I INTRODUCTION

1.1 Introduction of the Problem

The design of many hydraulic detention structures requires hydrologic data such as of streamflow. The hydrologic records, especially rainfall, evapotranspiration and runoff, are of fundamental importance in a wide variety of water resources systems analysis. Currently, the only basis for predicting yields and flows are these streamflow records compiled over many years, but unfortunately streamflows are scarce and often unavailable in small watersheds.

Central Vietnam is one of the places having the potential to construct reservoirs useful for irrigation, hydropower and other purposes needed for local developments. However, because of the repercussive effects of the prolonged, notorious Vietnam War, a vast majority of the basins of Central Vietnam are ungaged. Some of these water resources systems have been designed based on short-term data or even inadequate hydrologic records. As a result, the systems are either over or under designed, incurring extensive losses and unnecessary costs. In such a case, there is always a fundamental conflict of interest between the data collector and the water resources system designer. The data expert wishes to improve the accuracy of his product which may require more years of observations while the designer wishes to get on with his work.

The development of the science of hydraulics has been a significant aid to the hydrologist because it has given a basis for some quantitative estimates of flow rate in streams - a far better guide to judgment than merely the visible cross-section area. As a special case study , the main purpose of the present research is to establish an orderly discipline to account for the changes in the observed hydrologic variables and to predict the hydrologic behavior of watersheds where observed data are incomplete or unavailable. This study will discuss streamflows modeling which consider some relations between climatic data, which are readily available for longer periods than runoff and the physiographic factors. Data-scarce subcatchments of Thubon basin in Quangnam - Danang province, Central Vietnam will be considered for the estimation of rainfall and runoff model as a case study.

1.2 Objective and Scope of Work

The main objectives of this study are :

a) To examine the applicability of the rainfall-runoff model to predict runoff from small, ungaged basins in Central Vietnam by using a rainfall-runoff model developed by Danish Hydraulic Institute, that is NAM. In this context, ungaged basin means that the streamflow data is not recorded.

b) To calibrate the NAM model for a gaged basin located in the same regions and possesses similar hydrologic characteristics as the ungaged basins.

c) To extend the applicability of the calibrated model parameters to other ungaged basins of the same region.

The scope of this study is limited to the basins of Quangnam - Danang province in Central Vietnam. The data are taken from this region where some hydraulic projects will be constructed to meet the demands of irrigation, water supply, power consumption and flood protection.

1.3 General Approach

It is the intent of this study to determine the applicability of the NAM model to streamflow estimation in this region. In this regard, the model is tested on basins in Central Vietnam. Their relative performance in the calibration and verification periods are evaluated and compared with one another.

The final set of parameters of the NAM model applied to gaged basin of the same region are applied to other ungaged basins after being adjusted according to the characteristics of ungaged basins without calibration. Finally, the sensitivity analysis of certain parameter values for modeling will be done.

II LITERATURE REVIEW

2.1 Probable Approaches

For developing a model that can predict the flow from an ungaged basin, the values of the process parameters may be determined in either of the following ways :

(a) Developing a model for a gaged basin (whose basin characteristics are similar to that of the ungaged basin) and utilizing the values of process parameters of this model for developing a new model for an ungaged basin.

(b) Values of process parameters are calculated from the measured basin characteristics and available records.

Alternatively, a parameters model can be developed for an ungaged basin and calibrating this model with a physically based deterministic model and it can be used for the prediction of flow from the basin. Some of the attempts are reviewed here.

2.2 Different Models

The first approach to a short-term rainfall-runoff model devised from the combined work of SHERMAN (1932) and HORTON (1932). Sherman observed that the hydrographs of floods from a given basin were remarkably similar in the shape when caused by reduced by rainfall of similar durations and that, if the hydrographs were to a unit volume, they tended to be approximately identical. This concept, called the unit hydrograph, was the first expression of the linear system approach in flow simulation. SNYDER (1938) presented a procedure for developing systematic unit hydrographs when the necessary flow data were not available. Not long thereafter, CLARK (1945) demonstrated the equivalence of the linear routing procedure known as the Muskingum method and showed that a unit hydrograph could be constructed by routing of the time-area diagram of the basin converted to flow units. Many other persons - too numerous to mention here - contributed to the general development of the unit hydrograph concept.

ROCKWOOD (1961) reported the use of a digital computer in 1958 to route flows through the Columbia River Basin. Rainfall excess and snowmelt runoff were computed separately and input to the model which was a routing model adjusting for basin, channel and lake storage. Investigations based on the premise of a linear input-output relationship have been done by DOOGE (1959), NASH (1959), and GRAY (1961), to name a few. GRAY (1961), in his work, applied the unit hydrograph principle on small ungaged watersheds. He developed an equation to describe the graph based on two parameters which can be evaluated from measurable topographic characteristics.

SUGAWARA (1961) has proposed a model consisting of a complex cascade of linear storages for simulating the entire runoff process from rainfall to streamflow. His model has called the Tank model which is a series of storage-type model to account for the non-linear relationship between rainfall and runoff. The storage type is based on the hypothesis that runoff and infiltration are the functions of the amount of water stored in the ground. Sugawara does not, however, relate the various storages to features of the natural process, and hence, his model can be fitted to a real watershed only by a very length trial-and-error process. The applicability of the model to various conditions have been studied by SUGAWARA (1961,1974), himself, UDDIN (1977),LORIA (1979), PHIEN and PRADHAN (1983) and ANA (1984) on a daily and monthly basis.

NIELSEN and HANSEN (1973) developed the NAM model. It is a relatively simple model for the prediction of runoff from rural catchments. A mathematical hydrological model like NAM is a set of linked mathematical statements describing, in a simplified quantitative form, the behavior of the land phase of the hydrological cycle. Numerous hydrological models exist. The NAM model is a so-called deterministic, conceptual, lumped type of model with moderate input data requirements. NAM model has been applied to a number of catchments in different climatic regions of the world. The applicability of this model in Thailand have been reported by KHALIQ (1986).

MANLEY (1977, 1978) developed HYSIM, i.e. Hydrologic Simulation Model. Here the model parameters were derived from the measured basin characteristics without reference to the gauged flow. The model has 17 hydrologic parameters to describe the nature of the basin. Basin area and correction factors are used to compensate for systematic errors in precipitation, potential evapotranspiration and snowfall measurements.

HOSSAIN (1980), in his thesis research, applied the HYSIM model for predicting flow from an ungauged basin - Nam Mae Chan in Thailand. In his results, the correlation coefficients vary from 0.77 to 0.87 for daily flows and 0.88 to 0.99 for monthly flows. Comparison of measured and calibrated parameter values demonstrated the viability of this approach.

INTAHVONG (1983) calculated the long-term average annual runoff for ungaged small rivers in Vientainne, Laos. He presented a simplified technique of estimating runoff using the Pearson type III (P3) distribution. This concept availed of the functional relationship of the skewness coefficient with the coefficient of variation of the precipitation data.

MINIKOU and RAO (1983) presented a simple deterministic black box regression model for monthly rainfall simulation. This simple regional model is applicable to basins with either linear or non-linear rainfall-runoff behavior. The structure of the model depends on two key parameters which are estimated from the monthly rainfall-runoff relationships. They pointed out that simplicity of the model does not necessarily mean the model may be inaccurate.

SINGH and AMINIAN (1984) presented a rainfall-runoff model for basins with scarce discharge data or having no discharge data at all. The rainfall-runoff model, designated hence as Watershed Hydrological Simulation (WAHS) model based on the conceptual framework, consists of a number of component models.

JOKO (1989), in his thesis research, applied the WATBAL model in two catchments in Central Java, Indonesia. This model has been developed by the Danish Hydraulic Institute, DHI (1986). It is a semi-distributed rainfall-runoff model and represent an intermediate approach as compared with the two alternative rainfall-runoff model, i.e. lumped and fully distributed model. The main purpose of the WATBAL model is to predict the continuous streamflow for an ungauged watershed.

Many other models now exist but time does not permit a detailed discussion of each of them. Intercomparison of models have been carried out to provide information and guidance on the applicability of such models in various simulations and in terms of accuracy requirements. The World Meteorological Organization (WHO, 1975) undertook a project on the intercomparison of conceptual models used in operational hydrological forecasting from 1968 to 1974. LINSLEY (1981) made an overview the Rainfall-Runoff models. His paper traced the historical development of hydrology and the types of models currently in use. It discussed the purposes of modeling and the properties of model required to serve these uses.

III THEORETICAL CONSIDERATIONS

3.1 Concept of Water Balance Equation

The interdependence and continuous movement of all forms of water provide the basis for the concept of the hydrological cycle. Because the total quantity of water available to the earth is finite and indestructible, the global hydrologic system may be looked upon as closed. The hydrologic water budget for a certain system can be represented by the following general water balance equation:

Inflow = Outflow ± Storage Change

If a rainstorm or precipitation input is applied, an out put is determined as actual evapotranspiration and surface runoff, the above equation can be written as follows :

Precipitation = Runoff+Actual Evapotranspiration ± Storage Change

The term precipitation as used in hydrology includes all forms of water deposited on the earth's surface and derived from atmospheric vapor. The principle forms are mist, rain, hail, sleet and snow. Unless otherwise specified, the term precipitation and rainfall are often used indiscriminately to apply to any or all of the forms included in this group. Potential evapotranspiration is defined as the amount of water that would evaporate if the water supply is ample. The actual water loss, called the actual evapotranspiration, is less than or equal to the potential evapotranspiration. The water flowing in a stream, understood as runoff, may have found its way into stream channel from several different sources, namely precipitation falling directly on the stream, surface runoff and groundwater flow finding its way through the soil into the stream.

Water balance equation is an important basis for developing the hydrological models. This concept has been applied on the estimation of rainfall-runoff models.

3.2 The NAM model

NAM is an abbreviation of the Danish "Nedbor-Afstromnings-Model", meaning precipitation-runoff-model. NIELSEN and HANSEN (1973) devised this model that gives total daily-hourly stream flows, actual evapotranspiration and soil moisture storage index and groundwater storage index based on daily-hourly rainfall, monthly potential evapotranspiration and temperature (only if the snow routine is used). During the past decade it has been extensively applied and modified by the Danish Hydraulic Institute in a large number of projects in different climatic regions of the world.

The NAM model is a so-called deterministic, conceptual, lumped type of model with moderate input data requirements. Being a lumped model, NAM treats each catchment as one unit. The parameters and variables are thus presenting average values for the entire subcatchment. A conceptual model like NAM is based on physical structures and equations used together with semi-empirical ones. Thus, some of the parameters can be evaluated from physical catchment data, but the final parameter estimation must be performed by calibration applying concurrent input and output time series.

The structure of the model is shown in Fig. 3.1. It is an attempt to make a simplified imitation of the land phase of the hydrological cycle. The calculations minic key hydrologic processes : water from the air falls down the surface, infiltration of water into the soil profile, surface runoff and flow along subsurface flow paths into the streams. Individual routines of NAM model are described in the following sections.

3.2.1) Snow Routine: The snow routine is optional and temperature data is only required if the snow routine is applied. Precipitation passing through the snow storage is controlled by temperature condition. When the mean temperature, T, is above the freezing point the snow remaining in storage is assumed to release an amount of melting water, P_s,:

$$P_s = C_{smelt} * T \tag{3.1}$$

C_{smelt} is the snow melt coefficient in mm/centigrade/day. When the snow option is not selected, this parameter has no importance and any "dummy" value can be specified.

To consider snow routine is not necessary for the tropical regions such as Central Vietnam.

3.2.2) Surface Storage: Moisture intercepted on the vegetation, as well as water trapped in depressions and in the uppermost cultivated part of the ground is presented as surface storage. U_{max} denotes the upper limit to the amount of water in surface storage, the moisture content U is continuously reduced by potential evapotranspiration, and interflow.

The soil moisture in the root zone, a soil layer below the surface from which the vegetation can draw water for transpiration, is represented as lower zone storage. L_{max} denotes an upper limit to the amount of water in this storage.

The amount of water, U, in the surface storage is continuously diminished by evaporative consumption as well as by horizontal leakage (interflow). When there is maximum surface storage, some of the excess water, P_n, will enter the streams as overland flow, whereas the remainder is diverted as infiltration into lower zone and groundwater storage.

Moisture in the lower zone storage is subject to consumptive loss from transpiration, and the moisture content controls the amount of water that enters the groundwater storage as recharge.

When the surface storage spills, i.e. when $U \geq U_{max}$, the excess water, P_n, gives rise to overland flow as well as to infiltration. QOF denotes the part of P_n which contributes to overland flow. It is assumed to be proportional to P_n and to vary linearly with the relative soil moisture content, L, of the lower zone storage.

$$QOF = \begin{cases} CQ_{OF} \dfrac{L - TOF}{L_{max} - TOF} P_n & \text{for } L > TOF \\ 0 & \text{for } L \leq TOF \end{cases} \qquad (3.2)$$

where, CQ_{OF} : overland flow runoff coefficient
TOF : threshold value for overland flow

The parameters, CQ_{OF} and TOF, are both positive and constant less than unity and without dimension.

3.2.3) Lower zone Storage: The proportion of net rainfall excess, P_n, that does not runoff as overland flow infiltrates into the lower zone storage representing the root zone. A portion, DL, of the amount of infiltration, (P_n - QOF), is assumed to increase the moisture content, L, in the lower zone storage. The remaining amount of infiltrating moisture, G, is assumed to percolate deeper and recharge the groundwater storages. G and DL are calculated from :

$$G = \begin{cases} (P_n - QOF) \dfrac{L - TG}{L_{max} - TG} & \text{for } L > TG \\ 0 & \text{for } L \leq TG \end{cases} \qquad (3.3)$$

$$DL = (P_n - QOF) - G \qquad (3.4)$$

where, TG is the root zone threshold value for groundwater recharge.

The interflow contribution, QIF, is assumed to be proportional to U and to vary linearly with the moisture content, of the lower zone storage.

$$QIF = \begin{cases} (CKIF)^{-1} \dfrac{L - TIF}{L_{max} - TIF} U & \text{for } L > TIF \\ 0 & \text{for } L \leq TIF \end{cases} \qquad (3.5)$$

where, CKIF : time constant for interflow
TIF : root zone threshold value for interflow

Evapotranspiration demands are at first attempted to be met at the potential rate from the surface storage. If the moisture content, U, in the surface storage is less than these requirements, the remaining fraction is assumed to be withdraw by root activity from the lower zone storage at an actual rate Ea. Ea, is proportional to the potential evapotranspiration, Ep.

$$Ea = Ep * L/L_{max} \qquad (3.6)$$

The capillary flux of water from the groundwater table to the lower zone storage is assumed to depend on the depth of the groundwater table below the ground surface, GWL, as well as on the relative moisture content, L/L_{max}, of the lower zone storage.

$$CAFLUX = (1 - L/L_{max})^{1/2} \left(\frac{GWL}{GWLFL_1}\right)^{-\alpha} 1 \text{ mm/day} \qquad (3.7)$$

$$\alpha = 1.5 + 0.45\ GWLFL_1$$

The parameter $GWLFL_1$ is the groundwater table depth at which the capillary flux is 1 mm/day when the lower zone storage is completely dry. Equation (3.6) gives a good fit to the theoretical relationship between the capillary flux, the depth to the water table and the soil moisture content proposed by Rijtema (1969).

The groundwater level is calculated from a continuity consideration accounting for recharge G, capillary flux, CAFLUX, net groundwater abstraction GWPUMP, and baseflow BF. The inclusion of GWPUMP which is optional, can be done by specifying monthly net abstraction rates. The baseflow is calculated as the outflow from a linear reservoir with time constant CK_{BF}.

$$BF = \begin{cases} (GWLBF_0 - GWL)\ S_y\ (CK_{BF})^{-1} & \text{for } GWL \leq GWLBF_0 \\ 0 & \text{for } GWL > GWLBF_0 \end{cases} \qquad (3.8)$$

The parameter $GWLBF_0$ is the groundwater table depth which causes baseflow, while S_y is the specific yield of the groundwater reservoir.

The physical meaning of the parameter $GWLBF_0$ is illustrated in Fig. 3.2 (a) as the distance from the average ground level to the river water lever. Due to the variation in river water lever throughout the year $GWLBF_0$ can be given a significant annual variation as illustrated on Fig. 3.2 (b).

3.2.4) Time Constants for Routing: The interflow is routed through two linear reservoirs in series with time constant CK_1 and CK_2. The overland flow routing is also based on the linear reservoir concept with the two parameters CK_1 and CK_2, but with variable time constants.

$$CK = \begin{cases} CK_{par} & OF \leq OF_{min} \\ CK_{par} \left(\dfrac{OF}{OF_{min}}\right)^{-\beta} & OF > OF_{min} \end{cases} \qquad (3.8)$$

where,

OF	=	overland flow depth (mm/hour)
CK_{par}	=	model parameter, CK_1 or CK_2 (hour)
OF_{min}	=	lower limit for non-linear routing dynamics.
OF_{min}	=	0.4 mm/hour.
β	=	coefficient. β = -0.33 corresponding to the Chezy flow dynamics.

Equation (3.8) ensures in practice that the routing of real surface flow follows a kinematic flow dynamics, while subsurface flow being interpreted by NAM as OF (in catchments with no real surface flow component) is routed linearly corresponding to groundwater flow dynamics.

3.3 Model Parameters

A short description of the model parameters, their physical interpretation and importance is presented along with hints for parameter adjustments in the calibration phase.

3.3.1 Storage Capacities: U_{max}, L_{max} (mm) define the maximum water content in the surface and root zone storages respectively.

The value U_{max} should reflect that the surface storage is representing the interception storage (on vegetation), the surface depression storages and the uppermost few cm's of the ground.

L_{max} can be interpreted as the maximum soil moisture content in the root zone available for the vegetative transpiration. Ideally, L_{max} can then be estimated by multiplying the difference between field capacity and wilting point of the actual soil with the effective root zone depth. However, L_{max} represents the average value for an entire catchment; averages are for the various soil types and root depths of the individual vegetation types. Therefore, L_{max} cannot in practice be estimated from field data, but an expected interval, can be defined.

As the actual evapotranspiration is highly dependent on the water content of the two storages, U_{max} and L_{max} are the parameters to be changed in order to adjust the water balance in the simulations.

As a rule, $U_{max} = 0.1\ L_{max}$ can be used unless special catchment characteristics or hydrograph behavior indicate otherwise.

One important characteristic of the model is that the surface storage must be at its capacity, i.e. $U \geq U_{max}$ before any excess water, P_N, occurs. In dry periods, the amount of net rainfall that must occur before any overland flow occurs can be used to estimate U_{max}.

3.3.2 Overland Flow Runoff Coefficient: CQ_{OF} is the overland flow runoff coefficient. It is a very important parameter determining the extent to which excess rainfall runs off as overland flow and the infiltrating quantity.

CQ_{OF} is without dimension and has a value between 0 and 1. Physically, in a lumped way it reflects the infiltration and also to some extent the recharge conditions. Small values of CQ_{OF} are expected for a flat catchment having coarse, sandy soils and a large unsaturated zone; while, large CQ_{OF} values are expected for catchments having low, permeable soils such as clay or bare rocks. CQ_{OF} values in the range of 0.01 - 0.90 have been experienced.

3.3.3 Time Constant for Interflow: CKIF ($hour^{-1}$) is the interflow time constant. It determines together with U_{max} the amounts of interflow, as 1/CKIF is the quantity of U drained to interflow every hour. It is the dominant routing parameter of the interflow because CKIF >> CK_1, CK_2.

Physical interpretation of the interflow is difficult and will vary somewhat from one catchment to another. As interflow is seldom the dominant streamflow component, CKIF is not usually a very important parameter. Normally, CKIF values are in the range of 500 - 1000 hours.

3.3.4 Baseflow Parameters

a) $GWLBF_0$, baseflow is described as outflow from a linear reservoir, i.e. a reservoir from which the flow is proportional to the storage of water above the outflow level. The maximum groundwater depth for which baseflow occurs, $GWLBF_0$, represents the outflow level of the linear reservoir. This level may vary over the year as illustrated in Fig. 3.2.

b) S_y, the specific yield, depends on the soil type and has typically a value of 0.01 - 0.30.

c) CKBF, the time constant for groundwater flow, can be estimated from the hydrograph recession in dry periods.

3.3.5 Capillary Flux ($GWLFL_1$): $GWLFL_1$ is the depth of the groundwater table which gives an upward capillary flux of 1 mm/day when the moisture content of the upper soil layers are at wilting point, i.e. L = 0. Values of this parameter are given for different soil types in Chapter 3, NAM User's Guide.

3.3.6 Threshold Value: TIF, TOF, TG are all positive parameters less than the maximum soil moisture content, L_{max}. They work in Eqs. (3.2), (3.3), and (3.5) as threshold values so that no interflow, overland flow, or recharge is generated if L is less than TIF, TOF, or TG respectively.

The function of the threshold values is illustrated by the overland flow Eq. (3.2) in Fig. 3.3.

Physically,the threshold values should reflect the degree of spatial variability in the catchment characteristics, so that a small homogeneous catchment is expected to have large threshold values than a large heterogeneous catchment.

For catchments with alternating dry and wet periods, the threshold values determine the start times of the flow components in the periods where the root zone is being filled up. This can be used in the parameter estimation. For instance, TOF can be estimated on the basis of such situations where even very heavy rainfall does not give rise to the quick response of the overland flow component.

It should be noticed that the threshold values have no importance in wet periods when $L = L_{max}$. The importance of the threshold value varies from catchment to catchment and is usually larger in semi-arid regions.

The parameters are relatively easy to estimate through calibration.

3.3.7 Time Constants for Overland Flow Routing: CK_1 and CK_2 (hours) are the time constants for routing interflow and overland flow depend on the size of the catchment and how fast it responds to rainfall. The constants are found through calibration on peaks. Although CK_1 and CK_2 are two independent parameters it is recommended to set $CK_1 = CK_2$ so that only one time constant has to be assessed.

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3.4 Model Calibration and Verification

3.4.1 Model Calibration: Most hydrological models require adjustments to the process control parameters in order to "tune" the model to reproduce the catchment response. This procedure of adjusting parameters is called calibration.

Calibration of a model is one of the most important steps in model application. The accuracy of the whole study will depend on the level of calibration achieved. Calibration procedures vary from model to model but general alternatives can be listed :

(a) Trial and error calibration.
(b) Automatic calibration.
(c) Combination of (a) and (b).

In trial and error calibration, the user inputs all the parameters that can be based on physical observations, and provides estimates of the unknown parameters as a first trial. The model is run and the output compared to observed output from the prototype. Based on this comparison adjustments are made to one or more of the trial parameters to improve the fit between observed and simulated output. The comparison may be by means of visual pattern recognition of the simulated and observed flow hydrographs or may be based on some mathematical criterion of "goodness of fit". Repeated trial runs of the model are made until the required accuracy is achieved.

In automatic calibration, the model contains internal programming which will adjust the trial parameters in a step by step manner until the redefined goodness of fit criterion is satisfied. In this way the model will automatically calibrate itself and carry out the necessary number of trial runs until the best set of parameters is achieved.

The third alternative is to carry out trial and error manual adjustments of parameters until the model is almost calibrated, then to introduce the automatic search technique to refine the goodness of fit.

Calibration of NAM model is recommended by Danish Hydraulic Institute to use trial and error method.

3.4.2 Model Verification: Models generally are used in hydrology as predictive tools for assessing events either past, present of future such as missing streamflow record, possible magnitude of extreme events or the effect of catchment changes on streamflow.

It is the joint responsibility of the model developer, model user and the decision maker to ensure model validity is verified. It is not possible to completely validate a predictive "decision-aiding" model since the events being predicted are not represented by real-data. A model however can, over a period of prolonged use, be "well-proven" in assessing future events and hence be validated to the point where the decision maker has confidence in the results obtained from the model.

3.5 Data Requirements

The input requirements for the NAM model consist of :

a) model parameters.
b) initial conditions.
c) meteorological data, and
d) streamflow data for the calibration period.

The model parameters have been described in Section 3.3. The initial conditions are the water content in the storages, values of interflow and overland flow components, and the groundwater depth at the start of the simulation.

The meteorological data requirements are:

- rainfall
- potential evapotranspiration
- temperature (only if snow routine option selected)

The time resolution of the rainfall input depends on the objective of the study and on the time scale of the catchment response. In many cases daily rainfall values are sufficient, but in quick responding catchments with project focus on peak flows, rainfall input on a finer resolution may be required. Rainfall data with any (variable) time increments can be specified in the rainfall input. The NAM model will then make the necessary interpolations according to the simulation time steps.

For potential evapotranspiration monthly values are usually sufficient. Only minor improvements can be obtained by specifying, e.g. daily values instead of monthly values.

3.6 Test of Goodness of Fit

The test of goodness of fit between observed hydrograph compared to the simulated hydrograph can be obtained by computing and comparing certain statistical parameters of two series, or more directly by computing dimensionless coefficient of agreement, e.g. the coefficient of determination, relative errors and root mean square values.

3.6.1 Statistical Criteria: It is required that the range, mean and standard deviation of the computed flows agrees closely with the observed record.

a) Range

$$RQ_o = Q_{op} - Q_{ol} \qquad (3.8)$$

$$RQ_s = Q_{sp} - O_{ol} \qquad (3.9)$$

where, RQ_o is the maximum range of the observed daily flow
RQ_s is the maximum range of the simulated daily flow
Q_{op}, O_{sp} are the observed and simulated peak flows, respectively
Q_{ol}, O_{sl} are the observed and simulated low flows, respectively

b) Mean

$$\overline{Q_o} = \frac{\sum_{i=1}^{N} Q_{oi}}{N} \qquad (3.10)$$

$$\overline{Q_s} = \frac{\sum_{i=1}^{N} Q_{si}}{N} \qquad (3.11)$$

where, $\overline{Q_o}$ is the mean of the observed daily discharge
$\overline{O_s}$ is the mean of the simulated daily discharge
Q_{oi} is the daily observed discharge
Q_{si} is the daily simulated discharge
N is the number of days in one year

c) Standard Deviation

$$STDQ_o = \left[\frac{\sum_{i=1}^{N} (Q_{oi} - \overline{Q_o})^2}{N} \right]^{0.5} \qquad (3.12)$$

$$STDQ_s = \left[\frac{\sum_{i=1}^{N} (Q_{si} - \overline{Q_s})^2}{N} \right]^{0.5} \qquad (3.13)$$

where, $STDQ_o$ and $STDQ_s$ are the standard deviation of the observed and simulated daily discharge

d) Coefficient of Efficiency for Daily Flow

$$E_c = \frac{\sum_{i=1}^{N} (Q_{o\,i} - \bar{Q}_o)^2 - \sum_{i=1}^{N} (Q_{o\,i} - Q_{s\,i})^2}{\sum_{i=1}^{N} (Q_{o\,i} - \bar{Q}_o)^2} \qquad (3.14)$$

where, E_c is the coefficient of efficiency

The coefficient of efficiency is very commonly used for measuring the degree of association between observed and simulated flows. The value of $|E_c|$ tends to 1 for perfect agreement between observed and computed hydrographs.

e) Correlation Coefficient for Daily Flow

$$r = \frac{\sum_{i=1}^{N} (Q_{o\,i} - \bar{Q}_o)\,(Q_{s\,i} - \bar{Q}_s)}{\left[\sum_{i=1}^{N} (Q_{o\,i} - \bar{Q}_o)^2 \sum_{i=1}^{N} (Q_{s\,i} - \bar{Q}_s)^2\right]^{0.5}} \qquad (3.15)$$

Correlation coefficient r falls the range $-1 \le r \le 1$ and measures the linear relationship between computed and observed runoff values. A value of $r = -1$ indicates a perfect negative linear relation between them, while a value of $r = 1$ indicates a perfect positive linear association of them. If $r = 0$, then there is no linear relationship between them.

f) Daily Root Mean Square Error

$$RMS = \left[\frac{\sum_{i=1}^{N} (Q_{o\,i} - Q_{s\,i})^2}{N}\right]^{0.5} \qquad (3.16)$$

RMS can be regarded as a measure of absolute error between the simulated and observed hydrograph. RMS tends to be zero for perfect agreement.

3.6.2 Criteria for Peak Rate and Water Balance

a) Relative Error of Simulated Peak Discharge

$$REP = \frac{Q_{s\,p} - Q_{o\,p}}{Q_{o\,p}} * 100 \qquad (3.16)$$

b) <u>Relative Error of Simulated Annual Discharge Volume</u>

$$RE = \frac{\left[\sum_{i=1}^{N} Q_{s\,i} - \sum_{i=1}^{N} Q_{o\,i} \right]}{\sum_{i=1}^{N} Q_{o\,i}} * 100 \qquad (3.17)$$

RE can be regarded as a measure of the water balance simulation of the basin.

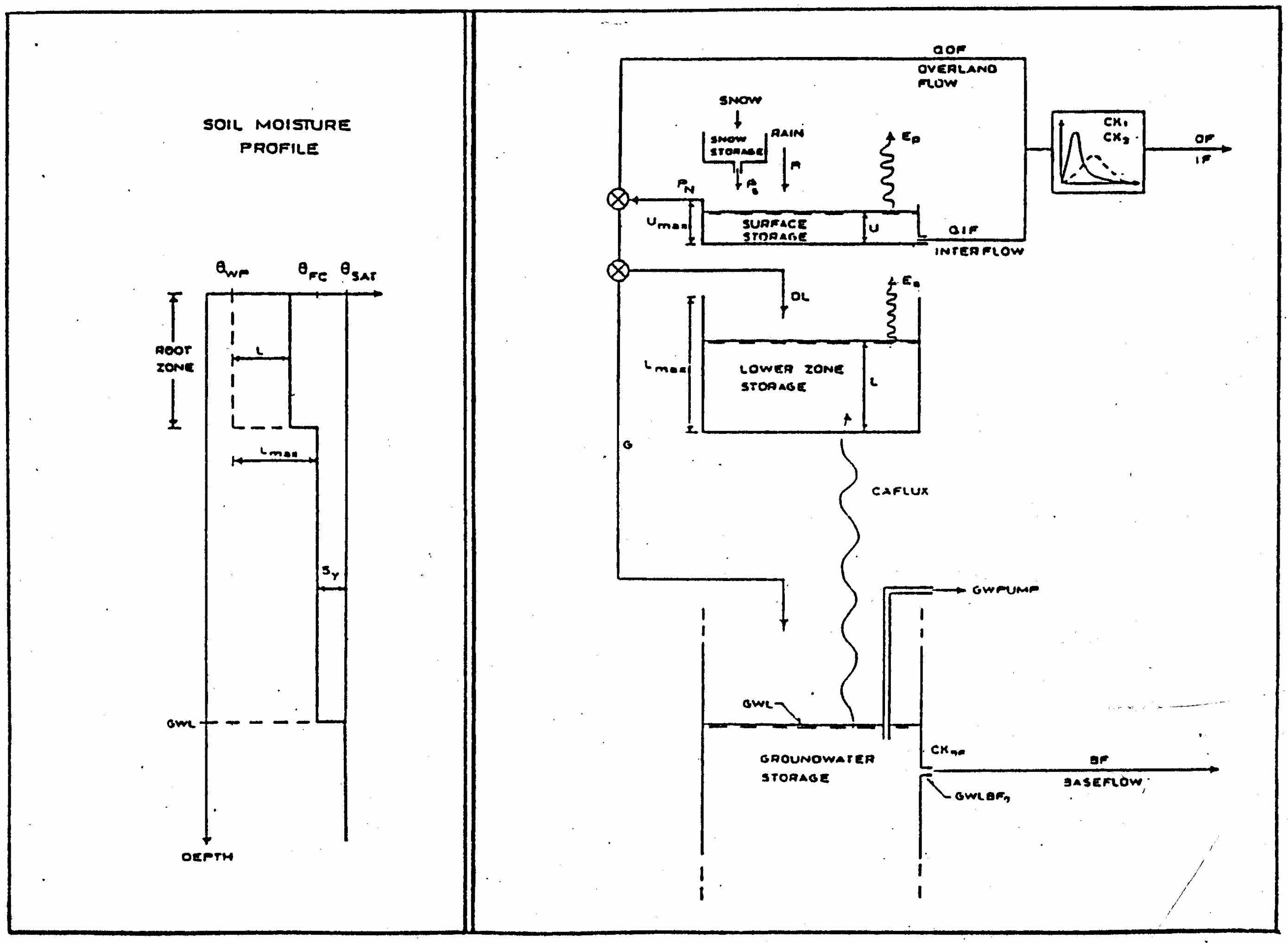

Fig. 3.1 Structure of the NAM Model

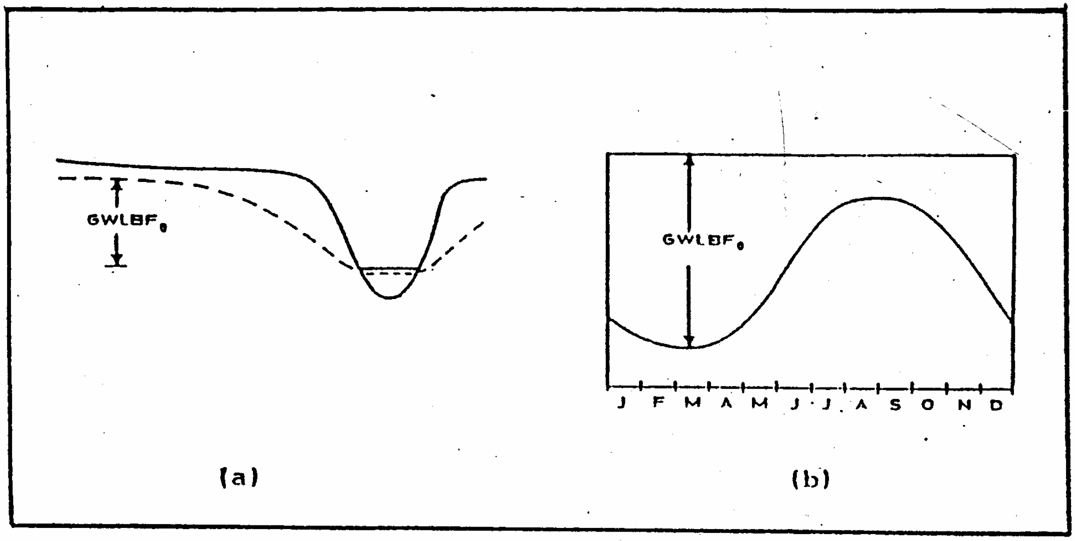

Figure 3.2 (a) The physical meaning of the parameter $GWLBF_0$ (= the maximum groundwater table depth which causes baseflow)
(b) The annual variation of $GWLBF_0$.

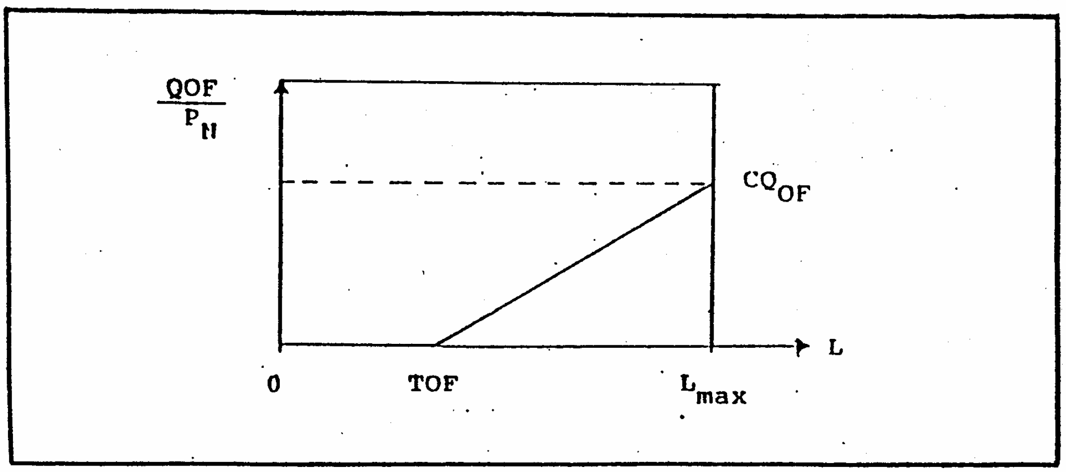

Figure 3.3 The function of the threshold value from the overland flow Equation (3.3).