiration from the catchment, the individual terms are summed. But the quantities E_1,E_2 and E_3 must first be multiplied by the portion of the catchment area over which these processes take place. This is called PAREA and is given by:

$$PAREA=1-(ADIMP+ PCTIM) \qquad ...(30)$$

where ADIMP=Portion of catchment area which is continuously impervious

PCTIM= Portion of catchment area which is continuously impervious (water covered area plus impervious surface directly adjacent to channel system)

Then ,the total evapotranspiration:

$$EUSE=PAREA(E_1 + E_2 \quad E_3) + E_4 + E_5 + E_6 \qquad ...(31)$$

The quantities E_4 and E_5, the evaporation from streamflow and riparian vegetation area must be subtracted from the quantities of water in the channel. This may be done after the runoff and drainage computations for the computational time period have been made, and the evaporation n subtracted from the runoff volume before it is applied to the temporal distribution function.

2.2.5 USGS model

The potential evapotranspiration rate(e_p) is assumed to vary directly with the daily pan evaporation rate,

i.e. $$e_p = K_p\, e \qquad \ldots(32)$$

where K_p = Constant of proportionality

During periods of no rainfall, the moisture content of the unsaturated zone will change due to redistribution of moisture in the soil profile or due to evapotranspiration. The moisture content m_o will continue to increase upto the value of field capacity m_c when vertical drainage d takes place from saturated zone. The moisture content will decrease if the evapotranspiration demand is not satisfied by the moisture supply from the saturated zone.

<u>when $d > 0$</u>

(i)

$$m_o(t + \Delta t) = m_o(t) + d\,\Delta t \qquad \ldots(33a)$$

when $m_o(t+\Delta t) < m_c$

(ii)

$$m_o(t + \Delta t) = m_o(t) = m_c \qquad \ldots(33b)$$

otherwise

<u>when $d = 0$</u>

(i) $$m_o(t + \Delta t) = m_o(t) - e_p * \Delta t \qquad \ldots(33c)$$

when $m_o(t) > e_p * \Delta t$

(ii) $$m_o(t+\Delta t) = 0 \text{ otherwise} \qquad \ldots(33d)$$

where $$e_p * \Delta t = e_p \Delta t-[i(t) - i(t + \Delta t)] \qquad \ldots(33e)$$

i= accumulated infiltration volume in wetted soil column during a period.

2.2.6 Leavesley model

This model considers an empirical relationship relating potential evapotranspiration(PET) with mean daily temperature and daily solar radiation as developed by Jenson and Haise(1963). The method was then modified by Jenson et al(1969) to better account variation in PET caused by changes in the aerodynamics of plant canopies,humidity and elevation. Basic equation used to computed PET is:

$$PET = CT * (TAVF - CTX) * RIN \qquad ...(34)$$

where

PET = Potential evapotranspiration in inches of water

CT = An air temperature coefficient and model parameter which is a constant for a given area

TAVF = Mean daily air temperature in ^{o}F

CTX = An air temperature constant for a given area in ^{o}F

RIN = Daily solar rediation expressed in inches of evaporation

CT and CTX values can be determined by calibration where accurate evapotranspiration data are available, if no data exists then they can be estimated empirically.

For forested area, CT is computed as:

$$CT = (C1 + (13.0 * CH)^{-1} \qquad ...(35a)$$

where

C1 = Elevation correction term

CH = Humidity index

C1 is computed as:

$$C1 = 68 - (3.6*(E/1000)) \qquad ...(35b)$$

where

E = Median watershed elevation in feet.

CH is computed by:

$$CH = (37.5 \text{ mm Hg})/(e_2 - e_1) \qquad (35c)$$

$$= (50 \text{ mts})/(e_2 - e_1)$$

where

e_2 = Saturation vapour pressure in mb for mean maximum air temperature for warmest month of the year

e_1 = Saturation vapour pressure in mb for mean minimum air temperature for warmest month of the year

This empirical procedure gives a reasonable estimate of the parameter CT,however, an optimum value for CT can be obtained for a given basin using the parameter fitting procedure:

CTX in ^{o}F is computed by

$$CTX = 27.5 - (0.25* (e_2-e_1))-(E/1000) \qquad ...(36)$$

Evapotranspiration values are computed monthwise.

Evapotranspiration value of a place varies depending on the condition whether it is snow free or snow covered.

2.3 Infiltration

Infiltration is defined as the entry of water from the surface into the soil profile. Infiltration is the key process at the land surface which must be carefully considered in models for describing the hydrology of a watershed. Water may infiltrate immediately from rainfall into the soil profile or it may flow into temporary storage and infiltrate later. Storage in the soil profile is large but direct infiltration into this storage occurs at relatively low rates. Delayed infiltration complements direct infiltration and occurs when water flows into temporary storages of limited capacity, such as surface depressions and soil fissures. This water will later infiltrate or evaporate. Horton (1931) defined infiltration capacity as the maximum rate at which a given soil in a given condition can absorb rain as it falls. It is the infiltration capacity of the soil that determines for a given storm, the amount and time distribution of rainfall excess that is available or runoff and surface storage. The rate of infiltration at which it occurs is influenced by many factors such as the type and extent of vegetal cover, the condition of the surface crust, temperature, rainfall intensity, physical properties of the soil and water quality.

The interaction of the direct and delayed processes of infiltration during rainfall is of major importance. As rainfall begins, flow enters soil fissures, lossely packed surface soil and surface depressions. Rate of infiltration at the begining of rainfall is high. As the rainfall continues the soil moisture also continues to increase till it reaches saturation value and the rate of infiltration continues to decline till it attains almost a constant value f_c. The decrease is primarily due to reduction in the hydraulic gradients at the surface but may also be affected by other factors such as surface sealing and crusting. The constant f_c is generally assumed to be equal to the saturated hydraulic conductivity K_o, but will actually be somewhat less than K_o due to entrapped air. In most cases f_c is more accurately approximated by K_s, the hydraulic conductivity at residual air saturation.

If water is applied at a constant rate, R to the surface the infiltration rate curve and the water content(θ) profile are shown in the figure 10 and 11 respectively. Infiltration at the early stages of events will be equal to R and is limited by the application rate rather than soil conditions and properties. These are shown by points 1,2 and 3 in infiltration rate curve and curves 1,2,and 3 as water content profiles.

θ = Initial soil moisture

θ = Soil moisture at saturation

As long as the application rate is less than the infiltration

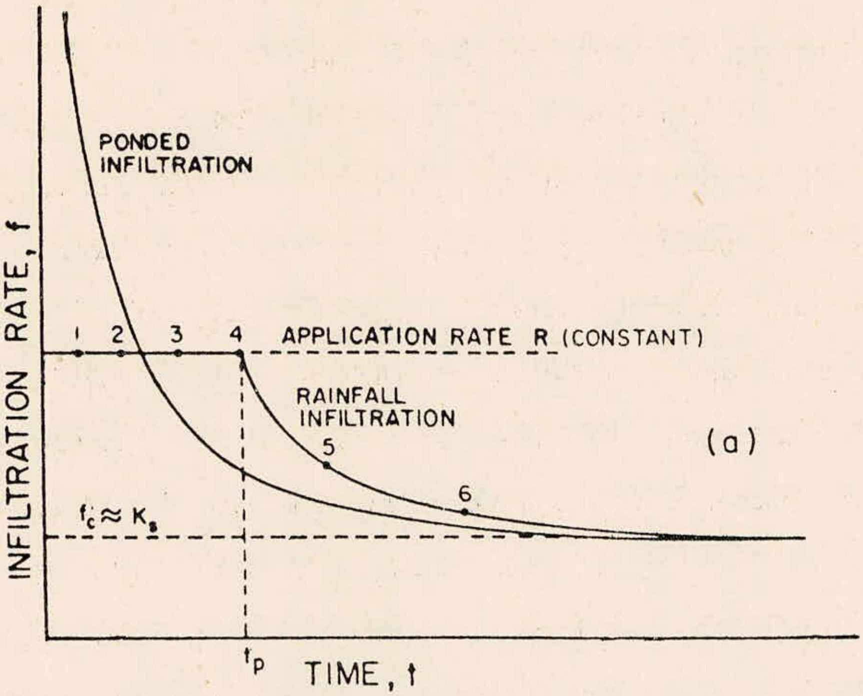

FIGURE - 10 INFILTRATION RATE CURVE
(Adopted from Haan 'Small Catchment Hydrology)

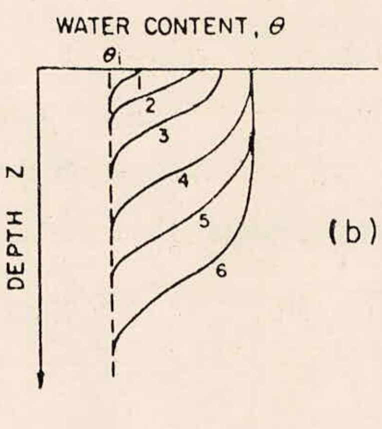

FIGURE 11 - WATER CONTENT PROFILES AT DIFFERENT TIMES
(Adopted from Haan 'Small Catchment Hydrology')

capacity, water will infiltrate as fast as it is supplied and the infiltration rate will be controlled by the application rate. At point 4, the infiltration capacity is equal to the application rate. After this, the infiltration capacity will be less than R. This is shown by points and 5 and 6 and corresponding water content curve 5 and 6. At this stage, water supplied in excess of the infiltration capacity will become available for surface storage and/or runoff. If the rate of infiltration is measured in a basin at different places by infiltrometers, then areal variation of infiltration capacities can be plotted in graphical form as shown in figure 12.

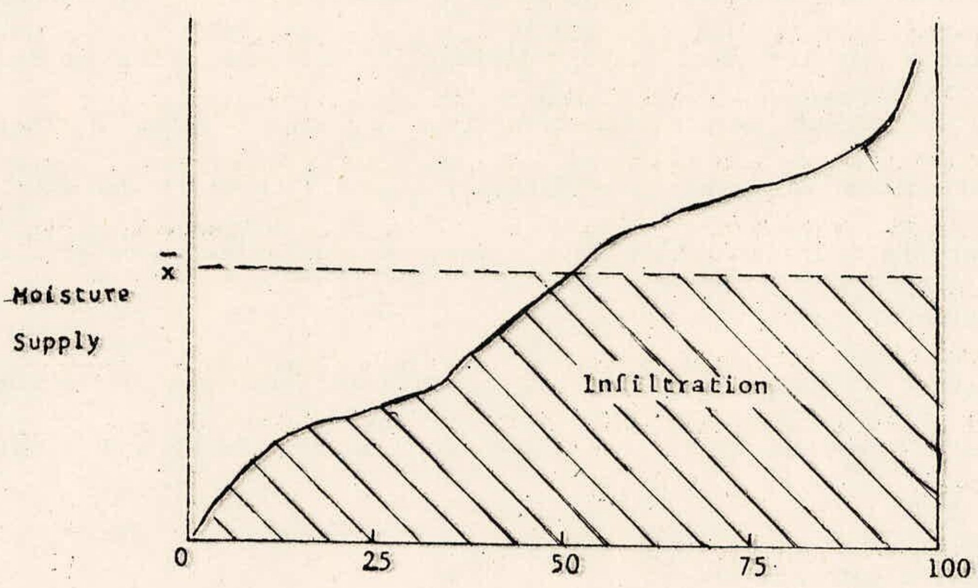

FIGURE - 12 APPROXIMATE LINEAR VARIATION OF INFILTRATION CAPACITY OVER A WATERSHED
(Adopted from SWM IV)

The plot shows % of area, with an infiltration capacity equal to or less than the indicated value. Since infiltration capacity changes with time, the curve is time dependent and will be applicable for some short time interval. X is the mean moisture supply value(in depth) during some time interval. Total volume of infiltration will increase as the moisture supply increases. The remaining area between the moisture supply line and infiltration capacity curve represents the volume of water that is free to move toward stream channel as overland flow.

2.3(a) Factors affecting infiltration

Infiltration rate depends on the following factors:

(i) Soil properties

The influence of shapes of soil and the hydraulic conductivity on infiltration was studied by Hanks and Bowers(1963). They showed that variations in the soil water diffusivity at low water content had negligible effect on infiltration from a ponded water surface. However, variations in either the diffisuvity or soil water characteristic at water contents near saturation have a very strong influence on predicted infiltration.

Infiltration rate and cumulative infiltration rate variations for different types of soils are shown in the figure 13 and figure 14 respectively.

(ii) Initial water content(θ_1)

This is one of the important factors that influences infiltration of water into the soil profile. Infiltration rates are high for drier initial conditions but the dependence on initial water content decreases with time. Infiltration rates are higher at low initial water contents because of higher hydraulic gradients and more available storage volume.

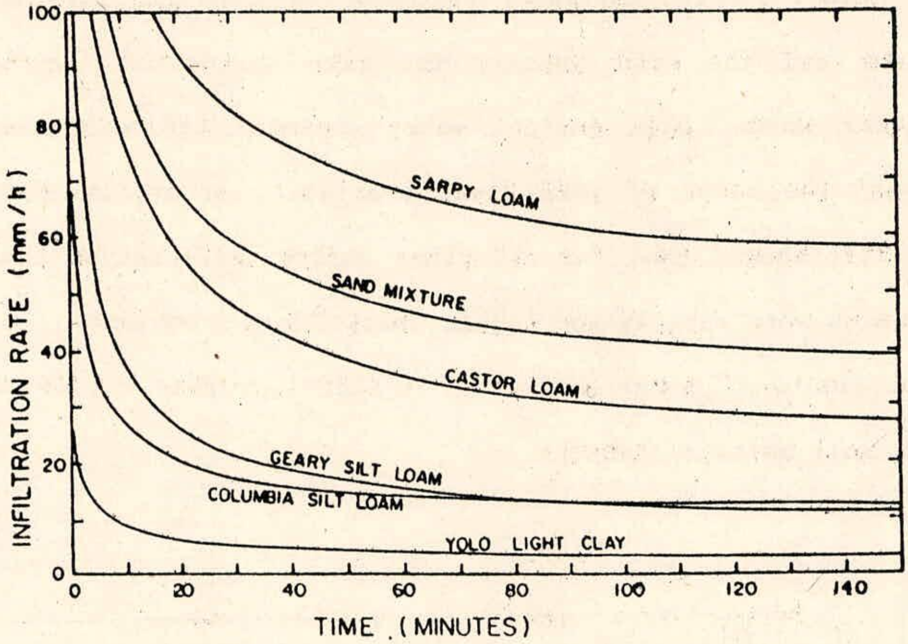

FIGURE 13 - INFILTRATION RATE CURVE OF SOME SOILS
(Adopted from Hann'Small Catchment Hydrology')

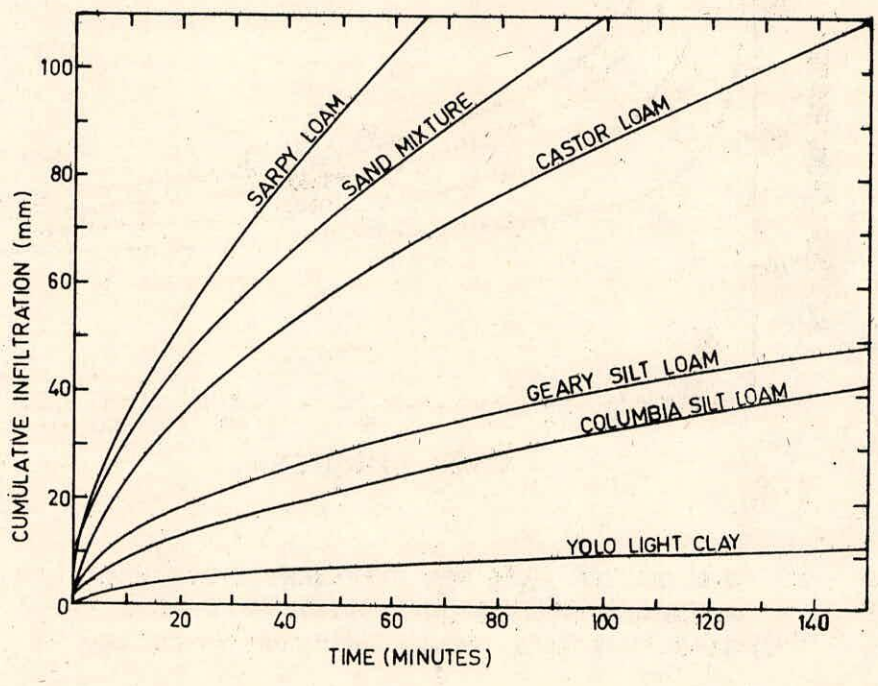

FIGURE 14- CUMULATIVE INFILTRATION RELATIONSHIP FOR THE SOILS
(Adopted from Hann 'Small Catchment Hydrology')

If infiltration is allowed to continue indefinitely, the infiltration rate will eventually approach K_s regardless of the initial water content. The higher the initial water content, the lower the initial infiltration rate and the more quickly the rate approaches the asymptote K_s. In other words, high initial water contents reduce the effective porosity and the range of pore sizes available for infiltrating water. Phillips(1957) showed that for all times during infiltration the wetting front advances more rapidly for higher initial water content.

The figure 15 shows different infiltration rate curves depending on initial soil moisture content.

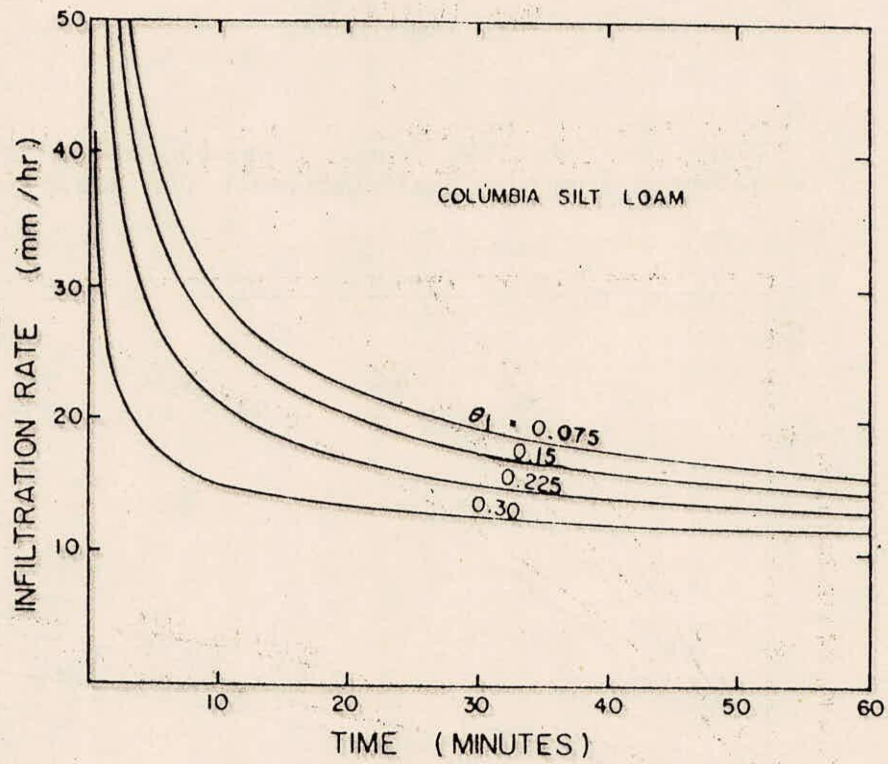

FIGURE - 15 INFILTRATION RATE FOR DIFFERENT SILTY LOAMY SOIL FOR DIFFERENT INITIAL SOIL MOISTURE CONTENT
(Adopted from Hann 'Small Catchment Hydrology ')

(iii) Rainfall rates

Infiltration depends on rate of water application as well as soil conditions. If the rainfall rate R is less than K_s for a deep homogeneous soil, infiltration may confine indefinitely at a rate equal to the rainfall rate without ponding at the surface. The water content of the soil in this case does not reach saturation at any point but approaches a limiting value which depends on rainfall intensity. For soils with restricting layers, infiltration at $R < K_s$ will not always continue indefinitely without surface ponding. When the wetting front reaches the restricting layer, water contents above the layer will increase and surface ponding may result even though the rainfall rate is less than K_s of the surface layer. Whether or not surface ponding and runoff occurs under such conditions, infiltration depends on the soil properties of the restricting layer, its initial water content and lower boundary condition as well as the rate of drainage in the lateral direction. Detailed investigations of rainfall infiltration have been conducted by Rubin and Steinhardt(1963,1964) Rubin et al(1964) and Rubin(1966).

(iv) Surface sealing and crusting

The soil matrix or skeleton though generally is considered as rigid but actually the hydraulic properties at the soil surface may change dramatically during application of water. Such changes on the surface cover influences the rate of infiltration. Edward and Larson (1969) used the theory of soil-water movement to investigate the influence of surface of seal development on infiltration of water into a tilled soil.

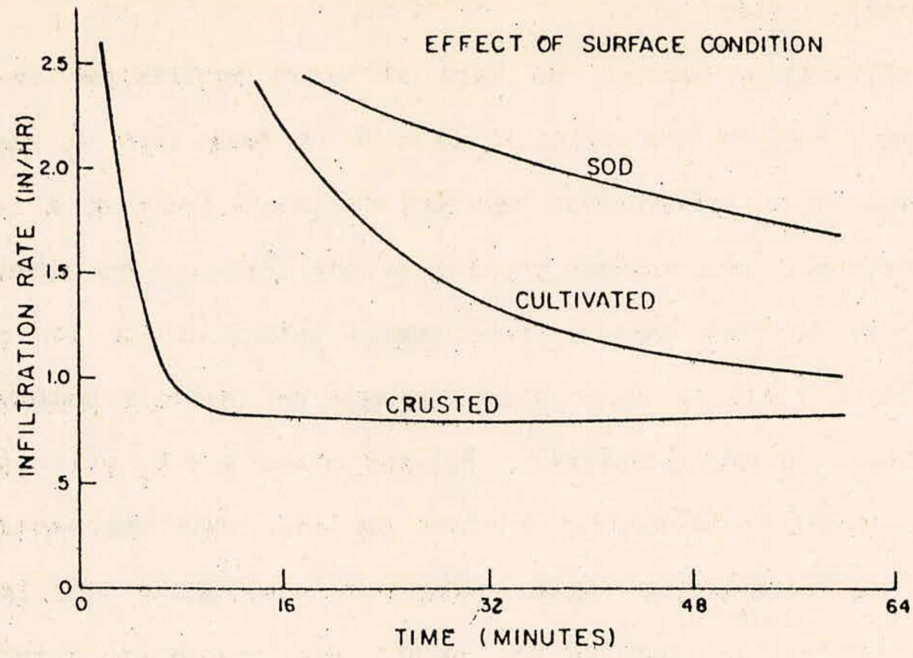

FIGURE 16- EFFECT OF SURFACE SEALING AND CRUSTING DUE TO RAINFALL IMPACT ON INFILTRATION RATE
(Adopted from Hann 'Small Catchment Hydrology')

(v) Layered soil

When water flows down through the layered soil, distribution of water content becomes discontinuous because of the difference in the soil water characteristics of the two soils. For a coarse soil layer over a fine soil, infiltration proceeds exactly as for a coarse soil alone until the wetting front arrived at the boundary between the two layers. Then the progress of wetting front slows down a positive pressure head develops in the top layer and the infiltration rate approaches that predicted for fine soil alone. Whisler and Klute(1966) worked on infiltration through different layered soil.

(vi) Movement and entrapment of soil air

Generally constant air pressure is assumed under which infiltration takes place. This assumption is usually justified by the fact that viscosity of air is small relative to that of water and air can escape through large pores that remain partially open during infiltration. While these assumptions may hold in some instances, there are numerous cases where air is trapped by infiltrating water causing an air pressure build-up in advance of the wetting front and a reduction of the infiltration rate. Entrapment of a certain amount of air within individual soil pores usually occurs during infiltration whether or not there is an air pressure build up in advance of the wetting front. Pores containing entrapped air are unavailable for the transport of water and result in a hydraulic conductivity K_s rather than K_o.

The difference in K_s and K_o depends on the number and size of pores blocked by entrapped air. Wilson and Luthin (1963) suggested that entrapment occurs primarily in larger pores. Slack(1978) presented a method for evaluating K_s for different amounts of air trapped in large pores.

2.3(b) Approximate infiltration models

Infiltration can be calculated by solving the governing differential equations under initial and boundary conditions using numerical methods. But procedure of such numerical solution of differential equations are elaborate and usually expensive due to computational requirements. Moreover, it is difficult to obtain required soil property data for such solution. So, such elaborate procedures are rarely used in practice. The numerical prediction methods are extremely valuable in analysing the effects of various factors of the infiltration process but due to above reasons the method is generally not applied in modeling watershed hydrology.

Simplified algebric equations in terms of time and soil parameters are attempted to calculate infiltration for field problems. Some of the approximate models have been developed by applying the principles governing soil water movement for simplified boundary and initial conditions. The parameters in such models can be determined from soil water properties, when they are available. Other models are strictly empirical and the parameters must be obtained from measured infiltration data or estimated using more approximate procedures.

(i) Kostiakov equation

One of the simplest infiltration equations was proposed by Kostiakov(1932):

$$f_p = K_k\, t^{-\alpha} \qquad \ldots(37)$$

where

$\bar{f}_p$ = Infiltration capacity

t = Time after infiltration starts

K_k and α = Constants which depend on the soil and initial conditions

The parameters of this equation have no physical interpretation and must be evaluated from experimental data.

(ii) Horton equation

Horton(1939,1940) presented a three parameter infiltration equation which may be written as:

$$f_p = f_c + (f_o - f_c)e^{-Kt} \qquad \ldots(38)$$

where

f_c = Final constant rate of infiltration capacity

f_o = Initial rate of infiltration capacity

f_p = Infiltration capacity at any time t

K = A constant dependent primarily upon soil and vegetation

t = Time from start of rainfall

e = Base of natural logarithm

Equation(38) parameters are usually evaluated from experimental infiltration data.

(iii) **Philip** equation

Philip infiltration equation from a ponded surface into deep homogeneous soil is expressed as:

$$f_p = \frac{S}{2} t^{-1/2} + C_a \qquad \ldots(39)$$

where

S = Sorptivity

C_a can be evaluated numerically using procedures given by Philip if the soil properties $D(\theta)$ and $h(\theta)$ are known.

Where $D(\theta)$ = Soil water diffusivity

$h(\theta)$ = Soil water characteristics

A regression fit to experimental data will tend to give $C_a=f_c$ Young(1968) showed that C_a could be approximated as $C_a = 2K_s/3$ and $S=(2MK_s S_f)^{1/2}$ where

M = fillable porosity

$= (\theta_s - \theta_1)$

S_f = Effective suction at the wetting front

(iv) Holtan equation

Holtan (1961) gave a empirical equation based on a storage concept as:

$$f_p = GI.a.SA^{1.4}+f_c \qquad \ldots(40a)$$

where

GI=Growth index of crop in percent of maturity

a=An index of surface connected porosity which is a function of surface conditions and the density of plant roots.

f_c= Constant or steady state infiltration rate which is estimated from the hydrologic soil group

In this method the initial soil water content θ_1 is measured or predicted and then the initial available storage SA is computed as:

$$SA=(\Theta_s-\Theta_1)d \qquad \ldots(40b)$$

where

d = Surface layer depth

The infiltrated water will reduce the value of SA, but this value will recover in part during the same time, due to drainage from surface layer at a rate of f_c upto the limit SA_o and by evapotranspiration(ET) through plant. That is, after a period of time Δt:

$$SA=SA_o - F + f_c\Delta t + ET\,\Delta t \qquad \ldots(40c)$$

where

F = The amount of infiltration during Δt

(v) Green-Ampt model

This is an approximate model based on Darcy's Law as proposed by Green and Ampt(1911). The original equation was derived for infiltration from a ponded surface into deep homogeneous soil with a uniform initial water content. Water is assumed to enter the soil as slug flow resulting in a sharply defined wetting front which seperates a zone that has been wetted from a totally unwetted zone. Application of Darcy's law gives the Green and Ampts equation as:

$$f_p = K_s(H_o+S_f+L_f)/L_F \qquad \ldots(41)$$

where

K_s = Hydraulic conductivity of the transmission zone

H_o=Depth of water ponded on the surface

S_f = Effective suction at the wetting front

L_F = Distance from surface to the wetting front

The Green-Ampt model assuming slug flow with a sharp wetting front between the infiltrated zone is shown in the figure 17.

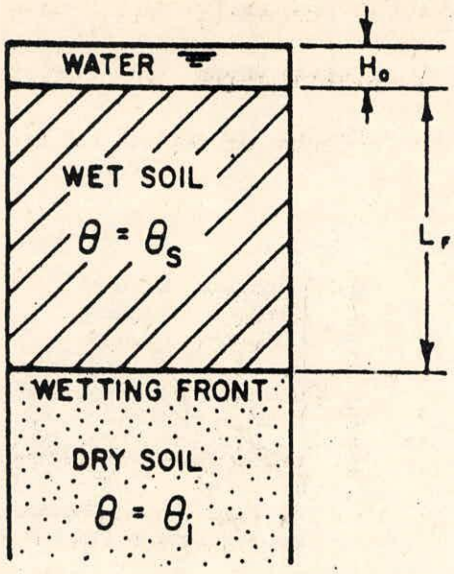

FIGURE 17- THE GREEN-AMPT MODEL ASSUMING SLUG FLOW WITH A SHARP WETTING FRONT BETWEEN THE INFILTRATED ZONE

2.3.1 Stanford Watershed Model (SWM-IV)

In this model infiltration is accounted continuously in terms of two components:

(i) Direct infiltration into soil profile, and

(ii) Delayed infiltration from temporary storages such as depression storages etc.

The moisture available is subject to operations that govern direct flows into long-term lower zone and ground water storages. That fraction of water determined to be remaining in surface detention after calculation of direct infiltration is disposed of according to operation of upper zone storages. The upper zone is designed to simulate the diversion of overland flows into depression storage, soil fissures and disturbed or dry surface soil. None of the soil moisture storages have fixed capacities. Addition to and losses from storages are determined from continuous dimensionless storage ratios to avoid discontinuous model response. The moisture supply available for infiltration in any time interval includes water in transient storage in overland flow.

Infiltration capacity will vary throughout the watershed and the cumulative distribution of infiltration capacity was introduced to simulate the effects of these variations on runoff and infiltration.

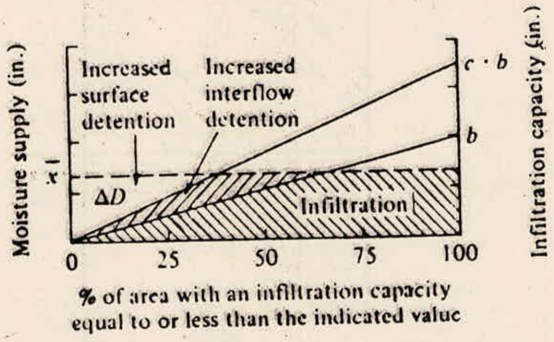

FIGURE 18 - ASSUMED LINEAR AREAL VARIATION OF INFILTRATION CAPACITY OVER A WATERSHED AS CONSIDERED IN SWM IV MODEL

The cumulative frequency distribution of infiltration capacity is assumed to be linear from zero to a maximum value as shown in the figure 18. Infiltration capacity is broken into two regions, one for lower zone and ground water storage, the other for interflow. In the region shown below the line 0 to b, all infiltrated water is assumed to move into the lower zone and groundwater storages. The region shown in between b and c,b is assumed to contribute to interflow. Thus, the tendency for infiltrating water to become interflow is assumed to be proportional to the local infiltration capacity.

Reaction of a watershed to a moisture supply X is shown in the figure 17.

where

$\bar{X}$ = Moisture supply available for infiltration

b = Maximum infiltration capacity

c= The parameter that controls the amount of water detained during the time increment.

The value of b and c depend on soil moisture ratio ($\frac{LZS}{LZSN}$), CB and CC.

CB= The index that controls the rate of infiltration and depends on the soil permeability and the volume of moisture that can be stored in the soil. The index varies from 0.3 to 1.2

CC= The parameter signifies an input value that fixes the level of interflow relative to the overland flow.

LZS = Current soil moisture storage in the lower zone

LZSN = A nominal storage level normally set equal to the median value of the lower zone storage(inches)

2.3.2 USGS model

In this model infiltration loss component is calculated using a modification of a method suggested by Philip (1954).

The Phillip equation:

$$\frac{di}{dt} = K\left[1+ \frac{P(\bar{m}-mo)}{i} \right] \qquad \ldots(42a)$$

where

i= Accumulated infiltration volume in wetted soil column during period$(t-t_o)$

K=Capillary conductivity of soil

P = Capillary pressure(suction) at wetting front in soil column

m_o=Initial moisture content of soil column at time t_o,and

$\bar{m}$= Moisture content,uniformly distributed through wetted column at time t.

The term $P(\bar{m}-m_o)$ is assumed to decrease linearly from a maximum$(r\ P_s)$,at the wilting point of the soil$(m_o=0)$ to a minimum P_s, at the field capacity of the soil$(m_o=m_c)$

Thus,

$$P(\bar{m}-m_o)\ rP_s-P_s(r-1)\frac{m_o}{m_c} \qquad \ldots(42b)$$

expressing

$$P_m = P(\bar{m}-m_o) \qquad \ldots(42c)$$

$$= rP_s-P_s(r-1)\frac{m_o}{m_c} \qquad \ldots(42d?$$

$$\frac{di}{dt} = K\left(1+\frac{P_m}{i}\right)$$

The four parameters of the soil K, P_s, m_c and r are defined by equation (42 c and 42d).

Infiltration occurs at varying rates over a basin, but equation (42a) describes infiltration at a point. Following a scheme of Crawford and Linsley, Dawdy, Lichty and Bergmaun schematically accounted for areal variability of infiltration as shown in figure 19 and avoided threshold effects.

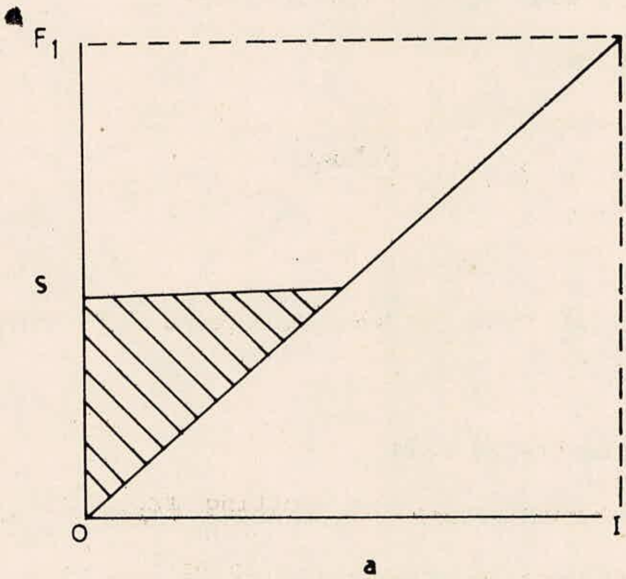

FIGURE 19 - AREAL VARIABILITY OF INFILTRATION CONSIDERED IN USGS MODEL
(Adopted from U S G S Model User's Manual)

With a varying infiltration rate over the area, the rainfall excess R_e during Δt is:

$$R_e = \begin{cases} \Delta t \; 1/2 \; \frac{S^2}{F1} & \text{when } S < F1 & \ldots(43a) \\ \Delta t \left[S - \frac{F1}{2} \right] & \text{Otherwise} & \ldots(43b) \end{cases}$$

The cumulative infiltration $i(t+\Delta t)$ at the time $t+\Delta t$ is:

$$i(t+\Delta t) = i(t) + t(S-R_e) \qquad \ldots(44)$$

During a period of uninterrupted rainfall the antecedent moisture content m_o at the start of rainfall is assumed to remain constant as the wetting front advances. During periods of no rainfall the accumulated infiltration will diminish due to evapotranspiration and vertical drainage.

2.3.3 SSARR model

In this model effect of infiltration into runoff is taken into consideration by soil moisture index and base flow infiltration index.

Soil moisture determined as time variable index of runoff effectiveness determine, in part, the amount of precipitation which contributes to runoff.

The processes(flow chart) that converts moisture input into runoff is shown in figure 20.

In this model rainfall input is divided into:

(i) Runoff

(ii) Soil moisture increase

(iii) Percolation into the groundwater system

(iv) Evapotranspiration losses

One of the most important parameter that affects the runoff hydrograph is the Soil Moisture Index(SMI). The SMI-Runoff Percent(ROP)

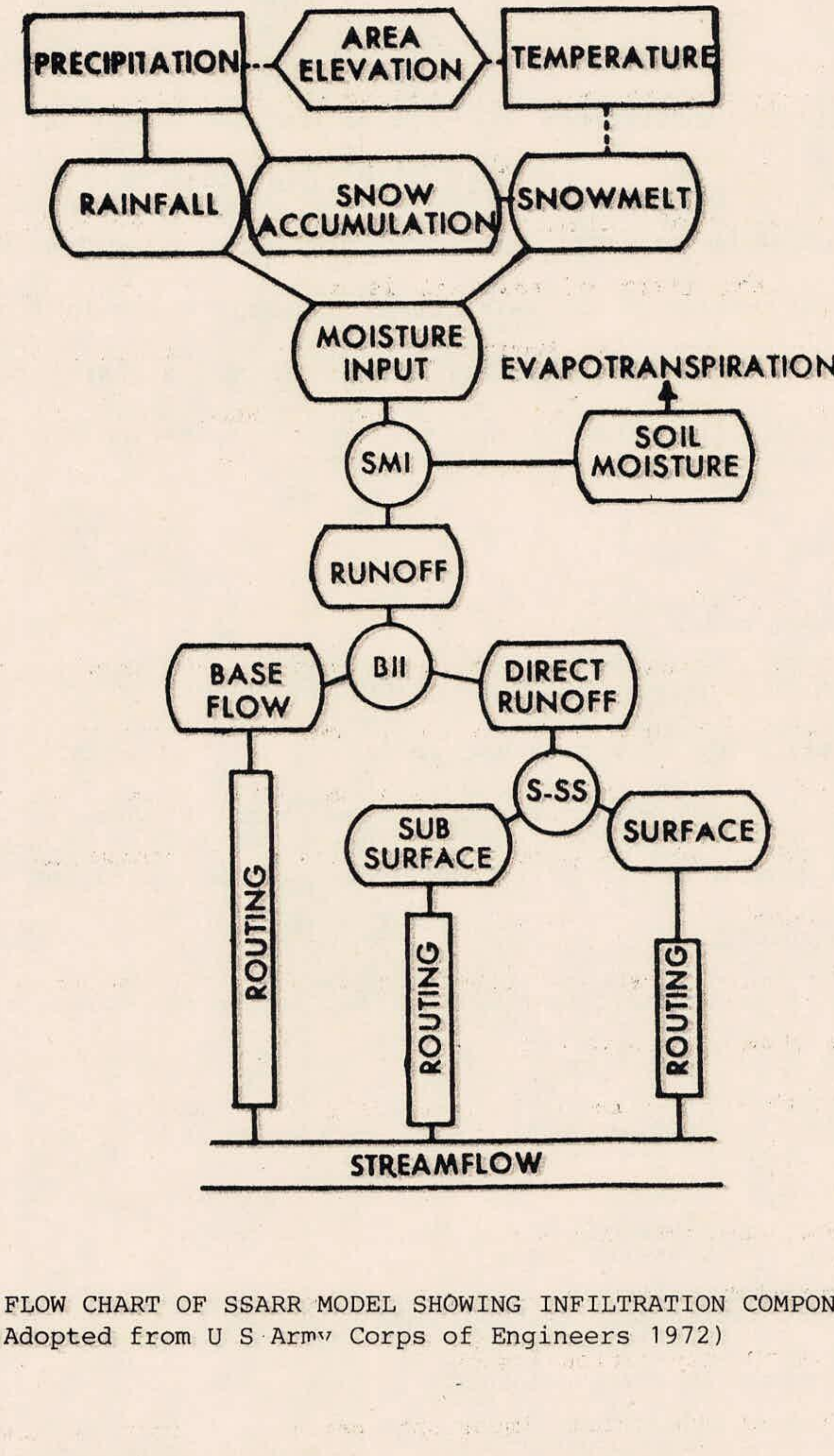

FIGURE- 20 FLOW CHART OF SSARR MODEL SHOWING INFILTRATION COMPONENT
(Adopted from U S Army Corps of Engineers 1972)

relationship as shown in figure 21 determines to a large extent, the volume of runoff and also affects the shape of hydrograph

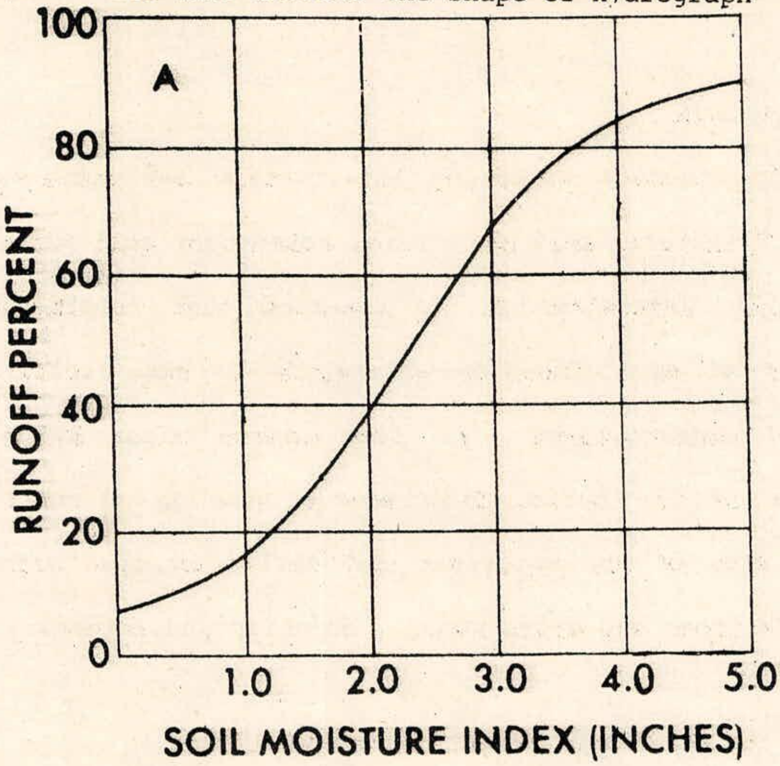

FIGURE 21- SMI-RUNOFF PERCENT(ROP) RELATIONSHIP(AS CONSIDERED IN SSARR MODEL)(Adopted from US Army Corps of Engineers 1972)

Baseflow Infiltration Index(BII) is used to seperate baseflow from observed stream flow volume.

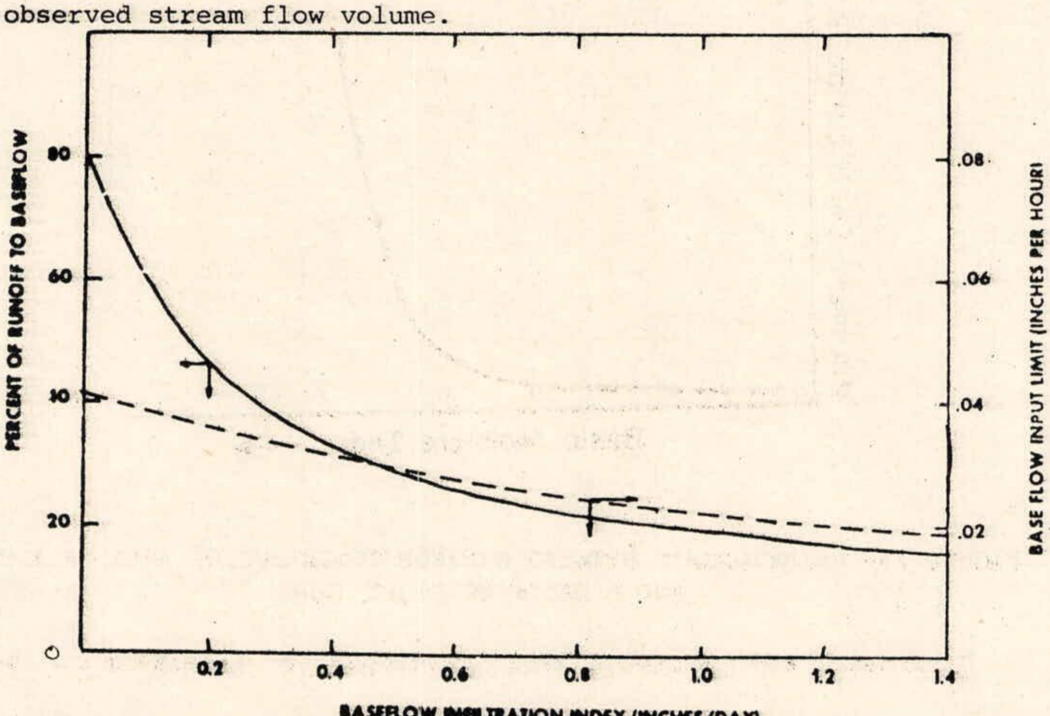

FIGURE 22 - VARIATION OF BASEFLOW PERCENT WITH BASEFLOW INFILTRATIO

Variation of baseflow percent with baseflow infiltration index is shown in figure 22.

2.3.4 Leavesley model

The major factors affecting infiltration and subsequent surface runoff are soil texture, soil structure, antecedent soil water conditions and water input intensity. It is observed that infiltration on the major portions of most forested watersheds is not limiting and that surface runoff contributions come from source areas lying along the stream courses of the basin. This source area is a small percentage of the total area of the watershed and varies in size with antecedent soil water conditions and storm amount, duration and intensity.

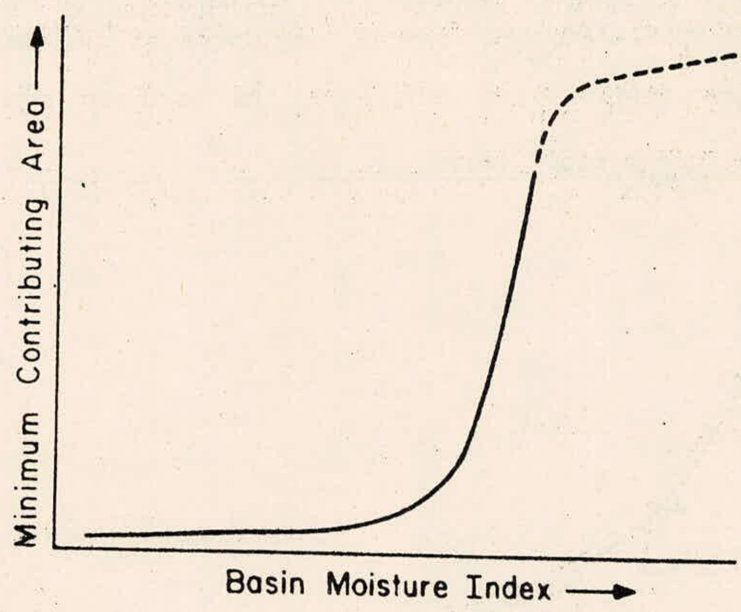

FIGURE 23- RELATIONSHIP BETWEEN MINIMUM CONTRIBUTION AREA OF A BASIN AND A BASIN MOISTURE INDEX

Dickinson and Whitely(1970) developed a relationship between minimum contributing area of a basin and a basin moisture index(which was a function of soil water and storm amount) as shown in figure 23. Contributing area remains small until some moisture index threshold is reached after which contributing area increases rapidly to some upper limit imposed by the basin.

In this model the contributing area concept is used to calculate the volume of surface runoff from rainfall events which occur on snow-free HRU. The percent contributing area of an HRU is assumed to be a linear function of the amount available soil water stored in the upper soil zone(SMAV) at the time of rainfall and of a maximum percent contributing area factor which is defined by the HRU variable SCT. Value of SCT may vary from 3 to 85% depending on soil and vegetation conditions.

Surface runoff contributing area (CAP) expressed as a percent of the total HRU area for a given storm is computed by:

$$CAP = SCT*(SMAV/SMAX) \qquad ...(45)$$

Volume of surface runoff is then computed as:

$$\text{VOL. Surface Runoff} = \text{Rainfall}*CAP*\text{ Area of HRU} \qquad ...(46)$$

where

CAP = Surface Runoff Contributing Area

SCT=Maximum percent contributing area factor

SMAV=Soil water stored in the upper soil zone

SMAX= Maximum value of soil water in upper soil zone

Units are:

depths in inches

area in acres

volume in acre-inches

The volume of surface runoff is removed from the effective rainfall reaching the soil surface and the remaining rainfall is assumed to infiltrate the upper soil zone replenishing any existing soil water deficit upto SMAX.

2.3.5 UBC model

In this model two types of situations are considered. Normal situation is that when runoff from moderate intensity rain and snowmelt events can be considered to be controlled by soil moisture levels. Second situation is that when runoff from high intensity events is controlled by the rate at which water can infiltrate into the soil system and these infiltration rates are relatively independent of soil moisture levels. For these high intensity rain events some of the precipitation infiltrates into the soil system and is subjected to the normal soil moisture budgeting. Intense snowmelt rates do not appear to be adequate to produce'Flash' runoff where as total rainfall of lesser amount but of high intensity may exhibit flash flood.

i) Normal soil moisture budgeting

(1-FLASHR)*(1-PMXIMP)*PRN

FLASHR=Flash Share Parameter

PMXIMP= Impermeable percentage of the particular watershed elevation band.

PRN = Daily precipitation

ii) The portion of total daily precipitation which flashes off.

(1-(1-FLASHR)*(1-PMXIMP)*PRN

2.3.6 Sacramento model

In this model, basin is considered to comprise of two types of basic areas(1) a permeable portion of the soil mantle, and (ii) a portion of the soil mantle covered by streams, lake surfaces, marshes or other impervious material directly linked to the streamflows network. The permeable area produces runoff when rainfall rates are higher than infiltration rates, while the second area produces direct runoff from any rain.

In the permeable portion of the basin, the model visualizes an initial soil-moisture storage identified as upper zone tension. This must be totally filled before moisture becomes available to enter other storages. Tension water is considered as that water which is closely bound to soil particles. Upper zone Tension represents that volume of precipitation which would be required under dry conditions to meet all interception requirements and to provide sufficient moisture to the upper soil mantle so that percolation to deeper zones and sometimes horizontal drainage can begin.

2.3.7 USDAHL-74 model

In this model Holtan expression of infiltration capacity is used, which is expressed as:

$$f = a\, S_a^{1.4} + f_c \qquad \ldots(47)$$

where

f = Infiltration capacity in inches per hour

a=Infiltration capacity in inches per hour per inch of avilable storage

S_a = Available storage in surface layer

f_c=Constant rate of infiltration after prolonged wetting in inches per hour

Gardner found that water entering the soil under positive heads through larger pores spreads to the smaller pores both vertically and horizontally by capillary action. The equation above estimates this slow capillary movement as a constant (f_c). The other term($a.S_a^{1.4}$) is an empirically derived expression of flow rates due to positive heads. It represents the sum of products of velocities and cross sections in flow tubes.

The infiltration process is quite complicated and varies both in space and time. It also varies on the rainfall intensity. Different equations have been developed by different persons such as Kostiakov, Horton, Phillip , Holtan, Green and Ampts etc.

Different watershed models have adopted different approaches and approximations in calculating infiltration Stanford model has adopted continuous accounting of infiltration in terms of two components direct and delayed and infiltration capacity is considered to vary linearly with the area, this also depends on upper and lower zone soil moisture. USGS model has adopted Phillip equation. In SSARR model calculation is based on accounting of soil moisture index and base flow infiltration index. Holtan equation is adopted in USDAHL-74 model. Leavesley model has adopted calculation of volume of surface runoff by calculating surface runoff contributing area, HRU and rainfall. In UBC model calculation is divided into two parts namely normal soil moisture budgeting and total daily precipitation that flashes out. In Sacramento model calculation is based on dividing the total area into permeable and that covered by water bodies.

2.4 Overland Flow

Solution of the overland flow problem is contained in the continuity and momentum equations and written in their respective form

$$\frac{\partial y}{\partial t} + \frac{\partial q}{\partial x} = r \text{ (or re)} \qquad \ldots(48a)$$

and
$$q = \frac{c}{\eta} S_o^{1/2} y^{5/3} \qquad \ldots(48b)$$

where

y = depth of flow

q=discharge per unit width

r_e=(r-I), rain fall excess where infiltration is taken into consideration.

r= rainfall intensity

I=rate of infiltration

c=a constant

η = roughness coefficient

s_o= bed slope

x= distance from the boundary of catchment and

t=time, measured from the onset of rainfall excess

Many investigators have dealt with the problem of overland flow under uniform, excess rainfall either by using the method of characteristics(Behlke 1957,Henderson 1964, Wooding 1965, Morgali and Linsley 1965, Abbott 1966, Brakensiek1966) or by solving the mass balance equation by assuming a linear relationship between outflow and storage (Horton 1938; Izzard 1944). Wooding (1965) has dealt with the problem of overland under a constant uniformly distributed rainfall of finite duration with an analytical solution for a hydraulic model based on the method of characteristics for flow over a plane which is part of V-shaped catchment.

In reality rainfall intensity is not uniform in space and time. Moreover, interaction between overland flow and infiltration need to be considered since both processes occur simultaneously. The infiltration rate is quite high at the begining of rainfall and the rate decay exponentially with time. Moreover rate of infiltration varies depending on the type of soil, soil texture, vegetation cover etc. The variation in rates of infiltration allow overland flow in areas with low infiltration while preventing overland flow in other areas. During overland flow water held in detention storage remains available for infiltration. Surface conditions like heavy turf or mild slope restrict the velocity of ofoverland flow, reduces the velocity of outflow and thus increases the volume of surface detention and there by increases the time for infiltration. Thus simulation of the infiltration-overland flow processes requires continuous estimates of detention storage as well as continuous outflow rates from overland flow.

Overland flow can be calculated by different methods. Rigorous methods of numerical solution of the governing partial differential equations,a the continuity and momentum equations requires substantial amount of computer time. In a natural watershed there are areal variations in the amount of runoff moving in overland flow due to areal variations in infiltration rates. Average value of the parameters like length and slope of overland flow are taken into consideration for the basin or the accuracy may be increased by dividing the basin into number of segments. Moreover analysis depends on condition of overland flow whether it is in laminar or in tarbulent range.

2.4.1 Stanford watershed model(SWM IV)

In this model overland flow is considered to be in natural watersheds tend to collect and move along a preferred path. Continuous surface detention storage is calculated in the model. Since the volume of surface detention was successfully used as a parameter for the rate of discharge for overland flows in the laminar range, the volume of surface detention was also selected as the logical parameter in this model to relate surface detention with overland flow. Some useful approximation to natural behaviour has been made.

Amount of surface detention is calculated as:

$$D_e = \frac{0.000818\, i^{0.6}\ n^{0.6}\, l^{1.6}}{S^{0.3}} \qquad \ldots(49)$$

where

D_e = Surface detention in ft^3/ft

i = Supply rate(rainfall) in inches/hour

S=slope in ft/ft

l=length of overland flow in feet

n= Mannings roughness coefficient.

The rate of discharge from overland flow based on the Chezy-Mannings equation is

$$q = \frac{1.486}{n}\ Y^{5/3}\ S^{1/2} \qquad \ldots(50)$$

where

q=discharge in $ft^3/sec/ft$

Y=depth in feet at outlet point

The depth Y is related to surface detention storage at equilibrium by:

$$Y = 8/5\ D_e/L \qquad \ldots(51)$$

For other conditions some approximations are needed.

The discharge hydrograph from overland flow takes place as shown in the figure 25. When rainfall continues at position'a' an equilibirum profile is established which continues upto point'b' where rainfall stops. From position 'b' the recession starts and follows the path bc. If at any time, D is the surface detention storage then minimum

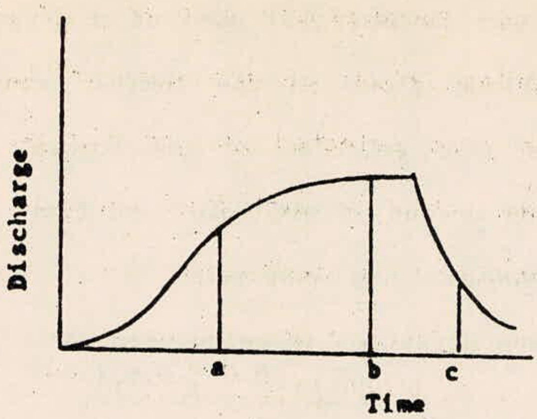

FIGURE 24 - DISCHARGE HYDROGRAPH FROM OVERLAND FLOW
(Adopted from SWM IV User Manual)

value of y will be(D/L).

The value of Y will be within the range

$$\frac{D}{L} \leqslant Y \leqslant \frac{8}{5} \quad \frac{De}{L}$$

The ratio (D/D_e) is used as an index to the distribution of water in the overland flow plane.

The model considers an empirical relationship found to be most suitable as

$$Y=D/L\ (1.0 + 0.6\ (D/D_e)^3) \qquad \ldots(52)$$

The rate of discharge from overland flow in ft^3/sec/ft is

$$q = \frac{1.486}{n} S^{1/2} \left[1.0 + 0.6 \left(\frac{D}{De} \right)^3 \right]^{5/3} \left(\frac{D}{L} \right)^{5/3} \qquad \ldots(53)$$

During recession the value of the ratio (D/D_e) is considered to be unity.

If the discharge q is expressed as inch per hour per unit area then

$$q = \frac{64200}{nL} S^{1/2} \left(\frac{D}{L} \right)^{5/3} \left[1.0 + 0.6 \left(\frac{D}{De} \right)^3 \right]^{5/3} \qquad \ldots(54)$$

The time at which the detention storage reaches an equilibrium is determined from

$$t_e = \frac{0.94\ L^{3/5}\ n^{3/5}}{i^{2/5}\ S^{3/10}} \qquad \ldots(55)$$

For each time intervalΔt, an end of interval surface detention D_2 is calculated from

$$D_2 = D_1 + \Delta D - \bar{q} \Delta t \qquad \text{....(56)}$$

where

D_1 = Initial value of storage at the begining of time interval

D= Surface detention storage added during t time

$\bar{q}$ = Average overland flow discharge that took place during Δt time interval

This gives the detention storage volume at any time.

2.5 Percolation

2.5.1 Stanford watershed model(SUM-IV)

The lower groundwater storage zone receives water from the net infiltration and from percolation. The portion of the upper zone storage which is not evaporated or transpired is proportioned to the surface runoff, interflow and percolation.

Percolation (upper zone depletion) from the upper zone to the lower zone occurs only when (UZS/UZSN) exceeds (LZS/LZSN).

When this occurs, the percolation rate in inch/hour isn determined from

$$PERC= 0.003(CB)\ (UZSN) \left[\frac{UZS}{UZSN} - \frac{LZS}{LZSN} \right]^3 \qquad \text{...(57)}$$

where CB= Index that control the rate of infiltration. The value ranges from 0.3 to 1.2 depending on the soil permeability and on the volume of moisture that can be stored in the soil

UZS,UZSN= Actual and nominal soil moisture storage amounts in the upper zone.

LZS,LZSN=Actual and nominal soil moisture storage amounts in the lower

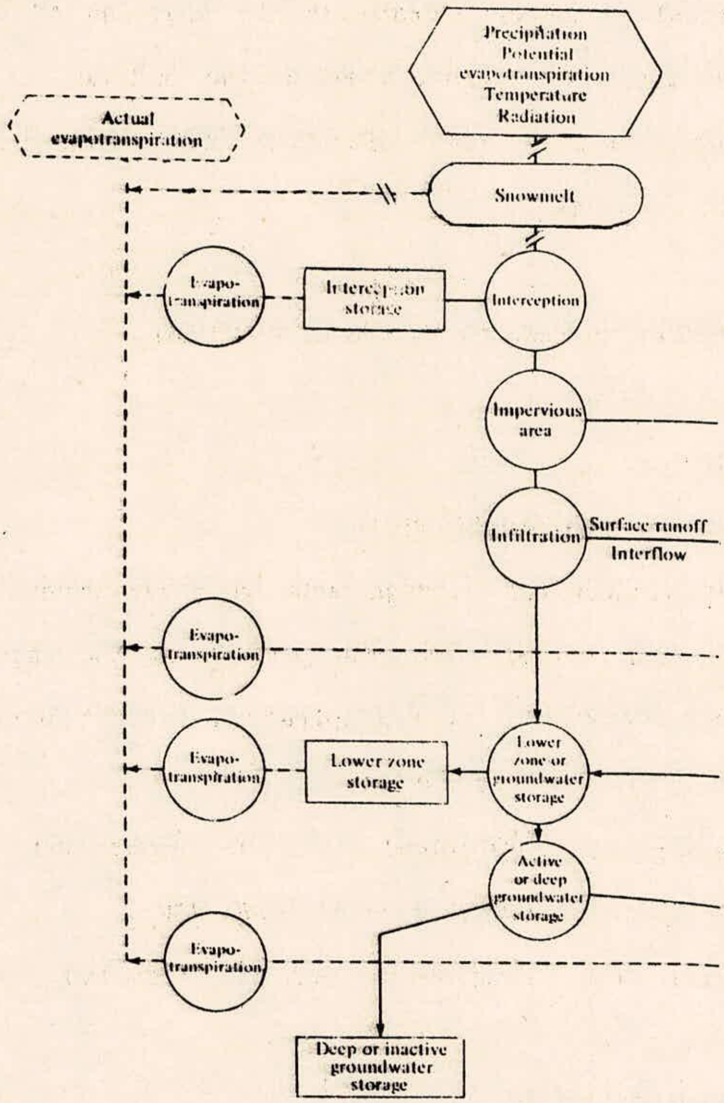

FIGURE - 25 FLOW CHART OF STANFORD WATERSHED MODEL FOR PERCOLATION COMPONENT
(Viessman et al 'Engg.Hydrology')

zone.

The nominal value of UZSN is approximately a function of watershed topography and cover and is always considered to be much smaller than the nominal LZSN value.

2.5.2 Sacramento model

In this model the mechanism of percolaotion is designed to correspond with observed characteristics of the motion of moisture through the soil mantle, including formation and transmission characteristics of the wetting front. Volume of water transfer from upper zone to lower zone is totally saturated, then percolation into the lower zone is limited to a value equal to that water which is deraining out of the lower zone. This limiting rate of drainage from the combined lower zone storage is expressed as:

$$PBASE = ((LZFM_S * LZSK) + (LZFM_p * LZPK)) \qquad ...(58)$$

where

$LZFM_S$=Lower zone free water maximum supplementary storage, which is the maximum storage capacity for faster draining base flow.

LZSK= Lower zone supplementary storage depletion coefficient.

$LZFM_p$=Lower zone free water maximum primary storage which is the maximum storage capacity for slower draining baseflow, and

LZPK= Lower zone primary storage depletion coefficient.

During dry period the percolation takes place at a much higher rate. Upper limit of percolation may be defined as

$$\text{Max.Percolation Capacity} = PBASE(1+Z) \qquad ...(59)$$

where

Z=multiplying value to increase percolation from the minimum PBASE to the maximum one.

Maximum percolation occurs when the upper zone is saturated and the lower zone is dry, then the percolation demanded by the lower zone can be stated as

Lower Zone Percolation Demand

$$=PBASE * [1+Z*f(\frac{\text{Lower Zone Deficiency}}{\text{Lower Zone Capacity}})] \quad ...(60)$$

It is assumed that the change in lower zone percolation demand is exponentially related to the ratio(Lower zone deficiency/Lower zone capacity), the equation for percolation demand with varying soil moisture is given by:

Percolation Demand

$$=PRASE[\ 1+Z\frac{\Sigma \text{ Lower Zone capacities less contents}}{\Sigma \text{ Lower Zone Capacities}}]^{REXP} \quad ...(61)$$

where

REXP=The exponent which defines the curvature in the percolation curve with change in the lower zone soil moisture deficiency.

But the actual percolation also depends on the supply of the available water, so the effective demand must be modified by a function of available supply of water from the upper zone in order to define the actual percolation.

Percolation = Percolation demand*

$$\frac{\text{upper zone free water content}}{\text{Upper zone free water capacity}} \quad ...(62a)$$

$$\text{or} \quad \text{Percolation} = PBOSE *[\ 1+Z(\frac{\Sigma \text{ Lower Zone Deficiency}}{\Sigma \text{ Lower zone capacity}})^{REXP}]\cdot$$

$$+ (\frac{UZFWC}{UZFWM}) \quad ...(62b)$$

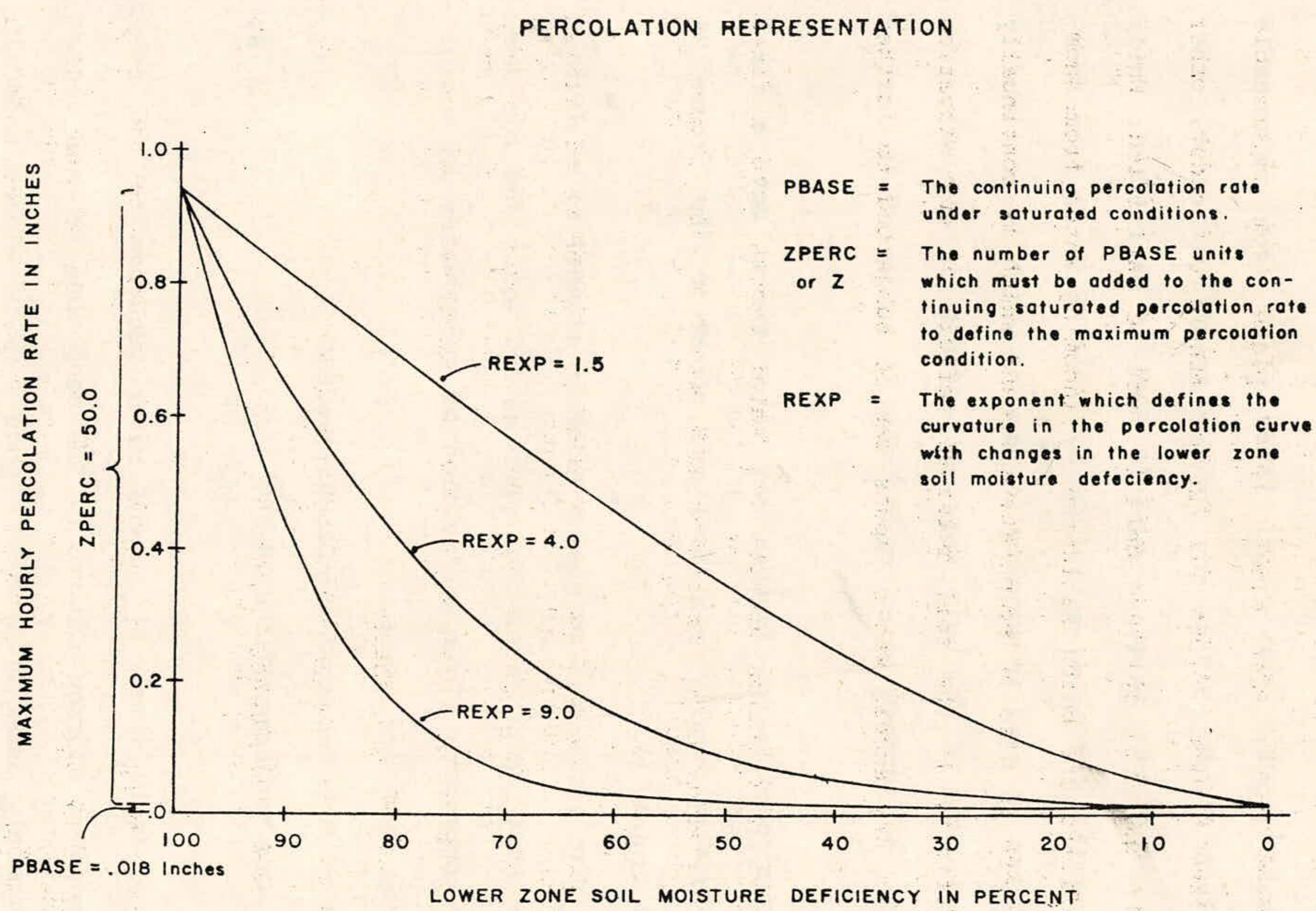

FIGURE - 26 PERCOLATION CURVES FOR SACRAMENTO MODEL FOR DIFFERENT LOWER ZONE SOIL MOISTURE DEFICIENCY
(Adopted from Burnas et al 1973)

2.5.3 UBC model

The snowmelt and rainfall input are divided in this model between evaporation loss and fast, medium, slow and very slow runoff. Soil moisture deficit is the main parameter which governs the subdivision of total watershed input. Fast runoff takes place from impermeable area and from flash floods. Before any further runoff can occur, other than fast runoff, the soil moisture deficit must be satisified. While soil moisture deficits are being satisfied by incoming water from snowmelt and rain, there is also an evaporative demand which is continually building up a deficit. If the soil moisture deficit of any elevation band reaches zero, any excess water inputs can be subjected to further priorities.

Ground water percolation accepts any water excess upto a fixed limit(GWPERC). Diagrametrically processes are shown in the figure 28 and flow chart in figure 29.

The water that percolates to ground water is assumed to be divided into two ground water components, the upper ground water and the deep zone groundwater components. This subdivision of groundwater is controlled by DZSHRE, the deep zone share.

So Upper Groundwater zone Recharge= (1-DZSHRE)*GWPERC ..(63)

Deep groundwater zone recharge=DZSHRE*GWPERC ...(64)

Where

DZSHRE=allocates a daily share of ground water recharge to the deep zone groundwater storage reservoir. The deep zone storage reservoir is commonly associated with a much slower release constant than the upper ground water reservoir.

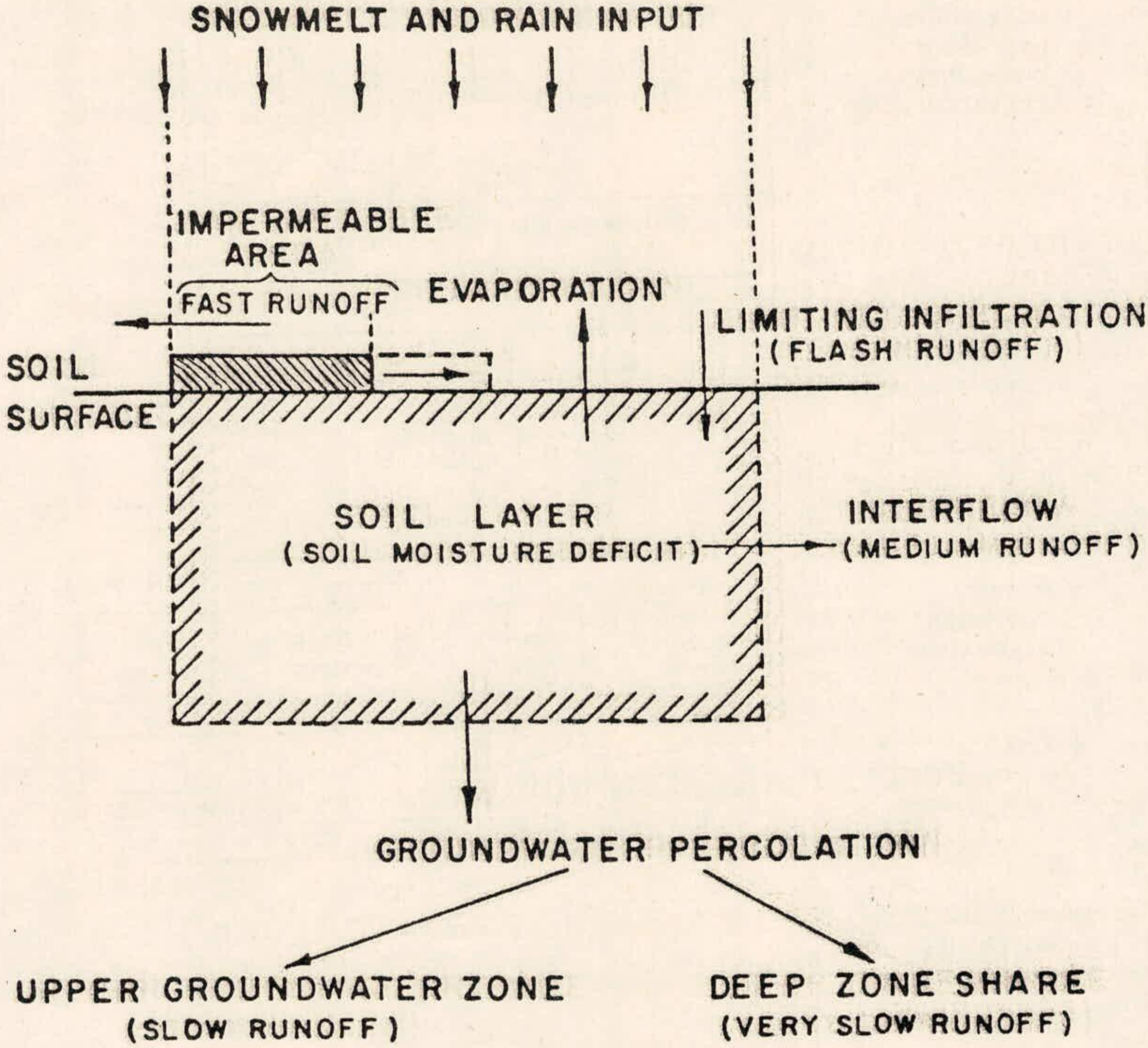

FIGURE 27- MODEL OF SOIL LAYER AND SUBDIVISION OF RUNOFF COMPONENTS OF UBC MODEL
(Adopted from UBC User's Manual)

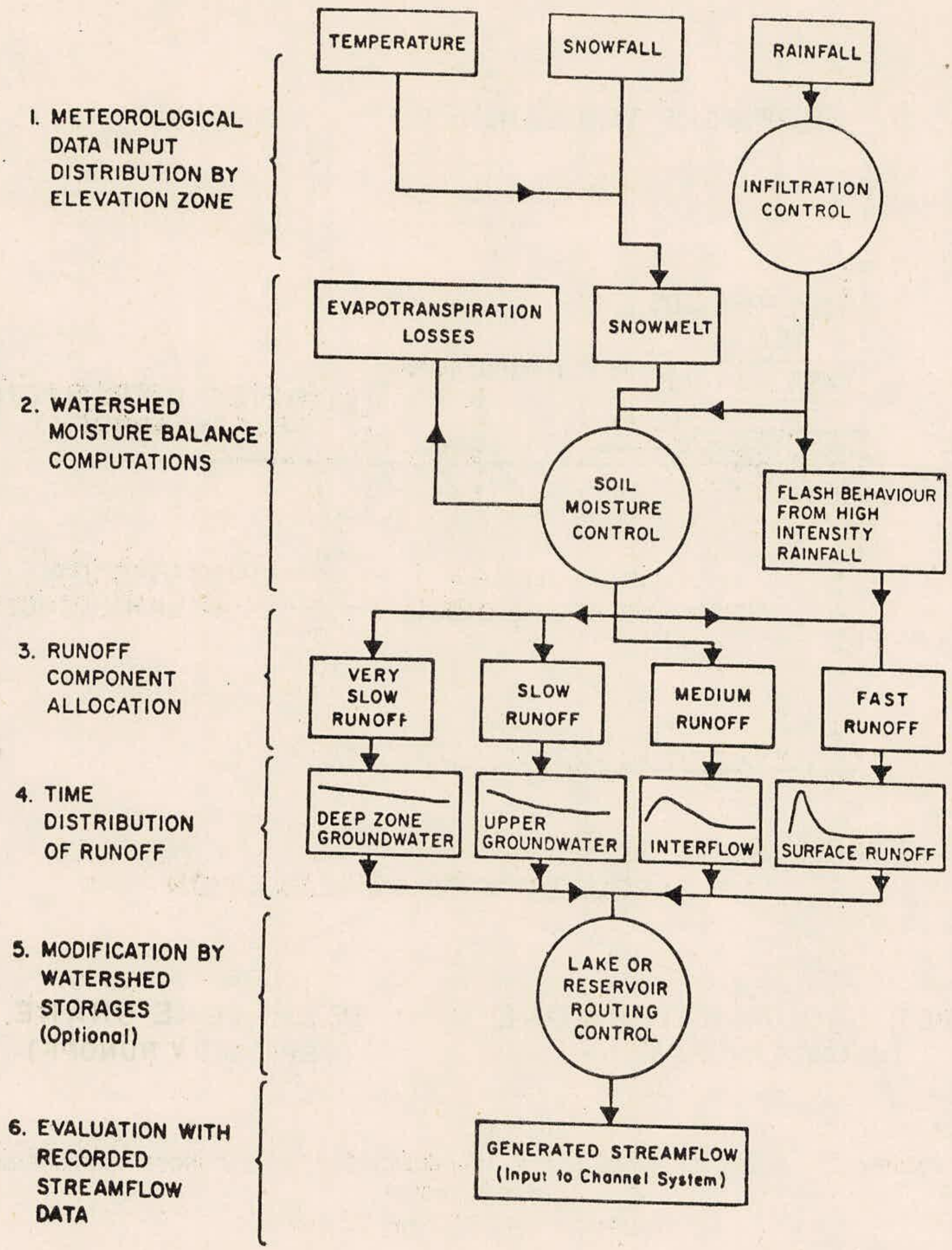

FIGURE 28- FLOW CHART OF UBC MODEL SHOWING SOIL MOISTURE CONTROL TO PERCOLATION PROCESS
(Adopted from UBC User's Manual)

2.6 Channel Translation and Routing

2.6.1 Stanford watershed model(SWM IV)

Channel time-delay histogram is developed for the basin. To construct the histogram, estimates of flow time in channels are needed. Some approximate estimation is done using empirical equations for steady open channel flow. Mannings equation is used to calculate flow time in hours for steady flow in a reach of wide channels

$$t = \frac{nL}{5370 \times Y^{2/3} S^{1/2}} \qquad \text{...(65a)}$$

or

$$t = \frac{n^{3/5} L \; W^{2/5}}{4560 \; S^{3/10} \; Q^{2/5}} \qquad \text{...(65b)}$$

where L = Length

S = Slope

W=width of channel

Q=discharge

n= Mannings n

The inflow ordinates in any time interval is multiplied by successive elements of the time-delay histogram to give a watershed outflow hydrograph neglecting storage attenuation. For each time interval, discharge neglecting storage attenuation is calculated as:

$$I_t = \sum_{x=0}^{x=z-1} R_{t-x} \, C_{x+1} \qquad \text{...(66)}$$

Where

I_t = Inflow to a hypothetical reservoir storage

R_{t-x}=Channel inflow at X time interval ago

C_{x+1}=element of the time delay histogram.

where

$$\sum_{X=0}^{X=Z-1} C_{x+1} = 1.0$$

and Z = total number of elements in the time-delay histogram.

The outflow hydrograph produced by channel translation calculation is routed through a storage system to simulate attenuation in the channel system. Outflow α storage

or $$Q = KS \qquad \ldots(67a)$$

$$\frac{dQ}{dt} = K \frac{ds}{dt} \qquad \ldots(67b)$$

since $$\frac{ds}{dt} = I-Q \qquad \ldots(67c)$$

so $$\frac{dQ}{dt} = K(I-Q) \qquad \ldots(67d)$$

where

I= Inflow

Q= Outflow

S= Storage

This can numerically be solved by considering a small time interval t from time t_1 to t_2.

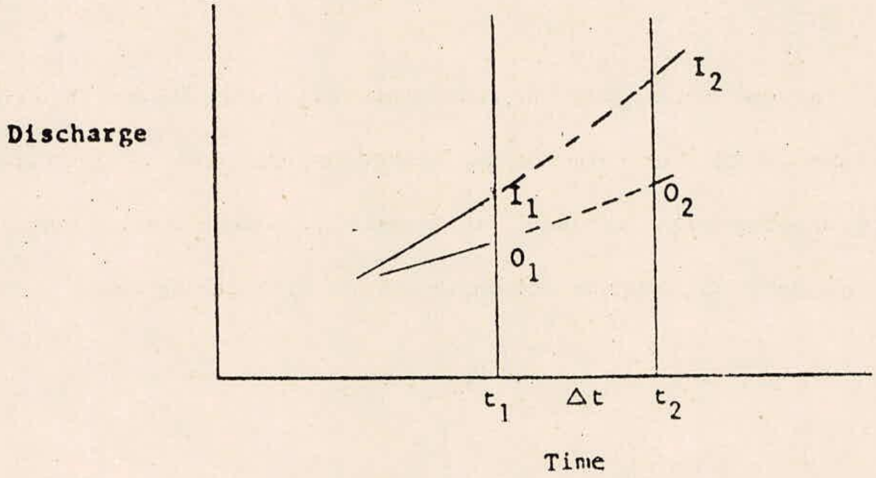

FIGURE - 29 DISCHARGES AT TWO SECTIONS ALONG A CHANNEL

At time t_1 inflow and outflowsare I_1 and Q_1

At time t_2 inflow and outflows are I_2 and Q_2

$$\frac{Q_2 - Q_1}{\Delta t} = K \left[\frac{I_1 + I_2}{2} - \frac{Q_1 + Q_2}{2} \right] \qquad \ldots(67e)$$

$$Q_2 = \frac{I_1 + I_2}{2} - \frac{(1/K + \Delta t/2)}{(1/K + \Delta t/2} \left[\frac{I_1 + I_2}{2} - Q_1 \right] \qquad \ldots(67f)$$

$$= I - KS1[I - Q_1] \qquad \ldots(67g)$$

where I = Average inflow, and

$$KS1 = \frac{(1/K - \Delta t/2\ )}{(1/K + \Delta t/2\ )} \qquad \ldots(67h)$$

2.6.2 UBC Model

In this model runoff is composed of four runoff components namely fast, medium, slow and very slow. Each runoff component is subjected to a routing procedure which produces a time distribution of runoff. Routing is done by considering the concept of linear storage reservoir. The fast and medium components of runoff are routed by considering a cascade of reservoirs identical to unit hydrograph convolution theory. The slow and very slow components of runoff are routed by considering single linear reservoir.

(a) Fast runoff routing

The routing is done by considering cascade of linear reservoir process as developed by Nash. The resulting outflow at time t from a unit impulse of inflow is given as

$$u(t) = \frac{1}{K^n} \qquad \frac{t^{n-1}}{(n-1)!} \, {}^{-}e^{t/K} \qquad \ldots(68)$$

where,

K=Linear storage constant for each of the reservoirs in the cascade n= number of linear reservoirs in the cascade.

t = time accounted after the input is given.

This gives the ordinate of Instantaneous unit hydrograph. The unit hydrograph is computed using the the equation expressed in the form of incomplete Gamma function. This U.H.is then used to calculate the fast runoff component using the relationship.

$$Q(J+N) = \sum_{R=1}^{N} P(R)\ UH\ (N-R+1) \qquad \ldots(69)$$

where $Q(J+N)$ = represent fast runoff on day(J+N)

$P(R)$= Input to fast runoff system on day R where $J < R \leq N$

N= base time of the fast unit hydrograph in days.

$UH(N-R+1)$= U.H.ordinate for $(N-R+1)$ day

(b) Medium runoff routing

This is basically the interflow component. This undergoes two stage routing process. In the first stage the medium runoff of each day enters a linear storage reservoir with a release constant(INTSTK). Daily release from the reservoir is a constant percentage of the total storage on any given day. This gives the same release as Nash model with one reservoir.

So Daily release QM=INTSTK*ST ..(70)

where ST= Total storage resulting from previous accumulations and depletions of medium runoff.

This daily release QM is then subjected to time distribution by convolution with a unit hydrograph which is similar to the fast unit hydrograph.

(c) Slow runoff routing

Groundwater flow or base flow component comprise this slow runoff component. This component is divided into slow-upper zone component and a very slow deep zone component by soil moisture model. Both of these components are routed using a single linear reservoir.

2.6.3 SSARR model

The basic routing procedure used in this model is similar to Wilson and expanded by Rockwood. Law of continuity in storage equation is followed:-

$$\left[\frac{I_1 + I_2}{2} \right] t - \left[\frac{Q_1 + Q_2}{2} \right] t = S_2 - S_1 \qquad \ldots(71)$$

where

I_1, I_2 = Inflow at the begining and end of the time period.

O_1, O_2= Outflow at the begining and end of the time period.

S_1, S_2= Storage at the begining and end of the time period

t=time duration.

2.7 Base Flow

Base flow is the flow to the channel of a watershed that comes from ground water or spring contributions and may be considered as the normal day to day flow. The base flow component is composed of the water that percolates downward until it reaches the groundwater reservoir and then flows to surface streams as ground water discharge. The ground water hydrograph during actual storm period may or may not show an increase. The release period of ground water accreted due to a storm depend on the size of the basin, for small basin it may be one deny and for large basin this may vary from a month to a year.

2.7.1 SSARR model

In this model runoff is divided between base flow and direct runoff. The portion of runoff that contributes to base flow is a function of base flow infiltration index(BII)

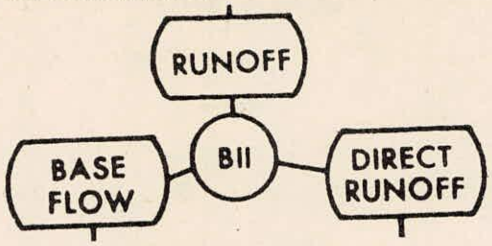

FIGURE-30 FLOW CHART OF SSARR MODEL SHOWING BASEFLOW COMPONENT

The relationship between BII, base flow percent(BFP) and base flow input limit may be given graphically as in figure 32.

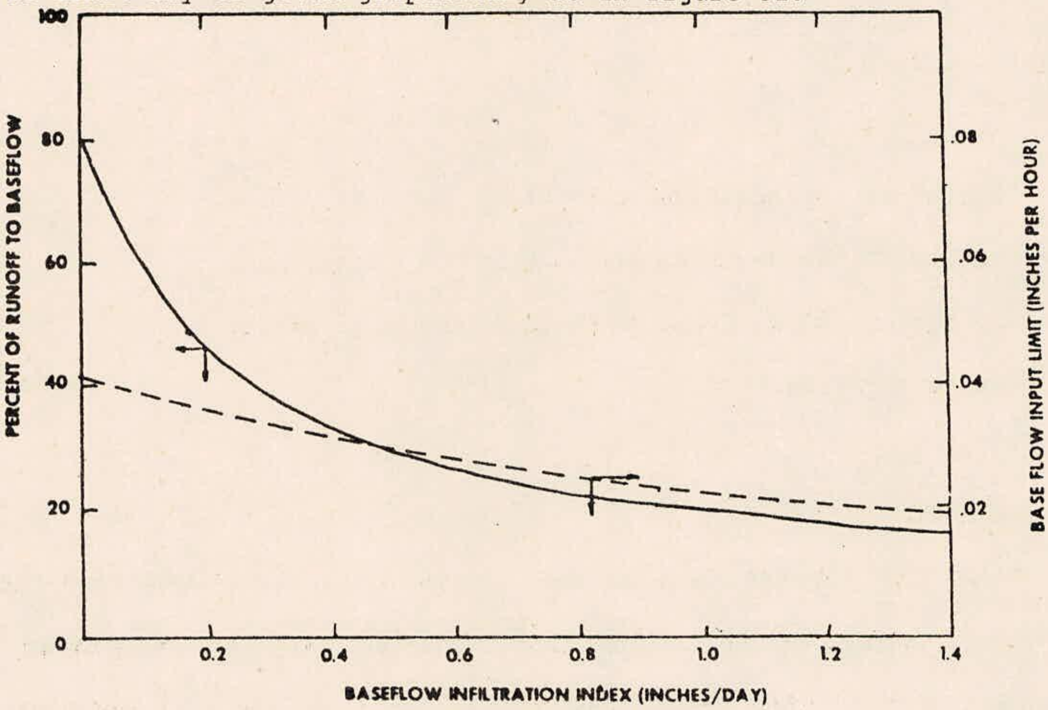

FIGURE 31- TYPICAL BASEFLOW INFILTRATION INDEX FUNCTION WITH CORRESPONDING BASEFLOW INPUT LIMITED.

Base flow infiltration index is computed for each period as

$$BII_2 = BII_1 + (24 * RG - BII_1)\left[\frac{PH}{TSBII+PH/2}\right] \qquad \ldots(72)$$

where

BII_1 = Base flow infiltration index(in inches per 24 hours) at the beginning of period.

BII_2=Base flow infiltration index(in inches per 24 hours) at end of period.

RG= RGP/PH = Runoff rate in inches per hour

RGP = Generated runoff for the period in inches.

PH= Period Length in hours

TSBII= Time delay or time of storage for calculation of change in BII. If desired, separate time of storage values may be designated for rising and falling flows.

Deep percolation takes place mainly from depression storage. So the base flow infiltration index may be thought of as an index of depression storage. The base flow component is computed as the product of base flow percent(BFP) and runoff rate(RG). ...(73)

Therefore, Base Flow Component=BFP*RG

2.7.2 Leavesley model

Baseflow takes place from ground water storage. The routing of ground water storage is done through one or more storage reservoirs in parallel. Output from storage reservoir is a linear function of the amount of water in storage. Input to ground water storage takes place from two sources. One is the seepage of a constant volume of water (SEP). from EXCS.

where

EXCS= Soil water in excess of SMAX, in inches

SMAX= Maximum available water storage capacity in upper soil zone

The other is a constant volume of seepage(RSEP) from its associated subsurface reservoir which occurs as long as melting snowpack exists to supply subsurface reservoir.

Schematically the ground water reservoir and base flow of the model is shown in figure 33.

When the upper soil zone remains in saturated condition SEP represents

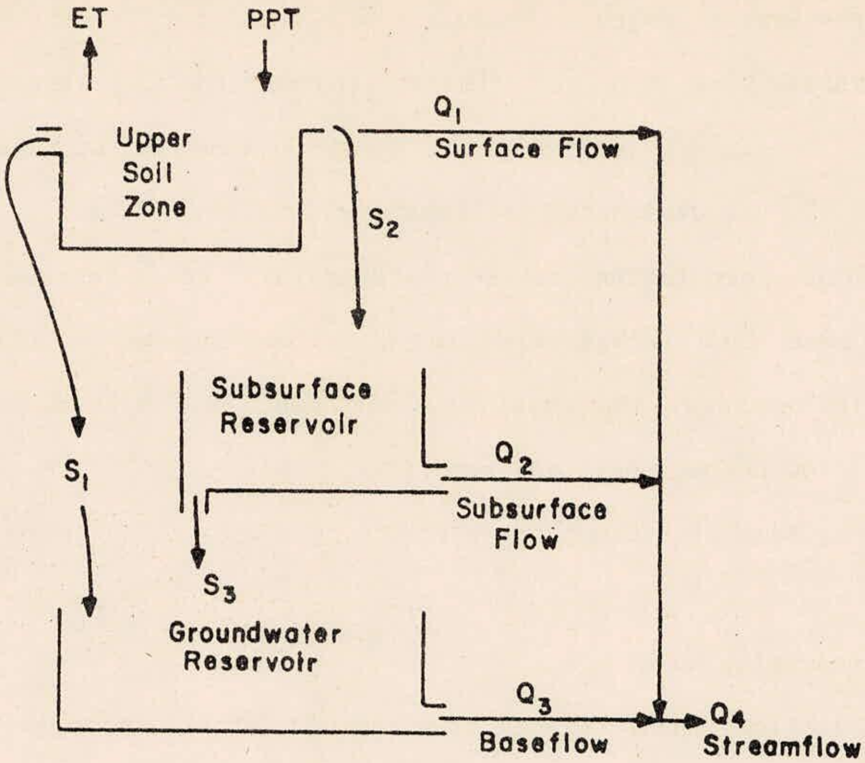

FIGURE 32- FLOW CHART OF LEAVESLEY MODEL SHOWING BASEFLOW COMPONENT(Adopted from Leavesley Model User's Manual)

percolation from upper soil zone to Ground Water storage. The value of SEP must be sufficient to restore ground water reservoir storage to that level which can supply the base flow for the subsequent low flow period.

GW = Ground water storage in inch

Base flow can be computed as

$$BAS = RCB*GW \quad ...(74)$$

where

BAS= Base flow in acre inch

RCB= a routing coefficient input for each reservoir.

The sum total of outputs from all ground water reservoirs give daily base flow component of stream flow.

The parameter RCB is the recession constant in the low flow period which can be obtained from the recession equation

$$q_t = q_o K^{-t} \qquad ...(75)$$

where

q_o, q_t= Mean daily stream flows at times 0 and t

K= Recession constant

t= time in days.

K value can be obtained from recession curve from October to March. This value of K can be used to equation

$$S_0 - S_1 = S_0(1-K) \qquad ...(76)$$

where

S_0 and S_1 = Reservoir storage quantities at 0 and 1 unit

The value of S_0 = GK in October

$$(S_0 - S_1) = BAS \qquad ...(77a)$$

$$(1-K) = RCB \qquad ...(77b)$$

2.7.3 Stanford watershed model(SWM-IV)

Base flow takes place from ground water storage. Active or deep ground water storage is divided between ground water storage and deep and inactive ground water storage. The flow charge is given in figure 34. Base flow from ground water storage at any time is proportional to the product of the Cross-Sectional area and the energy gradient of the flow.

Cross-sectional area of flow is assumed proportional to the ground water storage level.in the model. The energy gradient is estimated as a base gradient plus a variable gradient that depends on ground water accretion.,

The ground water outflow GWF at any time is

GWF= LKK4*(1.0+KV*GWS)*SGW ...(78a)

where

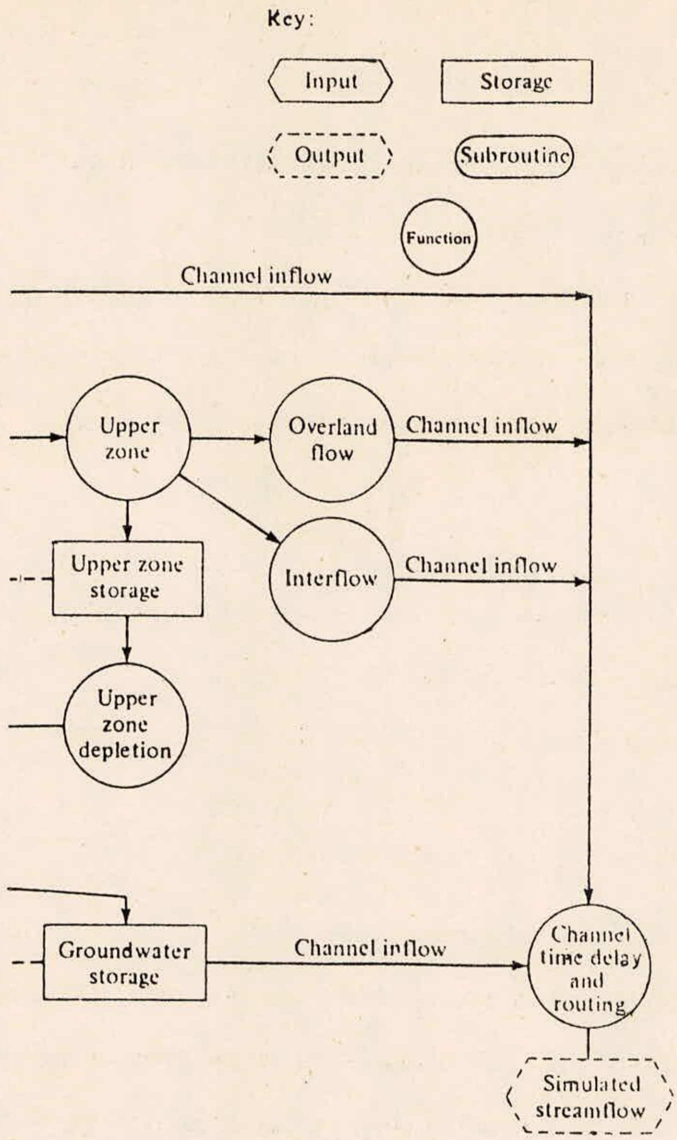

FIGURE 33- FLOW CHART OF SWM MODEL SHOWING BASEFLOW COMPONENT

$LKK4=1.0-(KK\ 24)^{1/96}$ ...(78b)

KK24 = minimum observed daily recession constant of ground water

flow= $(GWF)t/\ (GWF)\ t-24)$ ...(78c)

GWS= an antecedent index based on inflow to ground water storage and is calculated as

GWS=0.97 (GWS + Inflow to G.W.storage) ...(78d)

KV = a parameter to allow variable ground water recession rates.

When KV = Zero

and inflow to GW storage = zero

semilog plot of discharge VS time is a straight line.

But when $KV \neq$ Zero

Semilog plot of Q VS t is not linear.

3.0 REMARKS

For interception parameter calculation Leavesley model method is quite suitable in places where precipitation takes place either as rain or snow or a mixture of both.

In SWM-IV model linear approximate relationship between P.E. & percentage of area with a daily evapotranspiration opportunity equal to or less than the indicated value is relatively simple for calculation of evapotranspiration for the whole basin.

The infiltration rate mainly depends on soil properties, initial water content, rainfall rates and movement and entrapment of soil air. Some approximate methods have been developed for calculation of infiltration such as Kastianov, Horton, Phillip, Holtan and Green and Ampts. Watershed models have used such approximate methods.

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