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Application of Topmodel to Malaprabha Catchment

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CONTENT

LIST OF FIGURES

LIST OF TABLES

Abstract

The TOPMODEL (TOPography MODEL) is a variable contributing area conceptual model in which the predominant factors determining the formation of runoff are represented by the topography of the basin and a negative exponential law linking the transmissivity of the soil with the vertical distance from the ground level. Though the TOPMODEL is a conceptual model, i.e. one in which the physical reality is represented, in a simplified manner, the TOPMODEL is frequently described as being 'physically based', in the sense that its parameters can be measured directly in situ.

In the present study, the TOPMODEL has been applied to Malaprabha catchment in Karnataka to simulate the daily flows at Khanapur. River Malaprabha is a tributary of river Krishna. The catchment area of Malaprabha upto discharge measuring site Khanapur is 520 Sq. Km. The model uses topographic index for the formation of runoff. The topographic index for Malaprabha catchment was derived by developing a Digital Elevation Model (DEM) by interpolating the contours in the basin at 300 m grid size.

The results indicates that the model can be used to simulate the flows in the catchment quite accurately (the efficiency of the model is 0.89 and 0.79 respectively in calibration and validation run). Also, model is able to simulate the timing and magnitude of the peak flows satisfactorily.

1.0 Introduction

The analysis of the factors affecting the formation of basin flow is still one of the key area of research in hydrology. This analysis, which began as a quantitative investigation in the 1930s with Hortons's earliest studies of the mechanisms of surface runoff formations, has permitted the identification, at least partially of the varies factors potentially involved in the rainfall-runoff transformation process and their interaction. At the basin level, these factors can be summarised as follows; 1) space-time distribution of precipitation and evaporation; 2) basin topography and specifically, direction, length and gradient of slopes; 3) morphology and resulting structure of the drainage network; 4) heterogeneity of the characteristics of the hydraulic conductivity, porosity and thickness soil layering and, therefore, corresponding storage capacity; 5) hydrogeological characteristics of the aquifer below the basin; 6) vegetation cover, soil use and farming methods. These factors come into play in breaking down precipitation into surface runoff and infiltrated water, the interactions which occur between these components both prior to reaching the drainage network and afterwards, between the drainage network and aquifer. Specifically, the formation of surface runoff occur, in most cases, as the result of a combination of causes including rainfall intensity exceeding the infiltration capacity (soil saturation from above) and the available storage capacity in the soil being exceeded by the total mount of rainfall (saturation from below). This latter condition is caused by the low hydraulic conductivity in the first few soil layers or by the accumulation, at the foot of the slopes, of interflow; the area of the basin affected by these processes is not constant but evolves during the precipitation event (Franchini, 1996).

The literature contains many works which summarise the current level of understating of the physics of the complex process of rainfall-runoff transformation, and still more focus on the state of the art of the possible ways of schematising the whole process so as to develop mathematical models (Todini, 1988). In fact, the representation of runoff formation processes has been accomplished, over the decades, with methods which vary according to the purpose and application of the model. These range from the simple (in a manner of speaking) calculation of design discharge to the two-dimensional representation of the various processes, based on suitably conditioned mass balance, energy and momentum equations, and to the three-dimensional representation of all the exchanges. Taken together these latter kinds of model comprise the broad category of distributed differential models (Todini, 1988); they are frequently referred as 'physically based models' to highlight the fact that their respective parameters are reflected in the field measurements. Given their nature, they are mainly used in investigations and research as a mathematical support for the interpretation of physical reality. Yet, recently, major criticisms have been levelled against these complex and ambitious models (Beven,1989), calling into question the following: 1) the value of using equations which are undoubtedly valid for laboratory construction flow systems with clearly define boundary conditions and properties, but which do not afford equally acceptable guarantees for systems with markedly heterogeneous physical properties in which the boundary conditions remain uncertain, as occurs in a 'system' such as the one represented by a basin; 2) the meaning of the word 'physical' when assigned to a parameter which by its very nature has a local meaning but which instead is held to be representative of a loosely defined 'mean' value, which in turn is characteristic of the system's space discretisation grid.

Lying between the extreme categories of model indicated there are the distributed integral models (Todini, 1988), which certainly incorporate the large family of models commonly referred to as conceptual models. In these, the many factors listed above are only partly represented and, what is more, this is done by means of simplified schematisations whose basic purpose is to reproduce total flow measurable at the basin outlet.

The variable contributing area concept was introduced by Hewlett in 1961, and further clarified by Dunne and Black(1970). Subsurface flow was gradually recognized as a major storm-flow-generating processes, by itself as 'return flow' contributions to overland flow and by its strong influence on saturation overland flow. More recent research have been oriented towards the integration of all these concepts in a continuum of subsurface processes. Increasing the complexity of conceptual perception with respect to hydrological processes appeared to be an adequate answer to an enlarged need for accurate hydrological modelling of a land management impacts, water quality or climatic change assessments. Although lumped conceptual rainfall-runoff models claim to incorporate in their structures most of the process of the hydrological cycle, they do not provide a sound scientific basis for analysing the above mentioned modelling problems. Many parameters are required if all the processes involved are to be represented. However, most of these parameters cannot be related successfully to physical catchment characteristics and must be estimated by calibration using observed hydrographs. Therefore most models show serious drawbacks in parameter identification because of their conceptual structure and the data used for calibration.

1.1. Variable Contributing Area Concept

Runoff may occur in a uniform basin in at least four major ways,

a) Rainfall intensity exceeds infiltration or storage capacity resulting in overland flow all over the basin. This is the classical version of Horton's model and is thought to have considerable relevance in areas of low vegetation cover and high rainfall intensity. However, in humid temperate areas with a vegetation cover, the measured infiltration capacities of soils are generally high in comparison with normal rainfall intensities (Krikby, 1969; Freeze, 1972). In this case the Horton model of basin response is not applicable.

b) Rainfall intensity exceeds infiltration or storage capacity on a variable area of near-saturated soils. This is the basis of Betson's (1964) partial area conceptual model in which it is recognised that the spatially variable nature of infiltration capacities and differences in moisture status at the soil surface, caused by downslope flow of water, will result in some parts of the basin being far more likely to produce infiltration excess overland flow than others. Engman & Rogowski (1974) have produced a relatively simple physically based model founded on this concept.

c) Rain falling on stream channels and completely saturated soils. Where the latter are adjacent to stream channels this source of overland flow contributes directly to the storm hydrograph (Dunne & Black, 1974). The zone of soil saturation may extend completely from bedrock or may build up above a relatively impermeable layer within the soil.

d) Downslope lateral flow of saturated or unsaturated soil water. Most of this flow will be within the soil, but it may locally exceed the soil storage capacity and return to flow over the surface at much higher velocities ('return flow', Musgrave & Holton, 1964; Dunne & Black, 1970). Subsurface flow velocities are commonly too slow to contribute appreciably to the peak of the storm hydrographs although in volume terms subsurface flow may dominate the overall response of the basin in providing the hydrograph tail and low flows (Knisel, 1973).

In small humid temperate basins mechanisms (b) and (c) appears to be the critical sources of storm flow, with subsurface flow making a highly significant contribution in setting up the soil water conditions prior to further storm rainfall. These processes and their characteristics are thought to explain the observed nonlinearity of runoff in response to rainfall, and any simple physically based hydrograph model must reflect this general conceptual knowledge of the mechanisms involved (Beven & kirkby, 1979).

The TOPMODEL (Beven 1986 a) is a variable contributing area conceptual model in which the predominant factors determining the formation of runoff are represented by the topography of the basin and a negative exponential law linking the transmissivity of the soil with the vertical distance from the ground level. In this model the total flow is calculated as the sum of two terms: surface runoff and flow in the saturated zone. The surface runoff, in the most recent versions of the model, is in turn the sum of two components, the first generated by infiltration excess and the second, referring to a variable contributing area, by saturation excess. Though a conceptual model, i.e. one in which the physical reality is represented in a simplified manner, the TOPMODEL is frequently described as being 'physically based', in the sense that its parameters can be measured directly in situ (Beven and Kirkby, 1979). This definition is somewhat optimistic, in view of the doubts and uncertainties encountered even in defining the parameters of the 'physically based models', as already mentioned.

TOPMODEL performs what is called an 'upward search for conceptualisation' from the soil column level to the catchment scale. Basin parameters are related to point estimates. The spatial variability of both soil water content and lateral drainage is related to that of soil and topographic characteristics by means of simple but meaningful assumptions. The model is also attractive because of its structural simplicity and parsimonious parameterisation. The TOPMODEL is one of the few conceptual models that accounts explicitly for the saturation excess overland flow mechanism and integrates the variable contributing area concept, both of which are essential to model the catchment accurately.

In the present study, The TOPMODEL has been applied to Malaprabha basin. Also, this study aims to study the influence of the parameters of TOPMODEL in simulating the runoff of the basin.

2.0. DESCRIPTION OF STUDY AREA

The river Malaprabha originates from Kanakumbi in the Western Ghat at an altitude of about 793 m and 16 km west of Jamboti in the Belgaum district of Karnataka state. The river flows first in on easterly and then in northerly direction and joins Krishna river at an elevation of about 488m and about 300 km from its source. The total catchment area of Malaprabha basin is 11,549 sq.kms. In the present study, the basin upstream of Khanapur has been considered with an catchment area of 520 sq.km. It is principal source of supply from Ghat section of the basin.

Location

The Malaprabha river lies in the western part of the Krishna basin. It extends between 64° 20' and 74° 30' E longitude and 15° 20' and 15° 40' N latitude along the border of Karnataka and Maharastra state.

Hydrometeorological Network

There are five raingauges stations and two hydrometeorological station in Malaprabha basin (Fig.1.). The river is being gauged at Khanapur.

Geology : Geologically the Malaprabha basin comprises of two main geological formations, viz, 1) tertiary basalts and 2) sedimentary formations of Pre-Cambrian age. The major part of (96% of the catchment) the the catchment is covered by the tertiary basaltic rock, whereas, the sedimentary rocks have confined to the south eastern part of the catchment.

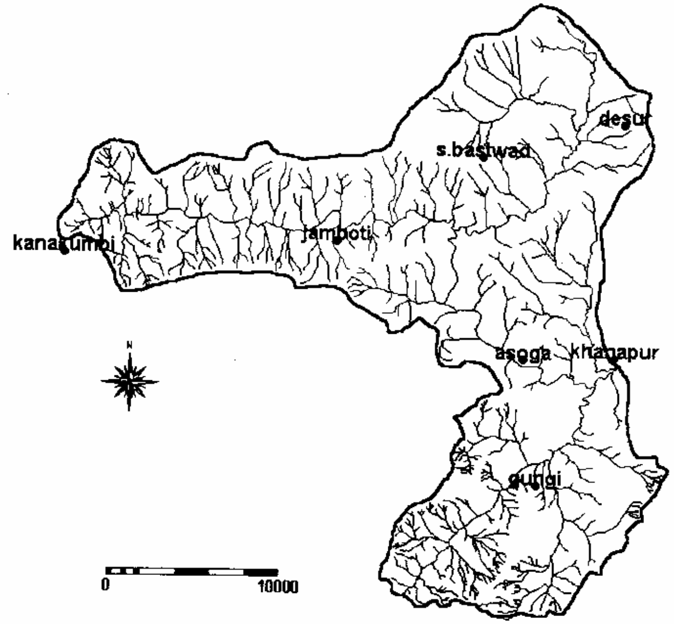


Fig. 1 Malaprabha at Khanapur showing drainage and raingauge stations

Soil : The depth of the soil in the Malaprabha catchment varies from 0.5 m to 10 mts. There are only two different type of soil found in the catchment. They are 1) red loamy and 2) medium black soils.

Land use Pattern : The vegetation cover over the study area is a complex one. The western and south-west part of the Malaprabha catchment is covered by the dry deciduous forest which accounts for about 62.65% of the total catchment area. Whereas the northern part of the catchment is being used for the agricultural (16.85% of the total area) purpose. The sloppy area of the catchment is covered by the shrubs (19.3% of the catchment area) and the rest is unused land.

Climatic condition over the basin

The Malaprabha basin has four distinct season in the year, such as cold weather, the hot weather and the south-west monsoon and post monsoon. The Malaprabha basin is mainly influenced by the south-west monsoon which normally sets on the mid June and recedes by the end of the September. This four month rainfall accounts for about 90.4% (Table 1) of the annual rainfall. Another 9.51% of the annual rainfall is spread over the rest of the year. The average rainfall in the catchment is about 2259 mm. This clearly indicates that, the generation of the runoff from the catchment is restricted to monsoon period with generating capacity of land to stream is 6 cumec/sq.km/day and flow will be there in the stream upto December. The temperature in the basin varies between 19.2° C to 29.5° C. The mean evaporation in the catchment is 1496.9 mm. Normally the climate over the basin is humid.

Table.1.Monthly budget of rainfall in Malaprabha basin.

Month	Rainfall in mm	Percentage of annual rainfall
Jan	0.646	0.02
Feb	0.995	0.04
March	3.46	0.15
April	20.26	0.89
May	48.08	2.12
June	496.82	21.98
July	790.4	34.97
Aug	586.6	25.95
Sept	171.67	7.59
Oct	107.6	4.76
Nov	31.71	1.40
Dec	1.531	0.06

3.0. DESCRIPTION OF TOPMODEL

3.1. Water Balance Component

Surface Runoff from Saturated Excess

In the TOPMODEL (Fig.2) 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 ;

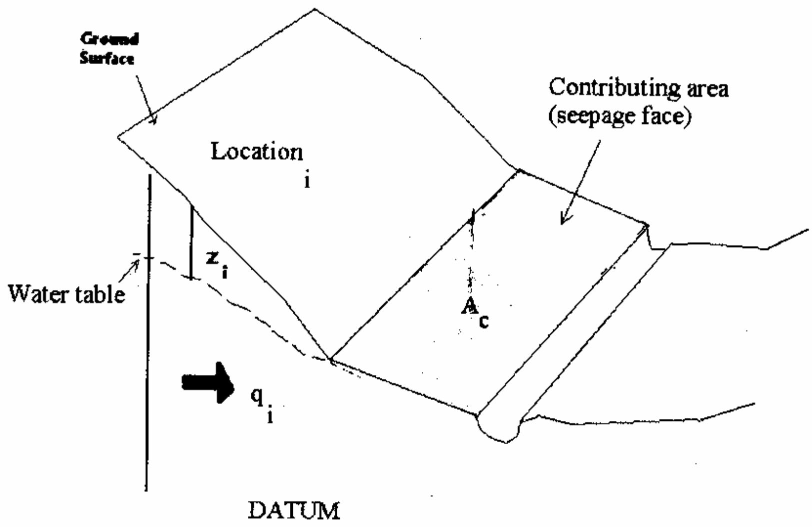


Fig. 2 Schematic representation of a valley and the formation of runoff according to the TOPMODEL. Ac, contributing area to surface runoff; qi, interflow corresponding to an area drained per unit contour length.

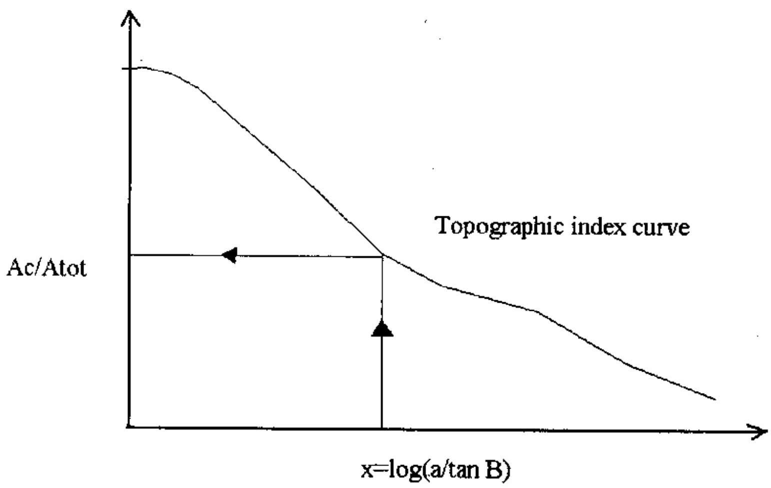


Fig. 3 Topographic index curve. Ac/Atot represents the fraction of the basin in saturated condition for a given topographic index value x=log(a/tan B).

$T_i(z_i)$ = 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 :

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

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;

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

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:

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

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

$$z_i = \frac{1}{f} Ln[\frac{a_i R}{T_0 \tan\beta_i}] \tag{6}$$

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

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

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 ;

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

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

lastly

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

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

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

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

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

Thus, eq (9) becomes:

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

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.3). 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'.

Surface runoff from infiltration excess

The infiltration excess computation is based on the Philip equation;

$$g = cK_0 + \frac{1}{2} S t^{-\frac{1}{2}} \qquad (13)$$

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.

3.2. 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' = Recharge rate of the saturated zone from the unsaturated zone over the time interval t, t+Δt (the symbol R instead of Q_V' has previously been used in the case of time-constant and space-uniform recharge);

Q_B' = 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' can be defined analytically as:

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

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

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

Since:

$\int_L a.dL$ = A (total area of the basin) then:

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

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):

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

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 $z_i \geq 0$.

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'.

$$Q_V^{1} = 0$$

$$Q_B^1 = Q_{ob}^1$$

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

$$Q_B^1 = Q_0 \exp\left(-f \overline{z}^1\right)$$

and therefore the initial state is

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

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

3.3. 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 $\tan\beta$ 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,

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

The amount of area A that contributes in each downstream direction I 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.

4.0. Application of TOPMODEL to Malaprabha Catchment

4.1. Deriving the Topographic Index and DEM analysis

An analysis of catchment topography is required in order to derive the a/tanβ distribution. In order to obtain discrete values of a/tanβ some sampling of topography is implied. The topographic index is estimates based on map information, of local slope angles, upslope contributing areas and cumulative areas. Beven & Kirkby (1979) outlined a computerised technique due 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 basin of dominant flow paths (inferred from lines of greatest slope). Calculation of a/tanβ was carried out for the downslope edge of each element. The Ln(a/tanβ) maps provide information which can be used to characterise catchment hydrological and hydrochemical behavior. The maps can be used to identify the source areas; these are potentially important in the control of the characteristics of the streams and sediment transport. The maps are also valuable in assessing the likely impacts of land-use changes and of herbicide or liming application. In addition, the maps can be used in combination with soil data, if available; Beven (1986) has shown that spatial variation in the transmissivity, T_0, can be taken into account in a combined topographic-soil index, Ln(a/ T_0 tanβ).

A DEM was produced for assessment of the Ln(a/ tanβ) index. This map was generated at 300m grid size by interpolation of the contours taken at a scale of 1:50,000 from Survey of India maps. The index values were calculated using a multi directional routing approach. The area, 'a', used in this index represents the area of catchment which drains through a given grid square. These areas are accumulated down the catchment until they reached the outlet.

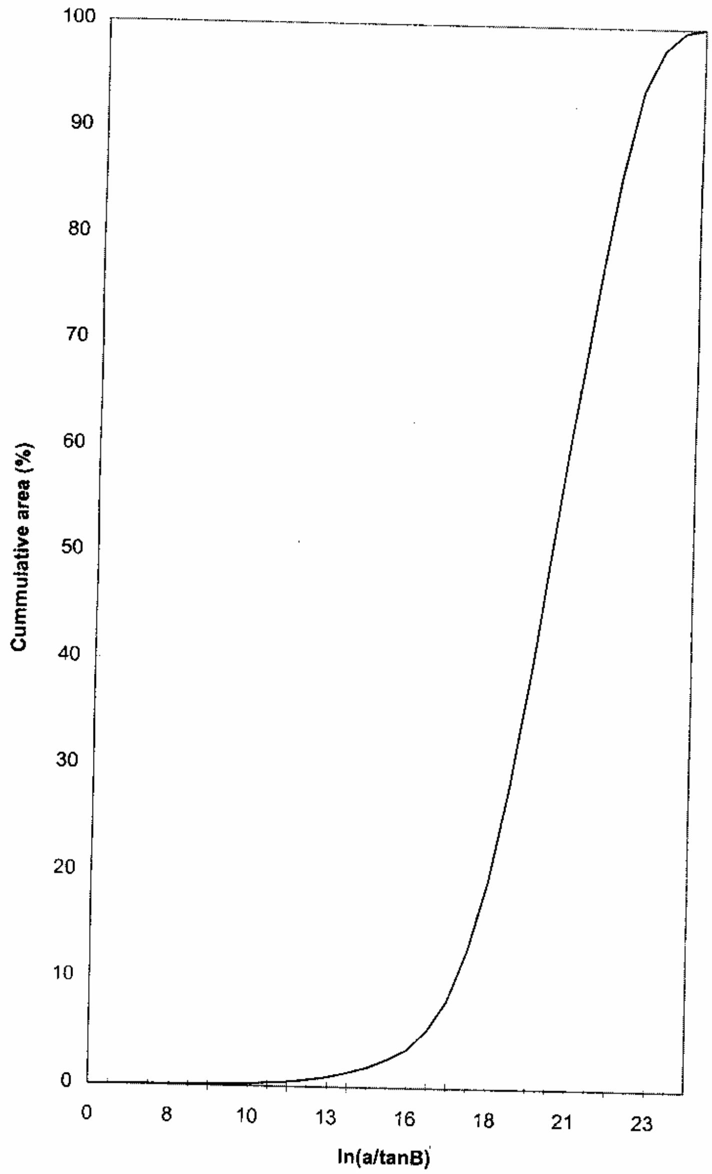
The Fig. 4. & Fig. 5. shows the cumulative frequency distribution of topographic index Ln(a/ tanβ) and spatial distribution of topographic index in the cathcment respectively. The distribution of these contributing areas well spread and nearly all the high index areas are located near the stream. These high index band widens towards the catchment outlet. In general, the index map correspond well to the catchment wetness. The catchment contains the dry rock areas on parts of the lower slopes; these agree well with the low Ln(a/ tanβ) values. Similarly, many of the wetter parts of the upper slopes show higher index values.

4.2. Model Calibration and Validation

The parameter values introduced in the TOPMODEL concepts have been given names 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.

It has been argued in recent years that the realism even of the most complex 'physics-based' hydrological models may be illusory (Beven, 1989). Once 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 within the physically agreeable limits. One consequence of these arguments is that the idea of a set of 'optimum' parameter values should be rejected in favor of a concept of equifinality of parameter sets (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. In fact, certain TOPMODEL, parameters may, in principle, be calibrated on the basis of field

Fig. 4. Cummulative frequency distribution of topographic index.



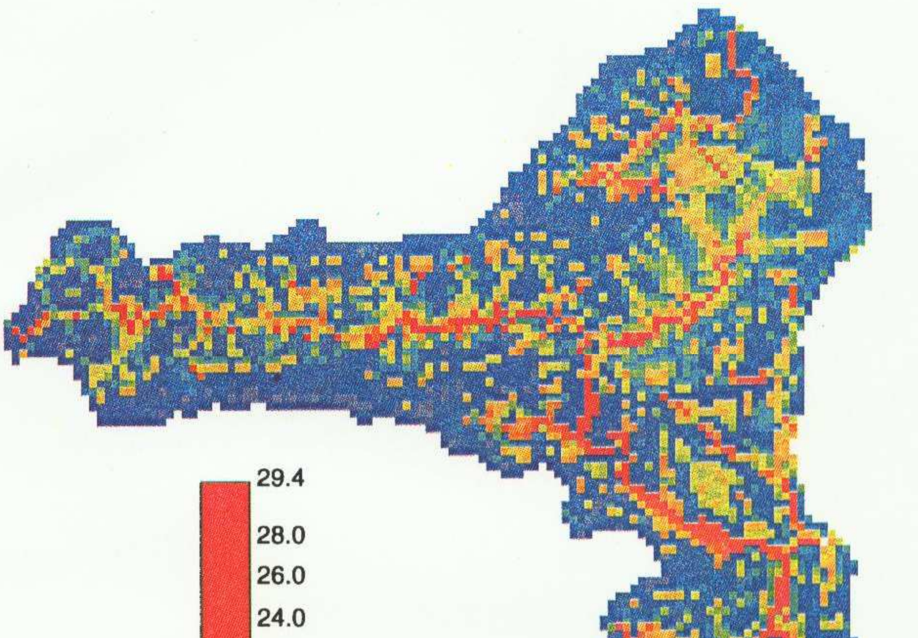


Fig. 5 Ln (a/tan B) distribution map of Malaprabha

measurements and some applications have been made using only measured and estimated values (Beven & Krikby, 1973; 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_{max}, and in larger catchment a channel routing velocity, v. 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 parameters it reinforces the idea that to simulate hydrological responses at the catchment scale.

The study made use of the rainfall, runoff and evaporation data at daily steps for a duration of 4 years which were available for Malaprabha catchment were used to calibrate the model. The data were divided into two groups, i.e., first two years of the data is used for the calibration and the remaining two years data was used for the model validation.

In the calibration run, the automatic optimisation procedure was implemented using the Rosenbroke algorithm. In the process of optimisation, the parameters were allowed to fluctuate within the physically acceptable range. The adjusted maximum transmissivity is 2.28 $m^2 h^{-1}$. This value do not seem unrealistic, owing to the presence fine textured of the soil (the major portion of the catchment is covered by red-loamy soil) in the catchment. The decay rate m, is 0.193 m, Where as the SR_{max} is 0.2612 m and the mean saturated deficit of the catchment T_d is 0.100 h. The model efficiency during the calibration is 0.8746. The Fig.6. shows the simulated and observed hydrographs for calibration period.

The validation was carried out for the period different than the one used for calibration. The optimised parameters were used to run the model. The efficiency of the model in the validation period was 0.7952. The observed and simulated hydrographs for the validation period are presented in Fig. 7.

Fig. 6 Observed and simulated discharge (1987-88) using TOPMODEL for Malaprabha catcment (Calibration)

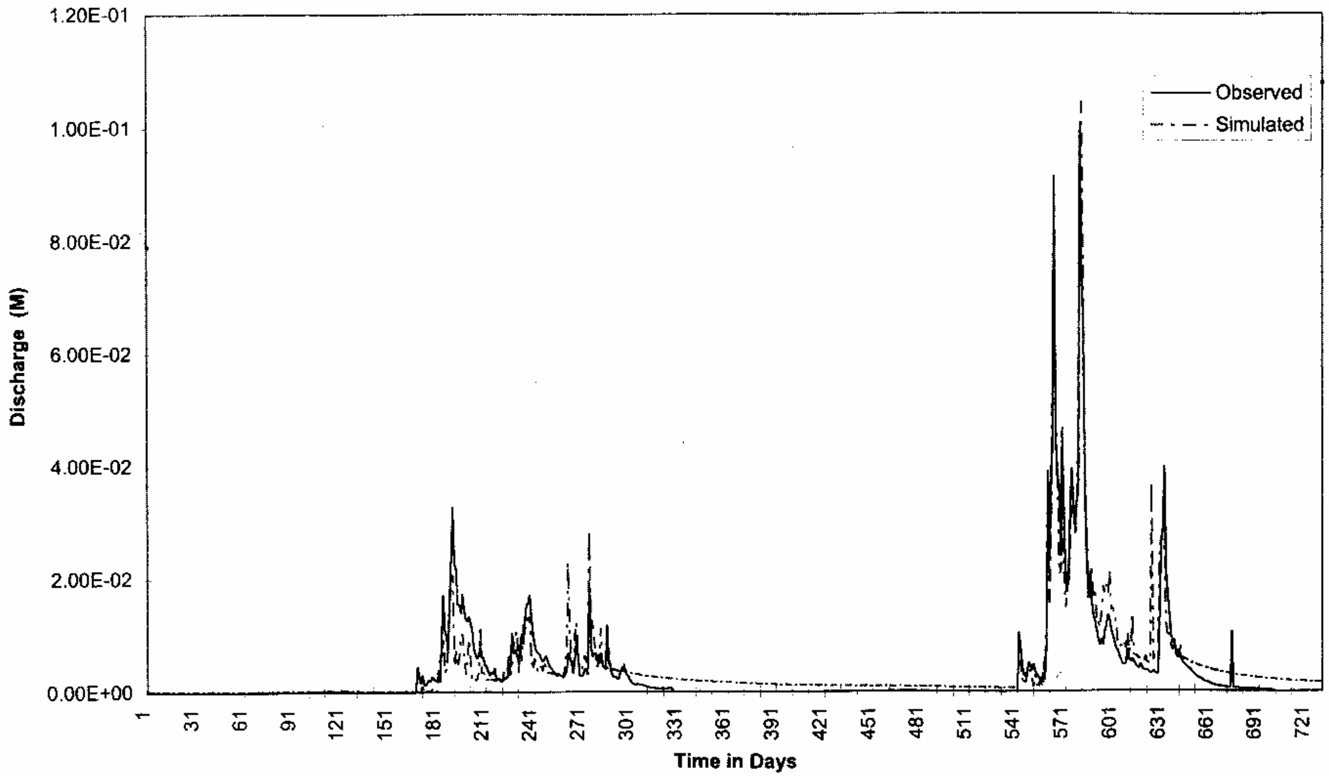
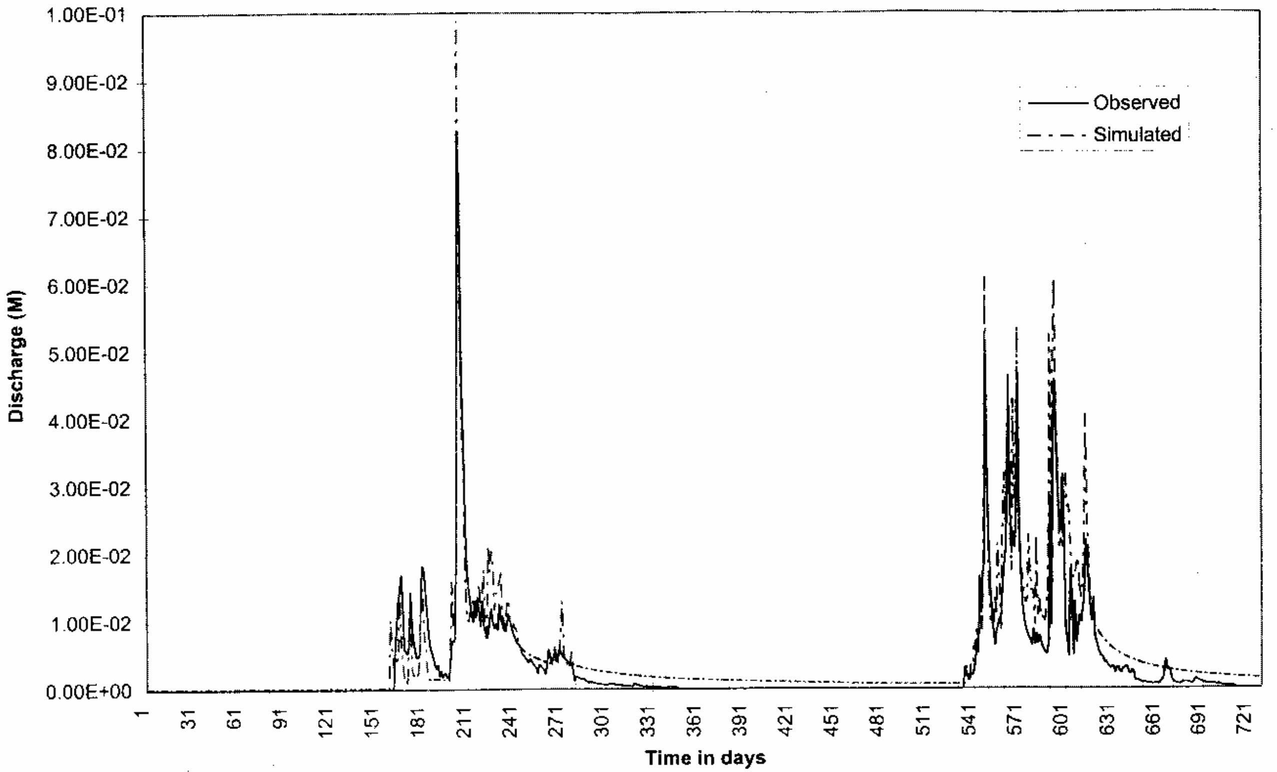


Fig. 7 Observed and simulated discharge (1989-90) using TOPMODEL for Malaprabha catchment (Validation)



5.0. RESULTS AND DISCUSSION

The comparison between the observed and simulated hydrographs for the catchment are presented in Fig. 6. In addition to a qualitative appraisal of the model's performance, the percentage efficiency (E) was used to provide a measure of the goodness-of-fit, where

$$E = 1 - \sum\left(Q_{obs} - Q_{sim}\right)^2 \Big/ \sum\left(Q_{obs} - \overline{Q}_{obs}\right)^2$$

The assumption that discharge results from a combination of quick response runoff generated by a dynamic contributing area is appears to be appropriate from the hydrographs of observed and simulated in the calibration period (Fig. 6.) for Malaprabha catchment. The success of the model in coping with a variety of hydrological conditions (dry and wet conditions, as the catchment is dry during the Feb to June) has been consistent, the efficiency obtained for this run is 0.8746. Further it is noticed from the hydrographs during the later part of the monsoon that the contribution from the groundwater sources to the hydrograph is not evident. This is because of the fact that the decay factor, m, and the transmissivity factor T_0 values are very low (as reported by the studies carried out for dry catchments elsewhere, Durand, et.al., 1992 and Rabson, et.al., 1993), hence the low infiltration into the groundwater regime. Also, it could be because of the fact that, the depth of soil varies from 0.5 m to 10 m and these soil profiles have been underlined by the massive hard rock strata with less permeability. The effect of lower values of m and T_0 is evident even in the validation period. Inspite of the model efficiency of 0.7952, it is observed that the contribution from the delayed flow (ground water contribution) is very less. This gives rise the doubt that the parameters like 'm' and 'T_0 ' may vary from season to season and consequently the use of the different values for different seasons (as the catchment is dry and wet) as seen in the hydrographs (Fig. 6.). Other wise the model performs quite significantly in predicting the magnitude and the timing of the peak flows and the volume of the total runoff.

It is also observed during the calibration run that the influence of the parameter like T_d (the unsaturated zone time delay per unit storage deficit in hrs) and SR_0 (initial value of root zone deficit, mts) on runoff simulation is negligible. This could be because of the fact that, the average catchment slope is 1.898 m/km. Also, it is mentioned that the major portion of the catchment is covered by massive hard rock strata with negligible permeability. This causes the less infiltration and more lateral flow, hence the time delay for unsaturated zone storage deficit in attaining the saturation. Also, the major part of the catchment is covered by the deciduous forest with the large moist holding capacity and more evaporation (the average annual evaporation is 1400mm). This leaves behind the less initial root zone deficit at the start of the next rainy season. Whereas the parameter Sr_{max} (the root zone available water capacity m) is active (As the previous study indicates that the soil moisture storage capacity of the catchment is 300 mm). And its effect on simulation of runoff is evident and it is true (the optimised value is 261 mm), as it is mentioned that the major part of the catchment is covered by the forest.

6.0. SUGGESTIONS

From the above discussion it is noted that

1. The TOPMODEL can be applied successfully for the drier catchment to simulate flows quite accurately (as the efficiency of the model is 0.87 & 0.79 respectively in calibration and validation run).

2. The study inplies that, there is a need for an indepth study to test and understand the use of different values for the sensitive paramenters like 'm' and 'T_0' for dry and wet seasons. As the catchment is experienceing a dry and wet season.

3. It is suggested that the model is to be tested for the catchment using more number of years of data before concluding that the model can be used for any hydrological purposes

4. Further, it is suggested to study the effect of land use change and morphology on the simulation of flows (as the model uses the topographic index in estimating the infiltration excess overland flow). Also the effect of DEM grid size on the simulation of the runoff.

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