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A Topography-based Rainfall-Runoff Model
(TOPMODEL)

by
B. Venkateshy

Organised by

Hard Rock Regional Centre
National Institute of Hydrology
Belgaum-590 001 (Karnataka)

A TOPOGRAPHY BASED RAINFALL-RUNOFF MODEL (TOPMODEL)

Venkatesh.B
Scientist.'C'
Hard Rock Regional Centre
National Institute of Hydrology
Belgaum-590 001 (Karnataka)

1.0 Introduction

TOPMODEL is a set of conceptual tools that can be used to reproduce the hydrological behaviour of catchments in a distributed or semi-distributed way, in particular the dynamics of surface or subsurface contributing areas. It is based on some simple approximated hydrological theory but recognises 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. The modelling concepts have always been kept simple enough that the model structures used can be modified to bring the predictions closer to the behaviour of a particular catchment.

Hydrological processes within a catchment are dynamically and heterogeneously distributed. The interaction of vegetation and rainstorm dynamics may lead to very non-uniform inputs to the upper boundary of the soil. Soil and bedrock heterogeneity may further complicate flow paths, causing phenomena such as linked saturated 'pockets', macropore and piped flow and flow 'fingering'. At the lower boundary of the soil, irregularities in the bedrock interface may affect the pattern of subsurface discharge to streams

The general tendency of water to flow downhill is, however, amenable to macroscale conceptualisation. For the case of saturated zone flow, for soils that are shallow relative to the hillslope scale, this gravitational control should result in a water table that is nearly parallel to the surface topography over much of the hillslope length, at least when conditions are wet enough for there to be lateral downslope saturated flow. Water may then be expected to flow down from steep to shallow slopes, and into areas of slope convergence.

It is clear, however, that surface and subsurface hillslope flow processes are very complex. Overland flow, generated by either the infiltration-excess or saturated -excess mechanisms, may take place through a vegetated mat, through organic litter, or in ephemeral rills and channels. Flow may also re-infiltrate locally as 'run-on'. Saturation-excess flow, which may have a contribution from 'return-flow', may be caused either by the water table rising to the surface or by local filling of near surface soil layers, leading to a form of perched water table. The area where the water table reaches the surface may vary over time, causing the source area for saturation-excess overland flow to be dynamic. Similarly, the area contributing to stormflow via subsurface pathways may vary in time. Unsaturated zone flow may be highly irregular with a complex geometry of preferential and other pathways. It has to be impossible to incorporate all the perceived heterogeneity of unsaturated flows in a set of relatively simple, macroscale concepts. Hydraulic soil properties are also frequently highly variable in the vertical, often with higher conductivities close to the surface.

The complexity of such a complex model mitigates against the formal conceptualisation of hydrological processes into the functional mathematical structures in a model such as TOPMODEL or any other hydrological model. Every model is an attempt to capture the essence of this complex nature in a manageable way, but it is important to recognise that this conceptualisation also involves a considerable degree of simplification.

In making such a conceptualisation TOPMODEL is premised upon two basic assumptions:

A1. that the dynamics of the saturated zone can be approximated by successive steady state representations;

A2. that the hydraulic gradient of the saturated zone can be approximated by the local surface topographic slope, $\tan\beta$.

These assumptions lead to simple relationships between catchment storage (or storage deficit) and local levels of the water table (or storage deficit due to drainage) in which the main factor is topographic index ($a/\tan\beta$). The topographic index represents the propensity of any point in the catchment to develop saturated conditions. high values will be caused by high values due to either long slopes or upslope contour convergence, and low slope angles. TOPMODEL in its original form, however, takes advantage of the mathematical simplifications allowed by a third assumption;

A3. that the distribution of down slope transmissivity with depth is an exponential function of storage deficit or depth to the water table:

$$T = T_0 e^{-s/m}$$

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 the model parameter (m).

2.0 Description Of Topmodel

2.1 Vegetation interception capacity

The vegetation interception capacity is represented by a reservoir with a capacity of Sr_{max}. The water is extracted from the reservoir at potential evapotranspiration rate; the net precipitation (represented by the difference between precipitation and evapotranspiration) in excess of the capacity Sr_{max} reaches the soil and forms the input for the subsequent model components.

The 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 specific directly. In the TOPMODEL, 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 draining zone is exhausted, evapotranspiration may continue to deplete the root zone store at the rate of E_a , given by

$$E_a = E_p\left(1 - \frac{S_{rz}}{SR_{max}}\right)$$

where the variables S_{rz} and Sr_{max} are, respectively, root zone storage deficit and maximum allowable storage deficit. For calibration it is only necessary to specify a value for the single parameter Sr_{max} , but the other forms may help in defining this value.

2.2 Water Balance Component

2.2.1 Surface Runoff from Saturated Excess

In the TOPMODEL (Fig.1) the saturated hydraulic conductivity of the soil follows a negative exponential lax versus depth:

$$K_s(z) = K_0 \exp(-fz) \qquad (1)$$

where:

z = depth into the profile (z-axis pointing downwards);
K_0 = hydraulic conductivity at ground surface, held constant over the entire basin;
f = decay factor of K_s with z, held constant over the entire basin.

It is then assumed that 'the water table is parallel to the soil surface so that downslope flow beneath a water table at a depth z_i is given for any point i by'

$$q_i = T_i(z_i) \tan \beta_i \qquad (2)$$

where :

$\tan \beta_i$ = slope of the ground surface at the location i ;
$T_i(z_i)$ = transmissivity at point i ;
q_i = discharge per unit width.

The value of $T_i(z_i)$ is given by integrating equation (1) over the vertical :

$$T_i(z_i) = \int_{z_i}^{Z} K_S(x)\, dx = \frac{K_0}{f}[\exp(-fz_i - \exp(-fZ)] = \frac{1}{f}[K_S(z_i - K_S(Z))] \qquad (3)$$

where Z defines the 'bottom' of the saturated zone. Generally it is possible to assume that the saturated hydraulic conductivity at large depth Z becomes negligible compared with the conductivity at depth z_i. Substituting (3) into (2) gives;

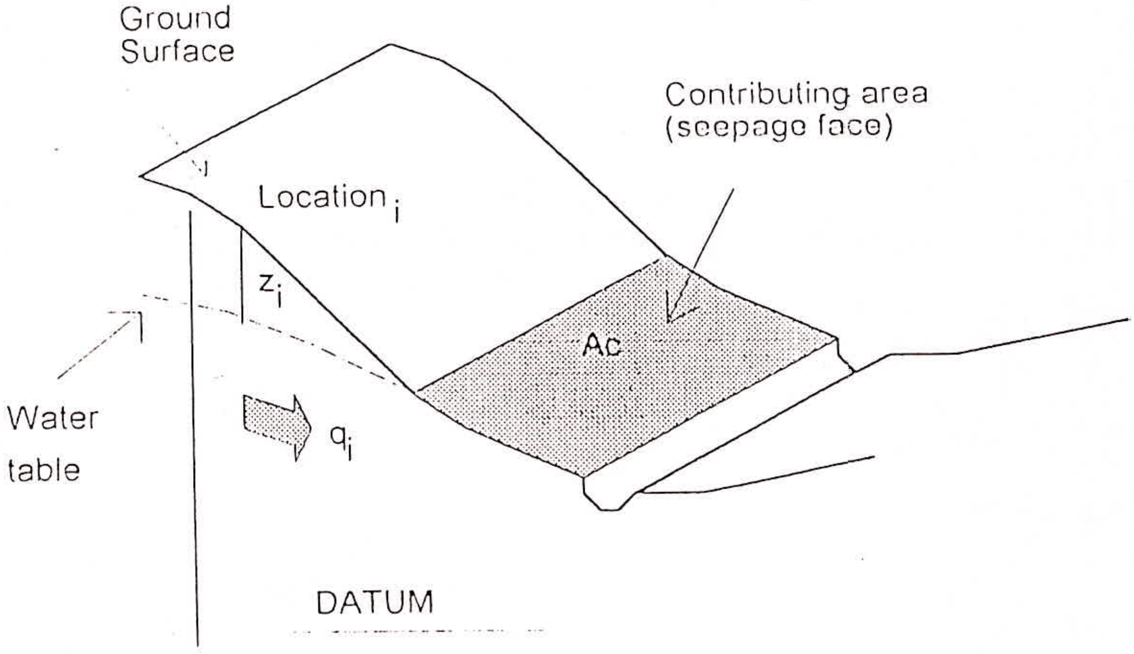


Figure 1 Schematic representation of a valley and the formation of runoff according to the TOPMODEL. A_c, contributing area to the surface runoff; q_i, interflow corresponding to an area drained per unit contour length.

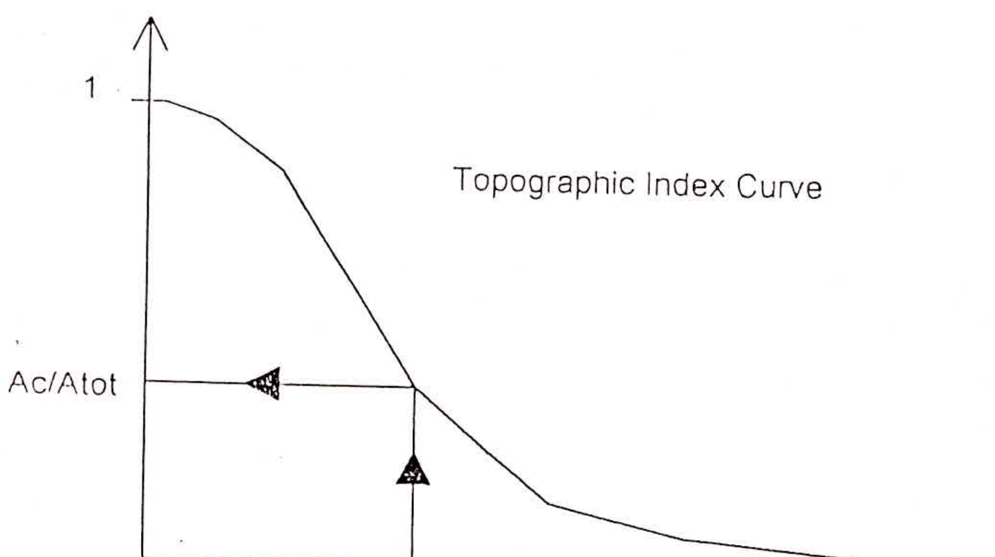


Figure 2 Topographic index curve. A_c / A_{tot} represents the fraction of the basin in saturated condition, for a given topographic index value $x = \log(a/\tan\beta)$

$q_i = K_0 / f \tan \beta_I \exp(-fz_i) = T_0 \tan \beta_I \exp(-fz_i)$ (4)

where $T_0 = K_0 / f$ is the transmissivity of the full saturated soil which, like K_0 and f, is assumed constant over the whole basin.

In steady state conditions the following expression also holds:

$a_i R = T_0 \tan \beta_I \exp(-fz_i)$ (5)

where ; R = spatially uniform recharge rate to the water table
a_i = area draining through location per unit contour length.

Making z_i explicit in (5) gives

$$z_i = \frac{1}{f} Ln\left[\frac{a_i R}{T_0 \tan \beta_i}\right] \quad (6)$$

Where
z_i = the depth of water table at any point in the basin under steady state condition.
R = Recharge to ground water body; assumed equal over the area.

By integrating over the entire area of the basin the mean values of the variable z_i is obtained as ;

$$\bar{z} = \frac{1}{A}\int_A z_i dA = \frac{1}{fA}\int_A \{-Ln[\frac{a_i}{T_0 \tan \beta_i}] - LnR\} dA \quad (7)$$

where equation (2) continues to hold for negative values of z_i. By combining eq.(5) with eq(7), the expression of $\bar{z}$ becomes

$$\bar{z} = \frac{1}{f}\left[-\frac{1}{A}\int_A Ln\left(\frac{a_i}{T_0 \tan \beta_i}\right) dA + fz_i + Ln\left(\frac{a_i}{T_0 \tan \beta_i}\right)\right]$$

and therefore ;

$$f(\bar{z} - z_i) = \left[Ln\frac{a_i}{T_0 \tan \beta_i} - \lambda\right] \quad (8)$$

where $\lambda = \frac{1}{A}\int_A Ln\frac{a_i}{T_0 \tan \beta_i} dA$

lastly

$$z_i = \bar{z} - \frac{1}{f}\left[Ln\frac{a_i}{T_0 \tan \beta_i} - \lambda\right] \quad (9)$$

It is worth noting at this point that, in the case of constant transmissivity, the expression for λ becomes

$$\lambda = \frac{1}{A}\int_A Ln\frac{a_i}{T_0 \tan\beta_i}dA = E\left(Ln\frac{a_i}{T_0\tan\beta_i}\right) = E\left(Ln\frac{a_i}{\tan\beta_i}\right) - LnT_0 \quad (10)$$

where the symbol E [ ] indicates the average value over the area of the basin.
It follows:

$$\text{(a)}\quad \lambda = \lambda^* - \text{Ln } T_0 \qquad \text{(b)}\quad \lambda^* = \text{E}\left(Ln\frac{a_i}{\tan\beta_i}\right) \quad (11)$$

Thus, eq (9) becomes:

$$z_i = \bar{z} - \frac{1}{f}\left[\ln\frac{a_i}{\tan\beta_i}(\lambda + LnT_0)\right] = \bar{z} - \frac{1}{f}\left[Ln\frac{a_i}{\tan\beta_i} - \lambda^*\right] \quad (12)$$

In other words the calculation of the depth z_i of the 'water table' is determined only by the parameter f and the topographic index x= ln(a/tanβ). In what follows, equation (12) will be referred to instead of eq.(9) since, as already stated, the transmissivity T_0 (i.e. the hydraulic conductivity $K_0 = T_0.f$) is considered constant over the whole basin.

Note now that if $z_i \leq 0$ then the ' water table' is, at least, level with the surface of the soil and therefore at this point -i- the saturation condition has been reached. All the points with $z_i \leq 0$ generate the basin fraction which is in a saturated condition where the rainfall produces direct surface runoff. The equation (12) shows that it is not the actual position of the i-th point which is important, but the value of the corresponding topographic index x= ln(a/tanβ); moreover from eq.(12) if x^* is the value of x which produces $z_i = 0$, then all the points with $x \geq x^*$ are in a saturated condition. The basin percentage with $x \geq x^*$ is then defined on the basis of the index curve which in turn represents the probability distribution of the variable x (Fig.2). A method used for computing the index curve, based on a Digital Elevation Model (DEM) which is described later.

Before concluding the model description, it is worth adding some considerations about the reference variable z_i and the use of the term 'water table'. In some other papers the reference variable is not the depth of the 'water table' z_i but the ' moisture deficit' S_i, which is nevertheless linked to the variable z_i through the equation $S_i = (\vartheta_s - \vartheta_r)z_i$, where ϑ_s and ϑ_r represents the moisture content in the saturated soil and the residual moisture content, respectively. The equations characterising the mode written in terms of S_i are entirely identical to the preceding ones except that z_i is replaced by $z_i = S_i/(\vartheta_s - \vartheta_r) = S_i/\Delta\vartheta$, or , as more frequently happens, z_i is substituted with S_i and the parameter $m = \Delta\vartheta/f$ is introduced. Below, unless otherwise specified, reference will nevertheless be made to the equations 1-20 written directly in terms of depth z_i and transissivity $T_0 = K_0/f$. This means that the porosity, or rather $\Delta\vartheta$, is taken as equal to 1.

Finally it should be noted that the use of the term 'water table' here considered is rather misleading since it suggests that a 'free surface groundwater table' is present, which, however, is completely absent in the TOPMODEL structure. The TOPMODEL 'water table' is indeed a 'perched water table' which defines the upper limit of a formation of a saturated zone in the upper part of the soil, which is clearly distinct from a real 'groundwater table'. This formation, which can be attributed to the reduction of hydraulic conductivity with depth, generates a saturated subsurface flow, i.e. an interflow which develops close to the surface where the driving head may reasonably be represented by the slope of the soil. On the basis of these considerations, reference will be made to 'interflow≡subsurface flow', 'saturated zone' and 'perched water table'.

2.2.2 Surface runoff from infiltration excess

The infiltration excess computation is based on the Philip equation;

$$g = c K_0 + \frac{1}{2} S t^{-\frac{1}{2}} \qquad (13)$$

where g is the potential infiltration capacity, S the 'sorptivity', and K_0 the saturation hydraulic conductivity at the soil level and c is a coefficient. The 'sorptivity' S is linked with K_0 as follows:

$$S = S_r K_0^{\frac{1}{2}} \qquad (14)$$

In the most recent versions K_0 may be allowed to vary randomly over the whole basin while S_r and c are regarded as constant coefficients.

2.3 Calculation of the flow in the saturated zone and the sequence of calculation in the TOPMODEL.

The equation (12) permits the estimation of the saturated basin fraction on the basis of the knowledge of the current value of $\bar{z}$. The value of $\bar{z}$ is updated at every Δt on the basis of the following equation :

$$\bar{z}^{t+1} = \bar{z}^{t} - \frac{\left(Q_V^t - Q_B^t\right)}{A} \Delta t \qquad (15)$$

where

Q_V^t = Recharge rate of the saturated zone from the unsaturated zone over the time interval t, t+Δt (the symbol R instead of Q_V^t has previously been used in the case of time-constant and space-uniform recharge);

Q_B^t = Outflow from the subsurface store into the channel over the time interval t, t+Δt ;

A = Area of the basin ;

Δt = Time interval.

The quantity Q_B^t can be defined analytically as:

$$Q_B^t = \int_L Q_{B_i}^t dL = \int_L T_0 \tan\beta . \exp\left[-f\, z_i^t\right] dL \qquad (16)$$

where L is twice the length of all stream channels. Bearing in in eq.(12), Q_B^t can be written as :

$$Q_B^t = \int_L T_0 \tan\beta \exp\left[-f\,\bar{z}^t - \lambda^* + Ln\frac{a}{\tan\beta}\right] dL = T_0 \exp\left[-f\,\bar{z}^t\right]\exp\left[-\lambda^*\right]\int_L a.dL$$

Since:

$$\int_L a.dL \quad = \text{A (total area of the basin)}$$

then:

$$Q_B^t = A.T_0.\exp\left[-\lambda^*\right].\exp\left[-f\,\bar{z}^t\right] = Q_0.\exp\left[-f\,\bar{z}^t\right] \qquad (17)$$

with : $Q_0 = A.T_0.\exp\left[-\lambda^*\right]$

The recharge Q_{V}^t can be represented as the sum of the contribution of all the grid square covering the basin (these grids are those of the Digital Elevation Model (DEM) used to define the index curve):

$$Q_V^t = \sum_{i\varepsilon A} Q_{V_i}^t = \sum_{i\varepsilon A} \alpha_i K_0 \exp\left[-f\, z_i^t\right] \qquad (18)$$

where α_i is the area of the i-th grid square. The equation assumes that the transfer from the unsaturated to the saturated zone is controlled by the conductivity at the depth of the 'perched water table', under unit vertical hydraulic gradient. Naturally eq.(18) holds good when the current water content in the unsaturated zone is not a limiting factor; otherwise the contribution is calculated on the basis of the actual amount of water available. Lastly, it is worth stressing that eq.(18) extends to all the grids where $z_i \geq 0$.

2.4 Initial conditions

The continuity equation (15) is initialised by assuming that the simulation begins after a long dry period; in other words the unsaturated zone is held to be totally dry and the flow observed at the basin outlet is deemed to have been generated only by the 'subsurface flow contribution'.

$$Q_V^1 = 0$$

$$Q^1_B = Q^1_{ob}$$

Recalleing eq.(17), Q^1_B may be written as:

$$Q^1_B = Q_0 \exp\left(-f \bar{z}^1\right)$$

and therefore the initial state is

$$\bar{z}^1 = -\frac{1}{f} Ln\left(\frac{Q^1_{ob}}{Q_0}\right) \qquad (19)$$

with eq.(12) it is possible to define the initial depth of the 'perched water table' in each grid square.

2.5. Description of a computational procedure for x=ln $(a/\tan\beta)$ in each grid square

In order to calculate x=ln $(a/\tan\beta)$ in each grid square the contributing area for that grid square must be calculated and then divided by the tangent of the slope relevant to that grid. Only the downward directions are considered below. If it is assumed that all the directions have the same water transportation probability, then the area drained by unit length of contour can be calculated as:

$$a = \frac{A}{nL} \qquad (20)$$

Where,

n = number of downward stream directions,

L = Effective contour length orthogonal to the direction of flow (L=$\frac{GS}{1+\sqrt{2}}$, where GS is the Grid Size of the DEM)

A = total area drained by current grid square (total upslope area)

One possible representation of tanβ is

$$\overline{\tan\beta} = \frac{1}{n}\sum_{i=1}^{n} \tan\beta_i \qquad (21)$$

Where $\tan\beta_i$ is the slope of the line connecting the current grid square with the furthermost grid square in the i-th downstream direction. Therefore,

$$\frac{a}{\tan\beta} = \frac{A}{L\sum_{i=n}^{n}\tan\beta} \quad \text{and} \quad Ln\left(\frac{a}{\tan\beta}\right) = Ln\left(\frac{A}{L\sum_{i=n}^{n}\tan\beta_i}\right) \qquad (22)$$

The amount of area A that contributes in each downstream direction is thus calculated as:

$$\Delta A_i = \left(\frac{A\tan\beta_i}{\sum_{i=1}^{n}\tan\beta_i}\right) \qquad (23)$$

The procedure is repeated on all the DEM grid square proceeding downstream.

2.6 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 of model output may be required. To this end, an overland flow delay function and a channel routing function might be employed within the TOPMODEL structure. Overland flow may be routed by the use of a distance-related delay. The time taken to reach the basin outlet from any point is assumed to be given by ;

$$\sum_{i=1}^{N}\frac{X_i}{v\tan\beta_i}$$

where X_i, is the length and $\tan\beta_i$ the slope of the i'th segment of a flowpath comprising N segments. The velocity parameter v(m/h) is assumed constant.

3.0 TOPMODEL Parameters and their Physical Significance

The concepts underlying TOPMODEL have thus far involved just the topographic index and five parameters: **M**, T_0, SR_{max}, T_d and SR_0. to transform the rainfall into runoff. Out of these five parameters, first 4 parameters i.e. **M**, T_0, SR_{max} and T_d play an important role in the transformation of rainfall into runoff. The physical significance of these three parameters is described below.

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 $\mathbf{T_0}$, 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 couple with a relative high $\mathbf{T_0}$, 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.

The parameter $\mathbf{T_d}$, the unsaturated zone time delay per unit storage deficit accounts for the changes in unsaturated zone fluxes with local unsaturated zone storage and depth to the water table (or storage deficit), and it is being related with the storage in the unsaturated zone and the local saturated zone deficit due to gravity drainage, and dependent on the depth of the local water table.

The vegetation interception capacity is represented by a reservoir with a capacity of $\mathbf{SR_{max}}$. The water is extracted from the reservoir at potential evapotranspiration rate; the net precipitation in excess of the capacity $\mathbf{SR_{max}}$ reaches the soil and forms the input for the subsequent model component.

The parameter $\mathbf{SR_0}$ represents the initial value of the root zone deficit, and it is related to the initial soil moisture available at the beginning of calibration period.

4.0 Model 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 physical descriptions at the hillslope scale is to minimise the number of parameter values to be calibrated. A number of studies have suggested that there is only enough informations in a set of rainfall-runoff observations to calibrated 4 or 5 parameters. In essence, TOPMODEL provides a non-linear functional model for fast and slow catchment responses and basic flow routing using a small number of parameters

In fact, certain TOPMODEL parameters may, in principle, be calibrated on the basis of field measurements and some applications have been made using only measured and estimated values. Each formulations of TOPMODEL may represent 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 m, the exponential decay factor, the saturated transmissivity values T_0, and the root zone parameter Sr_{max}, and in, larger catchments the channel routing velocity, v m/sec.

It has been noted 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 parameter 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 will 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. The measurements of these parameters can be made using available instruments (such as, infiltrometers). These measurements are, however, essentially 'point' measurement techniques, whereas the model requires either effective values at the model grid scale, or ideally distributions that reflect the spatial heterogeneity in the cathcment. 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 into the model predictions.

However, in most of the real world problems, these parameters are calibrated using the commonly available data such as rainfall, discharge and evaporation for any catchment. The parameters are changed to match the observed discharge and simulated discharge. The optimal values of these parameters are derived when the comparison of the observed and simulated discharge is very good.

5.0 Example Applications and Results

5.1 Simulation of Humid catchment responses

TOPMODEL was originally developed to simulate small upland catchment in the U.K. These studies showed that it was possible to get reasonable results with a minimum of calibration of parameter values. Since then there have been many applications to a number of other catchments in humid temperate regimes around the world. In India, the related works have been carried out for some of the catchment in Western Ghats, namely, Hemavathy and Malaprabha catchments. In all of these cases it has been found that, after calibration of the parameters, and broadly believable simulation of variable contributing areas. The calibrated model parameters are tabulated in Table.1. The observed and simulated hydrographs for Hemavathy and Malaprabha catchment for calibration and validation period are presented in Fig.3 and fig.4. respectively.

Table. 1. Calibrated TOPMODEL Parameters for Malaprabha and Hemavathy Catchment

Sl. No	Model Parameters	Calibrated Parameters for Malaprabha Catchment	Calibrated Parameters for Hemavathy Catchment
1	The parameter of the Exponential transmissivity function or recession curve (m) in mm	193.0	102.129
2	The Effective transmissivity of the soil when just saturated (T_0) in M^2/h	2.28	3.99
3	Time constant T_d in hrs	0.100	0.001
4	The soil profile storage available for transpiration (Sr_{max}) in mm	261.2	160.628
5	The initial storage deficit in the root zone (S_0) in mms	87.653	93.545

Experience with model calibration, however, has shown that the fitted parameter values may be difficult to interpret physically. In particular, calibration of the T_0 transmissivity parameter (to which the simulations tend not be very sensitive) often yield 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 fractured regolith. Certainly, Darcian flow simulations show that only with high values will any reasonably long slope show sensible drainage curves, otherwise the soils stays nearly saturated for long periods of time. Secondly, and possibly exacerbating this effect, the fast responses of TOPMODEL are governed by the distribution of the $a/T_0 \tan\beta$ 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, so that an effective a value, due to

Fig. 3a Calibration of TOPMODEL on Hemavati (June 1975-77)

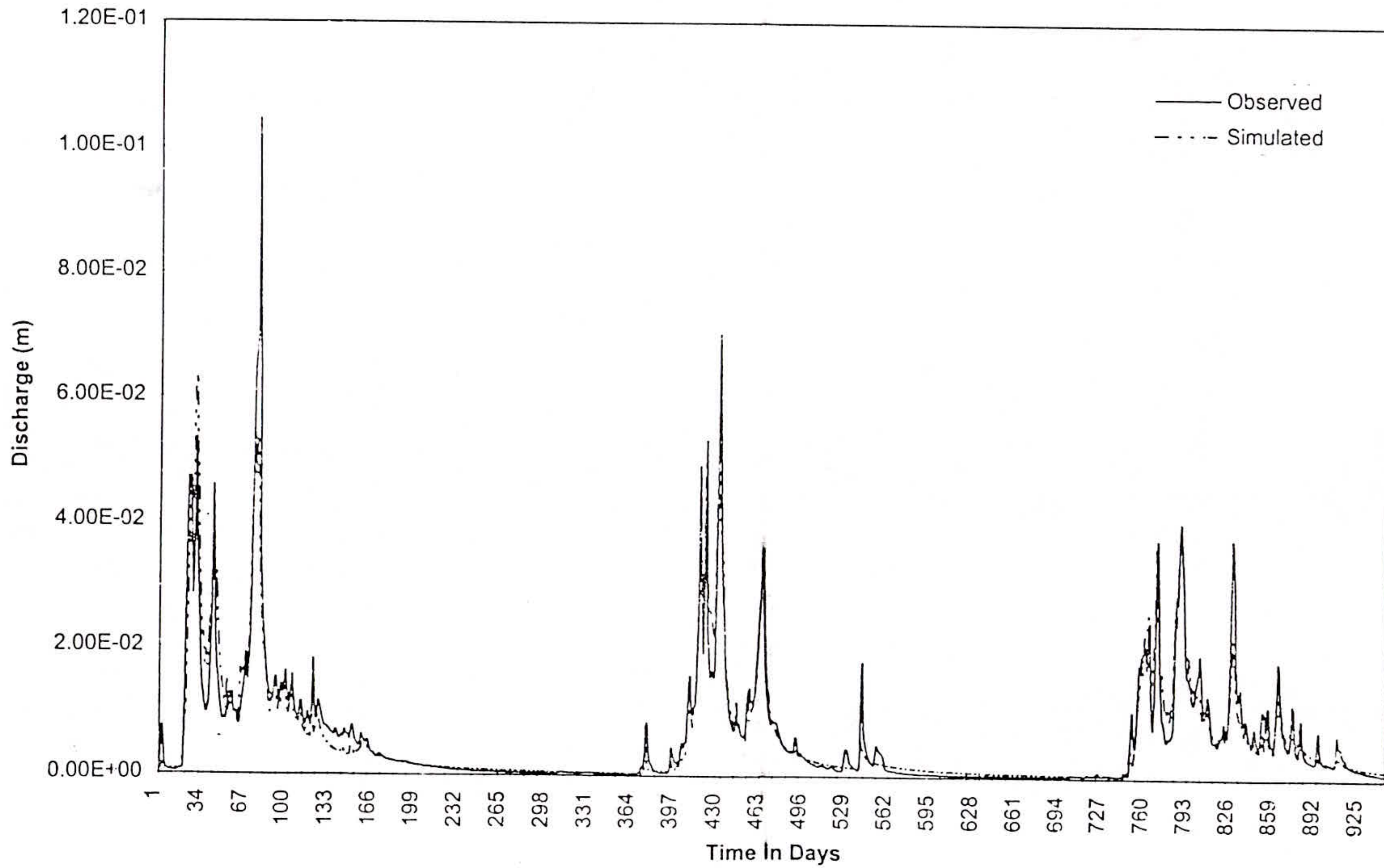


Fig.3b. Validation of TOPMODEL on Hemavati (1978-80)

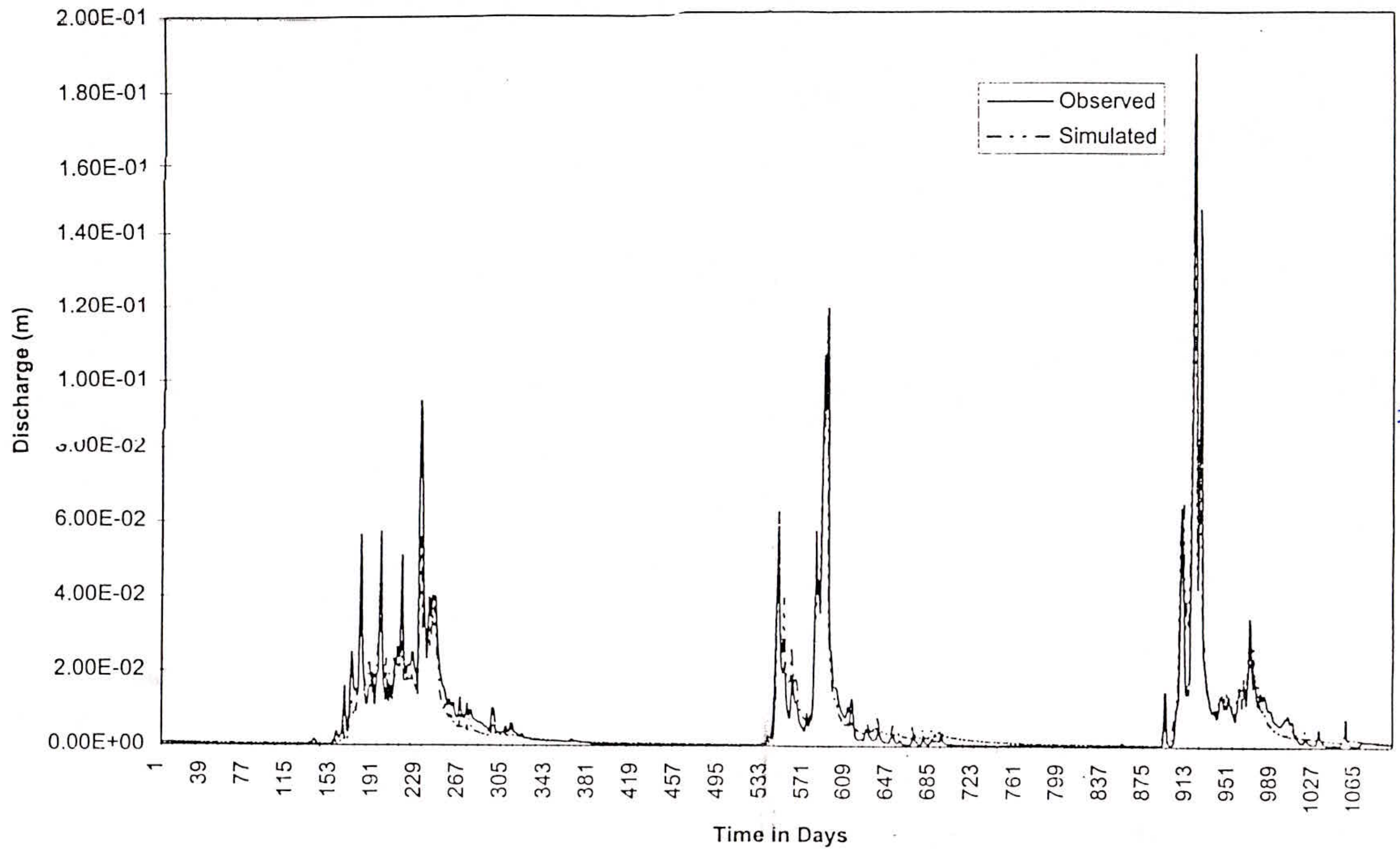


Fig.4a Observed and simulated discharge (1987-88) using TOPMODEL for Malaprabha catcment (Calibration)

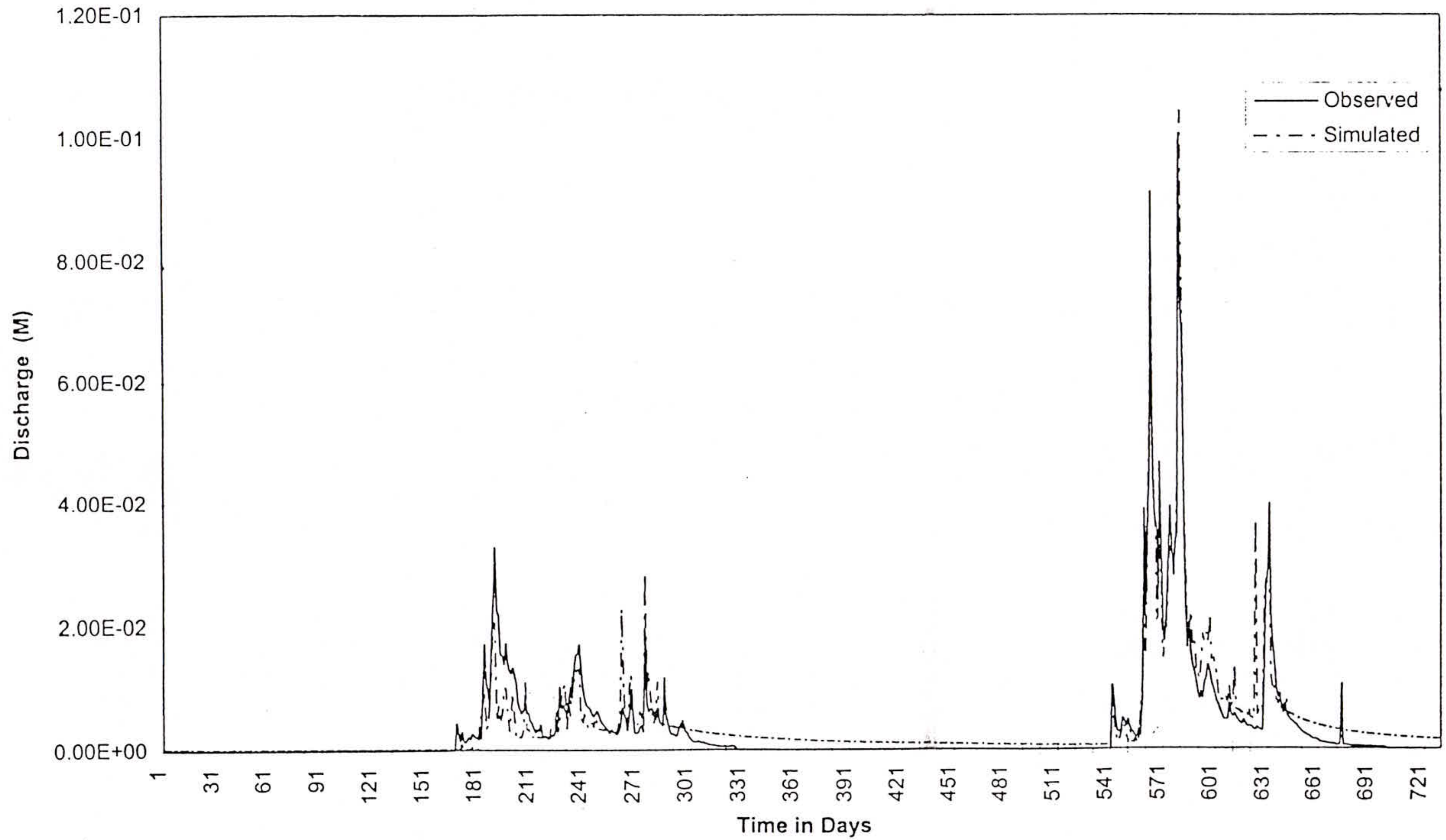
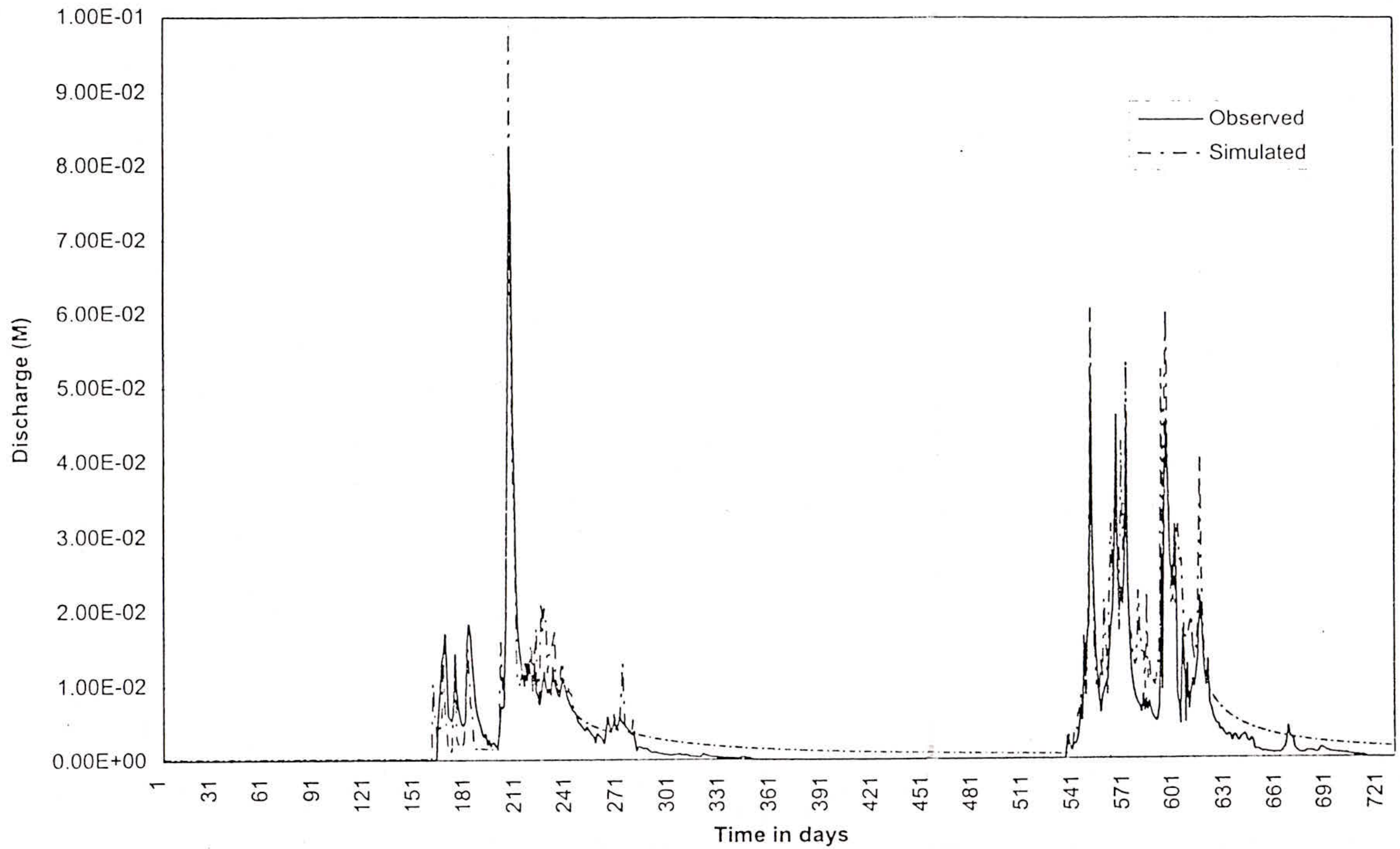


Fig.4b. Observed and simulated discharge (1989-90) using TOPMODEL for Malaprabha catchment (Validation)



variability in the upslope recharge rate, might be much smaller. A larger T_0 might then compensate for any overestimates in the a/tanβ 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.

5.2 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. However, Durand et al. (1992) have shown that TOPMODEL can successfully simulate discharge in such catchments at Mont-Lozere in the Cevennes, Southern France. They show that the runoff production function can be successfully used for flood forecasting purposes after calibration to a small number of storms and proved to be more robust in validation than other functions studied.

6.0 Remarks

The TOPMODEL may be seen as a product of two objectives. One is the development of a pragmatic and practical forecasting and continuous simulation model. The other is the development of a theoretical framework within which perceived hydrological processes, issues of scale and realism and model procedures may be researched. Parameters 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 artefacts of calibration exercise (although, as seen, 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.

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