TR(BR)-8/96-97

CATCHMENT MODELLING IN GIS ENVIRONMENT

NATIONAL INSTITUTE OF HYDROLOGY
JAL VIGYAN BHAWAN
ROORKEE - 247 667

ABSTRACT

Rainfall-runoff modelling helps in simulating and forecasting the flow from a catchment, and in determining the inflow series for the ungauged catchments on the basis of records of gauged catchments. Because of the scarcity of data, mostly lumped models have been used to model the rainfall-runoff process in a catchment. However, distributed models are more accurate and need to be adopted for modelling the complex processes at the scale of basins. With the development and wider availability of GIS tools, it is now easy to generate, store, manipulate, integrate and retrieve spatial data which can be used for distributed modelling of a basin. This has led to increased applications of distributed models for varied of water resources problems.

Topmodel is one such model having catchment representation in distributed functional form. Every such conceptualization also involves certain assumptions and the Topmodel is premised upon assumptions which lead to simple relationships between the catchment storage and local levels of the water table. The Topmodel is based on the concept of **Topographic Index.** This index represents as **ln(a/tanβ)**, where *a* is the area of the hillslope per unit contour length (m^2) that drains through a point and *tanβ* is the local surface slope. This index was first proposed by Kirkby and developed as a complete hydrological model by Beven and Kirkby (1979). The Kirkby index represents a theoretical estimation of the accumulation of flow at any point or the propensity of any point in the catchment to develop saturated conditions. The computational requirements of the model and quite modest. However, the processing of topographic data requires huge computer resources are if the catchment size is large and the pixel size is small.

The Topmodel has been implemented and applied to Kolar catchment of size 820 square km, located in Central India. The topographic data and DEM of the catchment were prepared from the Survey of India toposheets and the topographic data was processed through a GIS. The results of simulation show that a good match between the observed and simulated hydrographs was obtained despite handicaps in terms of data availability.

***

CONTENTS

CHAPTER - 1
INTRODUCTION

1.0 CATCHMENT MODELLING

In the broad sense, the term hydrological modelling implies rainfall-runoff modelling and a number of models have been developed for this purpose. Rainfall-runoff modelling helps in simulating and forecasting the flow from a catchment, and in determining the inflow series for the ungauged catchments on the basis of records of gauged catchments. The accuracy of the models increases from empirical to conceptual and from conceptual to physical. Because of the scarcity of data, lumped theoretical and conceptual models have mostly been used to model the rainfall-runoff process in a catchment. However, distributed models are more accurate and need to be adopted for modelling the complex processes at the scale of basins. With the development of GIS tools, it is now possible to use distributed models for varied water resources problems. Using GIS, it is possible to generate, store, manipulate, integrate and retrieve spatial data which can be used for distributed conceptual modelling of a basin.

The Topmodel is a conceptual model that has been developed to simulate the hydrological behavior of catchments in a distributed or semi-distributed way, in particular the dynamics of surface or subsurface contributing areas. It is based on simple approximate hydrological theories but recognizes that, because of the lack of measurements of internal state variables and catchment characteristics, the representation of the internal hydrological responses of the catchment must necessarily be functional while introducing the minimal number of parameters to be calibrated (Beven, 1989a). The modelling concepts are simple and the model structure can be modified to bring the predictions closer to the modeler's perceptions of the behavior of a particular catchment. The distributed nature of the predictions can be a great aid in this respect.

The Topmodel was developed with two main objectives:

1. To develop a pragmatic and practical forecasting and continuous simulation model,
2. To develop a theoretical framework within which perceived hydrological processes, issues of scale and realism and model procedures may be researched.

The parameters of Topmodel are intended to be physically interpretable and their number is kept to a minimum to ensure that their values do not become merely the statistical artifacts of a calibration exercise. It is recognized that this problem can never be entirely avoided. The model, in practice, represents an attempt to combine the computational and parametric efficiency of a lumped approach with the link to physical theory and possibilities for more rigorous evaluation offered by a distributed model.

1.1 SCOPE OF THE REPORT

The basic objective of this report is to apply the Topmodel to one Indian catchment and to evaluate the response of the model using the data availability constraints in Indian conditions. A detailed description of the history and various versions of the model is given in Beven et al. (1995a). The background theory to the model is explained in detail in the next chapter which has also drawn from the notes of Topmodel Workshop held at the University of Lancaster in 1995 and other published literature.

The focus area for the application of the model has been chosen as the Kolar sub-basin lying in the Narmada basin.

* * *

CHAPTER - 2
THEORY OF TOPMODEL

2.1 GENERAL DESCRIPTION

Every model is a simplified representation of the reality and the modelling process involves a sequence of stages of simplification. Each hydrologist may have an individual "perceptual model". Topmodel is one such model having catchment representation in distributed functional form. Every such conceptualization also involves certain assumptions and the Topmodel is premised upon following major assumptions:

1. the dynamics of the saturated zone can be approximated by successive steady state representations,

2. the hydraulic gradient of the saturated zone can be approximated by the local surface topographic slope, $\tan\beta$,

3. the distribution of downslope transmissivity with depth is an exponential function of storage deficit or depth to the water table, and

4. a spatially homogeneous recharge rate r (m/h) enters the water table.

These assumptions lead to simple relationships between the catchment storage (or storage deficit) and local levels of the water table (or storage deficit due to drainage) in which the main factor is the **Topographic Index (TI)** which is represented as **$\ln(a/\tan\beta)$**, where *a* is the area of the hillslope per unit contour length (m^2) that drains through a point and *$\tan\beta$* is the local surface slope. This index was first proposed by Kirkby and developed as a complete hydrological model by Beven and Kirkby (1979). The Kirkby index represents a theoretical estimation of the accumulation of flow at any point or the propensity of any point in the catchment to develop saturated conditions. This index can also be used to visualize macroscale flow patterns within a catchment.

According to the third assumption, one can write:

$$T = T_0 \, e^{-s/m} \qquad \ldots(1)$$

where T_0 is the lateral transmissivity when the soil is just saturated (m^2/h), s is local storage deficit (m) and m is a model parameter (m). Beven (1984) suggests that an exponential decline in (vertical) soil conductivity with depth may be adequate to describe the vertical changes in the hydraulic properties of a wide range of soils and has shown (Beven, 1986a) that the equivalent exponential transmissivity function can be derived under the assumption of isotropy. In terms of water table depth, equation (1) can be written as

$$T = T_0 \ e^{-fz} \qquad \ldots(2)$$

where z is local water table depth (m) and f is a scaling parameter (m^{-1}). The parameters f and m are approximately related by $f = \Delta\theta_1/m$ where $\Delta\theta_1$ is an effective water content change per unit depth in the unsaturated zone due to rapid gravity drainage (down to field capacity). A physical interpretation of the decay parameter m is that it controls the effective depth of the catchment soil profile. This it does interactively with the parameter T_0 which defines the transmissivity of the profile when saturated to the surface. A larger value of m effectively increases the active depth of the soil profile. A small value, especially if coupled with a relatively high T_0, generates a shallow effective soil, but with a pronounced transmissivity decay. This combination tends to produce a well-defined and relatively shallow recession curve response in the model hydrograph.

Under the second assumption of an effective water table gradient and saturated flow parallel to the local surface slope $\tan\beta$, the downslope saturated subsurface flow rate q_i per unit contour length (m^2/h) at any point i on a hillslope may be described by the equation:

$$q_i = T_0 \tan\beta \ e^{-f z_i} \qquad \ldots(3)$$

where T_0 and $\tan\beta$ are local values at a particular point. Under first assumption that, at any time step, quasi-steady-state flow exists throughout the soil, the subsurface downslope flow per unit contour length q_i may also be given by:

$$q_i = ra \qquad \ldots(4)$$

By combining equations (3) and (4), it is possible to derive a formula for any point relating local water table depth to the topographic index at that point, the parameter f, the local saturated transmissivity and the effective recharge rate, r:

$$z_i = (1/f) \ln(ra/T_0 \tan\beta) \qquad \ldots(5)$$

An expression for the catchment, lumped or mean, water table depth ($\bar{z}$) may be obtained by integrating equation (5) over the entire area of the catchment (A) that contributes to the water table. In the following, this areal averaging is expressed in terms of summation over all points (or pixels) within the catchment:

$$\bar{z} = (1/A) \sum_i -(1/f) \ln(ra/T_0 \tan\beta) \qquad \ldots (6)$$

In spatially integrating the whole catchment, it is also implicitly required that equation (4) holds even at such locations where water is ponded on the surface ($z_i < 0$). Beven (1991) justified this assumption on the basis that the relationship expressed by equation (3) is exponential and that, for many catchments, surface flow is likely to be relatively slow due to vegetation cover. By using equation (5) in equation (6), if it is assumed that r is spatially constant, *ln r* may be eliminated and a relationship found between mean water table depth, local water table depth, the topographic variables and saturated transmissivity having the form:

$$\bar{z} = z_i - \frac{1}{f}[\gamma - \ln\frac{a}{T_0 \tan\beta}] \qquad \ldots (7)$$

where $\ln(a/T_0 \tan\beta)$ is the *Soil-Topographic Index (STI)* of Beven (1986a), and

$$\gamma = \frac{1}{A}\sum_i \ln\frac{a}{T_0 \tan\beta} \qquad \ldots (8)$$

A separate areal average value of transmissivity may be defined as:

$$\ln T_e = \frac{1}{A}\sum_i \ln T_0 \qquad \ldots (9)$$

Defining λ as a topographic constant for the catchment

$$\lambda = \frac{1}{A}\sum_i \ln\frac{a}{\tan\beta} \qquad \ldots (10)$$

equation (7) can be rearranged to give:

$$f(\bar{z} - z_i) = [\ln (a/\tan\beta) - \lambda] - [\ln T_0 - \ln T_e] \qquad \ldots (11a)$$

This equation (11a), may also be written in terms of storage deficit as:

$$(\bar{S} - S_i)/m = \{ \ln (a/\tan\beta) - \lambda \} - \{ \ln T_0 - \ln T_e\} \qquad ...(11b)$$

The equations (11) express the deviation between the catchment average water table depth (or deficit) and the local water table depth (or deficit) at any point in terms of the deviation of the local topographic index from its areal mean, and the deviation of the logarithm of local transmissivity from its areal integral value. The relationship are scaled by the parameters f or m. Given a value of $\bar{z}$ or $\bar{S}$, equations (11) may be used to predict the pattern of water table depths over a catchment, based on knowledge of the spatial distribution of the STI or, if it is assumed that the soil saturated transmissivity is spatially constant, the TI ($a/\tan\beta$). Equation (11) implies that every point having the same STI value ($a/T_0 \tan\beta$) behaves functionally in an identical manner. The ($a/T_0 \tan\beta$) variable can therefore be considered as an index of ***hydrological similarity***.

The spatial distribution of ($a/\tan\beta$) may be derived from analysis of a Digital Terrain Model (DTM) of the catchment (Quinn et. al. 1991). To calculate the surface (or subsurface) contributing area, the catchment topographic index is expressed in distribution function form as shown in Fig. 1. Discretisation of the ($a/T_0 \tan\beta$) distribution function brings computational advantages. Given that all points having the same value of ($a/T_0 \tan\beta$) are assumed to behave in a hydrologically similar fashion, the computation required to generate a spatially-distributed local water table pattern reduces to one calculation for each ($a/T_0 \tan\beta$) class; calculations are not required for each individual location in space. This approach is computationally more efficient than a solution scheme that must make calculations at each of a large number of spatial grid nodes, a potentially significant advantage when iterative calibration and parameter sensitivity procedures are carried out.

Of particular interest is the case in which equation (11) predicts that the local water table is above the surface ($z_i < 0$), or above a specified near-saturated capillary fringe ($z_i - \Psi_0$). Such areas are where saturated overland flow are predicted to occur, and their spatial distributions constitute the variable saturated source areas which generate the modelled surface runoff response. In the procedural model, any rainfall falling upon the saturated source area is taken to be runoff, along with rainfall in excess of that required to top fill areas where local water table depth z_i is small. Equation (11) may also be used to predict the pattern of subsurface stormflow contributing areas, or flow through different soil horizons (Robson et al., 1992) if these can be defined by some threshold value of water table depth.

2.2 ORGANIZATION OF STORES IN TOPMODEL

There are various ways of organizing the stores embodied in Topmodel. Essentially

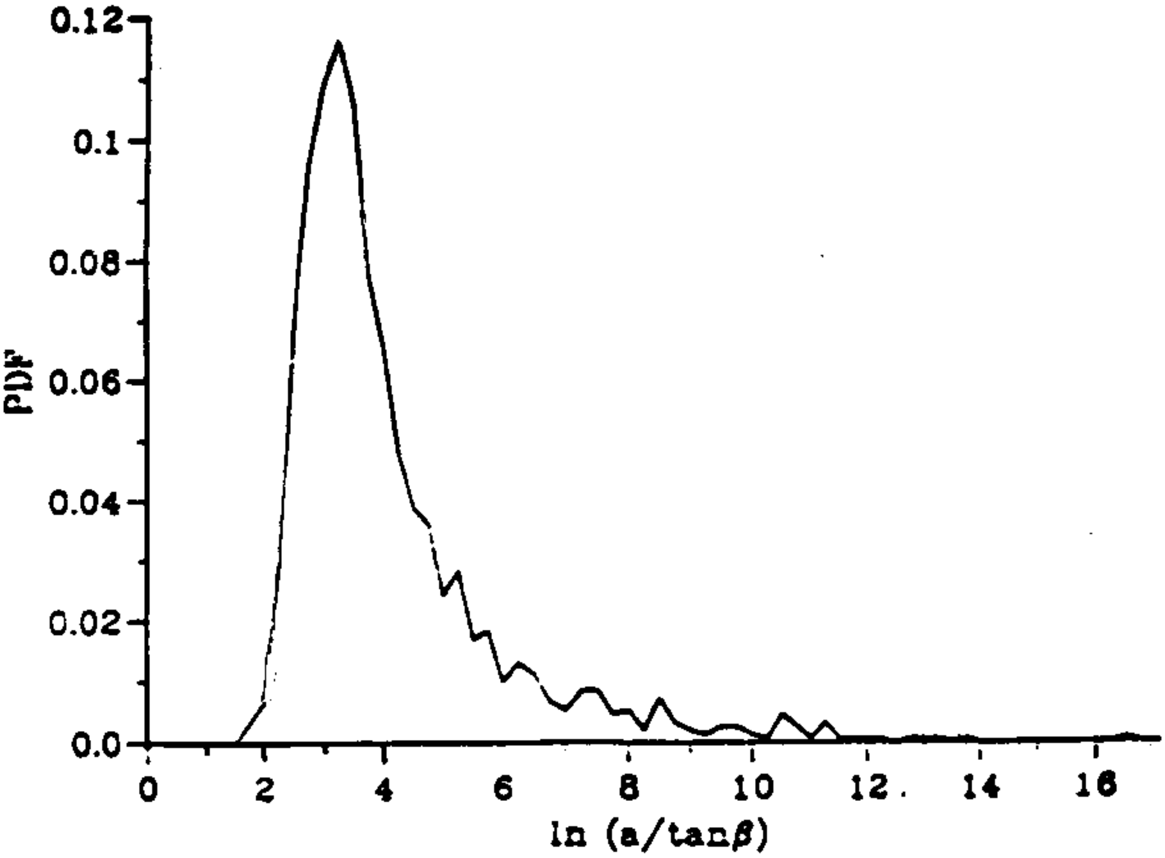

Fig. 1 A typical distribution function of ln(a/tanβ) index (after Beven 1995).

a simple series of stores is envisaged, routing water from the surface to the saturated zone. It is notable that, in any chain of stores, the form of the aggregate output will be most strongly controlled by the action of the least dynamic store Beven and Kirkby (1979). This applies to combinations of non-linear stores, such as those perceived to exist in catchment hydrology. Thus, it is important to accurately represent the non- linearity of the most slowly-responding store, whilst more dynamic stores may be approximated by simpler, linear representations. Apparently the saturated zone is typically the slowest to respond. Topmodel, therefore adopts a non-linear saturated zone store, but generally adopts simpler, linear representations of other, more dynamic stores. Storage, internal fluxes and inputs/outputs are generally expressed in terms of meters equivalent of water (per unit time).

Fig. 2 shows the schematic representation of storage elements within a discrete $\ln(a/\tan\beta)$ increment of a catchment area, showing the root zone store S_{rz}, vertical drainage store S_{uz}, and recharge to the saturated zone, q_v, for one increment, and area, a_i, draining through a particular point i. Beven and Kirkby (1979) applied Topmodel to the Crimple Beck catchment in the north of England using three stores: an interception and depression store, an infiltration store and a saturated zone store. The interception and depression store was required to be filled before infiltration could occur. Evaporation was allowed from this store at full potential rate until the store became empty. The infiltration store was formulated to allow for an Hortonian overland flow response, with a catchment average infiltration capacity calculated as a function of a maximum infiltration capacity and upper zone storage.

In a second application of Topmodel to the Crimple Beck site, Beven et. al. (1984) found that the earlier model structure, routing all infiltrated water immediately to the saturated zone, generated an overestimate of discharge which they attributed to an underestimation of evapotranspiration losses. It was considered that this problem reflected the model's inability to delay realistically the vertical unsaturated flow of water between the infiltration store and the water table. A notional "field capacity" was therefore, introduced to the model. This comprised a value of infiltration storage which had to be exceeded before flow to the saturated zone was permitted. The water thus, held from entering the saturated zone, remained available for evapotranspiration, reducing the model overestimation of storm flows following dry spells.

A formulation of the Topmodel stores presented by Quinn (1991), Beven (1991) and Quinn and Beven (1993) is shown in Fig. 3 for the case of saturation expressed in terms of depth to the water table. S_i is the local gravity drainage storage deficit, q_v is local recharge to the saturated zone, and ϕ_0 is the depth of the "capillary fringe". There will be one such

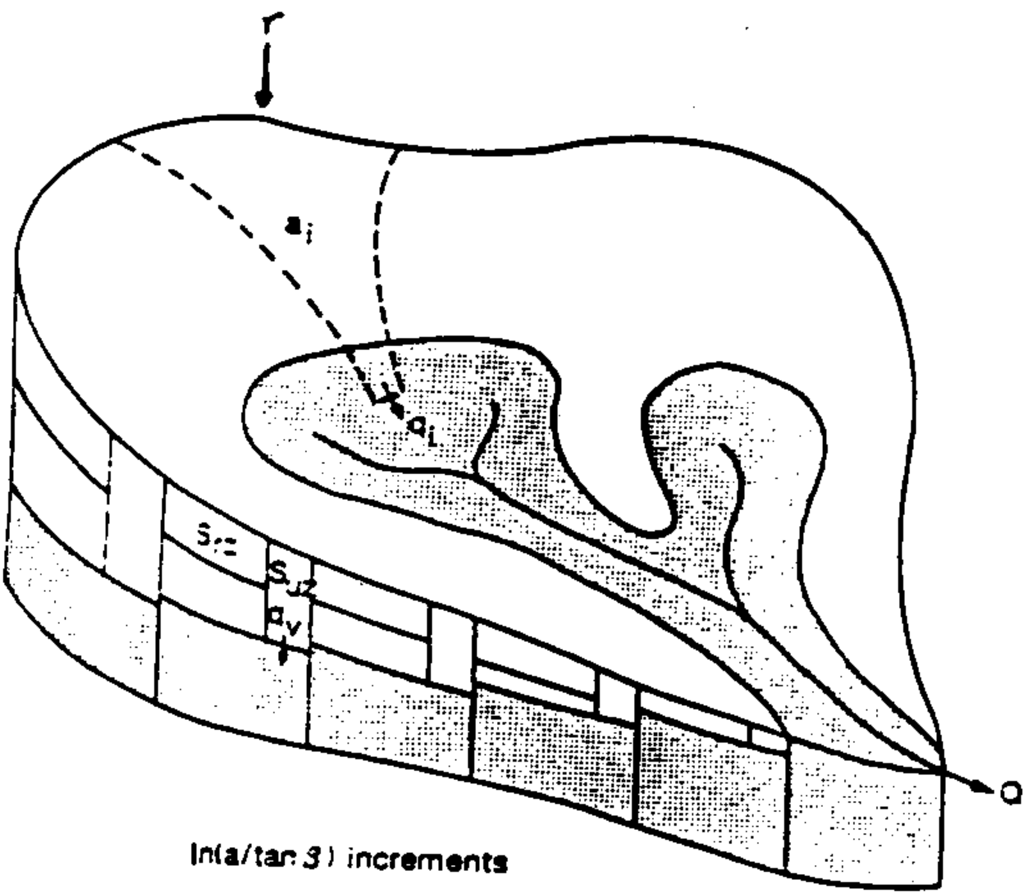

Fig. 2 Schematic representation of storage elements within a discrete ln(a/tanβ) index increment representing a catchment area. S_{rz} is the root zone store, S_{uz} the vertical drainage store, q_v recharge to the saturated zone, a_i is the area draining through point i and shaded area represents the area corresponding to saturation (after Beven 1995).

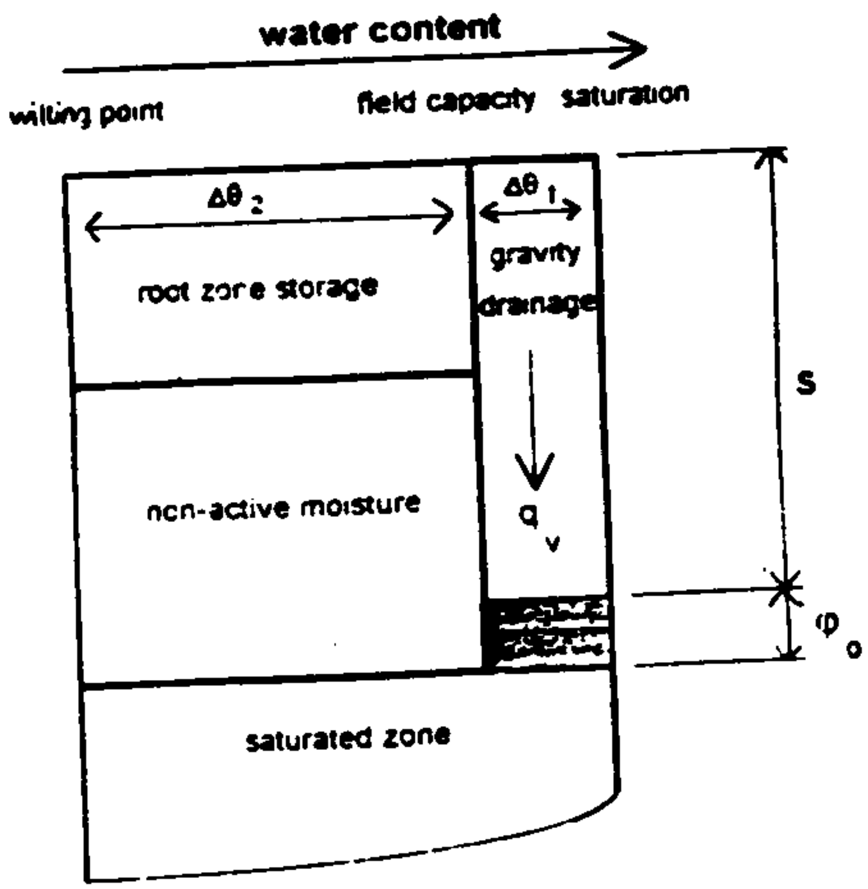

Fig. 3 The arrangement of storages in Topmodel. S_i is the local gravity drainage storage deficit, q_v is local recharge to the saturated zone and ϕ_0 is the depth of the capilarry fringe (after Beven 1995).

collection of stores for each class ($a/T_0 \tan\beta$) (Fig. 2), so that a proper account is taken of the predicted variation in depth to the water table in routing water through the unsaturated zone. Here, the "root zone" storage has the same function as the interception and evaporation stores of earlier Topmodel versions, although it may be necessary to add an additional interception and surface storage, particularly for forest canopies. The notion of a zone of non-active soil water reflects the "field capacity" concept. Vertical flux from the root zone to the water table may occur in the "gravity drainage" zone whenever field capacity is satisfied. The very simplest conversion between a storage deficit due to drainage and water table depth assumes that rapid gravity drainage affects only the largest pores, down to some "field capacity", and that the difference in storage between saturation and field capacity does not change with depth. This results in the simple linear scaling

$$S_i = (\theta_s - \theta_{fc})\ (z_i - \Psi_0) = \Delta\theta_1\ (z_i - \Psi_0) \qquad ...(12)$$

where θ_s is moisture content at saturation, θ_{fc} is moisture content at field capacity, $\Delta\theta_1$ is an effective drained porosity and Ψ_0 is an effective depth of capillary fringe assumed to be at saturation. Note that if saturation is expressed only as a storage deficit, it is not necessary to calibrate the parameters $\Delta\theta_1$ and Ψ_0 but any comparisons with observed water table levels must then be done separately.

2.3 MOISTURE ACCOUNTING: UNSATURATED ZONE FLUXES

The basic soil structure outlined in Fig. 2 may be used to accommodate a variety of unsaturated zone process. Beven (1991) explains the difficulty of predicting the spatial pattern of fluxes in the unsaturated zone. It is particularly difficult to account explicitly for the effects of local soil heterogeneity and macroporosity. No adequate mathematical description of unsaturated flow in structured soils with parameters, that can be identified at a practical prediction scale, is currently available (Beven, 1987a, 1989a,b, 1993) and if parameter values are to be determined by calibration, then only a minimal parameterisation is allowable.

A minimal parameterisation must account for the changes in unsaturated zone fluxes with local unsaturated zone storage and depth to the water table (or storage deficit), allowing for the fact that any functional relationships may be highly nonlinear. Two formulations that have been adopted in past Topmodel applications have assumed that the unsaturated zone flows are essentially vertical and have been expressed in terms of drainage flux from the unsaturated zone. These approaches have used either a simple storage deficit-dependent time delay or a conductivity-based flow equation. Both calculate a vertical drainage flux q_v for each topographic class. The formulations are:

1. Expressed in terms of storage deficit, Beven and Wood (1983) suggested that a suitable functional form for the vertical flux q_v at any point i is:

$$q_v = S_{uz} / S_i t_d \qquad \text{...(12a)}$$

where S_{uz} is storage in the unsaturated (gravity drainage) zone, S_i is the local saturated zone deficit due to gravity drainage, and dependent on the depth of the local water table. Parameter t_d is a time constant. Equation (11) is the equation of a linear store but with a time constant $S_i t_d$ that increases with increasing depth to the water table. This is the form used in the Topmodel software.

2. The second form was suggested by Beven (1986 a, b) on the basis of the Darcian flux at the base of the unsaturated zone which, for an exponential conductivity function, can be expressed as:

$$q_v = \alpha K_0 \exp(-f z_i) \qquad \text{...(13)}$$

where α, the effective vertical hydraulic gradient, is a parameter, K_0 is the saturated conductivity at the surface and z_i is the local water table depth. If α is set to unity, it implies that the vertical flux is equal to the saturated hydraulic conductivity just at the water table.

2.3.1 EVAPOTRANSPIRATION

Accounting for evapotranspiration with a minimal number of parameters poses a problem of similar complexity to that of unsaturated zone drainage. Topmodel follows generally-adopted practice in calculating actual evapotranspiration (E_a) as a function of potential evaporation (E_p) and root zone moisture storage for cases where E_a cannot be specified directly. In the Topmodel description of Beven (1991), evaporation is allowed at the full potential rate for water draining freely in the unsaturated zone and for predicted areas of surface saturation. When the gravity drainage zone is exhausted, evapotranspiration may continue to deplete the root zone store at the rate E_a given by:

$$E_a = E_p (1 - S_{rz} / S_{rmax}) \qquad \text{...(14)}$$

where the variables S_{rz} and S_{rmax} are respectively, the root zone storage deficit and maximum allowable storage deficit. If some effective root zone depth Z_{rz} can be assumed, then S_{rmax} can be calculated from $Z_{rz} (\theta_{fc} - \theta_{wp}) = Z_{rz} (\Delta \theta_2)$, where θ_{wp} is moisture content at wilting point. For calibration it is only necessary to specify a value for the single parameter S_{rmax} but the other forms may help in defining this value.

2.3.2 RECHARGE

The flux of water entering the water table locally at any time is q_v. This drainage is also a component of the overall recharge of the lumped saturated zone. To account for the catchment average water balance, all the local recharges must be summed. If Q_v is the total recharge to the water table in any time step, then

$$Q_v = \sum_{i=1}^{n} q_{v,i} A_i \qquad \ldots (15)$$

where A_i is the fractional area associated with topographic index class i as a fraction of total catchment area.

2.4 SATURATED ZONE FLUXES

Output from the saturated zone is given by the baseflow term, Q_b. This may be calculated in a distributed sense by the summation of sub-surface flows along each of m stream channel reaches of length l. Recalling equation (1), we may write:

$$Q_b = \sum_{j=1}^{m} l_j (T_0 \tan\beta)\, e^{-fz_j} \qquad \ldots (16)$$

Substituting for z_j using equation (5) and rearranging, it can be shown that:

$$Q_b = \sum_{j=1}^{m} l_j a_j e^{-\gamma}\, e^{-f\bar{z}} \qquad \ldots (16a)$$

Since a_j represents contributing area per unit contour length, then:

$$\sum_{j=1}^{m} l_j a_j = A \qquad \ldots (17)$$

Therefore:

$$Q_b = A\, e^{-\gamma}\, e^{-f}\, e^{-f\bar{z}} \qquad \ldots (18)$$

where A is the total catchment area (m^2). It is therefore, possible to calculate baseflow in terms of the average catchment water table ($\bar{z}$):

$$Q_b = Q_0 \, e^{-f z} \qquad ...(19)$$

where:

$$Q_0 = A \, e^{-y} \qquad ...(20)$$

is the discharge when $\bar{z}$ or $\bar{S}$ equals zero. This is the same form as that originally assumed by Beven and Kirkby (1979) if the linear relationship between storage deficit and depth to water table of equation (11b) holds. Solution of equation (19) for a pure recession, in which inputs are assumed to be zero, shows that discharge has an inverse or first order hyperbolic relationship to time as:

$$1/Q_b = 1/Q_0 + t/m \qquad ...(21)$$

Thus, if equation (19) is an appropriate relationship to represent the subsurface drainage of a given catchment, a plot of $1/Q_b$ against time should plot as a straight line with slope 1/m. Thus, given at least some recession curves, not greatly influenced by evapotranspiration or snowmelt processes, it should be possible to set the value of m which will then need minimal calibration. Fig. 4 shows master recession curves for three different UK catchments plotted as 1/Q versus time. A straight line plot indicates correspondence with equation (19) which may be an adequate representation for these cases.

The catchment average storage deficit before each time step is updated by subtracting the unsaturated zone recharge and adding the baseflow calculated for the previous time step, thus:

$$\bar{z}_t = \bar{z}_{t-1} + \Delta\theta_1 \, (Q_{bt-1} - Q_{vt-1}) \qquad ...(22)$$

A similar calculation can be made for a catchment average storage deficit formulation. Equation (19) can be used to initialize the saturated zone of the model at the start of a run. If an initial discharge is known and assumed to be only the result of drainage from the saturated zone, equation (19) can be inverted to give a value for $\bar{z}$ at t=0.

2.5 CHANNEL ROUTING AND SUB-CATCHMENT STRUCTURE

For many catchments, especially large ones, it may be inappropriate to assume that all runoff reaches the catchment outlet within a single time step. In such cases, some routing is required. To this end, Beven and Kirkby (1979) proposed that an overland flow delay function and a channel routing function might be employed within the Topmodel structure by the use of a distance-related delay. The time taken to reach the basin outlet from any point

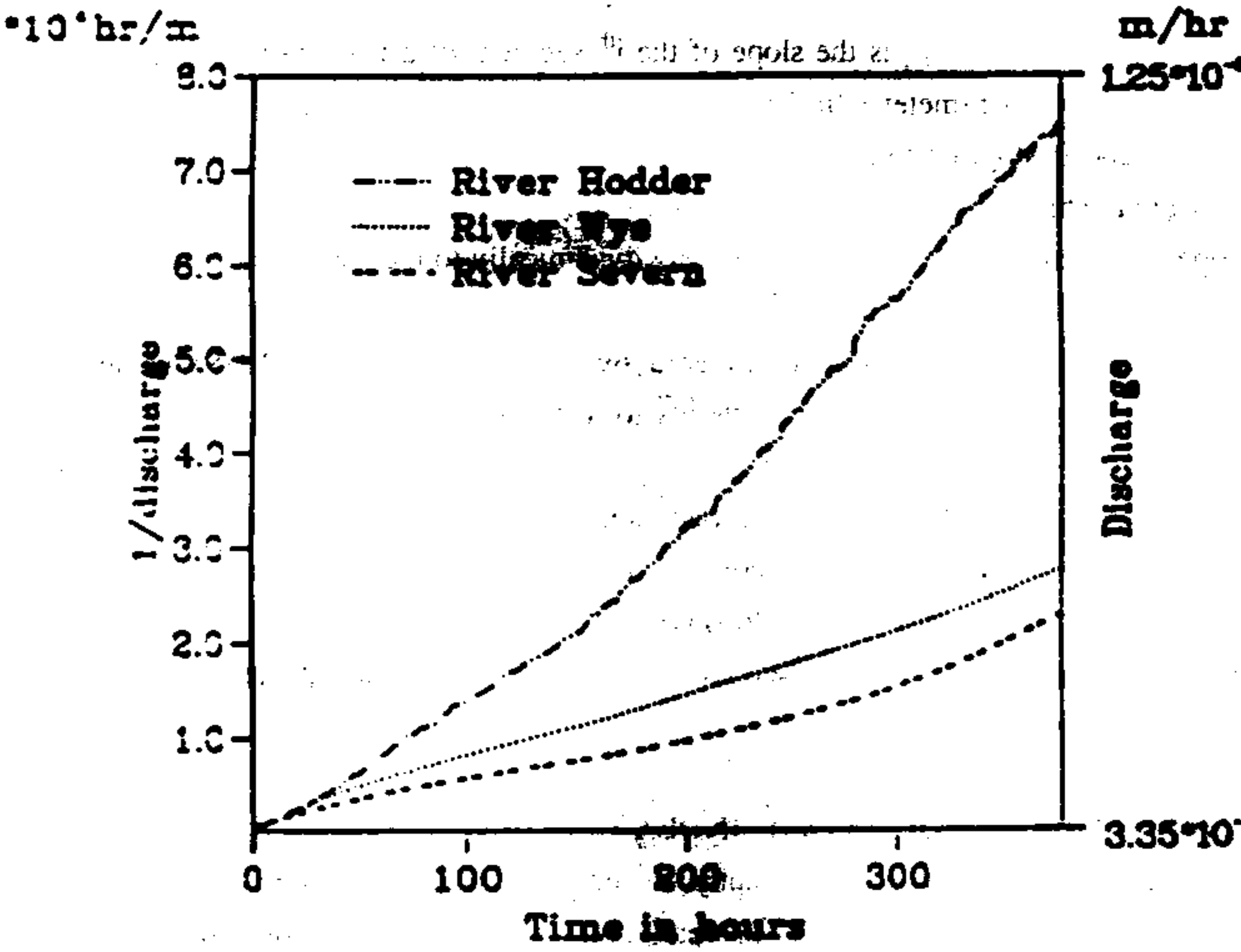

Fig. 4 Recession curves of three UK catchments.

is assumed to be given by :

$$\sum_{i=1}^{N} \frac{x_i}{\nu \tan\beta_i} \qquad \ldots (23)$$

where x_i is the length and $\tan\beta_i$ is the slope of the i^{th} segment of a flow path comprising N segments. The velocity parameter ν (m/h) is assumed constant. Given ν, equation (23) allows a unique time delay histogram to be derived on the basis of basin topography for any runoff contributing area. This is, in effect, a variation of the time-area routing method (Clark 1945), but developed so as to relate the runoff time delay dynamically to the size of the source area.

Channel routing effects were considered by Beven and Kirkby (1979) using an approach based on an average flood wave velocity for the channel network, this being related non-linearly to total outflow. The approach was an explicit approximation to a kinematic wave channel routing algorithm and is not recommended, since it is not always stable. Most applications have been based on a simple constant wave speed routing algorithm, equivalent to the channel width function based algorithms used by Surkan (1969), Kirkby (1976), Beven (1979) and Beven and Wood (1993), which has the advantage that it introduces only a single wave speed parameter.

2.6 DERIVING THE TOPOGRAPHIC INDEX

An analysis of catchment topography is required in order to derive the $a/\tan\beta$ distribution function (Fig. 1). To obtain discrete values of this function, some sampling of topography is implied. Early development of Topmodel relied upon the manual analysis, based on map information, of local slope angles, upslope contributing areas and cumulative areas. Beven and Kirkby (1979) and Quinn et. al. (1995) have outlined a computerized technique to derive the topographic distribution function and the overland flow delay histogram based on the division of the catchment into sub-basin units. Each unit was then discretised into small local slope elements on the basis of dominant flow paths (inferred from lines of greatest slope). Calculation of $(a/\tan\beta)$ is carried out for the downslope edge of each element.

However, given a DTM, more sophisticated computerized methods are now available. Quinn (1991) developed a group of Digital Terrain Analysis (DTA) programs, based on raster elevation data, with the aim of investigating their utility in deriving the topographic information required by Topmodel. Application of these techniques to catchment modelling studies has been described by Quinn et. al. (1991) and Quinn and Beven (1993). The functional nature of the index was discussed in terms of the quality of the representation of hydrologically significant topographic features by DTA methods. There are subjective choices to be made in the DTA. Quinn et al. (1991) described a multiple flow direction algorithm that is based on the distribution of area to all downslope grid elements. On the other hand, Wolock (1993) used an algorithm based on a single flow direction with the greatest slope. Most raster DTMs have been derived from contour data, digitized from existing maps. Moore et al. (1986) have used DTA techniques based on flowpaths derived directly from contour data to calculate the ($a/\tan\beta$) index.

Although these techniques offer great time savings in the application of Topmodel, they also raise two important questions. Firstly, how should "upslope contributing area" be calculated from topographic data and, secondly, what scale of resolution should be adopted for the catchment DTM? It may be noted that DTA methods, in the context of Topmodel, may have implications concerning the hydrological processes inferred to be related to topography and on the appropriate effective parameter values. The DTM must have a fine enough resolution to properly reflect the effect of topography on surface and subsurface flow pathways. Coarse resolution DTM data may, for example, fail to represent some convergent slope features. However, too fine a resolution may introduce perturbations to flow directions and slope angles that may not be reflected in the smoother water table surface. Though appropriate resolution depend on the scale of the hillslope features, however a resolution of 50 m or better data is normally suggested.

Wolock & Price (1994) studied the effects of DEM map scale and data resolution on the Topmodel. Higher values of $ln(a/Tan\beta)$ indicate greater potential for development of saturation. Higher values of this index curve at locations where large upslope areas are drained and where the gravitation local gradient is low (low value of $Tan\beta$). It was found that the spatial pattern of $ln(a/Tan\beta)$ distribution is affected more by DEM data resolution than by DEM map scale. All of the statistics of distributions, with the exception of the maximum value, were found to be affected by DEM data resolution and map scale. The Topmodel quotations were sensitive to the mean of the $ln(a/Tan\beta)$ distribution. It should not be concluded that the course resolution or map scale or DEM are inappropriate sources of topographic information for Topmodel. Quinn et al. (1991) found large pixels are unrepresentative of detail catchment form but are still useful for metro scale interpretation of moisture flux and prediction of hydrographs. Nested representative DTM with fine resolution should be used to internal state processors as 50 meter data are two coarse for point field data to validate. The shape and size of the catchment may alter if large cells are used in small sub-catchments. A large pixel resolution gives a bias to large *ln(a/Tanβ)* values. It was also reported that parameters are not transferable between grid resolution. This includes the optimum channel threshold values and final optimized parameter sets for the Topmodel.

It is worth noting that a parameterisation of the ln(a/tanβ) distribution may sometimes be useful. Sivapalan et al. (1990) introduced the use of a gamma distribution in their scaled version of Topmodel. Wendling (1992) also used a gamma distribution in a runoff production function for a flood forecasting model and Wolock (1993) gives details of a gamma distribution version for continuous simulation.

2.7 THE TOPMODEL CONCEPT AND HYDROLOGICAL SIMILARITY

The TI or STI, given acceptance of the simplifying assumptions of Topmodel, are indices of hydrological similarity at a point within the catchment. The response of any individual catchment, as predicted by Topmodel, depends upon the similarity in the distribution of the indices and in the input sequences to which the catchment is subjected including both, the time and space variability of rainfall rates and evapotranspiration losses. At small scales, say individual hillslopes, the pattern of variability may also be important, but at larger scales, it may be possible to have a sufficient sample of the possible variability of topography, soil and vegetation within the catchment to allow representation of that variability in terms of simple distribution functions. The scale at which this simplification may be valid, which they termed the **Representative Elementary Area (REA)**, has been investigated in hypothetical simulations by Wood et al. (1988, 1990). They used realistic topographic data together with stochastic spatial models of soil and rainfall to evaluate the variability of hydrological storm runoff responses predicted at different spatial scales. Both, infiltration excess and saturation excess runoff production mechanisms were included.

The results suggested a minimal REA, for the conditions simulated, of the order of 1 km^2. It is important to note that they did not suggest that this is the scale at which a lumped parameter description could be used; it may still be necessary to represent the heterogeneity of runoff production but at scales greater than the REA, this can be done by means of a distribution function or functions representing that variability. Sivapalan et al. (1987, 1990) have taken this concept further in producing a non-dimensional version of Topmodel in which different catchments can be compared in terms of various scaling coefficients. Similarity in these scaling coefficients should imply similarity in runoff production.

These results are, of course, dependent on an acceptance that the assumptions that underlie the Topmodel topographic index are realistic. These assumptions are, primarily, that the upslope contributing area extends to the hillslope divide and that the water table is parallel to the soil surface. Only then is the index easy to calculate from digital terrain analysis. It has already been shown above how an effective soil transmissivity can be used to modify the index but this requires additional information that is rarely available at the catchment scale. In addition, both assumptions can be relaxed (see below) but at the cost of requiring additional information.

There have been some tests of the hydrological similarity implied by the index. Troch et al. (1993a) have compared the Topmodel Conceptualization of Sivapalan et al. (1987) with field data and the output of a fully dynamic three-dimensional Richards equation based finite element model (although with quite different assumptions about soil conductivities). They showed that the quasi-steady state Topmodel representation was a reasonable approximation for the well measurements. In comparison with field data, Moore et al. (1988) found a good correlation with the topographic index and surface soil moisture measurements in a small fallow catchment, but Ladson and Moore (1992) found very poor correlations with gravimetric and remotely sensed (PBMR) surface soil moisture on the FIFE Konza Prairie site in Kansas, a natural prairie site with a dense grass cover at the surface. Further, little correlation was found between the topographic index and both gravimetric and surface soil moisture estimated by Synthetic Aperture Radar (SAR) for the agricultural Slapton Wood catchment in Devon, UK. Surface soil moisture is, of course, affected strongly by evapotranspiration as well as downslope flow and a better correlation might be expected with water table levels.

Burt and Butcher (1986) demonstrated that in some humid catchments with deeper subsurface and disconnected saturated flow systems, the topographic index (not allowing for a variable transmissivity) may have a limited utility as a predictor of water table depth. The Topmodel can give good simulations of discharges once the catchment has wetted up but it is very difficult to predict the wetting up sequence. Jones (1986) also discussed the limitations of using a static index, such as the $\ln(a/\tan\beta)$ index, for predicting catchment dynamic responses. Jordan (1992), working in a small catchment in Switzerland, found that for some dates, there was a very good correlation between topographic index and water table depth, while on other dates, relationship obtained was poor.

Another test variable is the extent of soil saturation under different conditions. Beven and Kirkby (1979) demonstrated similarity in pattern of measured saturated areas and the topographic index for a small subcatchment in Yorkshire, while Kirkby (1978) presented a similar analysis for the measured saturated areas.

***

3.1 TOPMODEL CALIBRATION

The parameter values introduced in the description of the Topmodel concepts have been given names based on physical descriptions of the flow processes. The descriptions used, however, are simplified physical descriptions, deliberately so since one response to the lack of detailed physieal descriptions at the hillslope scale is to minimize the number of parameter values to be calibrated. A number of studies have suggested that there is only enough information in a set of rainfall-runoff observations to calibrate 4 or 5 parameters (Kirkby, 1975; Hornberger et al., 1985). In essence, Topmodel provides a nonlinear functional model for fast and slow catchment responses and basic flow routing using a small number of parameters.

It has been argued in recent years that the realism, even of the most complex physics-based hydrological models, may be illusory (Beven 1987a, 1989, Grayson et al., 1992). One implication of this argument is that the physical interpretation of calibrated parameters may be difficult, since calibration of parameter values can compensate for model structural errors and there may be many different parameter sets that can give acceptable simulations and be physically plausible. One consequence of these arguments is that the idea of a set of "optimum" parameter values should be rejected in favour of a concept of equifinality of parameter sets (and possibly also model structures, see Beven, 1993). This concept allows that there may be multiple optima in the parameter response, interactions between parameters, errors in the observed data and, indeed, different "optimal" sets if different calibration periods are used.

Certain Topmodel parameters may be calibrated on the basis of field measurements and some applications have been made using only measured and estimated values (Beven and Kirkby, 1979; Beven et al., 1983). Each formulation of Topmodel may present an individual parameter set to be calibrated; however, there are invariably three or four critical parameters that most directly control model response. These are the saturated zone parameter f (or m in the original storage deficit formulation), the saturated transmissivity values T_0, and the root zone parameter s_{rmax} and in larger catchments, a channel routing velocity, ν.

It has already been mentioned that the saturated transmissivity decay parameter, m, may be derived from an analysis of catchment recession curves. Since this is one of the most important model parameters, it reinforces the idea that to simulate hydrological responses at the catchment scale, the most useful measurements will be made at the same scale especially in ungauged catchments. The most useful measurement needs to be stream discharge followed by a good estimate of the rainfall input, especially if it varies spatially.

Other parameters that can be estimated in the field include soil transmissivity profiles and root zone available water capacity. For early applications of Topmodel, Beven and Kirkby (1979) described a sprinkling infiltrometer method for calibrating the interception, infiltration store and overland flow velocity parameters. These techniques were used by Beven et al. (1984) with reasonable success in applications to three U.K. catchments. These measurements are, however, essentially "point" measurements whereas the model requires either effective values at the model grid scale, or ideally distributions that reflect the spatial heterogeneity in the catchment. The difficulties of sampling and the lack of a theory of effective parameter values mean that such heterogeneity (as well as the simplifications and errors inherent in the model structure) will inevitably introduce uncertainty in the predictions.

At any time step, it is possible to derive a probability density function for the whole range of model simulations weighted by their likelihoods. From this function, uncertainty bounds may be drawn defining the likely range of model predictions. No "optimum" simulation is recognized; however, as more calibration data is considered, or as more combinations of parameters are tested, simulation likelihoods may be updated using a Bayesian approach, thus modifying the model's uncertainty bounds. If these bounds become narrower, then the model calibration may be deemed to have improved. The case where the predictive uncertainty bounds do not envelope an observed response suggests a fundamental failure of the model or an inappropriate likelihood function. If the likelihood distribution changes so as to widen the uncertainty bcunds, then a reduction in the predictive power of the model is indicated.

An important feature of the uncertainty estimation procedure is that it allows simulation likelihoods to be defined according to different types of calibration data. For instance, borehole levels may be combined with observed discharge in calculating the power of a model to predict both water table depths and streamflow (Binley and Beven, 1991). This should give an improved indication of the realism of a model/parameter set combination. Field experience may be accommodated within this framework. For a catchment where overland flow was thought not to occur, all model simulations exhibiting overland flow could, for example, be given a likelihood of zero. It will be realized that traditional optimization corresponds here to one set of parameters having a likelihood of one, with all others given zero. One of the consequences of estimating uncertainties associated with predictions is that it raises the question of what type of data would be most valuable in constraining that uncertainty (see Beven and Binley, 1992), especially with respect to the distributed predictions of the model.

3.2 APPLICATIONS OF TOPMODEL

There have been a number of applications of Topmodel to a variety of problems. Some of these are summarized below:

3.2.1 Simulation of Humid Catchment Responses

Topmodel was originally developed to simulate small upland catchments in the U.K. (Beven and Kirkby, 1979; Beven et al., 1984). These studies showed that it was possible to get reasonable results with a minimum of calibration of parameter values. Since then, there have been applications to a number of other catchments in humid temperate regions in the eastern USA (Beven and Wood,1983; Hornberger et al., 1985), New Zealand (Beven, 1993), Scotland (Robson et al., 1993). In all of these cases, it has been found that after calibration of the parameters, Topmodel provides good simulations of stream discharges, and broadly acceptable simulations of variable contributing areas (see Quinn and Beven, 1993).

It has, however, been realized that the fitted parameter values may be difficult to interpret physically. In particular, calibration of the transmissivity parameter T_0 (to which the simulations tend not to be very sensitive) often yields very high values. This parameter controls the drainage rate from the saturated zone. There could be two reasons for this. One is that effective lateral downslope transmissivity values may be much higher than might be expected on the basis of small scale measurements of vertical hydraulic conductivity because of the effects of preferential flow pathways or zones of fractures. Secondly, the fast responses of Topmodel are governed by the distribution of the $\ln(a/T_0 \tan\beta)$ index. In the analysis of the catchment topography, the upslope drainage area, a, is assumed to extend to the divide. This might be an overestimate in many cases, especially in drier catchments, where an effective value, due to variability in the upslope recharge rate, might be much smaller. A large T_0 might compensate for any overestimates in the $(a/\tan\beta)$ index.

Catchments with deeper groundwater systems, or locally perched saturated zones may be much more difficult to model. Such catchments tend to go through a wetting up sequence at the end of the summer period in which the controls on recharge to any saturated zone and the connectivity of local saturated zones may change with time.

3.2.2 Simulation of Drier Catchment Responses

A model that purports to predict fast catchment responses on the basis of the dynamics of saturated contributed areas may not seem to be a likely contender to simulate the responses of catchments that are often dry. Durand et al. (1992) have applied the Topmodel to successfully simulate discharges in such a catchment.

3.2.3 TOPMODEL in a GIS Framework

As a distributed model that can make use of data on topographic, soil and vegetation information, Topmodel is well suited to implementation within a Geographical Information Systems (GIS) framework. Although many GIS systems do not have the facility to couple interactive modelling and visualization of distributed simulation results, a number of such Topmodel implementations have now been made. In UK, Topmodel has been linked to the Water Information System (WIS) which can directly access the 50 m topographic database for the UK, together with rainfall and discharge data for gauging sites. The Topmodel implementation can, for a selected point on the river network, determine the appropriate topographic index distribution, carry out parameter calibration and sensitivity analysis and display the resulting distributed predictions.

Recent work has focused on investigating model performance in terms of equifinality of parameter sets and uncertainty, rather than optimization of fit (Binley et al. 1991; Beven and Binley 1992; Beven, 1993; Freer et al., 1996). A methodology has been developed whereby the calibration of a hydrological model generates a set of uncertainty bounds defining the range of expected model responses. This procedure has been termed Generalized Likelihood Uncertainty Estimation (GLUE). Many model simulations are performed, each associated with a randomly chosen set of parameter values. The uncertainty methodology accords each parameter set a likelihood weight, defined by some subjectively chosen 'likelihood function' or performance index. This likelihood weight may be defined in terms of specific assumptions about the nature of the model residuals. Non-acceptable simulations are given a likelihood weight of zero. The cumulative likelihood of all retained simulations can be scaled to sum to unity.

Other Topmodel implementations within a GIS framework have been using the SPAN Modelling system by Stuart and Stocks (1993) and using GRASS by Chairat and Delleur (1993b).

3.2.4 TOPMODEL and Land Surface-Atmosphere Interactions

One limitation of the current generation of atmospheric general circulation models (GCM) is that they take little account of the variability of surface characteristics and hydrology in their calculation of land surface to atmosphere fluxes. Topmodel provides a way of incorporating such heterogeneity into soil-vegetation-atmospheric models in a way in which the surface availability of water for evapotranspiration can be allowed to vary dynamically. This possibility has been explored by Famiglietti and Wood (1991), Famiglietti et al. (1992) and Quinn et al. (1994). These papers show that the hydrological variability simulated by Topmodel may have an important effect on predicted evapotranspiration fluxes under certain circumstances. However, it must be remembered that the assumptions of Topmodel will not be met everywhere within a GCM grid square. A distribution function approach might still be appropriate but it may be necessary to consider a wider index of hydrological similarity.

3.3 ASSUMPTIONS AND LIMITATIONS

Like any other model, Topmodel may be expected to perform best when applied to catchments where its assumptions are met, in particular those of an exponential saturated zone store, a quasi-parallel water table and a topographic control on water table depth. Such catchments might be expected to have relatively shallow, homogenous soils and are likely to be generally quite wet, exhibiting the variable source area runoff mechanism which Topmodel simulates. These conditions may not be met everywhere, particularly in Indian catchment experiencing monsoon climate. Some Topmodel assumptions may be violated over certain ranges of behavior. Another instance where model assumptions may be invalidated is for those catchments where an infiltration-excess runoff mechanism is thought to be important; this mechanism is either ignored or modelled only crudely in Topmodel. These criticisms do not invalidate the Topmodel approach, but serve to demonstrate that the model has been somewhat restricted by its specificity to circumstances that accord with its assumptions. There may be restrictions on the geographical scope of Topmodel applications.

A fundamental assumption of Topmodel theory is that the saturated zone behaves as an exponential store, equations. Catchment recession behavior does not always conform to this model. Some work has now been completed on relaxing the exponential assumption within a Topmodel framework. Ambroise et al. (1996 a, b) have demonstrated the generalizations of the Topmodel theory required to use linear and parabolic transmissivity functions rather than, the exponential assumption. The alternative transmissivity functions lead to different recession curve shapes and different indices of hydrological similarity. For a linear transmissivity, the resulting index is $(a/\tan\beta)$ and for the parabolic case $\sqrt{(a/\tan\beta)}$.

It has been shown how the topographic index $(a/\tan\beta)$ is used to control the position of the water table in Topmodel. This procedure requires the assumption that the water table surface is everywhere approximately parallel to the ground, a condition which may not always be met, especially for deep soils. Quinn et. al. (1991) described a method whereby the assumption of a parallel water table may be relaxed. In basic Topmodel theory, the depth of the water table is defined with respect to the soil surface. Quinn et. al. (1991) introduced a 'reference level', based on a characteristic water table shape which may then deviate from being parallel to the soil surface. This reference level was used in order to calculate an *effective* ($a/\tan\beta$) distribution function for use in Topmodel. Calculated local water table depths were then adjusted according to the local reference level, requiring unsaturated zone calculations for each ($a/\tan\beta$) class to be distributed over a range of water table depth values.

A further assumption implicit in the use of the topographic index, is that the upslope contributing area, a, should be constant for any point. This may be a valid assumption in a moist climatic region, but for more arid catchments, there may be considerable evapotranspiration losses from farther areas of the catchment. These may be sufficient to prevent some parts of the catchment from contributing any water at all to the water table, especially during dry periods. The *effective contributing area* of the catchment may thus be variable. Topmodel assumes that the effective area of the catchment is defined solely by the watershed. One way of allowing for such variable upslope contributing areas is to use a fully distributed model, with calculations made for every grid element.

Topmodel uses simplified representations of the unsaturated zone and to improve the representation of evapotranspiration, the effects of macroporosity and heterogeneity of infiltration rates without adding a number of parameters to the model. The choice of an appropriate level of simplification for a given modelling problem is a matter of subjective choice within the constraints of available data, understanding of processes, and parameter identifiability. The Topmodel concepts are a compromise, making use of simplified physical reasoning to create a model capable of distributed predictions without the use of a large number of parameters. Most hydrological models can achieve reasonable predictions of stream discharges. The important feature of Topmodel is that the distributed nature of the predictions, the maps that are produced, allow a greater level of scrutiny of the results in relation to perceptions and measurements of the internal responses of the catchment under study. Such predictions may not be correct in spatial detail but major or unexpected differences in the nature of the spatial pattern of the response should stand out. The Topmodel concepts will not be appropriate in many catchments, especially those with deeper groundwater systems. However, it may be possible to reason as to how to modify the model structure to more closely reproduce the patterns of behavior.

***

CHAPTER - 4
APPLICATION OF TOPMODEL

4.1 THE STUDY AREA

For the present study, the data of the Kolar sub-basin in the Narmada basin, India, were used. The Kolar catchment is located in the latitude range of 22^0 40' to 23^0 08'N and longitude 77^0 01' to 77^0 29'E. During the course, the Kolar river drains an area of about 1350 sq. km. In the present study, only the catchment area of 820 sq. km up to the Satrana gauge- discharge site was considered. The average annual rainfall in the area is of the order of 1200 mm and the south-west monsoon accounts for about 90% of this amount. In the Kolar catchment, elevations vary from 600 m to 300 m. The rainfall in the mountainous areas is about 10% more compared to the plain areas.

4.2 DATA REQUIREMENTS FOR TOPMODEL

As the Topmodel is a topography based model, a topographic map with contours at close interval is needed to implement the model. The topographic map is used to create a digital elevation map (DEM) of the catchment and to compute the $\ln(a/\tan\beta)$ index. The hydro-meteorological data needed to run the Topmodel are rainfall, potential evapotranspiration and discharge at the outlet. These data are required at short time interval, say hourly. The important model parameter required to be given as input include the value of transmissivity (m^2/h), the unsaturated zone time delay per unit storage deficit (h), the main channel routing velocity (m/h), internal subcatchment routing velocity (m/h), the root zone available water capacity (m) and the exponent m in eq. (1). These parameters can be adjusted during a calibration run.

4.3 DATA AVAILABILITY

The base map of the Kolar sub-basin was prepared using the Survey of India toposheets (No. 55 E/4,8 and 55 F/1,5,6) at a scale of 1:50,000. The topographic sheets have a contour interval of 20m.

The daily rainfall data were available at four stations, namely, Rehti, Brijesh Nagar, Jholiapur and Birpur, for the years 1983 to 1988 and were considered for this study. In the present study, the weighted average rainfall for the catchment was used. The discharge data were available at the Satrana gauge site for the same period. An index map of the study area is given in Fig. 4.

4.4 PROCESSING OF SPATIAL DATA

4.4.1 GIS Software Used

The ILWIS (Integrated Land and Water Information System) GIS software was used in this study. The ILWIS was developed at the Computer Center of International Institute of Aerospace Survey and Earth Sciences (ITC), The Netherlands, for use on a PC platform. It provides the user with the capabilities of data gathering, data input, data storage, data manipulation and analysis and data output. It merges and integrates various conventional GIS procedures with image processing and can handle both vector and raster graphics data.

4.4.2 Creation of Spatial Database

The process of creation of database for the basin through the ILWIS involved collection of relevant data, converting these data in digital format through digitization, error checking and correction, polygonization of segment files and finally conversion of data acquired in vector structure to raster format.

The topographic map of the catchment was prepared using the Survey of India toposheets at the scale 1:50,000 having a contour interval of 20 m. The contours and the stream network were traced on a transparent sheet from the toposheets and these were then digitized on a PC through the ILWIS GIS package. A DEM of the catchment was prepared from the digitized data. The grid size of the DEM can be selected by the user. As a larger grid size is chosen, some information is lost due to discretization.

Quinn et al. (1995) have pointed out that grid sizes of around 100m are considered too large for application of the $\ln(a/\tan\beta)$ index which requires a finer grid resolution to depict the topographical form of the individual hillslopes. Large grid cells exhibit a bias towards larger index values. DEMs with fine resolution should be used to test internal state processes as 50m data are too coarse for point field data to validate. Small channels are ***hidden*** within the large-scale grid cells and a loss of resolution gives a loss of boundary information. A grid size of about 20m or less is considered necessary to obtain a realistic simulation. However, it is important to note that :

1. Fine grid scale digital terrain models are usually not available.
2. A reduction in grid size leads to huge increase in volume of data to be handled, even for a small catchment.
3. Interpolation errors may be introduced while creating fine grid maps from the data of coarse grids.

4. More pixels involve considerably more processing time.

In the present case, initially the grid size adopted in the horizontal plane was 50m*50m. Later, grids of smaller sizes were created. However, if the grid size was reduced, the number of grids became very large, as pointed out above. This required handling of very big data files as the catchment size was big, thus necessitating more computer resources. Moreover, as the contour interval was 20m, a smaller grid size did not lead to improved representation of topography in terms of elevation differences. Therefore, the grid size of 50m*50m was finally used.

4.4.3 Computation of Topographic Index

After the DEM was generated, an ASCII file of grid elevation was created from it using the format conversion module in which the elevations of the pixels were written. In a natural catchment, unless there is a lake inside the catchment, all the points should send the flow to the outlet. However, due to large contour interval of the topographic map and a large grid size used, local *sinks or pits* where the elevation is below the elevation of all surrounding pixels, are generated when DEM is created. The calculation of the $\ln(a/\tan\beta)$ index requires computation of contributing area at a point (a). Thus, it is necessary that these sinks are removed so that index is computed correctly. The sinks can be removed by artificially increasing the elevation of the sinks such that elevation of the sink pixel becomes higher than the elevations of the surrounding pixels.

A program named SINK is available to remove the sinks. This program runs in two modes: automatic and manual. In the manual mode, which requires graphical display support, the sink with elevation of surrounding pixels is displayed and the user is prompted to modify the elevation values. All the sinks are covered in this fashion. In the automatic mode, the user specifies an elevation increment and the maximum elevation change. The program raises the elevation of each sink node by the specified increment and then re-identifies the number of sinks. The iterations are continued till either all the sinks are eliminated or the maximum allowable elevation change is reached.

Next the $\ln(a/\tan\beta)$ index is to be calculated from the regular raster grid of elevations data of the catchment. A program named GRIDATB, which is based on multiple direction flow algorithm of Quinn et al. (1991), was used. The program reads in the raster grid data file. The output is a map file of $\ln(a/\tan\beta)$ index values and distribution function for the $\ln(a/\tan\beta)$ index which is needed by the Topmodel. The $\ln(a/\tan\beta)$ index for this catchment has been plotted in Fig. 5.

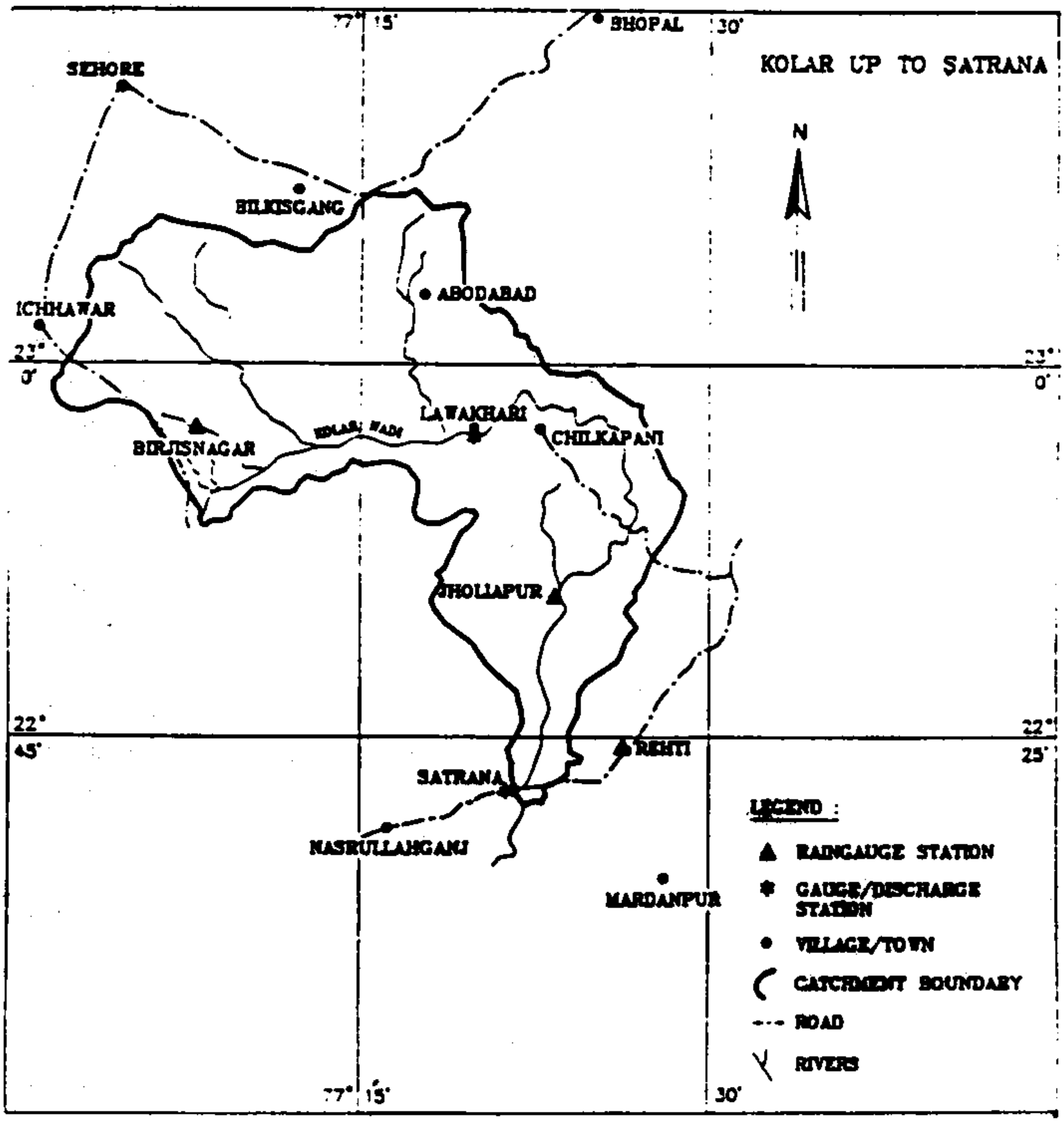

Fig. 5 The index map of the Kolar basin showing the location of rainguages and GD site.

4.5 SIMULATION OF CATCHMENT RESPONSE USING TOPMODEL

Besides the topographic data and the distribution function of the $\ln(a/\tan\beta)$ index, other data required to run the model were compiled in the appropriate format. The model was run on an IBM-compatible Pentium PC. Though Topmodel is capable of providing the results which can be mapped in space to verify their correctness in a spatial context. However, in the present case, measurements were not available to do this verification. The only variable which could be verified was the discharge at outlet. The model parameters were systematically varied to obtain a *good* match between observed and simulated discharge.

The values of various model parameters used in the final run are given in Table 1. A graph showing the observed and simulated hydrographs for an event of monsoon season of 1983 is given in Fig. 7. It can be seen that the match between the observed and simulated hydrographs is quite good. The efficiency of the model computed using the Nash formula turned out to be 0.837 for the data set. The major peak has been simulated rather well and most of the smaller peaks have also been reproduced well.

Table 1 : Topmodel parameters for Kolar catchment

Parameter m	Transmissivity (m^2/h)	UZ time delay (h)	Main channel routing velocity (m/h)	Internal routing velocity (m/h)	Root zone available water capacity (m)
5.5	5.0	10.0	180.0	100.0	0.10

The results of simulation for the data set of an event for the year 1984 have been plotted in Fig. 8. In this case also the major peaks show a good match. The model efficiency in this case was 0.725. It can be seen from the graph that the first small peak has not been reproduced properly. This could be due to in-appropriate initial condition at the beginning of simulation. It may be pointed out that a complete verification of the results of this kind of models requires cross checking of more than one variable. Therefore, besides discharge, if measured values of some other variable, for example, soil moisture are available, these should be used to verify the results. This type of information is usually not available in India, except for a few micro catchments used for research purposes.

For improvement of the results, the catchment can be subdivided in to smaller subcatchments and rather than using the average rainfall for the whole catchment, average rainfall for each subcatchment should be used.

***

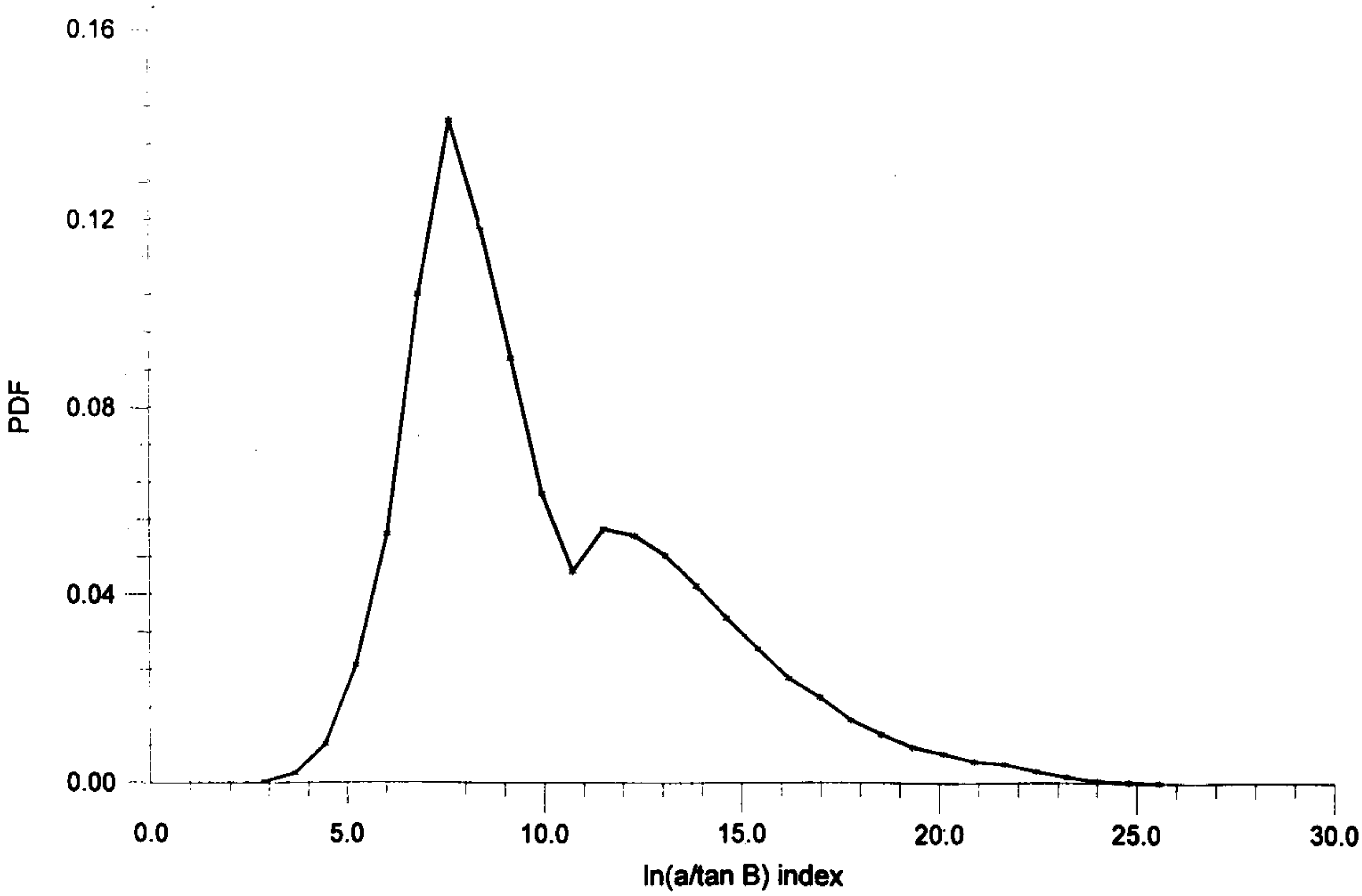

Fig. 6 The ln(a/tan B) index for the Kolar catchment

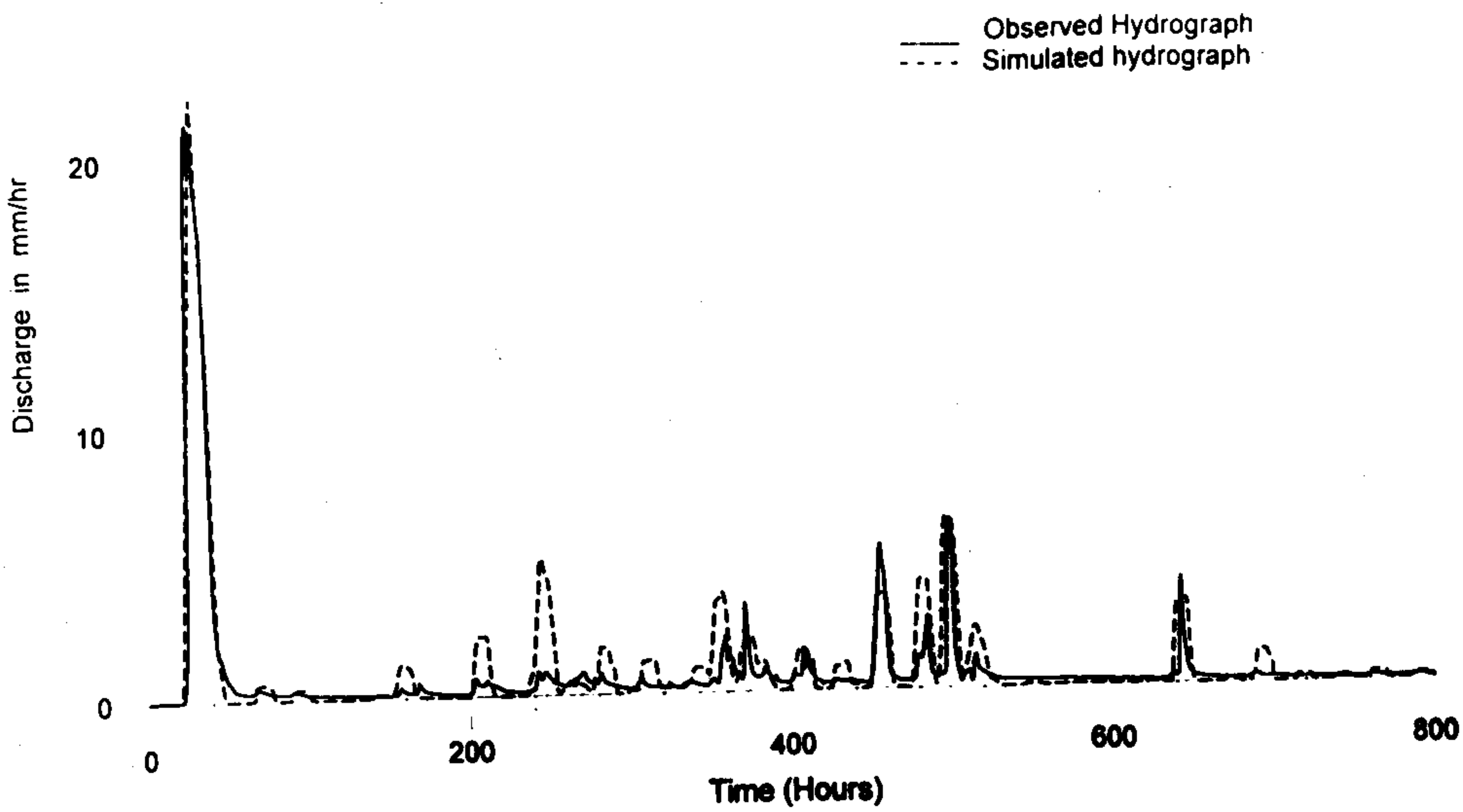

Fig. 7 Observed and simulated hydrographs for 1983 season

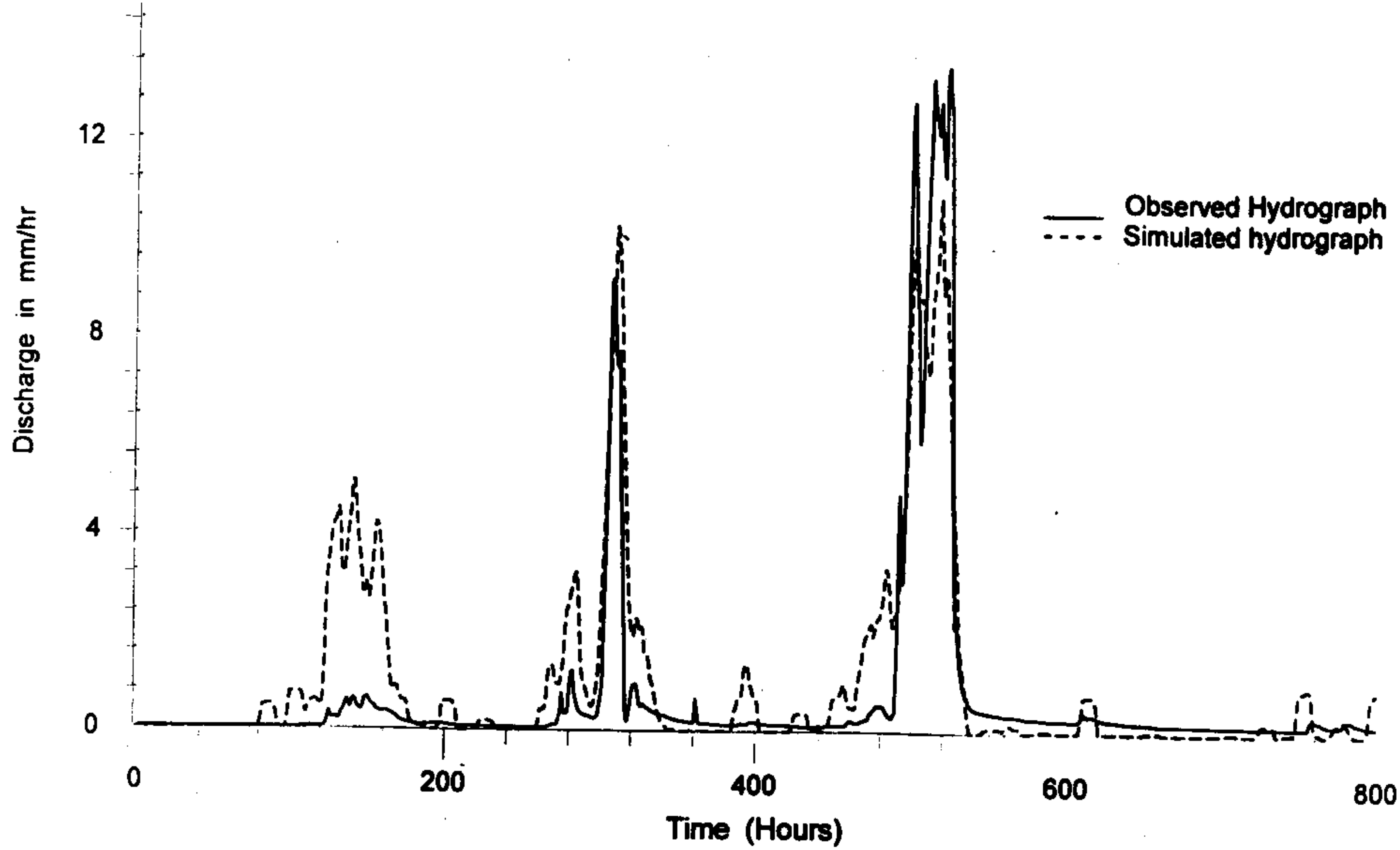

Fig. 8 Observed and simulated hydrograph for 1984 season

5.0 CONCLUSIONS

The Topmodel has been implemented and applied on a catchment of size 820 square km, located in Central India. The highlights of the study are :

1. The topographic data and DEM of the catchment were prepared from the Survey of India toposheets. The topographic data was processed through a GIS.

2. The Topmodel is distributed catchment model. It has the capability of providing spatial distribution of variables of interest, say, soil moisture, at the desired time. This could be very useful in understanding the basin response mechanism.

3. The computational requirements of the model and quite modest. However, the processing of topographic data can be a problem if the catchment size is large and the pixel size is small. In this case, the size of data file becomes very large and huge computer resources are required.

4. The topographic maps with small contour interval (of the order of a few meters) are required to get unbiased values of topographic index.

5. The results of simulation show that a good match between the observed and simulated hydrographs was obtained despite handicaps in terms of data availability.

6. Further application studies of this model may be carried out using data of a catchment for which better contour maps and hydrological data base is available.

***

REFERENCES

Ambroise, B., Beven, K., and Freer, J., 1996. Toward a generalization of the TOPMODEL concepts: Topographic indices of hydrological similarity. Water Resources Research, 32(7): 2135-2145.

Ambroise, B., Freer, J., and Beven, K., 1996. Application of a generalized TOPMODEL to the small Ringelbach catchment, Vosges, France. Water Resources Research, 32(7): 2147-2159.

Beven,K.J. 1984. Infiltration into a class of vertically non-uniform soils. Hydrological Sciences Journal. 29, 425-434.

Beven, K J. 1986a. Hillslope Runoff Processes and Flood Frequency Characteristics. In A. D. Abrahams (ed.) Hillslope Processes, 187-202, Allen and Unwin, Boston.

Beven, K J. 1986b. Runoff production and flood frequency in catchments of order n: an alternative approach, in V K Gupta, I Rodriguez-Iturbe and E F Wood (Eds.), Scale Problems in Hydrology, Reidel, Dordrecht, 107-131.

Beven,K.J. 1989. Changing ideas in Hydrology:The case of physically-based models, Journal of Hydrology, 105, 157-172.

Beven, K J. 1993 Prophesy, reality and uncertainty in distributed hydrological modelling, Advances in Water Resources, 16, 41-51.

Beven,K.J. 1995. Notes of The Topmodel Workshop, Lancaster University, UK.

Beven,K.J., and Kirkby,M.J. 1979. A physically based variable contributing area model of catchment hydrology. Hydrological Sciences Bulletin. 24(1), 43-69.

Beven,K.J. and Wood,E.F. 1983 Catchment geomorphology and the dynamics of contributing areas. Journal of Hydrology, 65, 139-158.

Chairat, S. and Delleur, J. W., 1993. Effects of the topographic index distribution on predicted runoff using grass. Water Resources Bulletin, 29(6): 1029-1034.

Durand, P., Robson, A., and Neal, C., 1992. Modelling the hydrology of submediterranean montane catchments (Mont- Lozere, France) using TOPMODEL. Initial results. Journal of Hydrology, 139(1-4): 1-14.

Franchini, M., Wendling, J., Obled, C., and Todini, E., 1996. Physical interpretation and sensitivity analysis of the TOPMODEL. Journal of Hydrology, 175(1-4): 293-338.

Freer, J., Beven, K., and Ambroise, B., 1996. Bayesian estimation of uncertainty in runoff prediction and the value of the data: An application of the GLUE approach. Water Resources Research, 32(7): 2161-2173.

Kirkby, M J, 1976, Tests of the random network model and its application to basin hydrology, Earth Surface Processes, 1, 197-212.

Ladson, A R and Moore, I D, 1993, Soil water prediction by microwave remote sensing and topographic attributes, J. Hydrology, 144.

Moore, R. D. and Thompson, J. C., 1996. Are water table variations in a shallow forest soil consistent with the TOPMODEL concept?. Water Resources Research, 32(3): 663-669.

Quinn, P F, Beven, K J and Culf, A. 1995, The introduction of macroscale hydrological complexity into land surface-atmosphere transfer function models and the effect on planetary boundary layer development, J. Hydrology, 166, 421-444

Quinn, P F, Beven, K J and Lamb, R, 1995, The ln(a/tan β) index: how to calculate it and how to use it in the TOPMODEL framework. Hydrol. Process, 9, 161-182.

Robson, A. J., Whitehead, P. G., and Johnson, R. C., 1993. An application of a physically based semi-distributed model to the Balquhidder catchments. Journal of Hydrology, 145(3-4): 357-370.

Sivapalan, M, Beven, K J and Wood, E F. 1987 On Hydrological Similarity: 2 A Scaled Model of Storm Runoff Production, Water Resour. Res. 23(12); 2266-2278.

Sivapalan, M, Wood, E F and Beven, K J. 1990 On Hydrological Similarity: 3. A dimensionless flood frequency distribution, Water Resour. Res., 26, 43-58.

Troch, P. A., Smith, J. A., Wood, E. F., and de Troch, F. P., 1994. Hydrologic controls of large floods in a small basin: Central Appalachian case study. Journal of Hydrology, 156(1-4): 285-309.

Wolock, D. M., 1995. Effects of subbasin size on topographic characteristics and simulated flow paths in Sleepers River watershed, Vermont. Water Resources Research, 31(8): 1989-1997.

Wolock, D. M. and McCabe, G. J., Jr., 1995. Comparison of single and multiple flow direction algorithms for computing topographic parameters in TOPMODEL. Water Resources Research, 31(5): 1315-1324.

Wolock, D. M. and Price, C. V., 1994. Effects of digital elevation model map scale and data resolution on a topography-based watershed model. Water Resources Research, 30(11): 3041-3052.

Wolock, D. M., Hornberger, G. W., Beven, K. J., and Campbell, W. G., 1989. The relationship of catchment topography and soil hydraulic characteristics to Lake Alkalinity in the northeastern United States. Water Resources Research, 25(5): 829-838.

Wood, E. F., Sivapalan, M., Beven, K., and Band, L., 1988. Effects of spatial variability and scale with implications to hydrologic modeling. Journal of Hydrology, 102(1-4): 29-47.

STUDY GROUP

Director	:	Dr. S. M. Seth
Scientists	:	Dr. S. K. Jain Mr. M. K. Goel
Scientific Staff	:	Mr. P. K. Agarwal
Typing	:	Mrs. Mahima Gupta