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STATUS REPORT
ON
CATCHMENT HYDROLOGY

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PREFACE

Hydrology is an earth science, it encompasses the occurrence, distribution, movement and properties of the waters of the earth and their environmental relations. Closely allied fields include geology, climatology, meteorology and oceanography.

A catchment or drainage basin is an any area of land determined by topographic features in which both surface and subsurface water drains to its lowest level. Catchments are separated from each other topographically hills slopes and ridges, It is conserved and promtting a balance between the incoming and out going fluxes. Catchment is a logical planning unit with readily identifiable boundaries and characteristic patterns of water movement, its resources and readily suited to coordinated planning and management.

Our ability to manage water resources is influenced by the today's environmental problems. To understand the hydrologic systems that exist in the catchments are very much concerned to manage to a large extent.

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Satish Chandra

(SATISH CHANDRA)

DIRECTOR

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1.Ø INTRODUCTION

Hydrology is the science of the behaviour of water under natural conditions. It is a physical science and deals with the movement of water from the sea to the atmosphere by evaporation, the transport of water vapour by wind, the precipitation on the land and the subsequent movement, storage and retention of the water in surface channels, in the soil, in the saturated zone beneath the soil and the transference of the water between these phases and its eventual disposal by receivers to the sea or by evaporation from the earth, the leaves of vegetation or open water surfaces.

Hydrology as a science has thus many components and in the broadest sense would include the movement of water into, in and from the atmosphere but these processes are often considered to be within the domain of their sciences such as meteorology, climatology and soil science. The influence of the vegetation is obviously also within the domain of botany.

In the narrower sense Hydrology deals with the movement of water from its precipitation on the earth to return to the sea as a river discharge or to the atmosphere as evaporation. It deals with the movement of water on the surface, in sheet flow and in open channels, infiltration to and retention in the soil, movement within the soil and into the zone of saturation, the vegetation soil moisture relationship as its affects.of the water flux, the process of evaporation from open water, bear soil and vegetation and movement within the zone of saturation to wells and stream channels which penetrate that zone.

The physical scale of the hydrological cycle is very great, eg; water evaporated near the equator may be removed by winds to become precipitation in temperate zones; fortunately the scale for most hydrological processes in the narrower sense is very much smaller. If we can isolate on the surface of the earth an area in which the precipitation having contributed to evaporation and storage is eventually returned to the sea through a single channel, the area represents a CATCHMENT within which water is conserved promoting a balance between the incoming and outgoing fluxes.

A catchment or drainage basin can be defined as any area of land, determined by topographic features in which both surface water and groundwater drains to its lowest level. For groundwater flow the catchment may be larger or smaller than is apparent from surface relief. Catchments are separated from each other topographically by hills slopes and ridges. The divide or boundary being known as the watershed.

Catchments are naturally occurring units of the landscape,which contain complex array of interlinked and inter dependent resources and activities, irrespective of political boundaries. A catchment is a dynamic and integrated social economic and physical system. A typical catchment is likely to include.

1. Forest and trees to protect the land surface, provide timber and shelter and act as wind breaks.

2. Storage dams for rural and urban purposes and for irrigation and industry.

3. Erosion control works such as contour ploughing contour banks, flume, strip cropping, minimal cultivation and stubble retention to reduce or control run off.

4. Natural areas such as parks, reserves and remnants of bushland on private property.

5. Towns, communication and transport links.

Catchment is a logical planning unit with readily identifiable boundaries and characteristic patterns of water movement, its resources and is suited to coordinated planning and management.

Geomophology is science of landforms. It describes landforms and attempts to explain evolution of landforms in terms of lethology, structure, and climate. The various parameters render the investigation of Geomorphologists are the landforms, the slopes, the stream network and the natural process. To quantify the geometry of a basin, the fundamental dimensions of length, time and mass are used. Many drainage basin features that are important to the hydrologists can be quantified in terms of lenght, lenght squared, or leght cubed. Geomorphology being the study of landforms has direct influence on the flow process Geomorphological studies has become much more important in the ungauged catchments since different geomorphological parameter help in regionalisation of hydrological models dealing with the runoff estimations.

The physical process of catchment hydrology which mainly depends as follows.

1. The primary process ofinterception, evaporation, transpiration by vegetation, infiltration and percolation.
2. Surface run off including flow in the open channels.
3. Groundwater hydraulics including flow in the unsaturated zone.

The rainfall run off process in a catchment is a complex and complicated phenomenon, in general 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 process such as interception, detention, evapotranspiration, overland flow, infiltration, interflow, percolation, sub-base flow etc and emerges as runoff at catchment outlet.

Each catchment comprises of different types of soil cover, vegetation landuse, topography, drainage pattern and density, slopes etc. The process is which take place are not uniform throughout the basin, more over 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 till rainfall continue. Similarly infiltration rate varies in space and time and also depends on initial soil moisture condition. To simplify analyses of these complex process, different catchment or watershed 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 process. In the catchment modelling, it can have four general types of mathematical models which are, i) Deterministic ii) Probabilistic iii) Conceptual iv) Parametric. The type of mathematical catchment model components can be found in variety of hydrologic applications. For example, the kinematic wave routing technique is deterministic being founded on basic principles of mass and momentum conservation, The Gumble method for flood frequency analyses is typical example of the probabilistic methods in hydrology. the Gumble method is statistical since the parameters of the frequency distribution are evaluated from measured data. Stochastic methods have been used primarily in synthetic generation of hydrologic time series such as daily stream flows from mid size catchments. The cascade of medium reservoirs is a typical example of a conceptual model in the case the physical process of run-off concentration and run-off diffusion are being simulated in the mean by the diffusion inherant in the mathematical solution of linear reservoir. Regional analysis is typical example of the parametric approach to hydrologic catchment modelling. In this case statistical regression technique are used to develop predictive equations having regional applicability. The component process that have been modelled in the review are interception, evapotranspiration, overland flow, infiltration percolation, interflow and base flow.

Determination of the water yield is the solution for number of water resources problems such as design of storage facilities, determination of minimum amount of water available for agricultural, industrial or muncipal use, estimation of future dependable water supply for power generation under varying pattern of rainfall, adjustment of long records of runoff for varying rainfall regimes for study of time trends in water yield, planning irrigation operation and design of irrigation projects and so on. With this concern, water yield aspect has been considered and available model reviewed in this report.

2.Ø CATCHMENT AS A GEOMORPHIC UNIT

Catchment area may be defined as the area which contributes water to a particular channel or set of channels. It is the source area of the precipitation eventually provided to stream channels by various paths. It forms a convenient unit for the consideration for the process determining the formation of specific landscape in the various regions of the earth. It provides a limited unit of earth surface within which basic climatic quantities can be measured and characteristic landforms described and system within which balance can be struck in terms of inflow and outflow of moisture and energy. The amount of precipitation that falls over a given catchment area can be measured and, the quantity of water that flows out of the catchment in stream channels, the change in groundwater storage, and evaporation and transpiration by the plants can be estimated. In addition rates and kinds of denudation may be measured in the catchment.

The drainage net is the pattern of tributories and master streams in a catchment as deliniated on a planimetric map. The net includes all the minor rills which are definite watercourses, even including all the ephemeral channels in the furthermost headwaters. The network, of drainage has been described as trellis or palmate and by other terms descriptive of veination of different sorts. The nature often reveals a simplicity of pattern reflects the operation of a few dominant physical process which, when unraveled, can be reduced to known physical and chemical laws.

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Geomophology is science of landforms. It describes landforms and attempts to explain evolution of landforms in terms of lethology, structure, and climate. The various parameters render the investigation of Geomorphologists are the landforms, the slopes, the stream network and the natural process. It is very much possible to apply the geomorphological data which was collected by a geomorphologist for different prospective areas just because of strong relations which geomorphology has with the other disciplines like geology, pedology, hydrology, forestry, civil engg., landuse, archaeology etc. Geomorphology being the study of landforms has direct influence on the flow process. Geomorphological studies has become much more important in the ungauged catchments since different geomorphological parameter help in regionalisation of hydrological models dealing with the runoff estimations.

The part of precipitation runs over the ground as surface runoff which the portion of it enter the earth's surface,either becomes subsurface water and joins the main stream after some time or percolates into the bottom layers and becomes groundwater. The apportioning of the precipitation to surface runoff which collects in the streams and infiltration which contributes to the water table is affected primarily by the form of the earth's surface, which is accountable through various geomorphological aspects like linear, and relief parameters. Parameters like channel gradients, circularity ratio, elongation ratio, length of overland flow play important role in the runoff process and effecting peak runoff. The larger the drainage area the greater the amount of rain intercepted and the higher the peak discharge that results.

The drainage density expresses the number of kilometers of channel maintained by a square kilometer of drainage area. Basin shape is very important factor influencing the peak flow and other hydrograph characterstics such as steepness of rising and recession limbs, time spread of hydrographs etc.

Variables involving relief aspects of the basin are the most significant parameters in the hydrological studies of the watershed. The slope is related to the rate at which the potential energy of the waters at the higher elevation of the catchments is converted in kinetic energy. Losses in various forms occurs in the process. Water is held in storage and the travel time in the hydrologic system is in general inversely related to the slope.

The various geomorphological parameters established in the study can be broadly classified into three categories.

i. Linear Aspects of the channels

ii. Areal Aspects of the basin

iii Relief Aspects of the catchment and channel network.

2.1 Linear Aspects

Various parameters which involve length of channels in different ways are grouped under this category. Linear aspects of channel network are listed and explained below:

a) Length of the main channel (L)

b) Length of the channel between the outlet and a point nearer to C.G. (Lc)

c) Total length of channels

d) Wandering Ratio

e) Fineness Ratio

f) Watershed Eccentricity

a) Length of the main channel (L)

This is the length along the longest water course from the outflow point of designated sub-basin to the upper limit to the catchment boundary.

b) Length of channel between the outlet and a point nearer to Centre of gravity (Lc)

It is the length of the channel measured from the outlet of the catchment to a point on the stream nearest to the centroid of the basin.

c) Total length of channels:

Total channel length is the sum total of the lengths of channels of all the orders in the basin. This parameter is important as it gives an idea of over land flow and channel flow a simple power function.

d) Wandering Ratio:

Wandering ratio is defined as the ratio of the main stream length to the valley length. Valley length (Lv) is the straight line distance between outlet of the basin and the farthest point on the ridge.

e) Basin Perimeter:

It is measured along the divides between basins and may be used as in indicator of basin size and shape.

f) Fineness Ratio:

The ratio of channel length to the length of the basin perimeter is fineness ratio, which is a measure of topographic fineness.

g) Watershed eccentricity:

The watershed eccentricity is given by the expression

$$\Gamma = \frac{(|(Lc^2 - Wc^2)|)}{Wc}$$

where Γ = Watershed Eccentricity, a dimensionless factor

Lc = Length from the watershed mouth to the centre of mass of the watershed in the same unit.

Wc = Width of the watershed at the centre of mass and perpendicular to Lc

2.2 Areal Aspects of the Basin

The parameters which are governed mainly by the area of the drainage basin are classed as Areal aspects of the basin. Areal aspects as follows,

a) Drainage Area

b) Drainage Density

c) Constant of Channel Maintenance

d) Channel Segment Frequency

e) Circularity Ratio

f) Elongation Ratio

g) Watershed Shape Factor

h) Unity Shape Factor

These parameters have been discussed in detail and are explained below:

a) Drainage Area:

Drainage area is defined as collecting area from which water would go to a stream or river. The boundary of the area is determined by ridge separating water flowing in opposite directions. This parameter is hydrologically important because it directly affects the flood hydrograph and the magnitude of flood peaks in mountainous areas. The larger the size of the basin, the greater is the amount of the rain intercepted and higher the peak discharge that results.

b) Drainage Density.

Drainage density is defined as the ratio of the total length of channels of all orders in a basin to the area of the basin. It should therefore be measured on the topomaps of large scales (1:5Ø,ØØØ) so that first order streams can also be taken into account. Drainage density is a textural measure of a basin which is generally independent of basin size. It is considered to be a function of climate, lithology, and stage of development. Numerically this ratio expresses the number of kilometers of channel maintained by a square kilometer of drainage area.

c) Constant of Channel Maintenance

Constant of channel maintenance is defined as the ratio between the area of a drainage basin and the total length of all the channels and is expressed in square feet per foot or square meters per meter. It is virtually the reciprocal of drainage density.

The importance of this constant is that it provides a quantitative expression of the minimum limiting area required for the developmental of a length of the channel.

d) Channel segment Frequency

Channel segment frequency or stream-frequency is defined as the number of streams per unit area in a drainage basin. Horton suggested that the composition of a drainage basin provided a more adequate characterization of a stream, than did drainage pattern. His "composition" was completely described using the two textural measures of drainage density and stream frequency.

e) Circularity Ratio

Basin circularity ratio is defined as the ratio of the basin area to the area of a circle having a circumference equal to the perimeter of the basin. The value of this ratio approaches unity as the shape of a drainage basin approaches a circle.

f) Elongation Ratio

Elongation ratio of a basin is defined as the ratio between the diameter of a circle with the same area as the basin and basin length. The value of the elongation ratio approaches unity as the shape of drainage basin approaches a circle.

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g) Watershed Shape Factor

Watershed Shape Factor is defined as the ratio of main stream length to the diameter of a circle having the same as the watershed.

h) Unity Shape Factor

Unity shape factor is defined as the ratio of the basin length to the square root of the basin area.

The indices such as circularity Ratio. Elongation Ratio. Watershed Shape Factor and unity shape factor are the measures to compare basin shapes. Basin shape is very important factor influencing the peak flow and other hydrograph characteristics such as steepness of rising and recession limbs, the time spread of hydrograph etc.

2.3 Relief Aspects of Catchments and Channel Networks

Relief Aspects are the functions of the elevation or elevation differences at various points in a catchment or along the channels. Contour lines on a toposheet are made use of while determining the relief aspects. Various parameters involving the relief aspects are as follows:

a) Basin Relief

b) Relief Ratio

c) Relative Relief

d) Ruggedness Number

e) Nash's measure of slope

Variables involving relief aspects of the basin are the most significant parameters in hydrological studies of the watershed. The slope is related to rate at which the potential energy of the water at high elevation in the headwaters of the catchment is converted into kinetic energy. Losses in various forms occur in the process. Water is held in storage and the travel time in the hydrologic system is in general inversely related to the slope. Mountainous catchments are characterised by the steep slopes and hence these parameters become still more important for mountainous catchments. These parameters, therefore, have been discussed in the following text.

a) Basin Relief

Relief of a basin is the maximum vertical distance from the stream outlet to the highest point on the dividing ridge. The total relief of a basin is a measure of the potential energy available to move water and sediment downslope.

b) Relief Ratio

The relief ratio is defined as the ratio between the basin relief and the basin length. In normally shaped basins the relief ratio is a dimensionless height length ratio equal to the tangent of the angle formed by intersection at the basin mouth of a horizontal plane with a plane passing through the highest point on the divide. This parameter permits comparison of the relief of two basins without regard to the scale of the topomaps used.

c) Relative Relief

Relative relief is defined as the ratio of the basin relief expressed in units of miles to the basin perimeter. Relative relief is an indicator of the general steepness of a basin from summit to the outlet. It has an advantage over the relief ratio in that it is not dependent on the basin length. When the main channel consists of two branches more or less of equal catchment the channel slopes are taken as the mean of the two values calculated separately and weighted with the appropriate catchment area.

d) Ruggedness Number

The product of drainage density and relief (in the same units) is termed as the ruggedness of a basin. Areas of low relief but high drainage are, therefore, as ruggedly textured as the areas of higher relief having less dissection.

e) Nash's Measure of Slope

Nash (1960) defined another measure of slope where the profile of the main channel having been plotted from the gauging site to the catchment boundary, a straight line was drawn through the gauging station and the vertical through the highest point of the main channel. Further the slope of the line being so chosen that the area of the triangle was equal to the area contained below the channel profile.

3.Ø REVIEW

3.1 Interception

Part of the storm precipitation that falls in a basin is intercepted by vegetation and other forms of the cover on the drainage area. Interception can be defined as that segment of the gross precipitation input which wets and adheres to above general objects until it is returned to the atmosphere through evaporation. Precipitation striking vegetation may be retained on leaves of blades of grass, flow down the stems of plants and become stem flow, or fall off the leaves to become part of the throughfall. Most interception loss develops during the initial storm period, thereafter, the rate of interception rapidly reaches zero. In studies of major storms and floods interception loss is commonly neglected. It may however, be significant in water balance studies.

The amount of water intercepted is a function of

i. Storm characteristics

ii. The species age and density of prevailing plants and trees.

iii. The season of the year.

The percentage of interception loss is more for smaller amount and less intense rainfall. Oak or Asper leaves may retain as much as 1ØØ drops of water. On an average for well developed trees, retention may be of the order of 2Ø drops per leaf. For light showers where $P < Ø.Ø1$ inch 1ØØ% interception losses may occur, where as for showers where $P > Ø.Ø4$ inch, losses occur in the range of 1Ø to 4Ø%.

Most interception loss develops during the initial storm period and the rate of interception rapidly approaches zero.

The total storms interception can be estimated by using

$$L_i = S_i + KEt$$

Where,

L_i = The volume of water intercepted (in).

S_i = interception storage that will be retained on the foliage against the force of wind and gravity (usually varies between Ø.Ø1 and Ø.Ø5 inchs)

K = The ratio of surface area of intercepting leaves to horizontal projections of this area.

E = Amount of water evaporated per hour during the precipitation period (value ranges from 1.1 to 9.2)

t = time (hr)

Total interception is directly related to the amount of foliage and its character and orientation. Interception loss function also varies with 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 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 of experimental plot strongly associated with a given locality.

3.1.1 Stanford Water-Shed model (SWM-IV)

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

grassland	0.10 inch
moderate forest cover	0.15 inch
Heavy forest cover	0.20 inch.

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

3.1.2 Leavesly Model

The factors like vegetation type, canopy density and precipitation type have been taken into consideration to compute interception of each hydrologic response of watershed cover density expressed as a percent of the hydrologic response unit, surface covered by a horizontal projection of the vegetation canopy, a canopy storage for rain in inches depth and canopy storage for snow in inches water equivalent depth are input for each hydrologic response unit.

For the precipitation events occurring as all rain or all snow and where,

i. Precipitation depth - PPT RNST or SNST

XIN = (RNST or SNST) X COVDIN (in)

Where,

RNST = Canopy storage for rain

SNST = Canopy storage for snow

XIN = interception of each Hydrologic Response unit of watershed.

COVDIN = Cover density

Where the precipitation occures as a mixture of rain and snow, it is assumed that rain occurs first followed by snow.

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

3.2 Depression Storage

Precipitation that reaches the ground may infiltrate flow over the surface or become trapped in numerous small depression from which the only escape is evaporation or infiltration. The retained water is called depression storage, when rainfall intensity exceeds infiltration capacity of the soil, the effective rainfall begins to accumulate, runs off the surface and begins to fill surface depressions. The nature of depression as well as their size is largely a functions of the original land form and local land use practice. Because of extreme variability in the nature of depressions and the paucity of sufficient measurements, no generalised relationship with enough specified parameters for all cases is feasible.

In order to subtract depression storage from rainfall and subsequently determine effective rainfall, we should determine the manner in which depression storage capacity or volume varied with time as well as with effective rainfall. To this, we consider a depression whose storage capacity V. The storage at anytime t is between 0 and V; $0 < S < V$ and storage remaining Sc to be filled is V-S.

$$\frac{dS_e}{dP_e} = -kS_e$$

in which Pc is the effective rainfall volume (gross rainfall (P) -Evaporation (E) -interception - infiltration (F)) and K is a constant or parameter).

$$S_e = C \exp[-kP_e]$$

Where C is constant integration using the condition that $P_e=0$, $S_e= V$ or $S=0$

$$S = V[1-\exp(-kP_e)$$

The above equation proposed by Linsley et al (1975) the parameter k can be estimated by noting that where $P_e = 0, dS_e/dP_e = 1$, implying that all the water goes to fill depressions and yielding $k = 1/V$. The typical value of V for pervious and impervious areas given below:

Land Cover	Surface Storage	Recommended Value
Impervious		
Large paved areas	0.05-0.15	0.1
Roofs a flat	0.10-0.30	0.1
Roofs slope	0.05-0.10	0.05

Pervious		
Lawn grass	02-0.50	0.30
Wooded area and open fields	0.2-0.60	0.40

3.3 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. Leopold 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 such as 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)m Knisel et.al.(1969) and Paramele (1972).

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

(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 Tanmner (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 result 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 behavior of E.T. values.

Schematically soil-plant-atmosphere system may be represented as shown in figure 3.1

Estimation of E T follows a vertical water budget within a system. It requires to consider three sets of variable (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 atmospheric determined quantity, which assumes that the E T flux will not exceed the available energy from the 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

$$PET = C_{ET} . E_P$$

Where

C_{ET} = a pan to PET 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.

(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.3.2

In this method vertical components may be expressed as,

$$Rn = A + LE + S + X$$

$$Rn = Rs - aRs + RI - RIr$$

Where

Rs = Incoming solar radiation (short wave)

aRs= Solar radiation reflected

RI = Incoming radiation (long wave)

RIr= Emitted long wave radiation

Rn = net radiation

A = Sensible heat air

LE = Latent heat of water vapour

S = Soil heat

X = Miscellaneous heat sinks like plant and air 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.

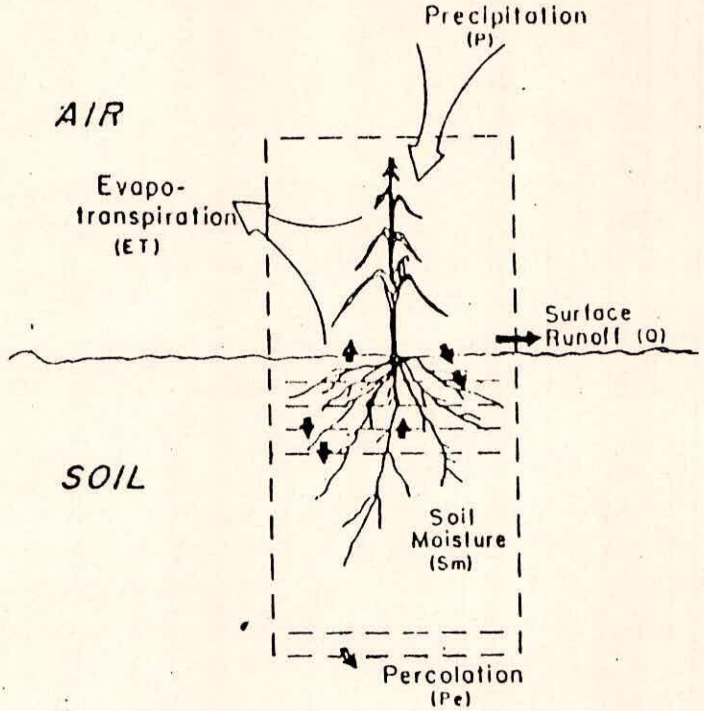

FIG: 3.1 SCHEMATIC REPRESENTATION OF SOIL-PLANT-AIR SYSTEM WATER BUDGET (Adopted from Haan'Small Catchment Hydrology)

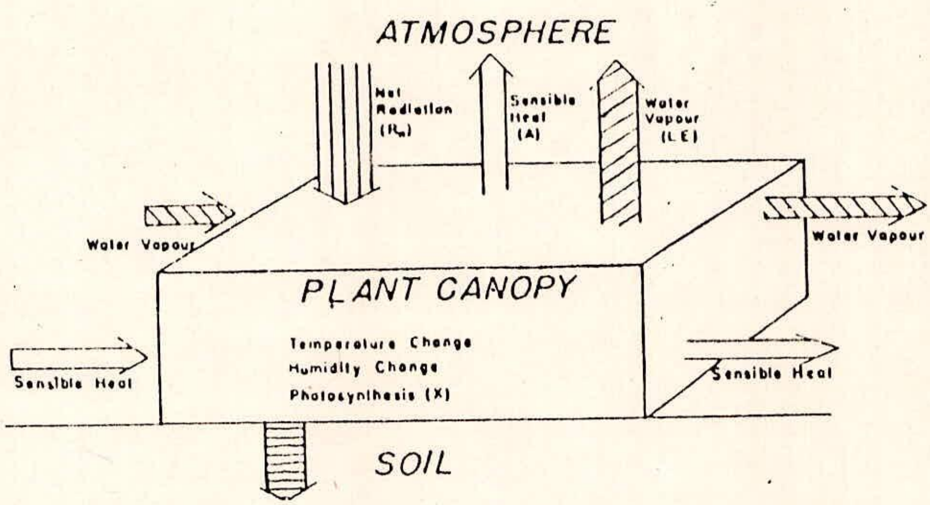

FIG: 3.2 ENERGY BALANCE OVER A VEGETATED SURFACE

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

$$U = \frac{Kp(45.7\ t+813)}{100}$$

U = Estimated monthly evaporation in mm

Kp= An empirical consumptive use coefficient

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

t = mean monthly air temperature in degree centigrade.

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 measurements and (b) Combination method, Penman (1948, 1956)

(f) 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 determine 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.

(g) 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.

(h) 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

$$ARI_i = [ARI_{i-1} + R_{i-1}]K$$

Where

ARI_i = Antecedent retention index for day i

R = Daily retention (infiltration)

K = 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

$$E = E_p(M/C)$$

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

$$AE_i = \sum_{j=1}^{n} K_j(S_i/S_j)Z_jPE_i\, e^{-w}\,(PE_i - \overline{PE})$$

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)

(iv) Holtan etal (1975) gave a similar simple equation:

$$ET = (GI)K\, E_p[(S-SA)/S]^x$$

Where

ET = Actual ET (mm/day)
GI = Growth index (percent)
K = ratio of pan evaporation for full canopy
E_p = Pan evaporation (mm/day)
S = Total soil porosity (percent)
SA = Available soil porosity(percent)
X = Exponent estimated to be 1.Ø

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 coefficientxInput value of 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)

$$K24E1 = \frac{\text{Total streams and lakes area}}{\text{Total watershed area}}$$

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.

$$K24EL = \frac{\text{Total area from which ET from the groundwater storage}}{\text{Total watershed area}}$$

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

The upper zone simulates storage in depressions and highly permeable surface soil while the lower zone 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 which is always less than E . 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 3.3.

The rate of evapotranspiration from the lower zone is determined from the shaded area or

$$E = E_p - E_p^2/2r$$

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

$$r = K_3(LZS)/LZSN$$

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 storage

K_3 = An input parameter that is a function of watershed cover

Table : -- Typical Lower Zone evapotranspiration

Watershed Cover	K_3
Open land	Ø.2Ø
Grass land	Ø.23
Light forest	Ø.28
Heavy forest	Ø.3Ø

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

The ratio 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.

3.3.1 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 3.4

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

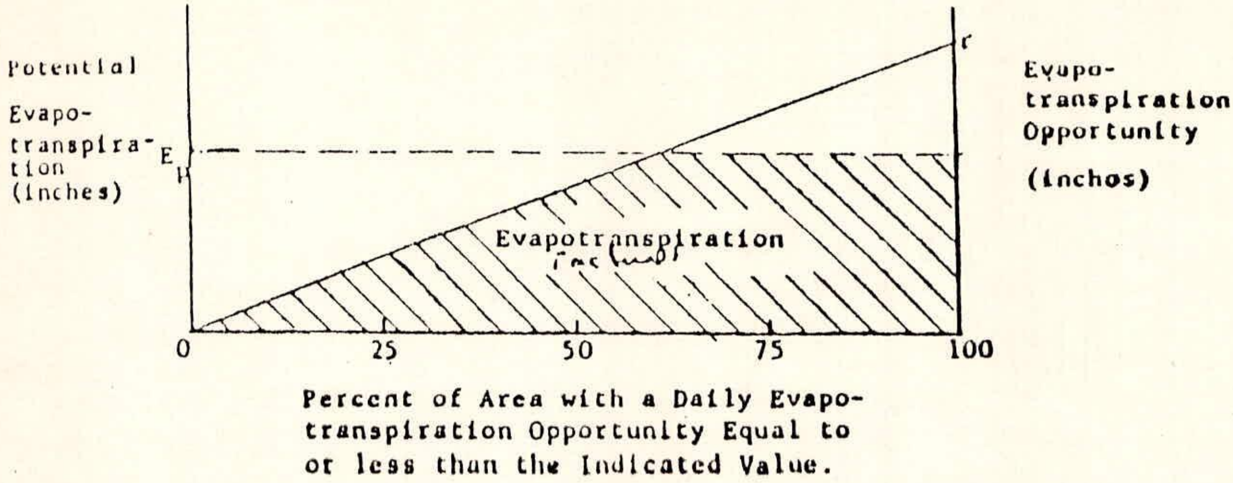

FIG: 3.3 ASSUMED LINEAR AREAL VARIATION OF POTENTIAL EVAPOTRANSPIRATION

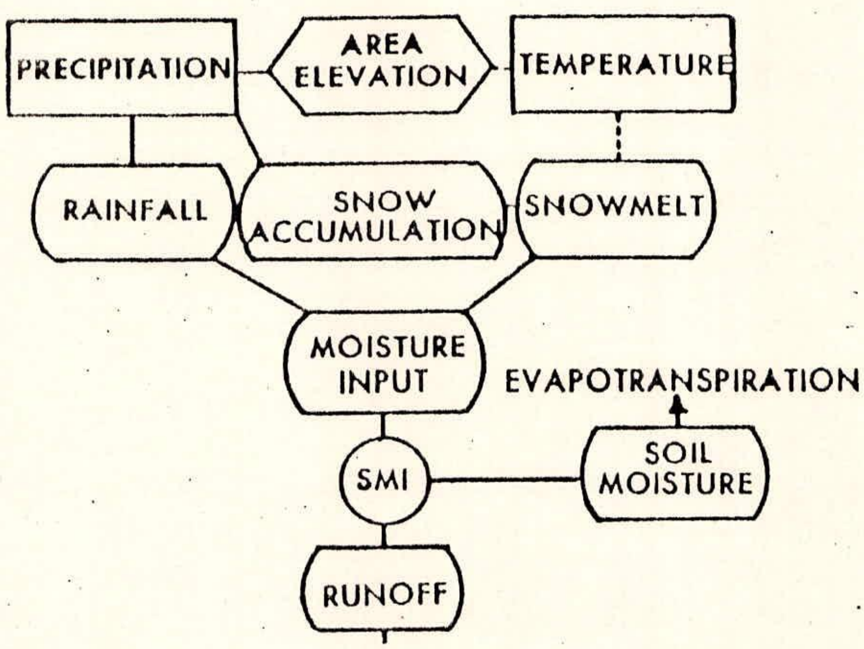

FIG: 3.4 FLOW CHART OF SSARR MODEL SHOWING EVAPOTRANSPIRATION COMPONENT (ADOPTED FROM U.S. ARMY CORPS OF ENGINEERS 1972)

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(i) in table form as much 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 tabular 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:

$$SMI_2 = SMI_1 + (WP - RGP) - (PH*KE*ETI)/24$$

where

SMI = Soil moisture index (in inches) at beginning of period

SMI = 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 versus rate of precipitation in inches per day.

For zero precipitation KE = 1.Ø when rainfall occurs, the amount of soil water depletion by ETI diminishes and follows a relationship as shown in figure 3.5.

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

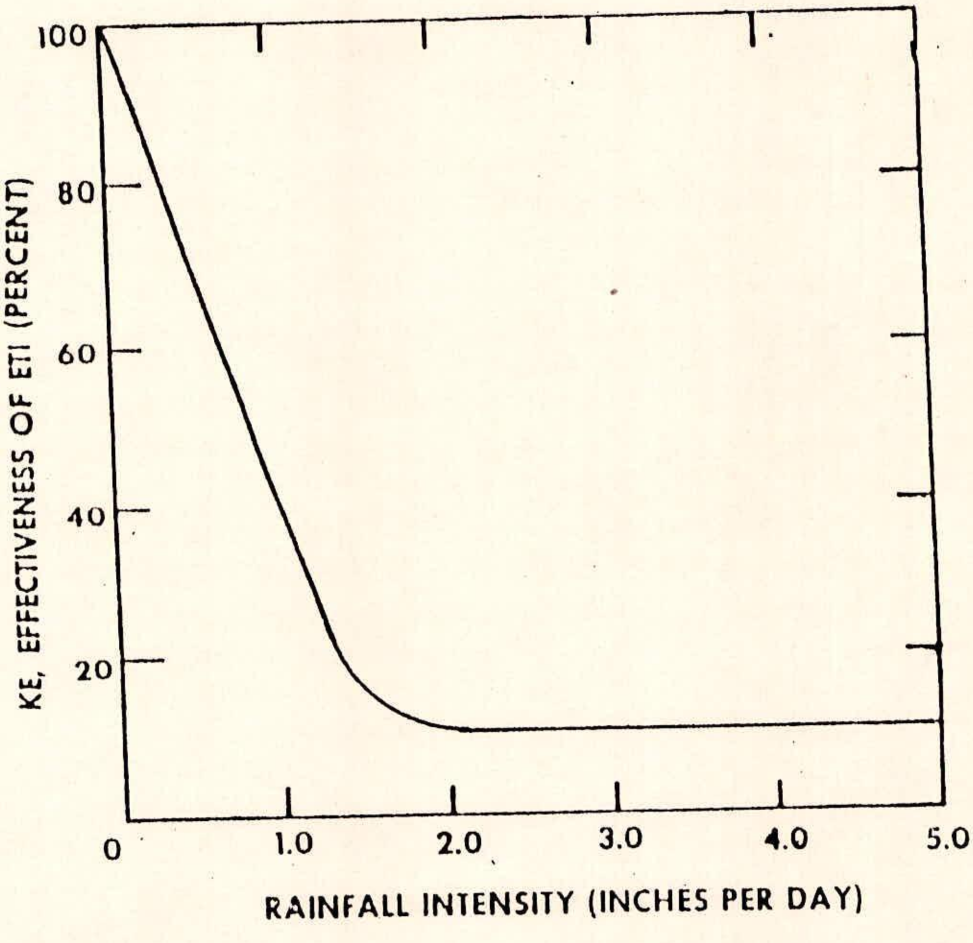

FIG: 3.5 SOIL WATER DEPLETION BY ETI WITH RESPECT TO RAINFALL INTENSITY (ADOPTED FROM U.S. ARMY CORPS OF ENGINEERS 1972)

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:

$$ETI_d = \frac{ETI_1 * Wt_1 + ETI_2 * Wt_2 + \ldots . ETI_n * Wt_n}{n}$$

Where

ETI_d = Weighted daily evapotranspiration index (inches)

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

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

When daily ETI values are used SMI is calculated as:

$$SMI_2 = SMI_1 + (WP - RGP) - (DKE * ETI_D)$$

Where

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

= 1.Ø when soil moisture is adequate and then ETI approximates potential evapotranspiration.

When soil is dry DKE value in percentage diminishes with SMI values this is shown in fig.3.6.

3.3.2 UBC Watershed Model

This is a daily rainfall runoff model which simulates daily runoff of a basin. The watershed model is designed primarily for mountaineous watersheds and calculates the total runoff

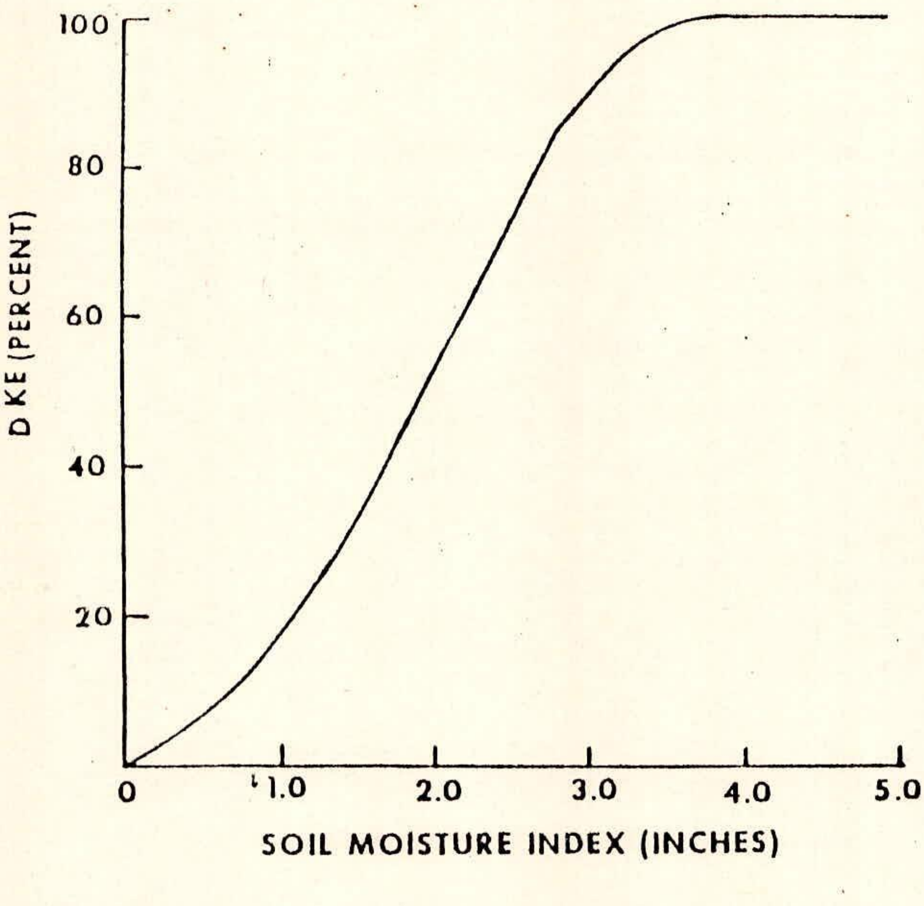

FIG: 3.6 RELATIONSHIP OF DKE VALUE IN PERCENTAGE WITH SMI
(ADOPTED FROM U.S. ARMY CORPS OF ENGINEERS,1972)

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 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)

$$EVAP = K*MK*\left[10\,\frac{(TX-14.50)}{64} - 10\,\frac{(TN-14.50)}{64}\right]$$

Where

K = Evaporation constant

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

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

40.

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

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

Where

TX(L) = Maximum Temperature

TEX = Maximum temperature for Meteorological Staion 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.

$AET = PET*10^{(-BSD/AGTGEN)}$

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:

$$BSD = BSD - PRN - BM = AET$$

Where

PRN = Rain input

BM = Snowmelt input

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

3.3.3 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 riparian 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:

$$E_1 = \frac{EMD\ (UZTWC)}{UZTWM}$$

Where

EDMD=The evapotranspiration demand is computed from pan evaporation or meteorological variables. The potential evapotransipration 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 (Ez)

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

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 :

$$\text{RATIO (T)} = \frac{\text{Tension water content}}{\text{Tension water capacity}} = \frac{\text{UZTWC}}{\text{UZTWM}}$$

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

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

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 withdrawal E_1.

$$\text{This excess} = [\text{ADIMC} - (\text{UZTWC} + E_1)] \text{ or}$$
$$= (\text{ADIMC} - E_1 - \text{UZTWC})$$

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

Then

$$E_6 = \frac{E_1 \text{ RED } [\text{ADIMC} - E_1 - \text{UZTWC}]}{\text{UZTWM} + \text{LZTWM}}$$

The quantity ADIMC is adjusted by subtracted E6 and of course E 6 may not exceed ADIMC. After the adjustment 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):

$$E_3 = \text{RED} \quad \left[\frac{\text{LZTWC}}{\text{UZTWM} + \text{LZTWM}}\right]$$

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:

$$\text{RATIO (T)} = \frac{\text{LZTWC}}{\text{LZTWM}}$$

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)

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.

$$RATIO\ (F) = \frac{[\ LZFPC + LZFSC - RFW\ ]}{LZFPM + LZFSM - RFW}$$

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:

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

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

(viii) Evaporating (EP_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

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

$$PAREA = 1 - (ADIMP + PCTIM)$$

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

$$EUSE = PAREA(E_1+E_2+E_3)+E_4+E_5+E_6$$

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

3.3.4 USGS Model

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

$$e_p = K_p\ e$$

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 profiles or due to evapotranspiration. The moisture content m_o will continue to increase up to 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.

When $d > \emptyset$

i)

$$m_o(t+\Delta t) = m_{P_o}(t) + d\,\Delta t$$
$$\text{when } m_o(t+\Delta t) < m_c$$

ii)

$$m_o(t+\Delta t) = m_o(t) = m_c$$
otherwise

when $d = \emptyset$

i)

$$m_o(t=\Delta t) = m_o(t) - e_p{*}\Delta t$$

$$\text{when } m_o(t) > e_p{*}\Delta t$$

ii)

$$m_o(t+\Delta t) = \emptyset$$
$$\text{where } e_p{*}\,\Delta t = e_p\,\Delta t - [\,i(t) - i(t+\Delta t)]$$

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

3.3.5 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 compute PET is:

$$PET = CT * (TAVF - CTX)*RIN$$

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 ·F

CTX = An air temperature constant for a given area in ·F

RIN = Daily solar radiation 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:

$$CT = (C1 + (13.0*CH)^{-1}$$

where
C1 = Elevation correction term
CH = Humidity index

C1 is computed as:

$$C1 = 68 - (3.6*(E/1000))$$

where

E = Median watershed elevation in feet.

CH is computed by :

$$CH = (37.5mm\ Hg)/(e_2-e_1)$$
$$= (50.0mts)/(e_2-e_1)$$

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 ·F is computed by

$$CTX = 27.5 - (0.25* (e_2 - e_1))-(E/1000)$$

Evapotranspiration values are computed month wise.

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

3.4 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 catchment. 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 in to 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

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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, losely packed surface soil and surface depressions. Rate of infiltration at the beginning of rain fall is high. As the rainfall continues the soil moisture also continues to increase till it attains almost a constant value fc. 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 fc is generally assumed to be equal to the saturated hydraulic conductivity K_o but will actually be somewhat less than K_o due the entrapped air. In most cases fc is more accurately approximated by K_s, the hydraulic conductivity at residual air saturation.

(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 diffusivity 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 type of soils are shown in the figure 3.7 and figure 3.8 respectively.

(ii) Initial water content ($\emptyset_i$)

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.

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 time during infiltration the wetting front advances more rapidly for higher initial water content.

The figure 3.9 shows different infiltration rate curves depending on initial soil moisture content.

(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

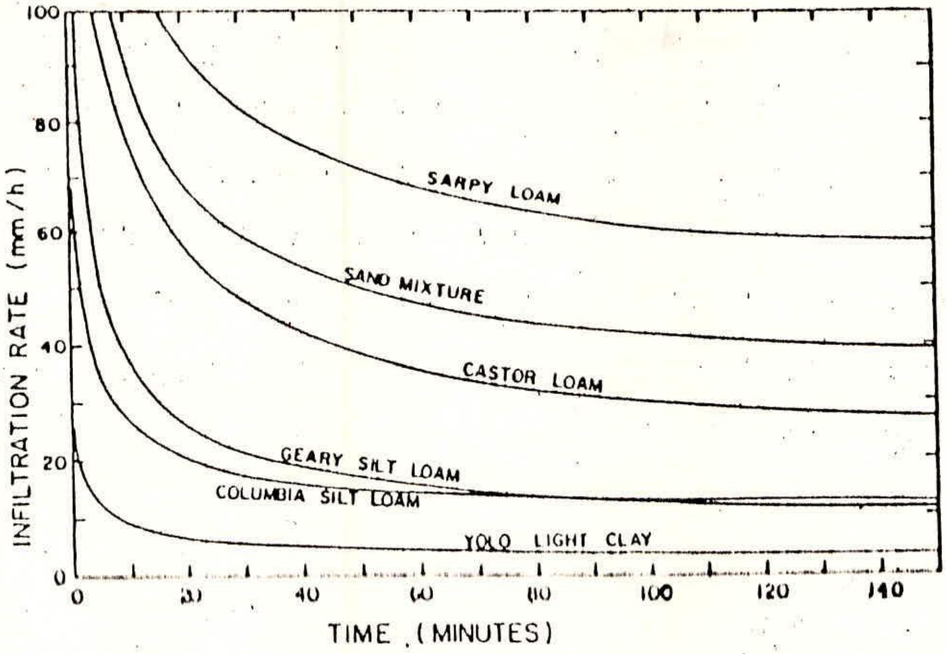

FIG: 3.7 INFILTRATION RATE CURVE OF SOME SOILS
(ADOPTED FROM HANN 'SMALL CATCHMENT HYDROLOGY')

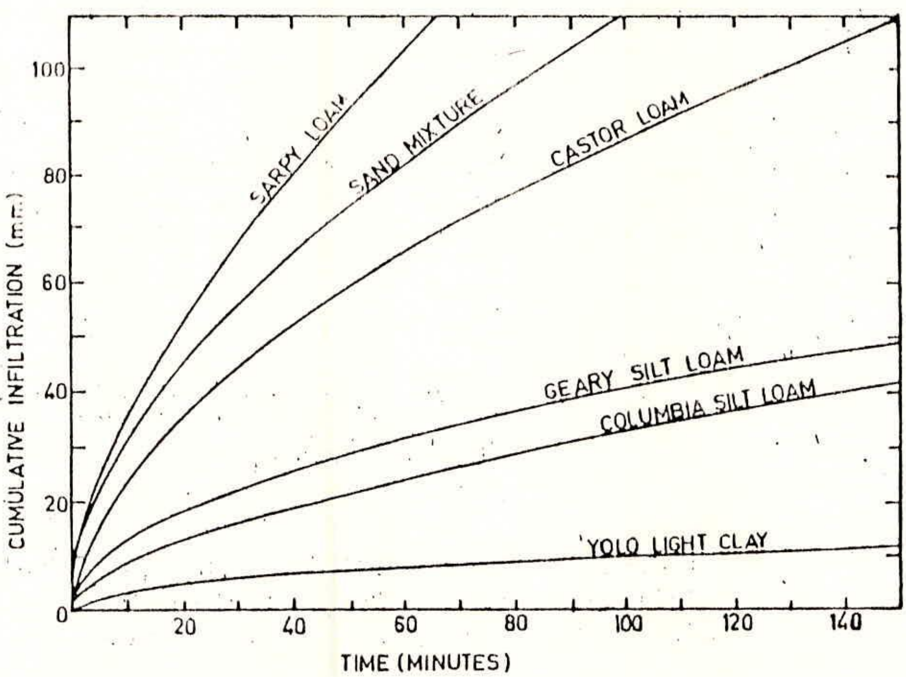

FIG: 3.8 CUMULATIVE INFILTRATION RELATIONSHIP FOR THE SOILS
(ADOPTED FROM HANN 'SMALL CATCHMENT HYDROLOGY')

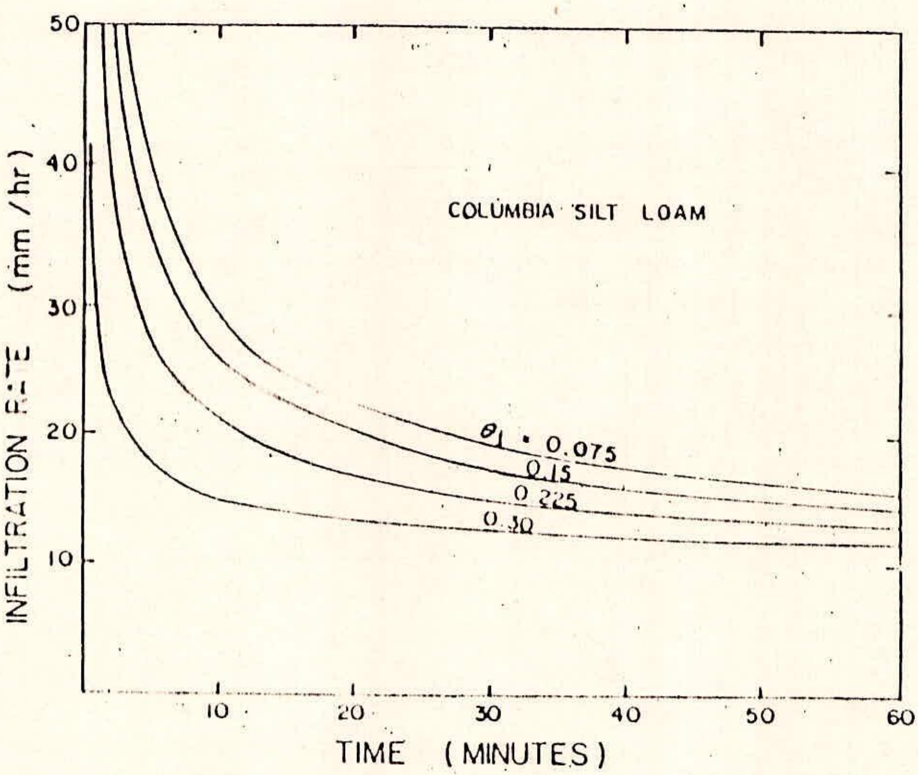

FIG: 3.9 INFILTRATION RATE FOR DIFFERENT SILTY LOAMY SOIL FOR DIFFERENT INITIAL SOIL MOISTURE CONTENT (ADOPTED FROM HANN 'SMALL CATCHMENT HYDROLOGY') deep homogeneous soil, infiltration may confine indefinitely at a rate equal to the rainfall rate without ponding at the surface. The water content of 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.

The effect of surface sealing crusting due to Rainfall impact in infiltration rate is shown in fig.3.10.

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 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 of evaluating K_s for different amounts of air trapped in large pores.

(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, much elaborate procedure 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 modelling catchment hydrology.

Simplified algebraic 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)

$$f_p = K_k\ t^{-\alpha}$$

where

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 :

$$f_p = f_c + (f_o - f_c)e^{-Kt}$$

where

f_c = Final constant rate of infiltration capacity

f_0 = Initial rate of infiltration capacity

f_p = Infiltration rate at any time of t

K = A constant dependent on primarily upon soil and vegetation

t = Time from start of rain fall

e = Base natural logarithm

(iii) Philip equation

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

$$f_p = S/2\ t^{-1/2} + C_a$$

where

S = sorptivity

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

where $D(\emptyset)$ = Soil water diffusivity

$h(\emptyset)$ = Soil water characteristics

A regression fit to experimental data will tend to give $C_a = f_c$

Young (1968) showed that Ca could be approximated as $S_a = 2K_s/3$ and $S = (2Mk_s S_f)^{1/2}$

where

M = fillable porosity

$= (\emptyset_s - \emptyset_1)$

S_f = Effective suction at the wetting front

(iv) Holton equation

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

$$f_p = GI * a * SA^{1.4} + f_c$$

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 is measured or predicted and then the initial available storage SA is computed as:

$$SA = (Ø_s - Ø_i)d$$

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 up to the limit SA_o and by evapotranspiration (ET) through plant. That is, after a period of time Δt:

$$SA = SA_o - F + f_c \Delta t + ET \Delta t$$

where

F = The amount of infiltration during ^t

(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 separates a zone that has been wetted from a totally unwetted zone. Application of Darcy's law given the Green and Ampts equation as:

$$f_p = K_s (H_o + S_f + L_f)/L_F$$

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 surfaces 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.3.11.

3.4.1 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 flow into long-term lower zone and groundwater storages. That fraction of water determined to be remaining in surface detention after calculation of direct infiltration is disposed off 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 included 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.

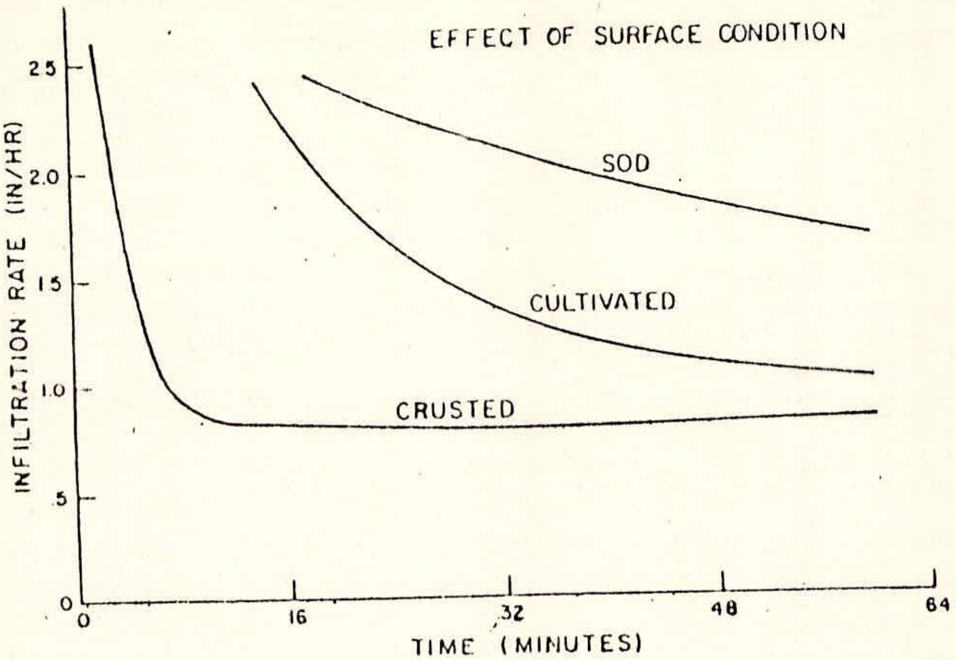

FIG: 3.10 EFFECT OF SURFACE SEALING AND CRUSTING DUE TO RAINFALL IMPACT ON INFILTRATION RATE

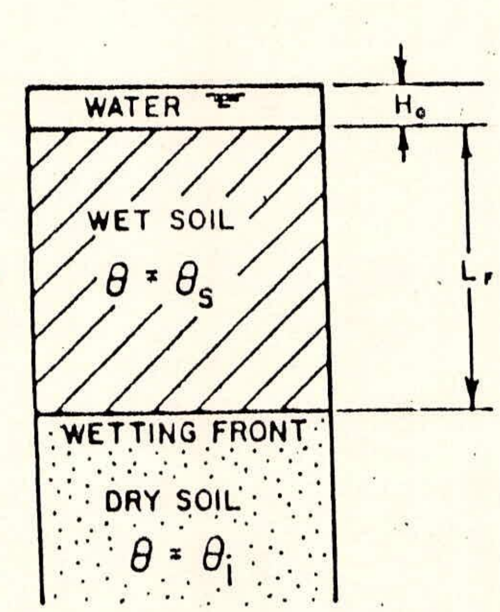

FIG: 3.11 THE GREEN-AMPT MODEL ASSUMING SLUG FROM WITH A WETTING FRONT BETWEEN THE INFILTRATED ZONE

The cumulative frequency distribution of infiltration capacity is assumed to be linear from zero to a maximum value as shown in the figure 3.12 Infiltration capacity is broken into two regions, one for lower zone and groundwater storage, the other for interflow. In the region shown below the line o 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 3.12.

where

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 (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 Ø.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)

3.4.2 USGS model

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

The Phillip equation:

$$\frac{di}{dt} = K[1 + \frac{P(m-m_o)}{i}]$$

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 t_o

m = Moisture content uniformly distributed through wetted column at time t

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

Thus,

$$P(m-m_o) = rP_s - P_s(r-1)m_o/m_c$$

$$P_m = P(m-m_o)$$

$$= rP_s - P_s(r-1)m_o/m_c$$

$$di/dt = K(1+P_m/i)$$

Infiltration occurs at varying rates over a basin, but equation describes infiltration at a point. Following a scheme of Crawford and Linsley, Dawdy, Lichy and Bergmaun schematically accounted for areal variability of infiltration as shown in figure 3.13 and avoided threshold effects.

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

$$R_e = \Delta t^{1/2}\, S^2/F_1 \quad \text{when } S<F_1$$

$$= \Delta t\,[S - F_1/2] \quad \text{otherwise}$$

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

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

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.

3.4.3 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 3.14.

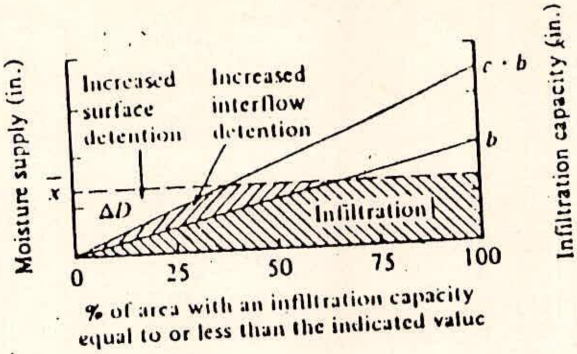

FIG: 3.12 ASSUMED LINEAR AREAL VARIATION OF INFILTRATION CAPACITY OVER A WATERSHED AS CONSIDERED IN SWM IV MODEL

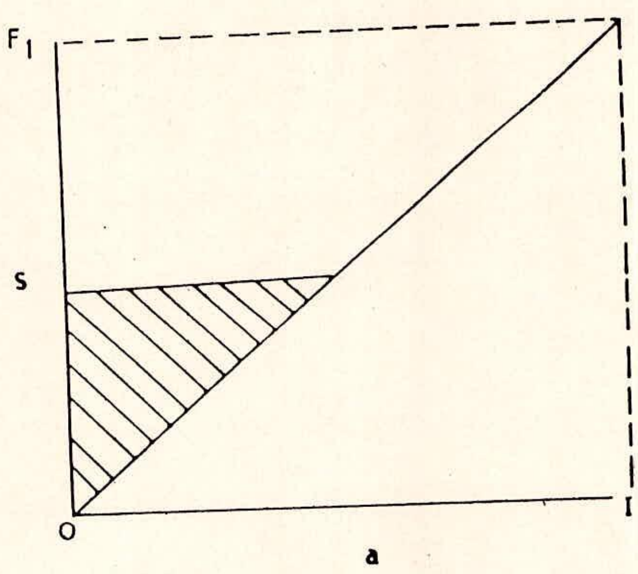

FIG: 3.13 AREAL VARIABILITY OF INFILTRATION CONSIDERED IN USGS MODEL (ADOPTED FROM USGS MODEL USER'S MANUAL)

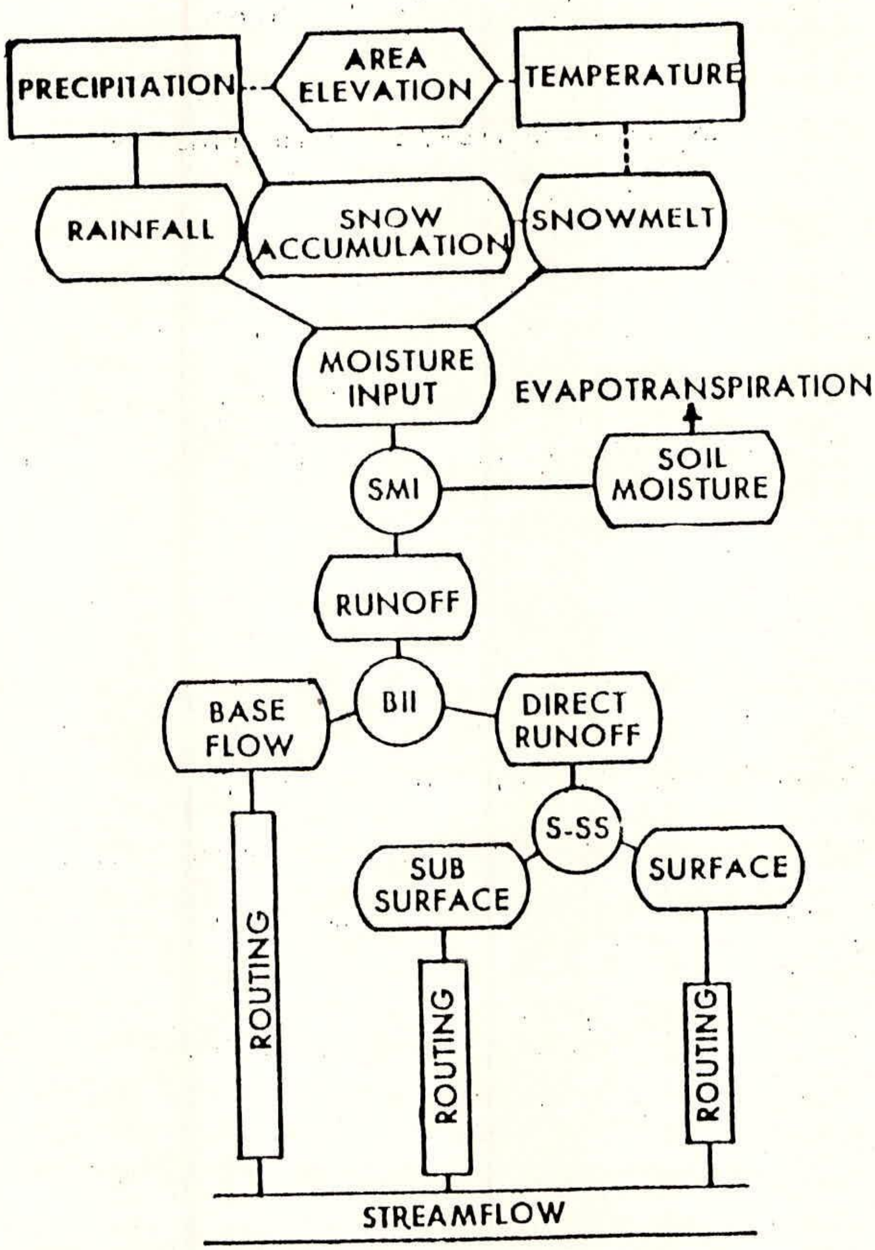

FIG: 3.14 FLOW CHART OF SSARR MODEL SHOWING INFILTRATION COMPONENT
(ADOPTED FROM U.S. ARMY CORPS OF ENGINEERS 1972)

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 (SMI). The SMI-Runoff Percent (ROP) relationship as shown in figure 3.15 determines to a large extent, the volume of runoff and also effects the shape of hydrograph.

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

Variation of baseflow percent with baseflow infiltration index is shown in figure 3.16

3.4.4 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.

Dickinson and Whitely (1970) developed a relationship between minimum contributing area of a basin and a basin moisture

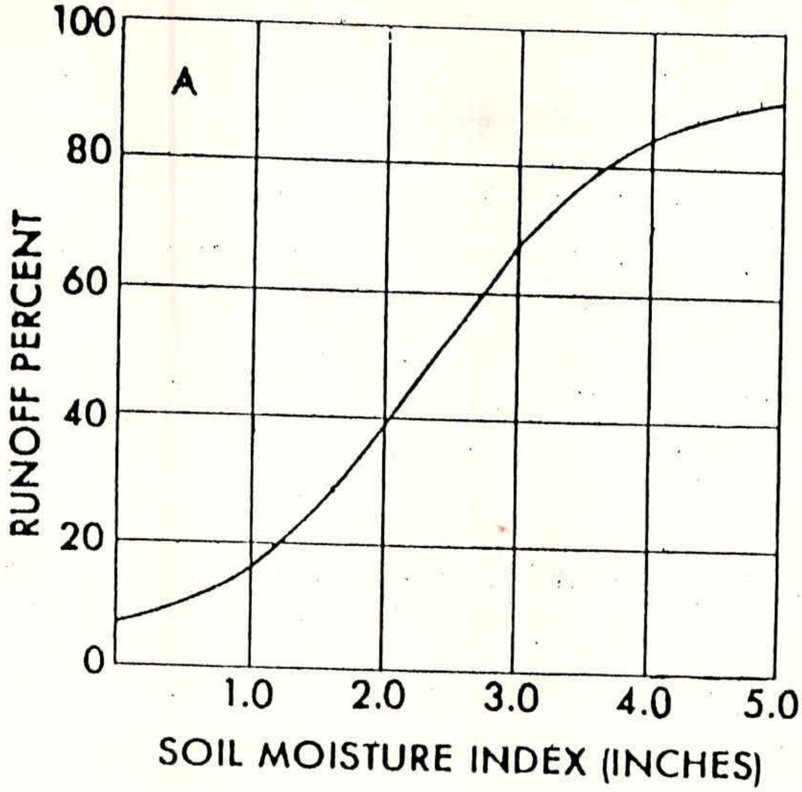

FIG: 3.15 SMI-RUNOFF PERCENT (ROP) RELATIONSHIP (AS CONSIDERED IN SSARR MODEL) (ADOPTED FROM U.S. ARMY CORPS OF ENGINEERS 1972)

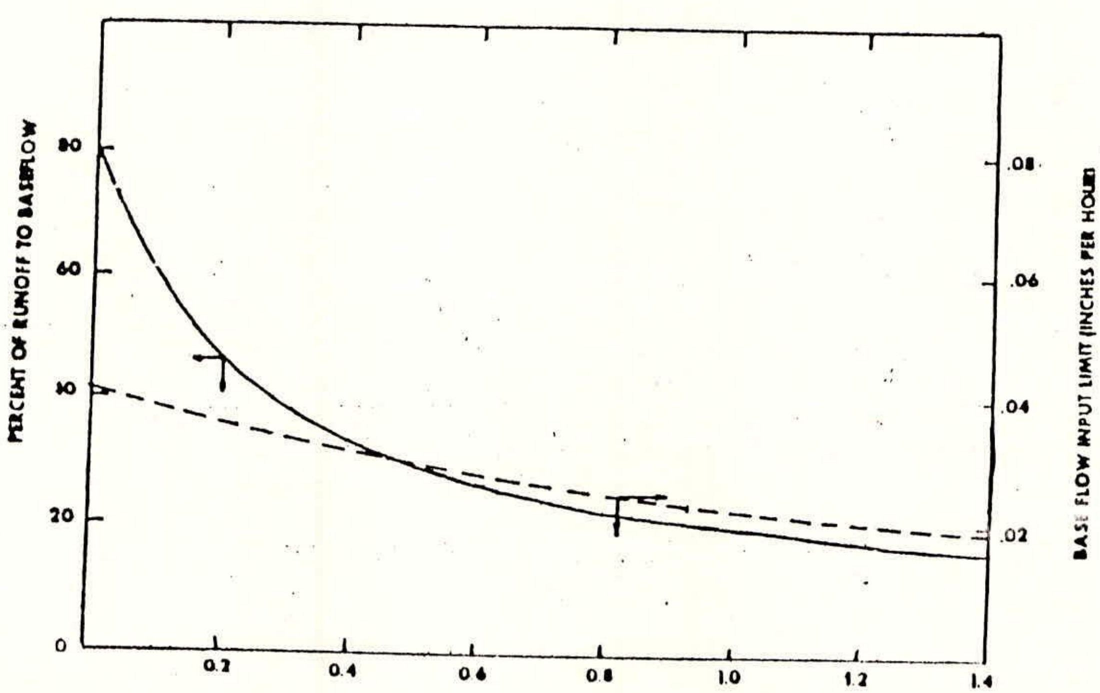

FIG: 3.16 VARIATION OF BASEFLOW PERCENT WITH BASEFLOW INFILTRATION index (which was a function of soil water and storm amount) where contributing area remains small until some moisture index threshold is reached after which contributing area increases rapidly to some upper limit imposed by the basin.

In this model the contributing area concept is used to calculate the volume of surface runoff from rainfall events which occur on snowfree 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:

$$CAP = SCT*(SMAV/SMAX)$$

volume of surface runoff is then computed as:

$$\text{VOL. Surface Runoff} = \text{Rainfall} *CAP* \text{area of HRU}$$

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

 70

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 up to SMAX.

3.4.5 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 parameters

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

3.4.6 Sacramento model

In this model, basin is considered to comprise of two types of basic areas (i) 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 streamflow 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.

3.4.7 USDAHL-74 model

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

$$f = a\, Sa^{1.4} + f_c$$

Where

f = Infiltration capacity in inches per hours

a = Infiltration capacity in inches per hour per inch of available 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.

3.5 Overland Flow

overland flow is surface runoff that occur in the form of sheet flow on the land surface without concentrating clearly in defined channels. This kind of flow is the first manifestation of surface runoff. After it reaches to channels it will become stream flow.

The interaction between overland flow and infiltration need to be considered since both processes occur simultaneously. The infiltration rate is quite high at the beginning of rainfall and the rate decay exponentially with time. The rate of infiltration varies depending on the type of 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 overlandflow water held in detention storage remains available for infiltration. Surface conditions like heavy turf or mild slope restrict the velocity of overlandflow, 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-overlandflow process requires continuous outflow rates from overlandflow.

Overlandflow can be calculated by different methods. Rigorous methods of numerical solution of the governing partial differential equations, the continuity and momentum equations requires substantial amount of computer time. In a natural catchment 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 overlandflow taken into consideration for the basin or the accuracy may be increased by dividing the basin into number of segments. The analysis depends on condition of ovrlandflow whether it is in laminar or in turbulent range.

The mathematical description of overland flow begins with the continuity equation of mass conservation. In one dimensional flow, this equation states that the change in flow per unit length in a control volume is balanced by the change in flow area per unit time.

$$\frac{dQ}{dx} + \frac{dA}{dt} = 0$$

This equation does not include source or sink. After inclusion of source or sinks which leads to;

$$\frac{dQ}{dx} + \frac{dA}{dt} = r$$

in which r = lateral inflow or outflow or net lateral flow per unit length.

In small catchment hydrology overland flow assumed to take place on the overland plane. This plane length L, slope So, and of theoretically infinite width. Therefore a unit width analysis is converted to ;

$$\frac{dq}{dx} + \frac{dh}{dt} = i$$

in which q = flow rate per unit width, h = flow depth i = lateral inflow per unit area,and t = time,measured from the onset of rainfall excess.

the solution of the overland flow is in the momentum equations can be written as;

$$q = \frac{c}{n} So^{1/2} h^{5/3}$$

in which q = flow rate per unit length, n = roughness coefficient, h = flow depth, c = a constant, and So = bed slope.

3.5.1 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 overlandflow in the laminar range, the volume of surface detention was also selected as the logical parameter in this model to relate surface detention with overlandflow. some useful approximation to natural behavior has been made.

Amount of surface detention is calculated as:

$$D_e = (0.000818 I\; n^{0.6}\; l^{1.6})/S^{0.3}$$

where D_e = Surface detention in ft^3/ft

I = Supply rate (rainfall) in inches per hour

S = Slope in ft/ft

l = length of overland flow in ft

n = Mannings roughness coefficient.

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

$$q = (1.486/n)\; Y^{5/3}\; S^{1/2}$$

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;

$$Y = 8/5 D_e/L$$

3.6 Percolation

3.6.1 Stanford water 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 wet zone occurs only when (UZS/UZSN) exceeds (LZS/LZSN). when this occurs, the percolation rate in inch/hour is determined from

$$PERC = 0.003(CB)(UZSN)\,[UZS/UZSN - LZS/LZSN]^3$$

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 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.

3.6.2 Sacramento model

In this model the mechanism of percolation 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 draining out of the lower zone. This limiting rate of drainage from the combined lower zone storage is expressed as:

LZFM=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= 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

Maximum percolation capacity = PBASE(1+Z)

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 (Lower zone Deficiency/Lower zone Capacity)]

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

$$= PBASE\left[1+Z \left[\frac{\text{Lower zone capacities less contents}}{\text{Lower zone capacities}} \right]^{REXP} \right.$$

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.

3.6.3 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 satisfied. 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 up to a fixed limit (GWPERC). Diagrammetically processes are shown in the figure 3.17.

The water that percolates to ground water is assumed to be

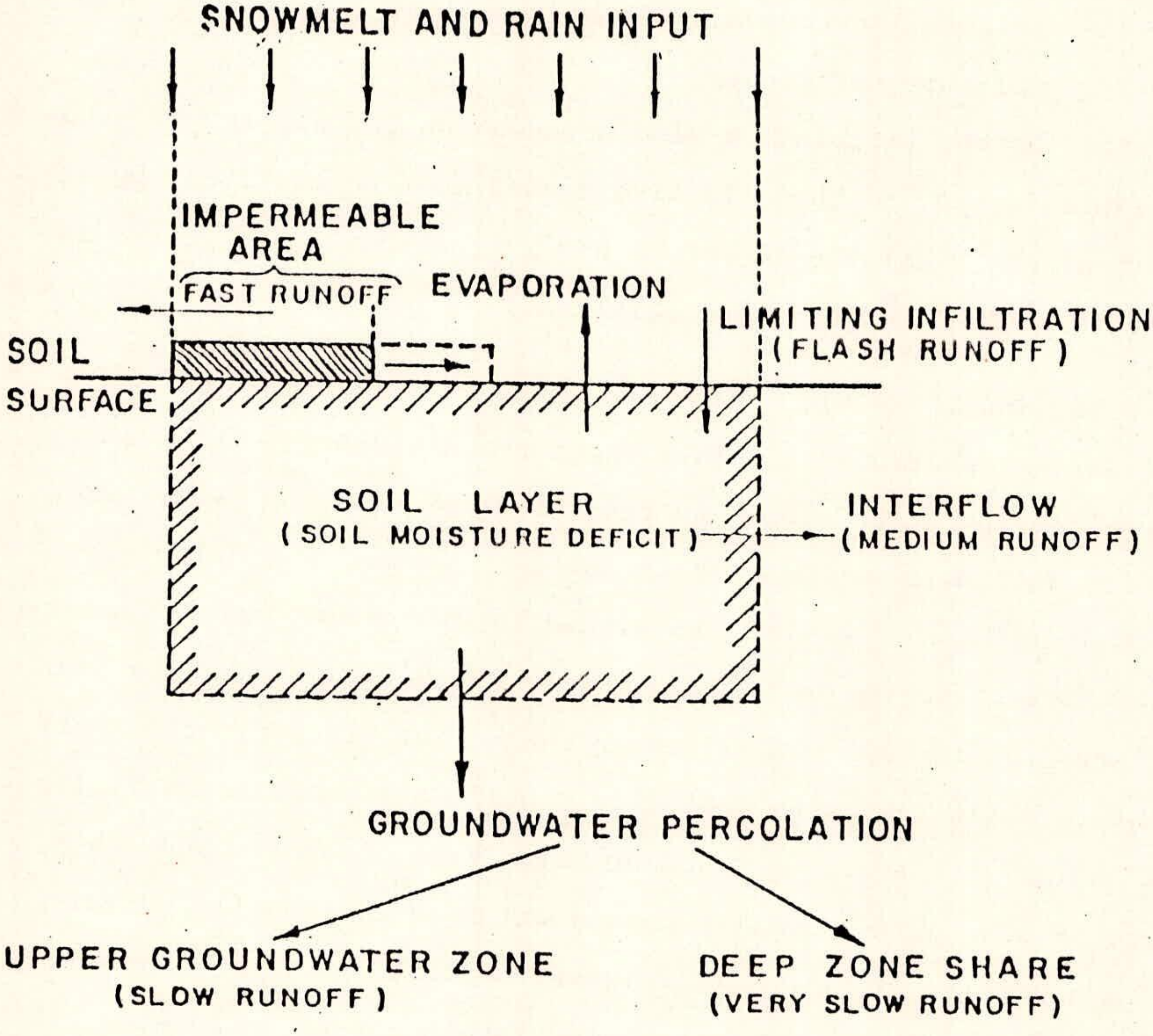

FIG: 3.17 MODEL OF SOIL LAYER AND SUBDIVISION OF RUNOFF COMPONENTS OF UBC MODEL (ADOPTED FROM UBC USER'S MANUAL) divided into two groundwater components, the upper ground water and the deep zone groundwater components. This subdivision of groundwater is controlled by DZSHRE, the deep zone share.

So upper Groundwater Zone Recharge = (1-DZSHRE)*GWPERC

Deep groundwater zone recharge=DZSHRE*GWPERC

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.

4.0 CATCHMENT HYDROLOGY IN OTHER COUNTRIES

Catchment hydrology has been studied for many decades elsewhere. The out come of the studies can be classified as two categories such as event-based studies and continuous-based studies. The event-based studies have been used to simulate for individual storms needed to solve the problems like design of hydraulic structures such as dams, culverts, bridges , spillways, urban and drainage, planning of flood control works evaluation of environmental impacts of landuse and management practices and planning of soil conservation works. Continuous -based studies have been applied for extending stream flow forecasting, supplementing of stream gauging programme, evaluation effect of landuse practices on catchment response, designing urban drainage; reservoirs etc., water quality modelling, flood mitigation and irrigation planning and management.

Large number of models have been developed and used all over the world. Some of the important models used has been discussed in this report.

4.1 HEC-1 Model

The HEC-1 flood hydrograph model was developed by the Hydrologic Engineering Center (1981) to simulate the direct runoff hydrograph (DRH) to precipitation by representing the watershed with interconnected hydrologic and hydraulic components. In addition, the model has options for multiplan-multiflood analysis, dam-break simulation, economic assessment of flood damage, and optimal sizing of flood control systems. This model has been extended to determining discharge-frequency relationships for ungaged watersheds (Hydrologic Engineering Center, 1982. This is perhaps the most comprehensive EBS model. Many elements of simulation are modeled using several options. Infiltration is estimated using four options (a) initial and uniform loss rate, (b) exponential loss rate,(c) SCS curve-number method, and (d) Holtan's infiltration equation. The DRH is estimated using the unit hydrograph method (with Clark, Snyder, and SCS dimensionless UH methods as options) and the kinematic wave method. The conic method, normal-depth storage and outflow, and modified Puls method are used for storage flow routing, whereas the lag and route and Muskingum methods are used for channel flow routing. A univariate search technique is employed to determine optimal model parameters. This model is one of the most commonly used models in the United States and can be used for hydrologic analyses under a wide variety of conditions (Feldman, 1981).

4.2 Soil Conservation Service Technical Report-20 (SCS TR-20) Model

The SCS TR-20 model was developed by the soil Conservation Service (1973) for inclusion of hydrologic process in project formulation. The primary objective was to improve the quality of watershed projects and, at the same time reduce overall costs by providing a means of analyzing alternative systems of structural measures. The model uses the SCS dimensionless hydrograph method to estimate surface runoff resulting from any synthetic or natural rainfall, which then is routed through stream channels using convex method and through reservoirs using storage indication method. It combines the routed hydrograph with the hydrographs from other tributaries and produce the discharges, their times of occurrence, and their water-surface elevations at any desired cross section or structure. The model provides for continuous analyses of nine different storms over a watershed under existing conditions and with various combinations of land-treatment flood water-retarding structures and channel improvement. These routings can be performed for as many as 120 reaches and 60 structures in one continuous run. The model has the flexibility to accommodate other aspects of watershed planning, provision of input data and use of engineering judgement (Kent. 1966).

4.3 USGS Model

The USGS model, developed by Dawdy et al. (1972) is a parametric rainfall-runoff simulation model for estimation of flood volume and peak rates of runoff for small drainage basins. Point rainfall and daily potential evapotranspiration data are used as input to the model. If more rain gages are available, then their records can be combined by the Thiessen polygon method to produce mean areal rainfall. A soil-moisture accounting is employed, considering infiltration, soil moisture accretion, and depletion, to determine the effect of antecedent conditions on infiltration. The flood-routing method developed by Clark (1945) is used to develop the basin unit hydrograph. The model has eight parameters, which are optimized using Rosenbrock's optimization method (1960). The sum of squared errors in logarithms of streamflows is used as an objective function. Errors of prediction result from both errors of rainfall data and approximations of the model. These two sources of error seem to be of the same order of magnitude. Rainfall errors have a magnified effect on the simulated streamflow estimates are approximately 10 mi^2. The errors of streamflow estimates are approximately linearly related to errors of rainfall data. The limit of accuracy of prediction of flood peaks with a single rain gauge seems to be on the order of 25 percent. The model has been modified to accommodate urban watersheds (Dawdy et al. 1978)

4.4 Hydrologic Model (HYMO)

A problem-oriented computer language for building hydrologic models (HYMO) was developed by Williams and Hann (1972, 1973). The HYMO was designed for planning flood-prevention projects, forecasting floods, and research studies. The model transforms rainfall data into runoff hydrographs using a two-parameter gamma distribution like the Nash model, wherein the parameters are estimated from their relationships with watershed area, slope and length-width ratio. These hydrographs are routed through streams and valleys or reservoirs. The variable storage coefficient (VSC) method is used for streamflow routing, and the storage-indication method is used for reservoir flow routing. Mannings equation is used to compute the normal flow-rating curve used in the VSC method. The model requires that rating curves must be available at enough locations along a valley to describe adequately the hydraulics of stream and valley. Most of these rating curves are computed. The modified universal soil loss equation is used to compute the sediment yield for individual storms on watersheds. The modified universal soil loss equation is used to compute the sediment yield for individual storms on watersheds. The model is simple, efficient, and flexible for its scope is limited to flood routing.

4.5 Storm Water Management Model (SWMM)

A storm-water management model (SWMM), developed by Metcalf and Eddy, Inc., et al (1971), was originally designed to represent urban storm water runoff for purposes of assisting administrators and engineers in the planning, evaluation and management of overflow abatement alternatives. The model has since been modified to accommodate rural watersheds. It represents storm-water runoff, both quantity and quality, from the onset of precipitation on watershed, through collection, conveyance, storage and treatment systems, to points downstream from outfalls that are significantly affected by storm discharges. The quantity submodels represent urban runoff, dry-weather sewage flow, infiltration, collection and main sewer system, storage simulation, and receiving water for river, estuary, or lake Zaghloul (1983) has studied sensitivity of these model parameters. The quality submodels are for runoff quality, dry weather flow quality, changing quality and routing during transport, receiving water quality, and treatment simulation. A cost effectiveness model include rainfall hyetograph, watershed characteristics, land use, gutter and pipe characteristics, street cleaning, storage facilities, inlet characteristics, treatment devices, and indexes for costs of facilities, For large watersheds, this may not be a suitable model due to its excessive detail.

4.6 Watershed Hydrology Simulation (WAHS) Model

The watershed hydrology simulation (WSHS) model, developed by Singh (1983,1987), is designed for prediction of the DRH for a specified rainfall event from an ungaged watershed. Rainfall hyetograph, observed at one or more points, constitutes input to the model. In addition, soil-vegetation-land use and geomorphic characteristics are needed to estimate model parameters. The model is a two-parameters. linear model, wherein the watershed unit hydrograph is determined using geomorphologic concepts involving one parameter-the watershed lag (Singh and Aminian, 1984, 1985)-estimated simply from watershed area. The direct runoff(DR) amount is obtained from the SCS curve number method. Then the effective runoff hydrograph (ERH) is estimated using the philip tow-term infiltration equation, where the steady infiltration parameter is obtained from soil characteristics and the sorptivity term comes from satisfying the continuity equation. If streamflow observations are available then the DR amount is obtained by baseflow separation. If needed information on soil characteristics is not available. the Rosenbrock-Palmer algorithm is provided to optimize model parameters based on minimizing the sum of squares of deviations between observed and computed DR peaks over a number of rainfall-runoff events. The model has been verified on more than 40 watersheds in the United States, Italy and Australia, with errors of less than 30 percent in predicted DR peak and its time. The model is simple, is suitable for ungaged watersheds, and can accommodate watersheds of varying sizes and land-use practices.

4.7 Rainfall-Runoff Routing (RORB) Model

The rainfall runoff routing model(RORB), developed by Laurenson and Mein(1983), is an interactive stream flow-routing program used for flood estimation, design of spillways and detention basins and flood routing. The model can be used for rural urban or partly urban catchments. Selvalingam et al (1987)applied it to an urban watershed in Singapur. With slight modification it can be used for flood fore-casting. Flood can be routed with single and multiple reaches,networks of streams,and lateral inflow and outflow. The model computes watershed losses and streamflow hydrographs resulting from rainfall events and other forms of inflow to channel networks(Mein et al.,1974). A loss model operates on rainfall to produce the effective rainfall. Rainfall and losses occur on the land surface before water enters the channel network. A flood routing model route flows through the channels received as direct run-off hydrograph,diversions,lateral inflows,and so on. variation of model parameters to obtain satisfactory fit of the calculated flows to the observed flows is not provided for by the model and has to be done by the user. The model is areally distributed and nolinear. It allows time variation of loss parameters and can model at any number of gauging stations. The model simple and efficient. Data requirements variable and matched to the particular problem at hand.

4.8 Watershed Bounded Network Model(WBNM)

The watershed bounded network model (WBNM) was developed by Boyd et al. (1979a, 1979b), and is, in general, similar to the RORB model. The main difference between these models is that the WBNM was two different types of storage which correspond to the two different types of subareas that comprise a watershed divided along watershed lines. These subareas are ordered basins and interbasin areas. The collection and drainage of water are different for these two types of subareas. The model comprises storage elements, each of which represents a watershed subarea, connected in the same arrangement as the stream network. The storage parameter of each element is related to the geomorphological and hydrological characteristics of the watershed, with the result that only one fitting parameter is used in the model. The model is simple, efficient, and applicable to ungaged as well as urban watersheds.

4.9 Flood Hydrograph Simulation model (FHSM)

The flood hydrograph simulation model (FHSM) was developed by Foroud and Broughton (1981) to estimate, taking into account storm and watershed characteristics, the design hydrograph and peak discharge for watersheds smaller than 400 Km^2. Antecedent moisture, rainfall loss, and runoff constitute the three main components of the FHSM. The antecedent moisture is accounted for by the use of antecedent precipitation index (API) which, in turn, determines the initial infiltration rate for a rainstorm. Kutilek (1982) has pointed out deficiencies in the API method. The ERH is determined using a modified form of Horton's infiltration equation and is then routed through a linear reservoir in conjunction with time-area curve. The model parameters are obtained by an optimization technique based on nonlinear least squares method. The model was applied to five watersheds and yielded less than 25 percent prediction error.

4.10 The Xinanjiang Model (XJM)

The Xinanjiang model (XJM) was developed by Zhao et al. (1980), Zhao (1985). This a conceptual model with distributed parameters corresponding to the various subwatersheds. The concept of runoff formulation is introduced to estimate the loss of rainfall to infiltration and then to estimate effective rainfall. The DRH is computed by the lag and route model and then routed through channels by the Muskingum method. The model has nine parameters, which are computed from observed hydrological data. Three of the parameters are somewhat insensitive, and the remaining four parameters need to be estimated carefully. The XJM has been widely used in humid and semiarid areas of the People's Republic of China.

4.11 The Guelph Agricultural Watershed Storm-Event Runoff (GAWSER) Model

The Guelph agricultural watershed storm-event runoff (GAWSER) model, developed by Ghate and Shiteley (1977, 1982), is a modified HYMO model incorporating areally variable infiltration. The model uses the Holtan infiltration equation; the time-area concentration curve; partitioning of infiltrated water to soil water storage, to subsurface runoff, and to groundwater storage and HYMO streamflow routing to produce runoff hydrographs from up to four soil groups on a watershed

The model has been used to produce estimates of surface and subsurface storm runoff generated by different types of soils for different storms, to assess pollution from land-use activities to evaluate effects of snow cover on runoff, and to study the effect of changes in land use and channel configuration on streamflow. The model has been tested thoroughly on three watersheds and has been found to be reasonably accurate.

4.12 The MIT MOdel

The MIT model, developed by Maddaus and Eagleson (1969), is a distributed linear model of direct runoff. Cascades of linear reservoirs, connected by linear channels and each having lateral input, are used to represent the watershed. Separate submodels of overland flow and streamflow allow simulation of the watershed response to spatially variable effective rainfall. The model parameters are related to physical features of the watershed. This model has the capability to handle spatial variability of rainfall and can be used to evaluate errors due to lumping of rainfall and to investigate the importance of inclusion of moving storms.

4.13 The Huggins-Monke (HM) Model

The Huggins-Monke (HM) model was developed by Huggins and Monke (1968,1970) and is reported by Huggins et al. (1975). This is a distributed model using the concept of sub-dividing the watershed into a definite number of small independent elements. The elements are assumed to be sufficiently small so that the hydrologically significant parameters are uniform within the element boundaries. The outflow from one element becomes the inflow for adjacent elements. Both interception and infiltration are subtracted from rainfall to determine the ERH. The ER satisfies depression storage or becomes surface runoff. All surface runoff is assumed to flow in the direction of each element's slope. Each element requires a definition of the interception six infiltration parameters, surface retention , hydraulic roughness, and slope direction and magnitude. Some of the model's parameters are determined from physical measurements of watershed characteristics. The model has been applied to both gaged and ungaged watersheds.

4.14 The Kansas Model

The Kansas model was developed by Smith and Lumb (1966) for large watersheds in Kansas. Areal nonuniformity of rainfall is handled by subdividing the watershed into Thiessen polygons. When daily precipitation amounts, hourly accounting is utilized. The subsurface component is modeled using a soil zone and a groundwater zone. The soil zone is divided into an upper zone and a lower zone. The groundwater zone is limited to the alluvial portion of the watershed. Subsurface drainage elsewhere in the watershed is treated as interflow in response to geologic and topographic considerations. Evapotranspiration is calculated at a potential rate based on mean daily temperature and discounted for moisture availability and depth. A lag and route procedure is used to develop the DRH. Based on limited testing, the model appears to simulate streamflow reasonably well.

4.15 The Institute of Hydrology Model (IHM)

The Institute of Hydrology model (HIM) is a physically based distributed model of watershed hydrology. In principle, the model parameters can be derived from experimental work or estimated a priori. Therefore, the model can be applied to ungaged watersheds or to predict hydrologic effects of land-use changes in gaged watersheds (Morris, 1980: Morris et al. 1980). The watershed is divided into hillslope areas represented by rectangular sloping planes and channel lengths represented by straight channels of constant cross section. Both channel and plane flows are modeled using one-dimensional form of the St. Venant equations for shallow water flow. Infiltration, throughflow and groundwater flow are treated together as saturated-unsaturated flow in porous medium described by Richards equation. In the IHM, evapotranspiration is determined using the Penman-Monteith equation and actual evapotranspiration in the root zone is calculated using the method of Feddes et al. (1976). The model has been applied to rural as well as forested watersheds. Rogers et al. (1985) undertook a sensitivity analysis of the IHM parameters and found that the model predictions were most sensitive to Chezy's roughness coefficient and saturated hydraulic conductivity.

4.16 Stanford Watershed Model IV (SWM)

The SWM, developed by Crawford and Linsley (1963, 1966) is perhaps the most widely accepted model for simulation of the land phase of the hydrologic cycle (Linsley and Crawford, 1960). The model has been applied to many watersheds throughout the world, and many modified versions of it have been developed (Fleming and

Bkacjm 1974; Liamas et al. 1980). Its applications have encompassed data extension, flood forecasting flood-frequency analysis, estimation of peak discharge (Clarke, 1968: Ligon and Law, 1972), sediment transport, effect of urbanization and land use practices (Cermak, 1979), and so on.

Hourly and daily precipitation, daily temperature, radiation, wind, monthly or daily evaporation, and a variety of watershed parameters constitute input to the SWM. Hourly or daily streamflow at the watershed outlet is the output. The time interval for calculation is 15 min. The model is a lumped parameter representation with 34 parameters. Most of these parameters are physically based, but four of them are obtained by using an optimization scheme, automatic or other wise. These four parameters pertain to infiltration, soil-moisture zones, and interflow. The remaining parameters (30) are evaluated from maps, surveys, or hydrometeorologic records. When snow melt simulation is not needed, the model parameters reduce to 25.

4.17 Ohio State University Model (OSUM)

Ricca (1972) developed the Ohio State University model (OSUM) by adding several significant capabilities to the SWM. Thus, OSUM is also a modified version of the SWM. It has the capability of using smaller than a 15-min time increment for streamflow routing and computing other components. The implication is that the OSUM can be applied to very small watersheds with a time of concentration of less than 15 min. Another modification is the addition of input parameters that control amount of water entering soil cracks and that recharge watershed swamps and marshes that may dry in summer. To accommodate geographic areas of stratified geology, several groundwater recession constants, compared for applications in regions with small amounts of snow and relatively little snow data. To facilitate the analysis of tabulated results, several subroutines are added for plotting of observed and simulated hydrographs, with rainfall hyetographs superimposed over selected storm hydrographs.

4.18 Streamflow Synthesis and Reservoir Regulation (SSARR) Model

The streamflow synthesis and reservoir regulation (SSARR) model was developed by U.S. Army Engineer Division, North Pacific (1975) for operational use in hydrologic engineering studies, reservoir regulation, and daily streamflow forecasting on large watersheds. The model has been successfully applied to several major projects on such large rivers as the Columbia River, the Mekong River, and the like. Both rainfall and snowfall events are considered in the model.

This is a lumped parameter representation model, with more than 24 parameters obtained by trial-and-error optimization. Daily rainfall, temperature, insolation, and snowline elevation constitute input to the model. Daily streamflow summary is the model input. The interval of calculation can be from 1 to 24 h. Unlike SWM, this model is much more simplified in representation of components. Based on an index, the moisture supply is divided into soil-moisture storage and runoff, which, in turn based on indices, is divided into baseflow, subsurface flow, and surface runoff. Major emphasis is placed on flow routing through channels, reservoirs, lakes, and so on. In a comparative study on a 2000 Km^2 watershed in north Ontario, Kite (1978) found significant differences between performances of NWSRFS and SSRR models.

4.19 Antecedent Precipitation Index (API) Model

The antecedent precipitation index (API) model is one of the simplest and oldest models (Sittner et al. 1969; Sittner, 1973). Four components are included in the model; an API rainfall-runoff connection, a unit hydrograph method, groundwater recession, and a method of relating groundwater flow to direct runoff. Precipitation volumes are input to the model, and both groundwater flow and DRH are output. The model must be calibrated on each watershed to obtain reliable simulated streamflows and hence cannot be applied to ungaged watersheds. It has been tested on watersheds of 68 to 817 M^2i in area.

4.20 Texas Watershed Model (TWM)

Another modification of SWM, called the Texas watershed model (TWM), was made by Claborn and Moore (1970). The TWM incorporates new parameters to describe evaporation, infiltration, and soil water movement. The authors argue that these changes help TWM better model small watersheds than SWM. The snowmelt component is deleted from TWM. The time of computation is considerably reduced and can be as small as 5 min. Rainfall for time increments from 1 min to 1 day, monthly evapotranspiration, and pan evaporation values, initial soil moisture, time area diagram, and watershed parameters constitute input to the model. If observed streamflows are supplied, the TWM makes several types of statistical analyses.

4.21 USDA Model

The U.S. Department of Agriculture Hydrograph Laboratory (USDAHL) model was developed by Holtan et al. (1975) primarily for agricultural watershed engineering by including the effects of soil types, vegetation, pavements, and farming practices on infiltration and overland flow. The model is a lumped parameter representation and has been applied to several small watersheds in the United States. A watershed is divided into as many as four distinct land-use or soil-type zones. The various components of the hydrologic cycle are computed for each zone. There can be as many as 41 parameters for each zone.

Input to the model is constituted by continuous records of precipitation, weekly averages of daily mean temperatures, weekly average pan-evaporation amounts, and data on soils, vegetation, land use, and agricultural practices. Runoff, return flow, and groundwater recharge form the model output. In addition to predicting streamflow, the model has been applied to simulate soil erosion, transport of chemical and environmental impact assessment.

4.22 Tank Model

The tank model was first introduced by Sugawara in 1961 and has since undergone a number of revisions until its final version reported by Sugawara et al. (1984). The main idea of the tank model is to represent the zonal structure of groundwater. Tanks rearranged in series and or parallel better to reflect watershed heterogeneity. Each tank has one or more outlets located at the bottom and sides. If a tank has only one outlet, then it is analogous to a reservoir, linear or nonlinear. The tank model has been applied to many watersheds in Japan and elsewhere and has been found to yield reasonably accurate results for both forecasting and simulation. The model parameters can be estimated either by trial and error or by an automatic optimization procedure, as done by it builders. Rainfall and discharge measurements constitute input to the model. Other information on daily evaporation, soil moisture, infiltration, and so on is helpful in model calibration. The time interval for computation can be taken to be as small as practically desired. A major problem with the tank model is that it requires some expertise to get a good model because the structure and parameters of the tank model must be determined subjectively by the analyst. Ozaki (1980) presented a mathematical formulation of tank model with a nonlinear feedback system.

4.23 Systeme Hydrologique European (SHE) Model

Three European organizations (the British Institute of Hydrology, the Danish Hydraulic Institute, and the French consulting company SOGREAH) jointly developed the Systeme Hydrologique Europeen (SHE) model, which has been reported by Abott et al (1986a 1986b), and Bathurst (1986a, 1986b). It is well suited for modeling the impact of people on land use change and water quality and other commercial uses. The model is physically based and considers spatial distribution of watershed parameters, rainfall, and hydrologic response. Its physical basis and flexible operating structure accommodate a broad range of data. Modeled either by finite difference representation of the partial differential equations of mass, momentum and energy conservation or by empirical equations derived from independent experimental research. The model has 18 parameters of which soil and resistance coefficients are most important. Several of the parameters have physical meaning and can be estimated from watershed characteristics. Rainfall, meteorological data, vegetation, and watershed characteristics constitute input to the model. The model is quite flexible in that it is relatively easy to add new components such as water quality and sediment yield. The results from watersheds in Europe and elsewhere have been reported to be very promising (Storm and Jensen 1984).

4.24 The Hydrological Simulation (HYSIM) Model

The hydrological simulation (HYSIM) model was originally developed for flood forecasting purposes using a minicomputer (Manly 1977, 1978a 1978b). The model has 17 parameters, most of which are determined from physically measurable watershed characteristics thus rendering the model suitable for application to ungaged watersheds. Precipitation and potential evapotranspiration data are input to the model. In addition snowmelt rates, sewage discharges, and river and groundwater abstractions can be used if available. The data can be monthly, daily, or of any shorter time interval. The model simulated daily flows 86 and monthly flow with the higher correlation coefficient, where the model parameters were estimated from watershed characteristics and without reference to measured flows.

4.26 Tennessee Valley Authority (TVA) Model

The Tennessee Valley Authority (TVA) model was developed as part of the Upper Bear Creek Experimental Project of the Tenessee Valley Authority (1972). It is a simple model with three main components runoff, soil moisture, and evapotranspiration. The model has 16 parameters, 5 of which are primary parameters the remaining 11 are secondary parameters. The primary parameters are linked to watershed measures but must be optimized The secondary parameters can be readily determined or in some cases may be computed. The model has been found to reproduce about 85 percent of the variation of daily flows when compared to observed flows and reproduces monthly and annual flow volumes within close limits of observed flows. The model has also been applied to simulate suspended sediment transport and potassium loads, to detect changes in water quantity and quality attributable to clear cutting of forest, and to quantify hydrologic effects of land use changes (Betson et al. 1980).

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5.0 CATCHMENT HYDROLOGY IN INDIA

5.1 Regional Unit Hydrograph Studies Conducted in India.

The Small Catchment Directorate of Central Water Commission (CWC) (1980) has developed a relationships between the 1-hour unit hydrograph parameters and physical characteristics of 22 catchments for Godavary basin subzone 3f.

CWC (1980) also developed the regional unit hydrograph relationships for Mahanadi basin subzone 3d analysing the data of 16 catchments of the basin.

Mathur and Vijay Kumar (1982) related the Physical parameters of 20 small and medium catchments with an objective to find out the most effective physical parameter representing the regional unit hydrograph relationships.

Huq et al. (1986) developed generalised synthetic unit hydrograph relationships analysing the data of 21 catchments in Lower Gangetic plains, Mahanadi Basin, Krishna Basin and Brahmaputra Basin. They have related the parameters of the representative unit hydrograph with a suitable combination of the physical characteristics of the catchments using regression analysis.

The small catchment directorate of CWC (1982) developed the regional unit Hydrograph relationships for Krishna & Penner basins relating the physical parameters of 21 catchments with their one hour representative unit hydrograph parameters.

The small catchment Directorate of CWC (1984) derived the relationships relating the physical parameters of the 23 catchments of upper Indo- Ganga plains with their representative 2-hour unit hydrograph parameters.

Singh (1984) of National Institute of Hydrology developed the regional unit hydrograph relationship relating the physical parameters of six catchments of Godavari basin subzone of with average parameters of Nash and Clark Model for each catchment are estimated using the average parameters obtained by taking the geometric means of the parameters of the unit hydrogaph for individual storm for the computal model approach. For the Collins and Least square methods, the standard averaging procedure is used to estimate the average unit hydrograph.

5.2.0 Regional Flood Frequency Analysis in India

Goswami (1972) carried out regional flood frequency analysis for Brahmaputhra basin in North East India, using USGS procedure. He analysed annual peak flood series data for 25 sites for 1955-70 for catchment areas ranging from 63 to 69230km^2. The mean annual flood Q (m^3/Sec) for 2.33 year return period was graphically related with catchment area A (km^2).

Thiruvengadachari et al (1975) carried out regional frequency analysis using USGS procedure and annual flood series data for 16 small and medium catchments ranging in size from 133 to 8500Km^2 in magnitude having exceedance probability of 0.43 as index flood $Q_{0.43}$(m^3/sec) which was related to catchment area and mean annual rainfall.

Seth and Goswami (1979) carried out regional flood frequency analysis for ten tributaries of river Brahmaputra in North Eastern India with available annual flood series varying in length from 11 to 25 years. Three techniques utilised in the study include:(a) using annual flood series of all stations in the region having more than 10 years of record, (b) extension of short records of some streams by developing suitable relationships with concurrent peak flood records of streams with long record and (c) adjustment of statistical parameters obtained from short records by means of statistical parameters obtained from longer records of neighbouring stations.

Jakhade et al (1984) have applied regional flood frequency approach of USGS to analyse the data of fourteen sites (having data for 10 or more years) in the Brahmaputhra valley divided into two different hydrometeorological zones.

Seth and Perumal(1985) of National INstitute of Hydrology have carried out regional flood frequency analyses for the region of subzone 3-d of Mahanadi basin with annual peak flood series data available at 18 stations for varying number of years. The following three methods were used for the analysis: (i) the index flood method (ii) method based on normalisation of peak flood data of different sites with reference to the respective site mean values and combine these normalised values to form a single series for regional analysis. and (iii) the method based on regional parameters of Wakeby distribution and James-Stein corrected means. Out of 18 bridge sites, the data of 15 sites were analysed and the data of remaining 3 sites has been used for verification of the results obtained from the anlysis.

Huq et al (1986) have attempted to evolve the frequency flood formulae for country wide application using the frequency storm rainfall and runoff models. For developing relationship for 50 year flood peak taking into account catchment area, statistical or equivalent slope of the stream , and 50 year return period 24 hour rainfall for use for ungauged catchments up to 5000 km^2 in size. The country was divided into four distinct categories of areas such as alluvial plains of Indus, Ganga and Brahmaputra river system with equivalent slope up to 1.5m/km ,for equivalent slopes above 1.5m/km, for remaining areas with statistical slope up to 3.5m/km and for statistical slope above 3.5m/km and evolved their respective flood formulae.

Seth and Singh (1987) of National Institute of Hydrology have further used frequency storm rainfall and runoff models with Wakeby distribution using data of catchments for three typical regions which are (i) Lower Godavari basin subzone 3f, (ii) Brahmaputra basin and (iii) Sub Himalayan region.

5.3.0 Application of SHE model in India

The following six tributary basins of Narmada has been applied for SHE model By the National Institute of Hydrology.

i . Narmada upstream of Manot (4980 km^2)

ii . Hiran upstream of Patan (4064 km^2)

iii. Sher upstream of Belkeri (1345 km^2)

iv . Marna upstream of Berely (1530 km^2)

v . Kolar upstream of Satrana (820 km^2)

vi . Ganjal upstream of Chhidgaon (1730 km^2)

In this application study, SHE model has been used for modelling entire landphase of hydrological cycle for above selected sub-basin of Narmada. Simulation has been done for the reproduction of streamflows volume, peaks, hydrograph and groundwater.

The computational grid network and channel system was set up for the basin, forming basis for the spatial distribution of topographic elevation, soil type, land use and rainfall stations in the data files. The basis network was composed of grid squares of 1kmx1km but for simulation work with grid size 2kmx2km. Four land uses were identified such as agricultural land, dense mixed forest and waste land. Soil has been categorised for low land, semi-hilly and hilly areas. The calibration was carried out for the period of 3 years and validation for 2 years.

5.4.0 Application of Tank Model in India

The Tank model developed by Sugawara for daily analysis has been used to simulate streamflow of Malaprabha sub-basin upto Khanapur and Malathi sub-basin upto Kalmane by K.S.Ramasastri of National Institute of Hydrology (1988-89).

The calibration of the model was carried out by observed flows for two years. The calibrated model has been used to simulate daily streamflow which were found to be within the 10% of observed streamflow . The overall performance of the model is found to be satisfactory.

6.Ø WATER YIELD

Water yield is the volume of the water available from a stream at a specified point area and at specified period of time. The water yield is determined at the point of outlet of catchment and the period of time a day or longer. The thrust is given on the volume of the flow rather than the instantaneous discharge and reflects the volumetric relationship between rainfall and runoff. There are many factors which affect water yield depending upon the period of the determination, meteorologic factors, and catchment characters. The most important meteorological factors are space time distribution of precipitation amount, intensity and duration and space time distribution of evapotranspiration. The important catchment factors include surface vegetation, soil moisture, soil characteristics, surface topography and drainage density.

The estimation of water yield will be helpful in number of water resources problems such as, design of storage facilities, determination of water availability for industrial and agricultural use, dependable water supply for power generation, planning irrigation operation and design of irrigation projects.

There are many approaches to determine the water yield. It may be classified as empirical approaches and continuous time simulation approaches which are employed in the determination of water yield. Models can be classified as (i) Volumetric storm rainfall-runoff models (ii) Monthly volumetric rainfall-runoff model and (iii) Yearly volumetric rainfall-runoff model. Several models have been developed to estimate the water yield, some of the models have been discussed in the report.

6.1.0 Volumetric Storm Rainfall-Runoff Model

These models have been developed to estimate direct runoff from storm rainfall. These models variously account for factors affecting storm rainfall- runoff relationship as the amount of rainfall, duration of rainfall, the antecedent soil moisture conditions and watershed storage.

6.1.1 Hamon Model

Hamon (1963) developed a model, based on the water balance approach, for estimating direct runoff from small watersheds employing storm rainfall V_P antecedent soil-moisture index M_I and the amount of rainfall retained V_r before the beginning of direct runoff V_Q The model utilized the water budget equation.

$$V_P = E + V_G + V_Q + \Delta S$$

in which, E represents evapotranspiration, V_Q is direct runoff, V_G is percolation loss to the ground, and ΔS is the change in soil moisture storage.

Antecedent Soil-Moisture Index M_I

The antecedent soil moisture index is defined as the predicted amount of moisture in the selected upper horizons of the soil profile in excess of the amount existing under extreme drought conditions. Hamon (1963) considered approximately 45 cm of the soil profile. The soil moisture content at this dry condition approximately corresponds to that under a tension of 15 atm.

In this model M_I is represented by two zones, one upper zone and one lower zone. The amounts of moisture to be retained in these zones are equivalent of their holding capacities, which are considered to be different for different parts of the year. When the amount of rainfall retained exceeds the holding capacity of the upper zone the excess is passed on to the lower zone. When the capacity of the lower zone is exceeded the excess goes to the groundwater. The moisture in the upper zone is available for evapotranspiration which is computed using readily available long term temperature records (Hamon 1961) as

$$E_p = kT_sD_v$$

where T_s is possible duration of sunshine in units of 12 h, D_v is saturated vapour density in grams per cubic meter at the daily mean temperature and k is a constant (equal to .00065 if E is in inches)

The moisture content in the lower zone is available for evapotranspiration at a reduced rate in proportion to the reduction in its amount. The reduction in ET is specified differently for different parts of the year.

Percolation Loss V_G

The daily percolation loss in centimeters is represented as

$$V_G = \left(\frac{2M_a - W}{W}\right)^2 .125W$$

Where $V_G = 0$ when $M_a \leq 0.5W$, M_a is actual initial moisture in excess of that at 15 atm tension, and W is total moisture minus the moisture at 15 atm tension. This Equation is used to compute percolation from both the upper and lower zones on the assumption that V_G from the layer at approximately 100 cm is representative of V_G from the upper soil layers.

Retained Rainfall V_r

A certain portion of rainfall is retained in the watershed before runoff begins. It includes the rainfall initially infiltrated and retained as interception and surface storage. V_r is related to M_I as

$$V_r = a - bM_I$$

where $a > 0$ and $b > 0$ are constants that depend upon watershed characteristics. This relationship is consistent with that obtained by Hartman et al (1960).

Runoff V_Q

An equation for computation of V_Q is developed from the basic concepts of the curve number method developed by Soil Conservation Service (1964). Accordingly,

$$\frac{V_Q}{V_p} = \frac{(V_p - V_Q)}{V_R}$$

Where V_R is the potential retention of rainfall and depends upon V_r available moisture in the upper zone and percolation into the lower soil profile. It is clear from the previous discussion that a certain amount of rainfall is retained before runoff begins. Then Substituting $V_P - V_r$ in place of V_p and simplifying.

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$$V_Q = \frac{(V_p - V_r)^2}{(V_p - V_r + V_R)}$$

with the condition that $V_Q = 0$ for $V_r \geq V_p$. This equation is used by Soil Conservation Service for predicting storm runoff, where V_r is replaced by $.2V_R$. Hamon suggested that,

$$V_R = c + kV_r$$

where c and k are parametres to be obtained for each watershed by curve fitting. Hence V_Q can be written as follows,

$$V_Q = \frac{(V_p - V_r)^2}{[(V_p - V_r) + (c + kV_r)]}$$

Like most models based on the water balance equation, the key component in the model is determination of M_I. First some initial conditions are known about the variables in equation. Then M_I is computed by daily accounting of V_p, V_Q, E, and V_G. For known M_I, V_Q can be computed. Hamon (1963) plotted a family of curves relating V_Q to V_p for various values of M_I which can be used effectively for application of the model.

6.1.2 Singh-Dickinson (S-D) Model

The S-D (Singh and Dickinson 1975a 1975b) also based on the water balance approach, is similar to the Hamon model in concept. The model was developed for small agricultural watersheds in southern Ontario , Canada.

Antecedent Soil Moisture Index M_I

The antecedent soil moisture index is expressed in terms of antecedent soil-moisture deficiency M_a which is obtained by using a soil moisture model (Singh 1970, 1971) Singh and Dickison 1975a). The moisture deficiency existing prior to the occurrence of rainfall is the amount of moisture necessary to bring available moisture in a soil profile to its field capacity. The soil moisture model estimates daily soil moisture from the upper 75 cm of the soil profile. The profile is represented by five zones of varying depths and water storage capacities . This representation is based on the soil characteristics of the watersheds in question, as enumerated by Webber and Tel (1966). The depth of soil profile and the number of zones representing it may vary from one watershed to another.

Fundamental to determination of soil moisture is simulation of wetting and drying phases of the soil profile. The wetting of zones depends on rainfall and occurs in a sequential order from the uppermost to the lowermost zone each filling to its capacity before discharging to the next-lower zone. There is a maximum amount of water that a given zone can hold. When rain occurs, the moisture content of the uppermost zone begins to fill. The amount of water in excess of the maximum capacity of the zone is percolated down. The wetting of other zones takes place in a similar manner. When all the zones have attained their maximum moisture contents, additional rain is treated as runoff.

Under saturated conditions, the soil moisture is depleted by evapotranspiration and vertical drainage. Evapotranspiration occurs at a potential rate. Under unsaturated conditions, the drying of zones is caused by evapotranspiration at actual rate as controled by available moisture.

Evapotranspiration is distributed over the various zones for simultaneous extraction of moisture from them. This distribution corresponds to the extraction coefficients assigned to the zones, which are considered to be a function of plant root development, soil characteristics, and climatological factors. The coefficients satisfy.

Retained Rainfall(V_r)

A certain portion of rainfall is retained by the vegetation and the surface depressions before runoff commences. This part of rainfall lumps interception loss and depressional storage and is called retained rainfall V_r is expressed as

$$V_r = a_0 + a_1 M_a + a_2 M_a^2 + a_3 M_a^3$$

Where M_a is antecedent moisture deficiency to be determined as explained before and a_i, $i=0,1,2,$and 3 are coefficients. The coefficients a_i are to be determined empirically. Singh and Dickinson (1975b) estimated them by selecting a number of rainfall events that did not produce any significant runoff and fitting a cubic polynomial between them and corresponding moisture deficiencies.

Runoff (V_q)

Using the work of Kohler(1963), V_q was determined by

$$V_q = \alpha[V(V_p - V_r)^\beta + M_a^{1/\beta} - M_a]$$

where α and β are watershed coefficients. α appears to account for such losses as not accounted for by retained rainfall. β

depends on M_d and is expressed as

$$\beta = a + bM_d$$

6.1.3 Multicapacity Accounting Model

The multicapacity accounting model was developed by Kohler and associates (Kohler and Richards 1963; Kohler 1963a. 1963b) principally for river forecasting. The concept of soil moisture accounting embodied in this model is incorporated in the U.S. Weather Bureau's model of river forecasting.

The principal hypothesis is that soil moisture capacity varies spatial within a watershed and its variation can be adequately expressed by several average values, for example 5,15,30, and 45 cm. The area represented by each capacity is not explicity considered in computations.

Soil moisture deficiency is the fundamental parameter in the multicapacity accounting model, which is estimated by maintaining a daily water balance. The mean deficiency for the watershed is the weighted average of the deficiency values associated with the various capacities. The weights are determined by correlation analysis to produce the best index to storm runoff.

The soil profile is divided in two zones one upper zone and one lower zone. The moisture in the upper zone is always depleted at the potential rate any deficiency in this zone must be satisfied before rainfall begins to recharge the lower zone. Depletion of moisture from the lower zone occurs only when that of the upper zone is completely depleted and is assumed to be proportional to the available moisture. Evaporation is computed with the help of nomogram (Kohler et al. 1955).

Total Runoff

Total runoff (direct runoff and groundwater runoff) is computed using the following relation

$$V_{QT} = (V_p^{\beta} + M_d^{\beta})^{1/\beta} - M_d$$

where

$$\beta = a_0 + a_1 M_d$$

in which a_0 and a_1 are constants

This model computes total runoff directly, which includes groundwater. The direct runoff can be determined by exclusion of groundwater from total runoff. This model is flexible and yields satisfactory results for river forecasting.

6.1.4 SCS Curve Number Model

The SCS model was developed by the soil Conservation Service (1964, 1973) for estimating direct runoff from storm rainfall on small ungaged watersheds. Because of its simplicity and reasonable accuracy, it is one of the most frequently utilized methods for runoff estimation (Hawkins 1973; Chiang, 1975, Hjelmfelt 1980; Ragan and Jackson, 1980; Chong and Teng, 1986). It employs rainfall and watershed data that are ordinarily available or easily obtainable. The model can be applied to large watersheds with multiple land uses.

The fundamental hypothesis of this model is two fold (1) runoff starts after an initial abstraction V_r has been satisfied.

This abstraction consists principally of interception, infiltration, and surface storage. (2) The ration of actual retention of rainfall to the potential maximum retention V_R and equal to the ratio of direct runoff to rainfall minus initial abstraction. Mathematically,

$$\frac{V_p - V_r - V_q}{V_R} = \frac{V_Q}{V_p - V_r}$$

This can be written as

$$V_Q = \frac{(V_p - V_r)^2}{(V_p - V_r) + V_R}$$

V_r can be expressed as a function of V_R. The Soil Conservation Service Stated $V_r = .2V_R$ Physically this means that for a given storm, 20 percent of the potential maximum retention is the initial abstraction before runoff begins. The remaining 80 percent is principally infiltration after runoff begins. Therefore,

$$V_Q = \frac{(V_p - .2V_R)^2}{V_p + .8V_R}$$

This equation is the SCS curve number model and is a one parameter model containing V_R as the parameter.

Determination of V_R

The parameter V_R depends upon characteristics of the soil cover complex and antecedent soil moisture conditions in a watershed. For each soil cover complex, there will be a lower limit and an upper limit of V_R. The Soil Conservation Service expresses V_R as a function of runoff curve number C_N.

$$C_N = 1000/\ V_R + 10$$

or

$$V_R = 1000/C_N - 10$$

C_N is related to soil cover complex conditions. Obviously when C_N equals 100, V_R becomes zero. This leads to $V_Q = V_P$ when V_R --> infinity, C_N tends to zero. This yields $V_Q = V_P$ for all V_P when $V_R = 0$, $C_N = 100$. The soil conservation Service has developed runoff curve number for various hydrologic soil cover complexes, as shown in Table 6.1. Slack and Welch (1980) and Ragan and Jackson (1980) have estimated curve number from Landsat data. Wood and Blackburn (1984) showed that the hydrologic soil groups classification system was inadequate for arid and semiarid range land.

Initial Abstraction (V_r)

An exact determination of V_r is almost impossible. However for practical purposes V_r can be related to V_R. Based on analysis of data from a large number of small watershed, the soil conservation service found V_r to be roughly equal to .2 V_R. It can also be estimated by relating to an antecedent soil moisture index, as done by Hamon (1963) and Singh and Dickinson.

Hence equation for V_Q can be written as,

$$V_Q = \frac{(V_P - 200/C_N + 2)^2}{V_P + 800/C_N - 8}$$

Runoff curve numbers for hydrologic soil-cover complex (after Soil Conservation Service, 1975)

(Antecedent moisture condition II, and $V_i = .2V_R$)						
Cover			*Hydrologic Soil Group*			
Land Use	*Treatment or Practice*	*Hydrologic Condition*	*A*	*B*	*C*	*D*
Fallow	Straight row	—	77	86	91	94
Row crops	Straight row	Poor	72	81	88	91
	Straight row	Good	67	78	85	89
	Contoured	Poor	70	79	84	88
	Contoured	Good	65	75	82	86
	Contoured & terraced	Poor	66	74	80	82
	Contoured & terraced	Good	62	71	78	81
Small grain	Straight row	Poor	65	76	84	88
	Straight row	Good	63	75	83	78
	Contoured	Poor	63	74	82	85
	Contoured	Good	61	73	81	84
	Contoured & terraced	Poor	61	72	79	82
	Contoured & terraced	Good	59	70	78	81
Close-seeded legumes[a] or rotation meadow	Straight row	Poor	66	77	85	89
	Straight row	Good	58	72	81	85
	Contoured	Poor	64	72	83	85
	Contoured	Good	55	69	78	83
	Contoured & terraced	Poor	63	73	80	83
	Contoured & terraced	Good	51	67	76	80
Pasture or range		Poor	68	79	86	89
		Fair	49	69	79	84
		Good	39	61	74	80
	Contoured	Poor	47	67	81	88
	Contoured	Fair	25	59	75	83
	Contoured	Good	6	35	70	79
Meadow		Good	30	58	71	78
Woods		Poor	45	66	77	83
		Fair	36	60	73	79
		Good	25	55	70	77
Farmsteads		—	59	74	70	77
Roads (dirt)[b]		—	72	82	78	89
(hard surface)[b]		—	74	84	90	92

[a] Close-drilled or broadcast

[b] Including right-of-way

TABLE: 6.1 RUNOFF CURVE NUMBERS FOR HYDROLOGIC SOIL COVER COMPLEX (AFTER SOIL CONSERVATION SERVICE 1975)

In this equation, C_N is a parameter that is to be determined.

6.1.5 The R-Index Method

The R-index method was developed by Hewlett et al (1977a, 1977b) using 468 rainfall runoff events from 11 small (forested and wildland < 50 Km^2) watersheds in the eastern United States. According to Hewlett and Moore (1976) the variable source area concept serves as the basis of this method. Only those events that produced 25.4 mm or more of storm rainfall were included. The hydrograph separation (separation of direct runoff from baseflow) was based on the approach of Hewlet and Hibert (1967), wherein a line of constant slope of 1.13 mm/da is projected from the beginning of the rising hydrograph to the point where it intersects the recession limb of the hydrograph. Storm rainfall was taken to include all rain falling up to 2 hr. before the initiation of the storm hydrograph and until the termination of direct runoff.

The R-index method hypothesizes that stormflow depends on three factors, namely storm rainfall, antecedent moisture conditions, and the hydrologic or inherent, storage capacity of the watershed. In metric version (Hope 1983, 1984), the R-index method can be written as

$$V_Q = 25.4\ b_1 R(\frac{V_P}{25.4})^{b_2}[1.0 + (0.0136 Q_o)^{b_3}$$

where V_Q is direct runoff in millimeter R is the R-index, V_P is storm rainfall in millimeters, Q_o is initial flow rate in 1/s/km, and b_1 b_2 and b_3 are fitting parameters. By examining a number of rainfall, storage, and watershed variables. Hewlett et al (1977a) fitted from above equation to the observed storm flows using a nonlinear least squares, method (Marquardt, 1963). The addition of 1.0 to $(.0136Q_o)^{b3}$ prevents an indeterminate V_Q when V_p approaches zero.

The R-index can be computed as follows;

$$R = 1/n \sum_{i=1}^{n} [V_Q(i)/V_p(i)]$$

where n is the number of observations and $V_p(i)$ is storm rainfall In the eastern United States, the optimal values of the parameters were found ((Hewlett et al., 1977a) to be. $b_1 = .4$, $b_2 = 1.5$, and $b_3 = .25$. In order to enhance practical applicability of this method. Hewlett et al. (1977a) suggested an alternative for the initial flow rate as

$$S_I = \sin(360D/365) + 2.0$$

where S_I is the sine-day factor and D is the number of the day counted from November 21 = 0. The term S_I replace (1.0 + .0136 Q_o) in the above equation.

Hope (1983,1984) extended the R-index method to six small humid and subhumid near Grahamstown in South Africa. The physical and environmental characteristics of these semiarid watersheds differ markedly from those of the watersheds where the method was developed originally. Hope (1983) introduced antecedent rainfall AP_n as an alternative to Q_o which better reflected the effect of antecedent conditions on stormflow. The quantity AP_n as an alternative to Q_o, better reflected the effect of antecedent conditions on stormflow. Thus equation becomes

$$V_Q = 25.4 b_1 R (V_p/25.4)^{b_2} [1.0 + (AP_n)^{b_3}]$$

The quantity AP_n computed by totaling rainfall for a given antecedent period (n days), and the total was expressed as 10^{-2} for humid and sub-humid watersheds, Hope (1983) computed AP_n for 5-, 10-, and 15-day periods. The parameter values were markedly different for different antecedent periods. Using AP_n resulted in the most accurate estimates of V_Q for four watersheds whereas AP_5 and AP_{10} each was the best for one watershed. For semiarid watersheds, Hope (1984) used 1-,7- and 10-da periods and found them somewhat comparable. He found that the watershed wetness index was important for small and intermediate size events but not for larger events.

6.1.6 Coaxial Graphical Correlation

The coaxial graphical correlation method was developed by Linsley et al. (1949) and is discussed in Kohler and Linsley (1951) and Linsley et al (1975). This method represents perhaps the earliest satisfactory attempt to estimate storm runoff from a given volume of rainfall.

A large number of factors, some of them being interactive, affect the storm runoff. A detailed catalogue of these factors is given by Chow (1964). In their analysis Linsley et al (1949) selected the following independent variables: (1) antecedent precipitation index (API) (2) season or week of the year (3) storm duration and (4) storm rainfall. They replaced the dependent variable, storm runoff, by basin recharge defined as the difference between rainfall and runoff. The reasons for selecting these variables are discussed by Kohler and Linsley (1951). Nash (1966) has discussed the motivation underlying this selection.

The variables to be selected for graphical correlation may change from one watershed to another. This is shown by Witherspoon (1961), where he selected storm runoff as the dependent variable and antecedent precipitation index, cover condition, duration of the effective rainfall, and the amount of the effective rainfall as independent variables.

The API is a measure of soil moisture deficiency existing prior to the occurrence of a storm. By assuming a logarithmic recession of antecedent moisture. API can be defined during period of no precipitation as

$$I_{at} = I_{ao}k^t$$

where I_{at} is the API at time t,is the initial API, and k is a constant. t is usually in days. If t equals unity,

$$I_{a1} = kI_{ao}$$

Thus, the API for any day is equal to that of the previous day multiplied by k. If it rains on a given day, then the amount of rain must be added to the API of that day. The value of k usually ranges from .85 to .90 over most of the eastern and central portions of the United States (Kohler and Linsley, 1951), although it depends upon basin physiography. API can be computed either from average daily rainfall values over the watershed or from daily rainfall recorded at various stations, which are then averaged. The usual practice to compute API is either to start computations at the end of a dry spell with an assumed low value thereof or to start computations 2 or 3 wk in advance of the first storm with an assumed value equal to the normal 10-day precipitation of the season (Kohler and Linsley 1951).

The effective rainfall is the rainfall volume per unit area that generates runoff. The rainfall volume is determined from only those rainfall intensities that are in excess of a specified value, say 1.5 cm/h. depending upon the infiltration characteristics, interception, and detention storage. The duration of this effective rainfall is the effective storm duration.

The cover conditions can be defined variously. Their specification by the Soil Conservation Service (1975) is one example. Witherspoon (1961) classified these conditions as poor, intermediate, and good. If we follow his classification, then the cover conditions are defined as follows. Poor cover condition is when the surface is not covered by vegetation or vegetative residues. This is a common condition in agricultural watersheds prior to planting or subsequent to harvesting. Intermediate cover condition is when the surface is partially protected by vegetation. This occurs after planting but before the crop reaches its height. Good cover condition is when the surface is fully covered by vegetation, such as good grass cover or a fully grown.

6.1.7 Derivation of Graphical Correlation

The graphical correlation consists of four graphs, designated as A B C and D as shown in fig. 6.1 The construction of these graphs may involve the following steps. For illustration we use the variable selected by Witherspoon (1951).

1. Construct graph A, which is a three variable relation. Plot storm runoff versus API labeling the points with cover condition and fitting a smooth family of curves representing the various cover conditions.

2. Construct graph B, which plots computed versus observed runoff such that computed runoff is on the vertical scale matching exactly the vertical scale of graph. A Label the points with the effective rainfall duration. The computed runoff is obtained from graph A by entering API and cover condition. A smooth family of curves is then drawn, which incorporate the effect of the effective rainfall duration on storm runoff. Graphs A and B, when combined, represent a graphical relation for estimation of storm runoff from API cover condition and effective rainfall duration.

3. Construct graph C by plotting computed versus observed storm runoff, such that computed runoff is on the horizontal scale matching exactly the horizontal scale of graph B.. Label the points with the effective rainfall amount. The computed runoff is obtained from graphs A and B in manner similar to that of step (2). Fit a smooth family of curves, which incorporate the effect of the effective rainfall on storm runoff. Graphs A, B and C represent the first approximation of the coaxial graphical

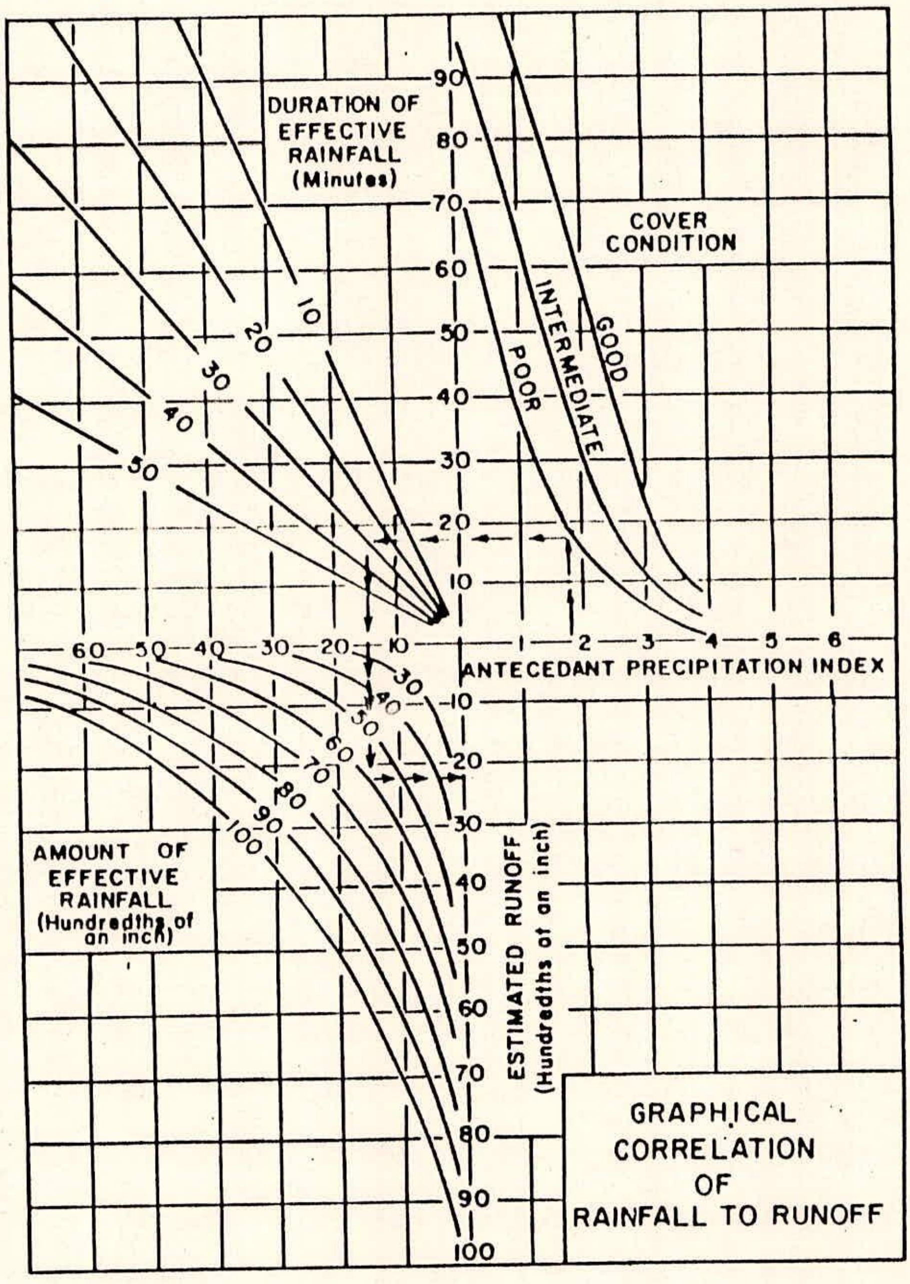

FIG: 6.1 COAXIAL GRAPHICAL CORRELATION OF RAINFALL TO RUNOFF (AFTER WITHERSPOON, 1961) correlation.

4. Construct graph D by plotting computed versus observed storm runoff. Computed storm runoff is obtained from graphs A B and C. This graph is an indication of the overall correlation.

5. Check the accuracy of the first approximation. First, graph A can be checked with the assumption that other graphs are correct. The vertical coordinate of an adjusted point and graph A can be obtained by first entering into graph C and then B the observed runoff, effective rainfall amount, and duration. The abscissa for the adjusted point corresponds to the observed API. Therefore, the cover condition curves must be revised first to the adjusted point so that the relation yields a computed value equal to the observed value.

Likewise, graphs B and C can be checked for the second approximation. Subsequent approximations are made in a like manner. In each case the points are plotted by entering the graph sequence from both ends with observed values to determine the adjusted coordinates.

The coaxial method is flexible and predicts storm runoff satisfactorily. Because the variable selected for estimation of runoff are considerably interactive, it is very difficult to develop a regression equation. The problem is circumvented by coaxial correlation.

On the other hand, the coaxial method has certain deficiencies. The method involves successive approximation and it is, therefore, time consuming. The selection of variables and their plotting in a preferable sequence requires considerable judgement. Only a limited number of variables can be used to keep the method a practically attractive tool. Rainfall intensity has usually been omitted in the analysis. The effect of rainfall variability on runoff was discussed by Fogel (1966). Since rainfall depth and duration are considered, average rainfall intensity becomes an integral part of the coaxial method. This, however, does not account for the effect of space time variations in rainfall intensity on runoff.

6.2.0 Monthly Volumetric Rainfall Runoff Model

The problem of relating long term say monthly or annual volumes of rainfall and runoff is relatively easier. Over a longer period of time, the averaging of a variety of rainfall storms tends to minimize the effect of rainfall intensity and antecedent moisture conditions on the volumetric relationship. Indeed, in many cases a simple plot or a linear relation may be adequate to define the relationship between annual volumes of rainfall and runoff if the water year is properly selected(Brakensiek, 1959), Smith (1973) tested three simple models for estimating monthly runoff from mountainous watersheds. In a similar vein, linear or nonlinear regression analyses have been carried out to relate monthly runoff to monthly rainfall and some other characteristics (Schicht and Walton 1961; Ledger 1975; Sharma and McCuen, 1983; Tsykin, 1985). More realistic models are based on water balance. Some of the simple models are discussed here.

8.2.1 Water Balance Model

A monthly water balance model was developed by Van Der Beken and Byloos (1977) and applied to Grote Nete basin, having an area of 533 km^2 , in the north of Belgaum. The governing model equation can be expressed as

$$AS = N - V_Q - R$$

where AS is the change in storage S,N is effective precipitation V_Q is streamflow, R is the net loss resulting from deep percolation (Loss) and seepage (gain), and the time period is 1 mo. Thus, theis model computes actual evapotranspiration, watr storage in the watershed, direct runoff, infiltration, baseflow and total stream discharge, deep infiltration into the underlying aquifer, and constant seepage from the canals.

Evapotranspiration

The storage at the beginning of a month has a value of S. Actual evapotranspiration E is assumed to occur at a rate equal to or less than the known potential evapotranspiration, E. Therefore,

$$E_a = E_p[1 - \exp(-a_1 A)$$

where a_1 is a parameter ($a_1 \geq 0$). The underlying hypothesis of the above equation is that $E_a = E_p$ until the soil moisture content drops to the wilting point. For the same values of S, a_1 should decrease when the soil texture is more sandy. An initial estimate of a_1 can be obtained as follows. For a sandy soil E_a/E_p may be taken as .75. If the average groundwater level in the region is I m above the river bed and the effective porosity of the sandy layer amount into 15 percent, then average active water storage becomes 150 mm. Substitution of these values into above equation gives an initial estimate of $a_1 = .01$; this can be used in a pattern search procedure for automatic optimization of model parameters.

Effective Precipitation

Effective precipitation minus actual evapotranspiration. When E_a exceeds V_p the storage in the next month is depleted by the effective precipitation $N = V_p - E_a$. On the other hand, if E_a is less than V_p part of N fills the storage reservoir and the other part goes to the river directly.

Streamflow

The streamflow has two components: (1) baseflow and (2) immediate runoff. The former depends upon the storage and the latter on the effective precipitation. Therefore,

$$V_Q = a_2 S + a_3 N$$

where $0 \le a_2 \le 1$, and $0 \le a_3 \le 1$. a_2 will increase when the soil texture is more sandy, whereas a_3 will increase with the degree of urbanization and the average watershed slope. An initial estimate of a_1 and a_2 can be obtained as follows.

If N is negligible, which is true in dry periods.

$$Q = a_2 S$$

This then leads to the exponential baseflow formula.

$$Q = Q_0 \exp(-a_2(t - t_0))$$

and therefore,

$$S = \int_{t_o}^{\infty} Q\,dt = Q_o/a_2$$

in which Q_o is Q at time $t = t_o$ (initial value). Van Der Beken and Byloos (1977) analyzed some baseflow periods and reported an initial estimate of a_2 to be 27.

The parameter a_3 is analogous to rational runoff coefficient. It represents the percentage of N that immediately becomes V_Q . Therefore, a_3 is related to the percentage of impermeable surface. From a study of a hilly basin that is 18 percent urbanized. From Van Der Beken and Byloos (1977) found an initial estimate of a_3 to be 2. They however, suggested that a_2 and a_3 are interrelated. In their model they recommended that

$$a_2 < a_3/(1-a_3)$$

This condition may hold in using initial estimates of a_2 and a_3 but can be used in optimization.

Deep Percolation and Canal Loss

R is composed of deep percolation, L_p, which is partially compensated for by canal losses L_c.

$$R = L_p - L_c$$

The deep percolation may be evaluated as a linear function of the water storage S. If S = 0, deep percolation must be zero. Then

$$L_p = a_2S$$

The canal seepage remains more or less constant;

$L_c = a_5$

Therefore,

$$R = a_4 S - a_5$$

where a_4 is a parameter relating to deep percolation and a_5 relates to canal loss.

An initial estimate of a_4 can be made from an estimate of the average deep percolation L_p. Van Der Beken and Byloos (1977) made an initial estimate for the Neogene basin as follows. The permanent groundwater flow in this aquifer depends on the slope of the aquifer. This slope can be evaluated at .25 percent in the north eastern direction for Neogene aquifer. The average thickness of the aquifer is about 170 m, and the average width of the drainage basin is 25 km. The hydraulic conductivity is K = 10^{-4} m/s. The transmissivity is then T_s = 1700 m^2/da. Thus, groundwater flow can be evaluated as

$$Q_g = 1062 \ m^3/s$$

If the groundwater recharge is uniformly distributed over the whole basin.

$$\overline{L_p} = Q_g/\text{area} = 5 \text{ mm}$$

Therefore,

$$a_4 = \overline{L_p}/S = 0.03$$

An initial estimate of a_5 can be made by determining long term average of N, V_q, and L_p. A value of a_5 = 3 mm was recommended by Van Der Beken and Byloos (1977).

Initial Storage

At the beginning of the period of computation S is normally unknown. It can be estimated by correlating with the long term average discharge volumes. As an initial estimate $S_i = 2\ S*$ is reasonable to use, where S* is average water storage.

6.2.2 Haan Model

Haan (1972) presented a four parameter water yield model for small watersheds. The central idea of the model is that moisture holding and moisture transmitting characteristics of the soil and underlying strata, along with rainfall intensities, are the most important factors governing the runoff volumes from small watersheds. This model relies heavily on the soil moisture model developed by Ligon et al. (1965, 1967).

Evapotranspiration

The soil moisture zone is divided into an upper zone and a lower zone. The moisture holding capacity of the upper zone is assumed to be 2.5 cm. The water of this zone M_u is readily available for ET. In other words, ET occurs from this zone at potential rate. The moisture holding capacity of the lower zone is M_c The water available M_L in this zone is less readily available to evapotranspiration, that is, ET occurring from this zone is at a rate less than potential.

Evapotranspiration occurs at a potential rate as long as water is available. It is reduced by the ratio of the amount of moisture available in the lower zone to its moisture holding capacity. On days when rainfall occurs, it is reduced by a factor of w account for cloudy conditions and low solar radiation. Thus, the daily ET is computed as follows;

$$
\begin{aligned}
E_a &= E_p; & V_p &= 0, & 0 < M_u \leq 2.5 \\
&= E_p\, M_L/M_c; & V_p &= 0, & M_u = 0 \\
&= 1/2\, E_p; & V_p &\geq 0.025, & 0 \leq M_u \leq 2.5 \\
&= 1/2\, E_p\, M_L/M_c\ ; & V_p &\geq 0.025, & M_u = 0
\end{aligned}
$$

The potential ET is estimated by the Thornthwaite method (Thornthwaite, 1948). V_p represents precipitation volume that occurred on the day in question.

Infiltration

Rainfall is partitioned into infiltration and surface runoff. The infiltration rate f is estimated as follows:

$$
\begin{aligned}
f &= f_o, & p &\geq f_o; & M_u < 1 \text{ or } M_L < M_c \\
&= p, & p &\leq f_o; & M_u < 1 \text{ or } M_L < M_c \\
&= 0 & M_u &= 2.5, & M_L = M_c
\end{aligned}
$$

where f_o represents maximum infiltration rate. All infiltrated water is stored in the upper soil moisture zone until its capacity is filled. Thereafter any additional infiltrated water automatically goes to the lower soil moisture zone. The entire rainfall is assumed to be runoff after both zones are filled to their capacity.

Surface Runoff

The surface runoff is determined from

$$
\begin{aligned}
V_Q &= (p - f)\Delta t, & p > f \\
&0, & p < f
\end{aligned}
$$

where Δt denotes the time increment.

Deep Seepage

Deep seepage S_e is determined from

$$S_e = S_m M_L/M_c$$

where S_m is the maximum possible seepage rate on a daily basis. Therefore, the volume of daily seepage V_{Se} is equal to S_e.

Return Flow V_r

A certain amount of return flow is permitted within the watershed, which is obtained from

$$V_r = \alpha V_s$$

where α is a constant defining the portion of seepage that appears as runoff.

Total Runoff V_T

The total runoff is the sum of V_Q and V_r

$$V_T = V_Q + V_r$$

Parameter Estimation

Average potential evapotranspiration and daily rainfall data constitute input to the model. The model has four parameters f_o, S_m, M_c and α. These parameters are estimated by an optimization procedure minimizing the sum of the squares of deviations between the observed and computed monthly runoff volumes. Thus, approximately 3 years of observed monthly flows are required for parameter optimization. The daily rainfall is broken into hourly rainfall . To account further for temporal variability of rainfall on runoff, hourly rainfall data are divided into ten periods of 6 minutes each and the rainfall intensity for each 6 min period is determined by proportioning the hourly rainfall by the rainfall distribution. Thus, this model provides for runoff that may occur when the two soil moisture zones are not filled to their capacity.

6.3.0 Yearly Volumetric Rainfall-Runoff models

The simplest linear models for volume take the form

$$V_Q = aV_p - b$$

where V_Q is annual runoff volume, V_p is annual watershed mean precipitation, a and b are coefficients estimated by linear regression analysis. This type of relationship has been used by several investigators (Sutcliffe and Rangeley, 1960 Pike, 1964: Rodier 1976 Marsh Littlewood 1978). Woodruff and Hewlett (1970) predicted and mapped V_Q/V_p for the eastern United States (281 watersheds, up to 200 M). Bosch and Hewlett (1982) reviewed the experiments conducted to determine the effect of vegetation changes on V_Q and evapotranspiration but found in conclusive results.

The difference between annual rainfall and runoff from a watershed comprises principally evapotranspiration, seepage, V_g and the change in groundwater storage ΔS;

$$V_p = V_Q + E + V_g + \Delta S$$

The term AS in this equation can be negative or positive. If the watershed is watertight and ΔS is negligible, then equation reduces to

$$V_Q = V_p - E$$

Equating these equations;

$$E = b + V_p(1 - a)$$

It is thus seen from these equations that water yield and evaporation increase linearly with precipitation. This assumption is reasonable in temperate and subhumid regions where precipitation is moderate and well distributed temporally. It was made by Sutcliffe and Rangeley (1960) in their water balance study in Iran.

On comparing of these equations it is clear that the key to determination of annual water yield is the accurate determination of annual evapotranspiration. Ayers (1962) suggested that the annual evaporation is approximately half the annual precipitation for the relationship expressed by equation . On the other hand, equation (10.65) gives a linear relationship between them. Several other types of evapotranspiration precipitation relationships have been proposed (Pike 1964; Solomon, 1967;. Majtenyi, 1972). For example Schreiber, referred to by Budyko (1948), suggested the following relationship in 1904;

$$E_a = V_p[\ 1 - \exp(-P_E/V_p)]$$

where E_a is the actual evaporation (land and lake evaporation as well as transpiration and P_E is the maximum possible value for actual evaporation in the given condition; this will amount to potential evaporation.

Another well known relation is given by Turc, cited by Pike (1964);

$$E = \frac{V_p}{[.9 + (V_p/L)^2\ ]^{1/2}}$$

where E is annual evapotranspiration in millimeters V_p is in millimeters $L = 300 + 25T + 0.05T^3$ in millimeters and T is the mean annual air temperature (·C). The above equation was developed using data from a large number of worldwide sites. Inclusion of temperature in this equation reflects indirectly the energy available for evapotranspiration as done by Thornthwaite (1948).

Pike (1964) suggested that equation (10.67) would be more realistic if L were replaced by Penman's estimate E_o (Penman 1948,1950), or adjusted pen data, and if the constant were made unity. Since E_o is an approximation of energy balance. It would be a more realistic estimate than L of the energy available for evapotranspiration. Therefore,

$$E = \frac{V_p E_o}{[V_p^2 + E_o^2]^{.5}}$$

Another formula for determining E is the Kritzki and Menkel formula (Kritzki and Menkel, 1949), which employs saturation deficit instead of temperature:

$$E = V_p[1 - K/(1+.5d)3.5]$$

Where K is a coefficient depending upon local condition (K< 1) and d is the saturation deficit, defined as the difference between the saturation vapour pressure at the recorded air temperature and the actual vapor pressure.

Employing Bouchet's concept of actual and potential evaporation (Bouchet, 1963; Morton, 1965), Solomon (1967) derived an expression for computation of E;

$$E + P_E = N_E$$

and

$$W_A = E(N_E - E)/[P_E(N_E - 2E)]^{.5}$$

where W_A is available moisture, P_E is potential evaporation and N_E is net energy. It may be noted that the variation of N_E is generally not significant from one year to another, and can be assumed to be a function of net solar radiation R_N. For the tropical equatorial zone, $N_E = R_N = (1 - \epsilon)R_I$, where R_I is incident solar radiation and ϵ is albedo. W_A approaches V_p for a water year starting at the lowest value of flow.

These models do not incorporate seasonal differences in evaporation demands that may deplete soil moisture storage. Consequently there may be significant errors in estimation of actual evaporation and soil moisture. As an alternative Glaspoole (1960) suggested a multiple linear regression for water yield utilizing seasonal precipitation values.

$$V_Q = a_0 + a_1 V_{ps0} + a_2 V_{pw0} + a_3 V_{ps1} + a_4 V_{pw1}$$

where a_i, i = 1,2,3 and 4 are regression coefficients, V_{psi} i=0, 1, is summer precipitation for the previous and current years, and V_{pwi} i=0,1 is winter precipitation for the previous and current years Glaspoole (1960) applied it to the Thames River of

England and was to able to account for 92.9 percent of variability. A similar study was conducted by Mustonen (1967) for finnish watersheds, which included fall, winter, and summer precipitations, average annual temperature, potential evapotranspiration in summer, frost depth, percentage of drainage area with coarse soils, volume of forest growing stock and change of soil moisture during water year in a multiple linear regression analysis to determine annual water yield. He found that 89 percent of the flow variability was explained by his analysis.

Along similar lines. Hann and Read (1970) correlated mean annual runoff with mean annual precipitation, watershed perimeter, P_e and watershed relief ratio R_r for small agricultural watersheds in Kentuky;

$$V_Q = a_o + A_1V_p + A_2P_e + a_3R_r$$

where a i=0,1,2,3, are regression coefficients Equation (10.73) explained 91 percent of variation in V_Q. On the other hand, Sutcliffe and Carpenter (1967) used annual precipitation and elevation in their correlation study on mountainous and semiarid area in western Iran and found

$$V_Q = a_o + A_1V_p + A_2P_e + a_3R_r$$

where this equation explained 66.4 percent of variation in V_Q

$$V_Q = a_o + a_1V_p + a_2E$$

Hawley and McCuen (1982) related V_Q to several physiographic characteristics from 605 watersheds in the western United States divided one for each region of hydrologic similarity, were derived. The equation explained from 57 to 95 percent of the variance in Vq. Sharp et al. (1960) and Harris et al (1961) reported multiple regression equations for estimation of Vq. Similar studies were conducted by Wang and Huber (1967) for ungaged watersheds.

7.0 PROPOSED STUDY DURING AND AFTER TRAINING

The estimation of water yield will be helpful in number of water resources problems such as, design of storage facilities, determination of water availability for industrial and agricultural use, dependable water supply for power generation, planning irrigation operation and design of irrigation projects.

Water yield is the volume of the water available from a stream at a specified point area and at specified period of time. The water yield is determined at the point of outlet of catchment and the period of time a day or longer. The thrust is given on the volume of the flow rather than the instantaneous discharge and reflects the volumetric relationship between rainfall and runoff.

During the tenure of training, it is proposed to work on developing water yield model for smaller catchments with the available data. These models may be design to extend it to the bigger catchments. There are many approaches to determine the water yield such as empirical approaches and continuous time simulation approaches to determine the water yield of a catchment. Models can be classified as Volumetric storm rainfall-runoff models, Monthly volumetric rainfall-runoff model and Yearly volumetric rainfall-runoff model. The model is proposed to develop on the basis of the hydrologic characteristics and hydraulic characteristics of the catchments.

Water yield models are quite essential for Indian conditions as in most of the regions, though the catchments receive enough rainfall , there is no provision for the control or to utilise the available water resources in a proper planned manner. This is due to the lack of knowledge on water yield estimation. Personnels involved in water resources planning should need training on problem related to water yield in the catchment. The present training will give a insight of the problem and trainee may be able to guide or train the local engineers and scientists with the specific problem.

In this region anomaly is found in estimating water yield or discharge of the catchment which leads to over or under design of water storage facilities. The concerned local authority have cited this problem to our institution to come out with proper methodology or model to estimate water yield from the catchment for their proper design of water storage structure.

We have selected two representative basins for hydrological studies in this region. After successful completion of training on catchment hydrology, the knowledge acquired during the training may be extended to apply it to the representative basins selected for the hydrological studies as initial application of the model developed during the training in abroad.

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