HYDROLOGICAL PROCESSES, VOL. 7, 425–448 (1993)

SPATIAL AND TEMPORAL PREDICTIONS OF SOIL MOISTURE DYNAMICS, RUNOFF, VARIABLE SOURCE AREAS AND EVAPOTRANSPIRATION FOR PLYNLIMON, MID-WALES

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ABSTRACT

For many years hydrologists have tried to build physically realistic models which are still simple enough to be fitted to a range of observations made in the field. This is an ongoing process which will become even more difficult as the quality and variety of field and remotely sensed data improves. Hence models must be able to predict soil moisture patterns in time and in space as well as the outflow hydrograph. The model presented here (TOPMODEL) aims to predict the nature of variable source areas in a way that reflects their dynamics over space and time. All component processes are described and shown in operation. As TOPMODEL and similar models have a growing popularity, this paper can be seen as a demonstration of the model's predictive capabilities. The model is applied to the catchments of Plynlimon, mid-Wales for 1984, 1985 and 1986 data sets. The model has been thoroughly tested and cross-validated against independent data sets for different time periods, for a separate catchment, for internal gauges and for wet and dry periods. The resulting predicted soil moisture patterns show a small, semi-permanent variable source area that has the ability during large storms to expand dynamically over short time periods. Spatial predictions of evapotranspiration are also shown which reflect the influence of soil moisture patterns on this process. The weakest component of the model is the representation of root zone evaporation and how this pre-sets the antecedent condition of the catchment during long dry periods.

KEY WORDS Soil moisture dynamics Runoff Evapotranspiration Hydrological models Plynlimon, Mid-Wales

INTRODUCTION

TOPMODEL is a simple but physical hydrological model that aims to represent the effects of catchment heterogeneity and, particularly, topography on the dynamics of hydrological response. The model was first described by Beven and Kirkby (1979) and has become an increasingly popular modelling approach (for example, in Beven *et al.*, 1984; Hornberger *et al.*, 1985; Beven, 1986; Wood *et al.*, 1988; Wolock *et al.*, 1989; Famiglietti and Wood, 1990; Quinn *et al.*, 1990; Robson *et al.*, 1992). Today there are a number of projects using TOPMODEL concepts across Europe and in North America, with similar work being carried out Australia (O'Loughlin, 1981; 1986; Moore *et al.*, 1986). The topics of interest to these scientists include catchment chemistry, pollution, ecology and geomorphology as well as hydrology. This paper is intended to provide a demonstration of the modelling capabilities of TOPMODEL. The paper will show how the model is parametrically simple, how the model can predict long flow records and how the model makes physical predictions in space and time.

At the heart of TOPMODEL are two simple distribution functions that attempt to reflect the most dominant components of hillslope flow. The first is a probability density function that describes the likely saturation potential at any point on a hillslope depending on its relative topographic position. It was first stated in work by Hewlett and Hibbert (1963) and by Hewlett and Troendle (1975) that a study of contour maps should lead to a criteria for mapping the likely position of variable source contributing areas. In the past spatial terrain analysis was extremely tedious. However, with the availability of digital terrain maps

0885–6087/93/040425–24$17.00
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Received 8 June 1992
Accepted 16 September 1992

(DTMs) this process has been largely automated (see, for example, Quinn *et al.*, 1991). Thus by digital terrain analysis (DTA) a topographic index for each DTM grid cell can be calculated *a priori* for any catchment. The cells are then used to build a single topographic distribution function to represent the range and occurrence of the calculated index. The second function is a depth–transmissivity relationship across the soil profile which can be used in essence to approximate to a storage–runoff relationship for a catchment. The storage–runoff relationship can be simply described by a study of recession hydrographs. Hence a simple parameterization for lateral flow in the soil, related to the moisture status over time, can be made. The topographic index and the soils component can be joined to make a combined soil–topographic distribution function.

With these two simple distribution functions the model can reflect our perceptual model of the dominant hillslope hydrological forces. Principally the model tries to do the following:

(1) To trace the position and evolution of variable source contributing areas (Dunne and Black, 1970), which dynamically expand and contract as flow is contributed to them by subsurface pathways.
(2) To provide a movement of 'old' water (Sklash and Farvolden, 1979) already in the variable source area to the stream by a mechanism often referred to as a 'piston-like' displacement.
(3) To reflect the role of macroporosity and flow bypass mechanisms that can increase the effective transmissivity of the soil (Beven and Germann, 1982).
(4) To allow for increases in discharge due to the build up of saturation wedges in the soil (Whipkey, 1965; Weymann, 1970).
(5) To represent high proportions of subsurface contributions to storm runoff by allowing subsurface stormflow (Hursh, 1936).

Infiltration excess can be modelled but for this application has been ignored. Infiltration excess is assumed to be a local process that does not directly contribute to channel flow. Any infiltration excess volume is assumed to reinfiltrate locally.

TOPMODEL tries to capitalize on the natural hydrological–geomorphological links shown by catchment form. It is also expected that the soils and the vegetation are strongly influenced by their hydrological–geomorphological niche. Thus analysis of one measurable data set (in this instance the topography) can be used to indicate (via an index) the operation of another process (that is, the hydrology). Emphasis is placed on the spatial representation of a process as opposed to complex physical descriptions at a point (which is a point raised by Moore *et al.*, 1991). The model gives internal state predictions as well as flow prediction and thus can be thoroughly investigated using all available observations. The model automatically controls antecedent conditions in the catchment before storm events begin.

The model aims to capture the essence of reality while only using three parameters to represent a variety of hydrological phenomena. Low numbers of parameters minimize optimization problems and are likely to make the final optimized values more physically meaningful (Sorooshian, 1991).

TOPMODEL should be thought of as a set of conceptual tools for hydrological analysis rather than as a fixed model structure. Its most important capability, as will be shown later, is the mapping of predictions back into the catchment space, so that the nature of the responses can be compared with observations and impressions of how a catchment is responding. It will be shown that the modelling concept appears to be appropriate for catchments in mid-Wales but may not work so well in different climates and geological conditions. Consistent with the model theory, the model is likely to work best where soils are thin, where there is moderate to steep topography and where runoff coefficients are moderate to high.

The catchments used in this study are the Institute of Hydrology research catchments at Plynlimon in upland Central Wales (Newson, 1976; Newson and Harrison, 1978; Gilman and Newson, 1980). The river Wye catchment ($10{\cdot}55\,km^2$), a grassy catchment, was the primary area of study while the river Severn ($8{\cdot}7\,km^2$), a coniferous forested catchment, was used to independently validate the model. These catchments were chosen because a number of intensive field studies had been carried out in that area, but primarily because of the availability of a continuous hydrological data record. Data were first downloaded from the Institute of Hydrology Oracle database. Available data included hourly flow and rainfall data and

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a daily estimate of Penman–Monteith potential evaporation. Finally a 50 m resolution DTM was constructed for the catchment using the original 1 : 10 000 Hunting Survey maps as the source.

ESSENTIAL TOPMODEL

A simple description of the TOPMODEL concept is given in the following but fully worked descriptions are given in Beven (1986) and Quinn *et al.* (1990).

Firstly, the topographic index $\ln(a/\tan\beta)$ is calculated for each gridded cell of the digital terrain map (Quinn *et al.*, 1991), where a is the accumulated upslope area that has drained into that cell. This is proportional to the volume of water likely to be moving through that cell under steady state flow conditions. β is the local slope angle which is used to approximate the hydraulic gradient acting on the cell, that is, a measure of the capability to move the volume of water in that cell.

Figure 1A is used to show the link between topography, the index and the soil moisture status. The figure shows that the gravity drained soil moisture deficit (which should be related to the water-table position or depth to capillary fringe) decreases with increasing index value for the typical case where $\ln(a/\tan\beta)$ increases downslope. The topographic index is naturally sensitive to flow convergence and divergence as well as breaks in slope. The resolution of the DTM must thus be small enough to depict those features which influence flow paths (this is discussed in Quinn *et al.*, 1991). A resolution of 50 m or smaller is strongly recommended for DTA, depending on the scale of the hillslope.

For each timestep and for each value of the index a prediction is made of the local soil moisture deficit.

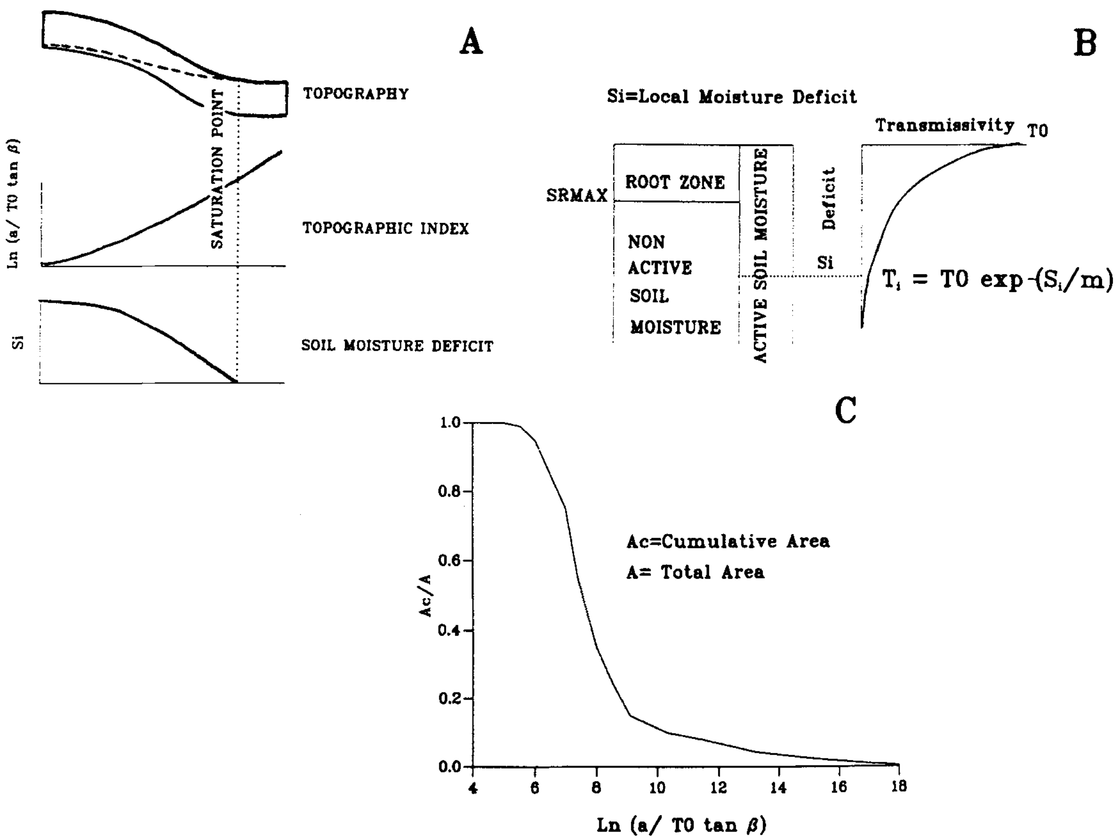

Figure 1. Essential TOPMODEL concepts. (A) Relationship of topography, topographic index and soil moisture deficit for a single typical hillslope (dotted line is the water-table position). (B) Soil profile conceptualization showing the partitioning of moisture zones and the relationship of transmissivity with depth. (C) Compilation of map 1C into a single discrete distribution function

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Typically between 15 and 30 subdivisions are made in the combined index distribution function to cover the range of values across the distribution. All values of the index falling in the same incremental step are assumed to respond in a hydrologically similar way.

Figure 1B shows the conceptualization of what happens in the soil within TOPMODEL. Three calibration parameters are required for the model.

(1) m is the factor by which the rate of the exponential decrease of lateral transmissivity with increasing soil moisture deficit is controlled. An approximate value can be derived from an analysis of recession hydrograph form. Recession flow after overland flow has ceased is in essence the integrated response of the catchment storage–transmissivity regime.

(2) T_0 is the value of lateral transmissivity when the water-table just cuts the surface. Thus the transmissivity at any point in the catchment (T_i) for any prevailing soil moisture deficit value is:

$$T_i = T_0\,\mathrm{e}^{-Si/m} \tag{1}$$

where Si = local soil moisture deficit at that point.

(3) SRMAX is the maximum storage capacity of the root zone and is used to control the simulation of evaporation.

In this application T_i is combined with the topographic index to give the combined soils/topographic index $\ln(a/T_0 \tan\beta)$ (see Beven, 1986). Figure 2 shows a map of the combined index for the DTM of the river Wye. The single values of the index for each cell are combined to build one probability density function. Figure 1C shows the distribution function used for the river Wye. Only one discrete distribution

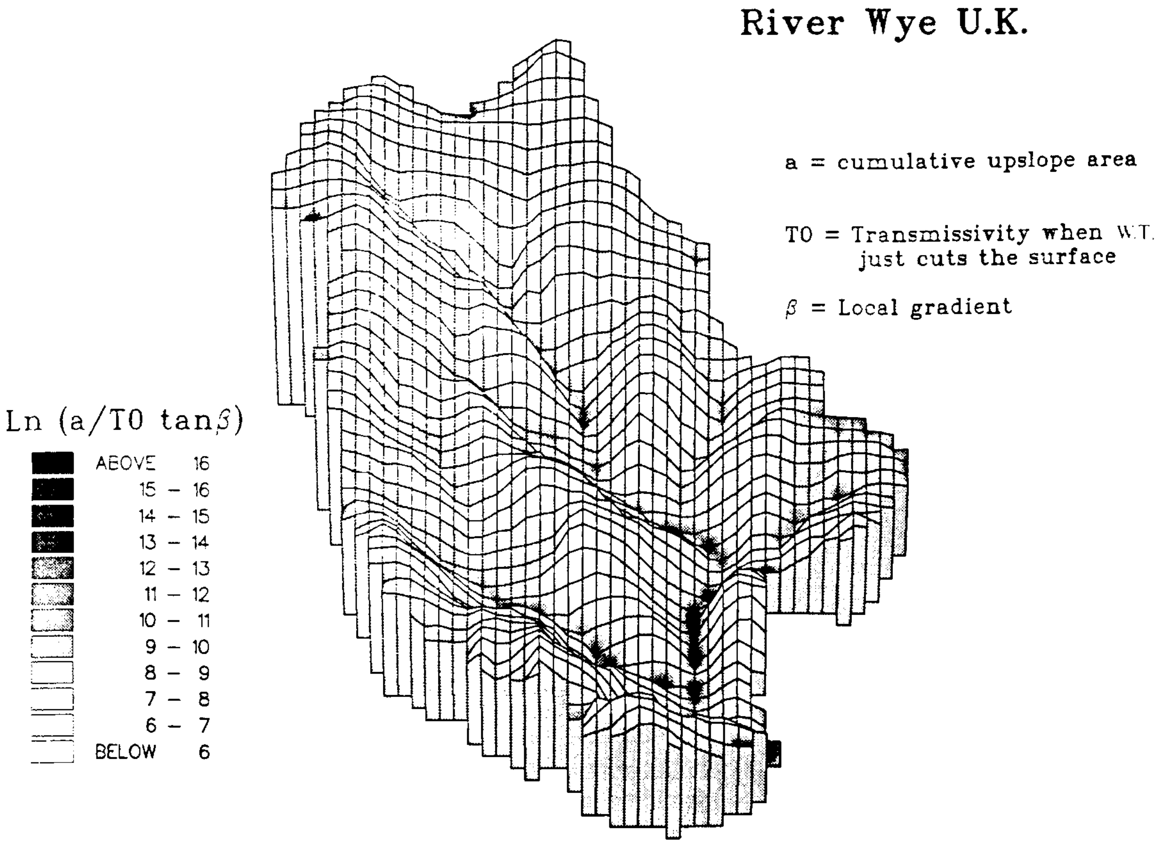

Figure 2. Topographical map for the river Wye with a map of the soil–topographic index superimposed

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has been used here (in contrast to multiple subcatchment simulation used, for example, in earlier studies of Beven *et al.*, 1984). Thus the water balance accounting is lumped at the catchment scale although the distribution function still allows the spatial variability of responses to be reflected in the model. The major advantage of this is in the speed of the calculations (one minute for simulation of one year at a one hour timestep on a Sun Sparc 2 machine).

Figure 1B shows T_0 falling off with depth at a rate controlled by the parameter m so the lateral flow at any point (q_i) is given by

$$q_i = T_0 \tan \beta_i \, \mathrm{e}^{-Si/m} \tag{2}$$

Figure 1B shows a root zone that controls the evaporation from the model. The soil is further subdivided into: (1) an active soil moisture zone to reflect vertical flow of water in the unsaturated zone — this implicitly includes the role of macropores and larger voids and is approximated by a gravity drainage regime; and (2) a non-active soil moisture zone which assumes that water is held in the bulk matrix at field capacity and is unlikely to change in status over long periods of time.

Based on an assumption of a quasi steady state recharge throughout the catchment (Beven *et al.*, 1984; Quinn *et al.*, 1991) the current local soil moisture deficit position can be calculated by determining the deviation of the local combined index from the catchment average of the combined index (γ) in the form

$$S_i = \mathrm{SBAR} - m[\gamma - \ln (a/T_0 \tan \beta)_i] \tag{3}$$

where SBAR = the mean storage deficit for the whole catchment (or subcatchment) and

$$\gamma = \frac{1}{A} \int \ln (a/T_0 \tan \beta)_i \, \mathrm{d}A$$

As it can be seen the relationship of SBAR to S_i is scaled by the parameters T_0 and m. Hence T_0 and m are powerful parameters as they control the rate of change of the soil moisture pattern over time.

High values of the combined index yield a negative S_i value and progressively more grid cells will have negative S_i as the value of SBAR decreases (that is, as the catchment gets wetter). A negative S_i value indicates that the soil is saturated, hence at any timestep the saturated area can be calculated. The saturated area is assumed to act as a zone of return flow and rainfall on this area will generate saturation excess overland flow after any root zone deficit is satisfied.

Areas where the soil moisture deficit is approaching zero can also yield saturation excess overland flow. If the amount of water overflowing from the root zone into the active drainage store during a storm is greater than the available storage, then the excess is assumed to add to the saturation excess overland flow.

Soil moisture accounting is carried out at each timestep to update the average storage deficit in the catchment (SBAR)

$$\mathrm{SBAR}_t = \mathrm{SBAR}_{t-1} + Qb - Qv$$

where Qb = the amount of subsurface flow reaching the channel network for timestep $t-1$ and Qv = the amount of recharge to the water-table from the active drainage store for timestep $t-1$, summed over the catchment.

Lateral transmissivity, or more importantly the range of lateral transmissivities, is difficult to determine in the field, therefore T_0 is calibrated as a lumped catchment parameter. This is at a scale above the representative elementary area (REA) scale as discussed by Wood *et al.* (1988). Thus a detailed pattern of the heterogeneity of T_0 is not needed and a simple distribution of transmissivity will suffice. It was shown by Beven *et al.* (1988) that a good description of the pattern of topography was of greater importance due to its much larger variance at the catchment scale and hence the distribution of T_0 will be approximated by its mean value.

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Evaporation is driven by the calculation of a Penman–Monteith potential evaporation value. This is based on data from the Institute of Hydrology automatic weather stations available in the catchments. A daily average value is altered to an hourly rate assuming a sinusoidal curve distribution centred on 12 noon, commencing at 0600 h and ending at 1800 h. Two components of evaporation are allowed in this implementation of TOPMODEL, both of which are controlled by SRMAX. A reduction from the potential evaporation value occurs depending on the moisture status of the root zone (after Beven *et al.*, 1984). When the root zone store is empty evaporation by this process is also zero

$$\mathrm{Esrz} = \mathrm{Epot} \times \mathrm{SRZ}/\mathrm{SRMAX} \tag{4}$$

where Esrz = the evaporation from the root zone storage, Epot = the potential evaporation (the I.H. estimate) and SRZ = the current moisture status in the root zone.

This is a useful and popular form of soil moisture driven representation of the actual evaporation. Similar forms using soil saturation, field capacity and the wilting point are also often used (for example in the Institute of Hydrology distributed model after Beven *et al.*, 1987). Equation (4) does not, however, allow for the case where water is redistributed to the root zone over time from upslope. This redistribution process is common where variable source areas can remain near the surface for long periods of time (especially in riparian areas). The distribution of vegetation types may reflect shallow water-tables, thus keeping evapotranspiration high. The redistribution processes will maintain evaporation during long interstorm periods and in catchments with hot climates.

Thus a second conceptualization is allowed in TOPMODEL to allow for water redistribution over time. Areas where gravity drainage moisture is moving back into the root zone are determined by Equation (3). Any unexploited potential evaporation [the residual from Equation (4)] is allowed to evaporate this gravity drainage moisture. The upper limit of moisture available to the residual evaporation potential is determined by the values of SRMAX and S_i. The evaporation process is thus spatial, depending on the local moisture status, yet the conceptualization is simple in that it is to be controlled by only one effective parameter (SRMAX). The model operates on a reasonable physical basis but avoids the problem of areal averaging of the Penman–Monteith value. No explicit vegetation model is included, but the model implicitly assumes the root zone is where plants access available moisture and continue to use moisture until SRMAX is exceeded. Note that in this application interception storage is assumed to be included in SRMAX and no call to an explicit calculation of interception is made. This reduces the number of parameters in the model.

Finally, in the current model, despite the prediction of certain cells as being saturated, evaporation is continued in the root zone and a deficit will build up during interstorm periods. This is to reflect a surface downwards drying effect. If this process is not modelled then storm predictions become too peaky after long interstorm periods.

Soil moisture values can be mapped back into space by using the original soil–topographic index map (Figure 2). Equation (3) only requires an SBAR value for the current timestep, the parameter m and the value of the local soil–topographic index to calculate the local value of S_i. Hence, spatial and temporal analysis of the model predictions can be carried out to produce: (1) maps of the spatial soil moisture extremes; (2) the cumulative frequency of saturation mapped over time to reflect the dynamic change in the variable source areas; (3) the fluctuation of soil moisture traced across a soil profile over time; and (4) spatial evaporation maps.

The drawback of these predictions is that they are still in reality lumped predictions as the grid cells are 50 × 50 m in size. As most internal state data are point values then local heterogeneity may still dominate the observations. The strength of the predictions is that the overall pattern of results can be matched. Integration of field measurements builds up our hydrological perceptions and it is against these perceptions that TOPMODEL can be evaluated. It is hoped that similar pattern analysis of hydrological phenomena will occur with the improvements in remote sensing, mapping of saturated areas and other field techniques.

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OPTIMIZATION AND MODEL EFFICIENCY

Optimization is carried out using the same code as that used by Hornberger *et al.* (1985). The procedure is a Rosenbrock hillclimbing method (Rosenbrock, 1960). Hillclimbing leading to the optimum solution is achieved by re-running the model with differing parameter values and minimizing an optimization function for each run. In this study a simple sum of squared error method was used

$$\mathrm{OPFUNC} = \sum_{j=1}^{T}(\mathrm{QOBS}_j - \mathrm{QSIM}_j)^2$$

where OPFUNC = value of the optimization function for current run of model, QOBS is an array of all the observed flow values, QSIM is an array of all the simulated flow values and T is the total number of timesteps.

A number of other optimization functions was available but only the sum of squared errors method was used here. Hornberger *et al.* (1985) had concluded that this method was equal to, if not better than, the more complex methods discussed in their paper.

In each run the success of the model is evaluated in terms of its model 'efficiency'. Efficiency is calculated as

$$E = 1 - \left(\frac{\text{variance of the residual errors}}{\text{variance of the observed data}}\right)$$

This value is converted to a percentage figure by multiplying by 100.

Local optimization minima are commonly found within the parameter space. These minima may not locate the global minimum point, hence the full parameter space should be tested to show the true model result. This is achieved by starting optimization at different initial positions that cover the parameter space.

The use of a sensible m value (as derived from recession hydrographs) can reduce optimization difficulties. T_0 in optimization tends to be fitted to produce flow peaks. SRMAX is optimized to fit antecedent conditions before storm events. Hence a wide range of storm conditions are needed to fully test for the most representative calibration parameter values.

CALIBRATION AND VALIDATION RESULTS

The calibration and validation procedure used in this paper are largely influenced by the work of Klemes (1986). Klemes described a thorough set of rules for testing models to catchment data. The Klemes guidelines have been followed and extended for this modelling exercise so as to include: (1) the use of long periods of data for calibration; (2) the use of long periods of data for validation; (3) the use of a second independent catchment for validation of the calibrated parameters; (4) the validation of the model using gauges inside the catchment; and (5) testing of the model on distinctly wet and dry periods, with cross-validation of the parameter data sets.

The model was first optimized on 1985 data for nine months (avoiding snow months as indicated by the Institute of Hydrology database). The resulting modelling efficiency was 85·24% and the three optimized parameters were: $m = 0{\cdot}0093\,\mathrm{m}$, $T_0 = 8{\cdot}2746\,\mathrm{m}^2/\mathrm{h}$ and $\mathrm{SRMAX} = 0{\cdot}0899\,\mathrm{m}$.

Figure 3 represents the hydrographs produced by the model. Figure 3A is the main input to the model, that is, the rainfall each hour. Figure 3B is a comparison of the observed and simulated hydrographs for the complete nine month period. Figure 3C and 3D are two windows onto Figure 3B to show in more detail the dynamics and goodness of fit of the model for individual events. The results have a reasonable quality of fit for most storms and their individual characteristics, especially in the recession portion.

To cross-validate these results other periods of data were used. The 1985 nine month parameter set was rerun on another two nine month periods for the river Wye, for 1986 and 1984. The model was also cross-validated against another nine month period in 1986 for the river Severn. The Severn is the forested

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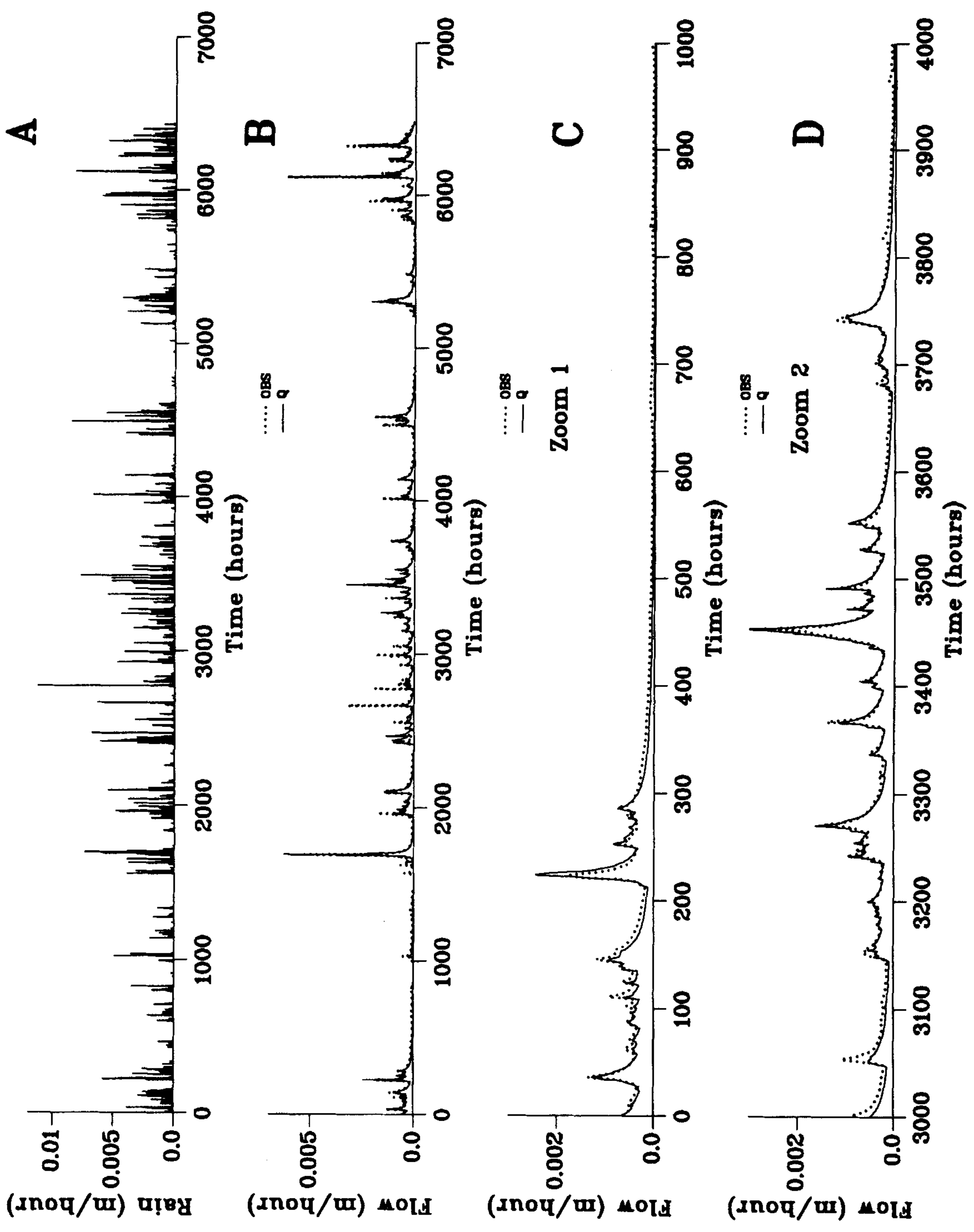

Figure 3. Calibration of TOPMODEL. (A) Rainfall for nine months for the river Wye 1985. (B) Observed and simulated flows for nine months. (C) Zoom-in on Figure 2B (about two months). (D) Second zoom-in on Figure 2B (about two months)

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catchment adjacent to the Wye which has similar topograph and precipitation, but has differing soil, vegetation and evaporation regimes and hence is an interesting test of the model.

Figure 4 shows two of the three validation runs. Some interesting points are raised by the results of the river Severn simulation. Firstly, the rainfall–runoff relation is similar to that of the Wye; that is, the runoff dynamics are similar, despite the change from a grassy to a forested vegetation. Close analysis of a number of storm peaks for the Severn showed the model to overpredict slightly; this is believed to be due to this version of the model not containing an explicit interception store. The work of Robson *et al.* (1992) on the Hafren (a subcatchment of the Severn) showed the inclusion of an interception store in TOPMODEL helps reduce this problem.

Table I shows a full description of the model components and the observed and predicted water balances for the above runs.

The 1984 results were reduced in efficiency by one period of poor storm reproduction seen in August and September of that year and may well be associated with the long dry period of summer 1984.

The most striking observations from Table I are: (1) the dominance of subsurface flow on runoff production; (2) the proportions of the runoff components are similar for all the simulations; (3) actual evaporation is close to the potential evaporation due to the model optimizing large SRMAX values and the overall wetness of the catchments; and (4) Esrz is dominant and Ered provides a much smaller proportion of evaporation.

The model was also calibrated to the river Severn 1986 and to the river Wye 1986 data to cross-validate the values of the parameters. Table II is a summary of these runs.

The calibration to the particular time periods changes the parameter values in only a limited way giving relatively small improvements in model efficiency. Conversely, the change in parameter values causes a relatively small deterioration in cross-validation efficiencies. This interchangeable nature of the parameter values is thought to reflect the hydrological similarity between both the data periods and the catchments.

Within the Wye catchment there are three internal gauges for which flow and catchment rainfall were available. These data were downloaded from the Institute of Hydrology database for the same nine month period in 1985. Each catchment has thus had separate rainfall and flow data and were run using the 1985 Wye catchment calibration parameter set. Table III is a summary of the results.

Figure 5 shows the nature of the results for a 500 hour period in April 1985. It is clear that the Cyff and the Gwy are acting in a similar manner to the complete catchment with some small local variation. The Nant Iago results with low model efficiency (00·71%), show that there is a general water balance problem across that period as well as poor timing for some of the events. The possible cause of such prediction errors could be the effect of extensive old mine workings which exist in the Nant Iago catchment, resulting in a different flow regime. The relatively small size of the catchment means that it has not significantly reduced the overall efficiency for the complete model. Similar problems in reproducing the internal subcatchment responses of the Wye catchment have been reported by Bathurst (1986).

COMPONENT PROCESSES

Accepting that the model is working with reasonable accuracy for the River Wye, an analysis of the component processes predicted by the model was made in both space and time. All the results shown were taken from the 1985 data set using the optimized parameters for this period.

Predicted subsurface stormflow (*Qb* on Figure 6B) can be seen to make up a large proportion of the total flow and clearly responds quickly during storm events. The pattern of change of the saturated area follows the flow as the two processes are directly linked (Figure 6C). The saturated area is predicted as varying from about 5% of the catchment area up to 20% of the catchment area within 10–20 hours (depending on the storm shape). Saturated areas recede with the same temporal form as the flow figures. The fundamental state variable of TOPMODEL, SBAR (the catchment average soil moisture deficit), can be seen to mirror the subsurface flow and the saturated area size (Figure 6D). This is because SBAR controls the magnitude of both these flow components through the non-linear soil–topographic and subsurface drainage function.

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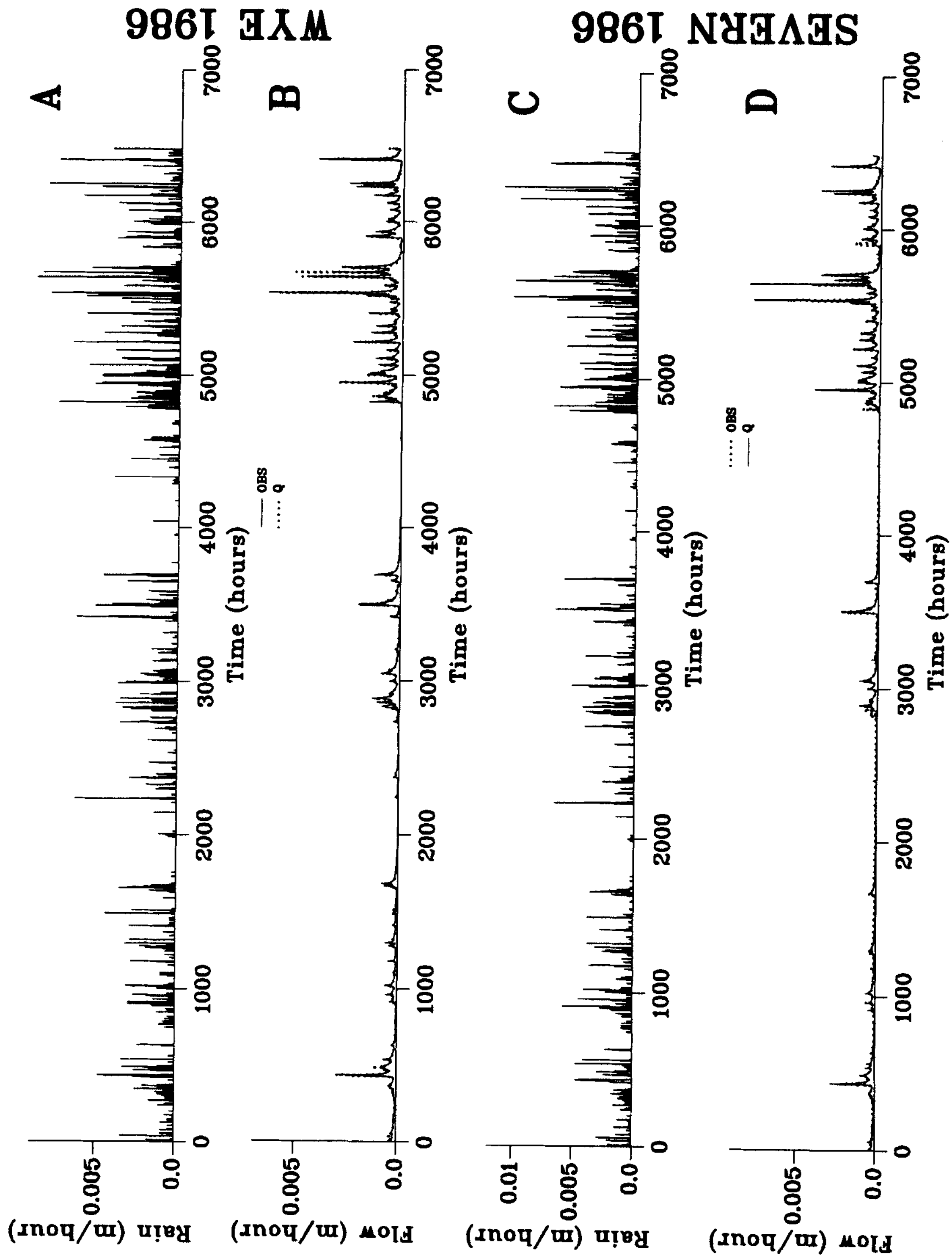

Figure 4. Validation of TOPMODEL using the calibration parameters for the rivers in 1985 (nine months). (A) Rainfall for nine months for the river Wye 1986; (B) Observed and simulated flows for nine months using Figure 3A. (C) Rainfall for nine months for the river Severn 1986. (D) Observed and simulated flows for nine months using Figure 3C

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Table I. Breakdown of all the observed and modelled water balance components for the calibration and validation runs

	Wye calibration	Severn nine months	Wye 1986 nine months	Wye 1984 nine months
Efficiency (%)	85·24	72·05	92·94	53·39
Totals (m)				
Rain	1·877	1·842	1·784	1·375
Observed flow	1·659	1·429	1·539	1·078
Simulated flow	1·445	1·402	1·325	0·986
Potential evaporation	0·449	0·437	0·453	0·417
Esrz*	0·419	0·381	0·404	0·333
Ered†	0·030	0·054	0·049	0·046
Subsurface flow	1·269 (87·8%)	1·201 (87·7%)	1·152 (86·9%)	0·862 (87·4%)
Saturation excess	0·175 (12·2%)	0·200 (14·3%)	0·172 (13.1%)	0·124 (12·6%)

Table II. Calibration and cross-validation of the model using two independent data sets

	1986 Wye	1986 Severn
Calibration result	94·02%	85·68%
Validated on		
1984 Wye	47·48	70·75
1985 Wye	84·28	73·11
1986 Wye	—	79·43
1986 Severn	68·51	—

Table III. Validation of the model using nine months of data for 1985 for the internal gauges of the Wye catchment

Catchment name	Catchment area (km^2)	Model efficiency (%)
Wye (all)	10·55	85·24
Cyff	3·15	91·58
Gwy	3·84	83·04
Nant Iago	1·05	00·71

Figure 7A and 7B are the rainfall average and the calculated Penman–Monteith potential evaporation (produced by the Institute of Hydrology), respectively. Evaporation was distributed into hourly timesteps using a sine curve to reflect the incoming short wave radiation pattern. Figure 7C is the cumulative evaporation generated by the model using the two processes described earlier. Esrz (the evaporation controlled by the root zone soil moisture status) dominates the evaporation; this is because the catchment is generally wet and also the optimized value of SRMAX is large (89·9 mm), hence the root zone is rarely depleted. Esrz should be dominant anyway due to the fact that the saturated area is only a small proportion of the total area at any one time (this is investigated later). Figure 7D is the status of the root zone which has a maximum of 89·9 mm of storage. After the start of a rainfall event the root zone fills to SRMAX and depletes by the potential given in Figure 6B.

Figure 8 shows the spatial characteristics of the predicted variable source areas. As Equation (3) showed, a knowledge of SBAR and the local and average soil–topographic index values means a prediction of S_i (the local soil moisture deficit) can be made anywhere in the catchment (noting that higher SBAR values

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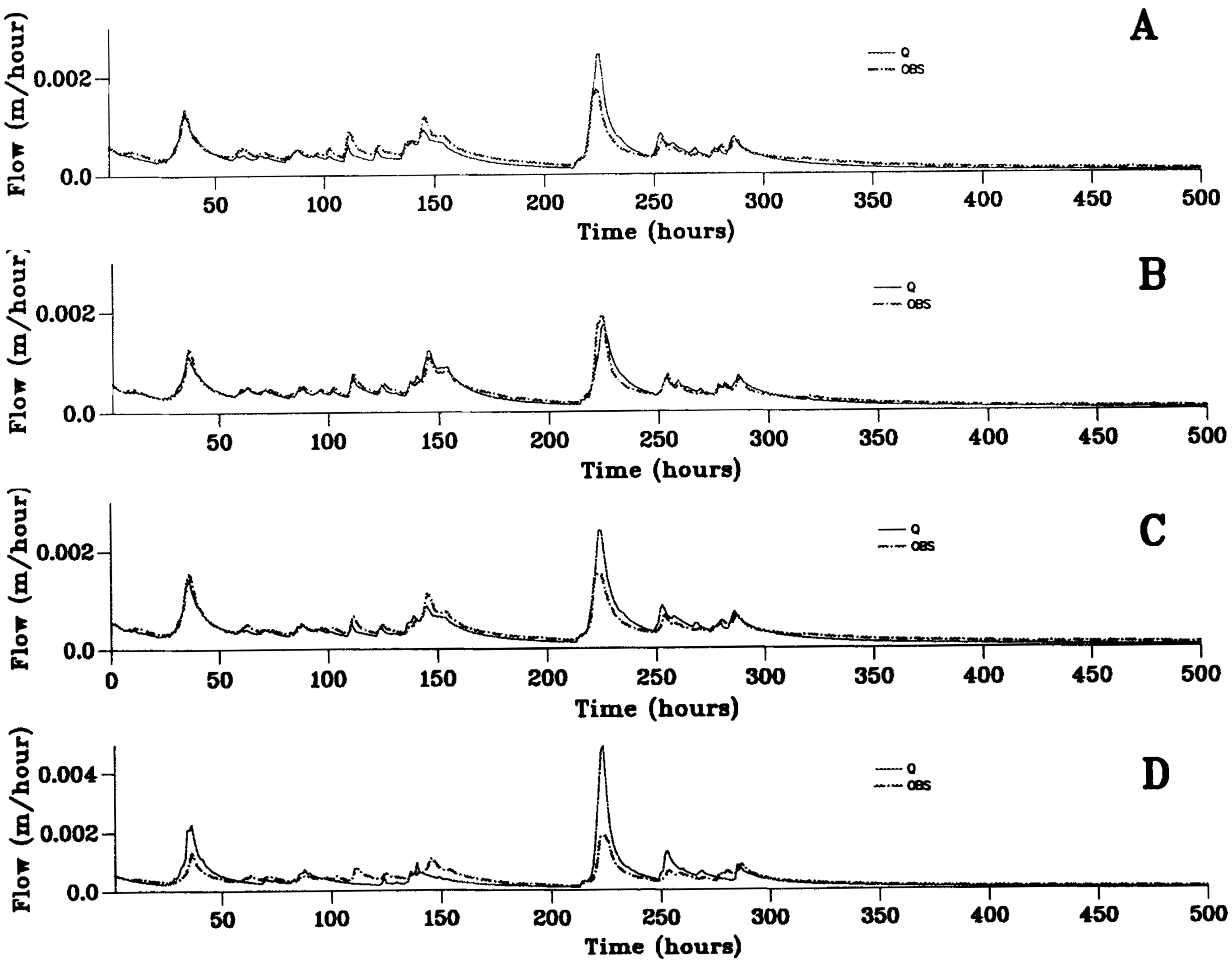

Figure 5. Validation using internal gauges (a window of 500 hours has been chosen for inspection). (A) Results for the complete river Wye catchment. (B) Cyff. (C) Gwy. (D) Nant Iago

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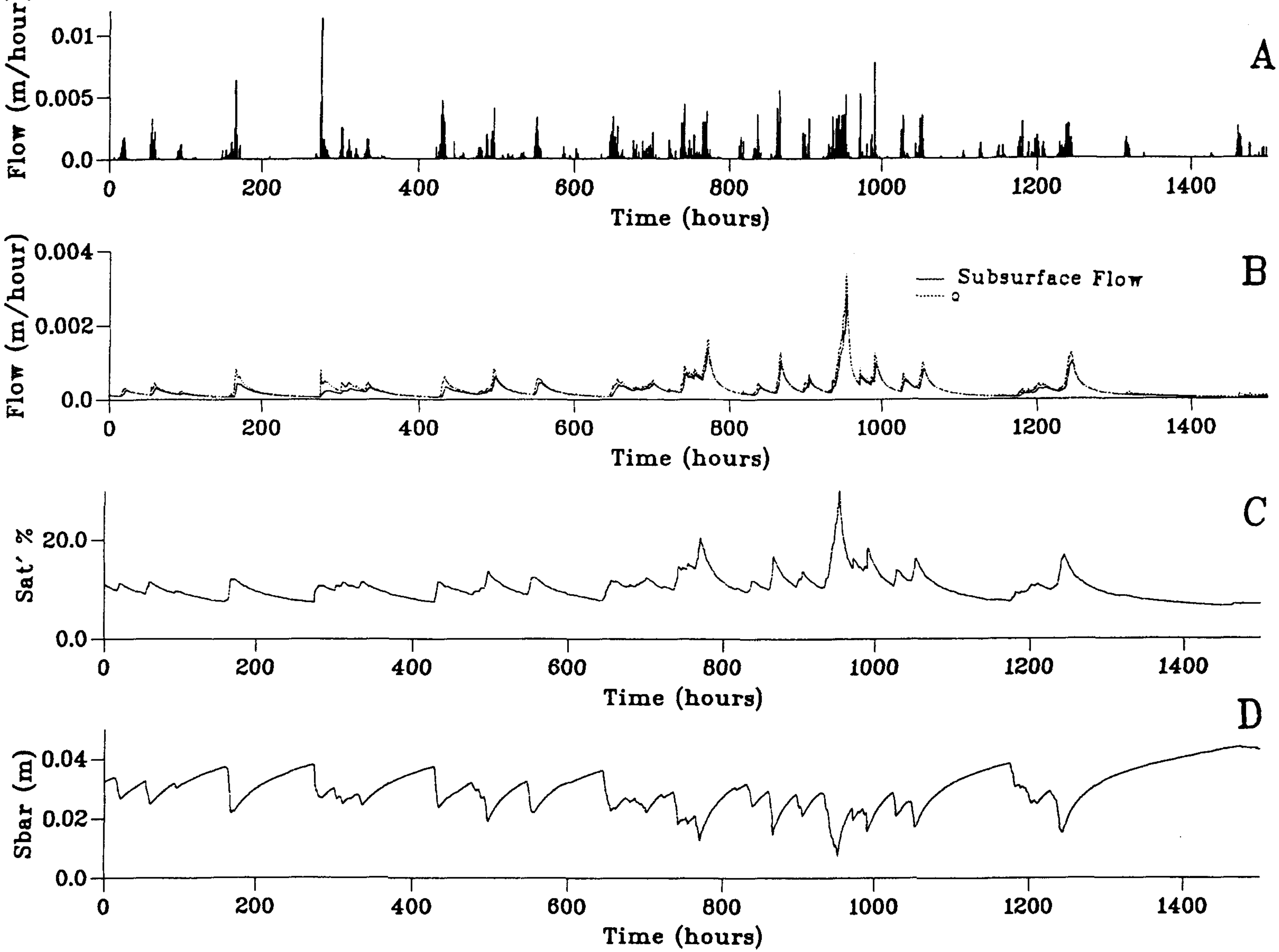

Figure 6. TOPMODEL component processes. (A) Rainfall for a three month period for 1985 for the river Wye. (B) Total flow and that made up by subsurface flow. (C) Percentage of the catchment saturated in each timestep. (D) Fluctuation of SBAR, the catchment average deficit

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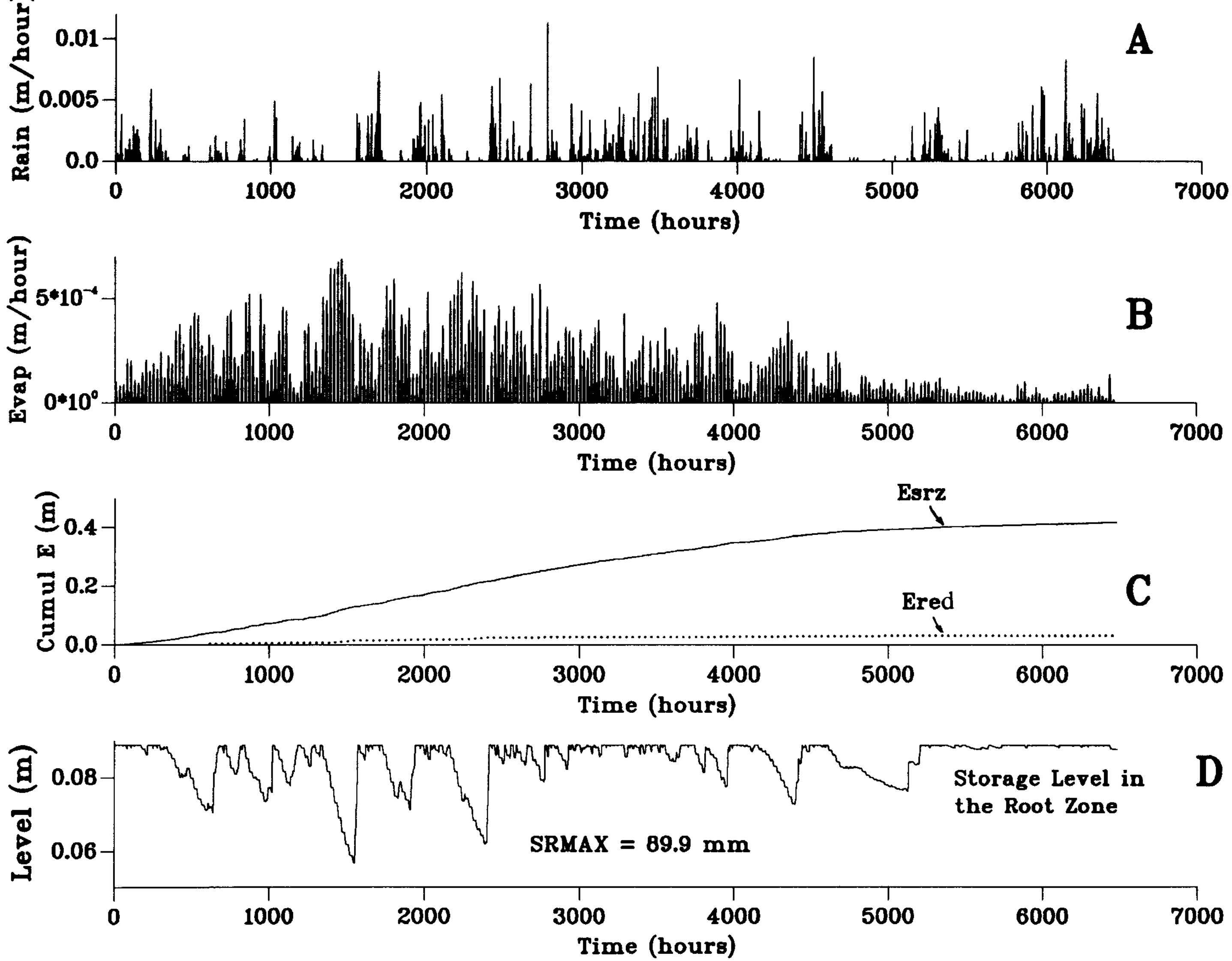

Figure 7. Temporal evaporation dynamics. (A) Rainfall for nine months for the river Wye. (B) Evaporation rate for nine months for the river Wye. (C) Two components of evaporation: Esrz is controlled by the root zone storage; Ered is controlled by the redistribution of water to the root zone. (D) Status of the root zone store in each timestep

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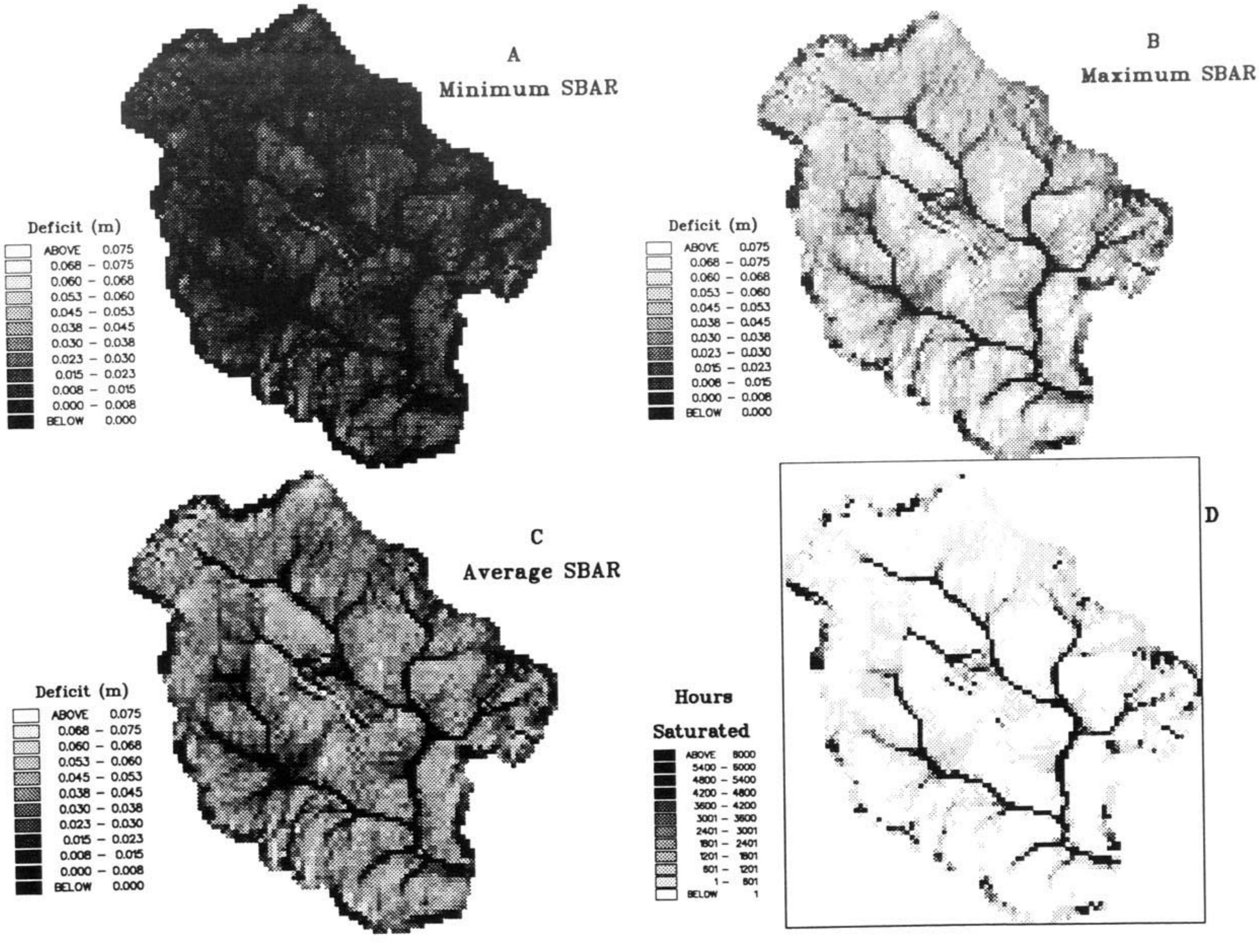

Figure 8. Variable source are dynamics: plan view. (A) Soil moisture map with the largest saturated area (1985). (B) Soil moisture map with the smallest saturated area (1985). (C) Soil moisture map with the average saturated area (1985). (D) Cumulative frequency of saturation for nine months in 1985

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give smaller saturated areas). A map of the predicted soil moisture deficits during the largest event of the year (Figure 8A) shows that the saturated area becomes fairly large (36% of the total area). The driest point in the year for 1985 shows a small saturated riparian area and some small saturated areas on the catchment divides (approximating to peat hags induced by the low gradients). Some saturated area is still present despite drainage of the catchment for several hundred hours. Using the average SBAR value for 1985 a further map of the variable source area was produced (Figure 8C), noting that this map resembles Figure 8B rather than 8A. On average the saturated area fluctuation is fairly small for most of the simulation, varying between 5 and 10% for most of the year. The final map is a summation of both the spatial and temporal dynamics of the variable source area for the whole nine month period (Figure 8D). The map shows the number of hours that each cell was saturated in 6480 hourly timesteps. This again shows the dominance of the small riparian type feature. The riparian feature is a topographically controlled phenomena that is maintained by downslope flows that are implicit in TOPMODEL. Storm events rarely expand saturation further than the riparian area, but occasionally stretch to fill the whole of the valley bottom feature.

The conclusion from these spatial predictions is that the variable source area for the river Wye is a small but permanent feature that has the dynamic capability to expand into a large area during major storms.

To study the fluctuation of the water-table within the soil profile a similar soil moisture deficit analysis was carried out, but only on a single section taken from the DTM. This section (chosen to span the river channels) is used to reinforce the links between the relative elevation of any grid cell, the soils–topographic combined index and the subsequent subsurface soil moisture dynamics experienced at that position. The topography for the section and the soil–topographic index value can be seen in Figure 9A and 9B, respectively. Obvious patterns exist, such as the high values corresponding with the channels and the lowest value associated with a single hillslope peak. Some higher values can also be seen near the catchment divides due to their low tangent β values. Figure 9C is a map of the soil profile over the cross-section. By creating on the y-axis a subdivision of the possible soil moisture deficit values calculated from Equation (3), it can then be determined where the position of S_i is for each timestep in each cell (Figure 9C). The figure is thus a cumulative frequency of saturation map for the complete nine months. The pattern is distinct, showing the high index values predicted as being permanently or near saturated for the nine months. A gradation of values is seen following the hillslope form, reaching its deepest position on the top of the single peaked hillslope. The small fluctuation of the larger deficits reflect the fact that the water-table tends to stay at that depth for most of the year. The 1–600 hours saturation class reflects the maximum rise in the water-table due to the larger storms.

WET AND DRY PERIOD SIMULATION

An attempt has been made to test the model on a distinctly wet and a distinctly dry period. Most hydrologists are more interested in modelling flow peaks as opposed to interstorm periods. However, the capability to model low flows is also a sound modelling goal. This is especially the case for general circulation modellers who are keenly interested in landscape–atmospheric fluxes over long time periods.

TOPMODEL predicts flow very well from a wet period to a dry period as recession simulation is one of its strengths, but the same is not true for the converse situation. If TOPMODEL wrongly predicts the antecedent storm conditions then discharge resulting from a particular storm will either be missed or overpredicted. For the chosen dry period (late April to early May 1985) the flows are small with a series of intermittent storms, hence will test the root zone formulation and the SRMAX value most vigorously. The wet period simulation also begins in the height of summer (late June to early July 1985) when the evapotranspiration values are at their highest.

The parameters calibrated to their respective periods are shown in Table IV.

The value of m is not significantly different for each simulation; values of 0·0079 m and 0·012 m will give hydrograph shapes that are similar. The T_0 parameter seems to fluctuate substantially but this value can change significantly without reducing the model efficiency dramatically. The SRMAX values are, however, much changed, optimizing to 5 and 7 mm compared with 89·9 m for the whole of 1985. This is due

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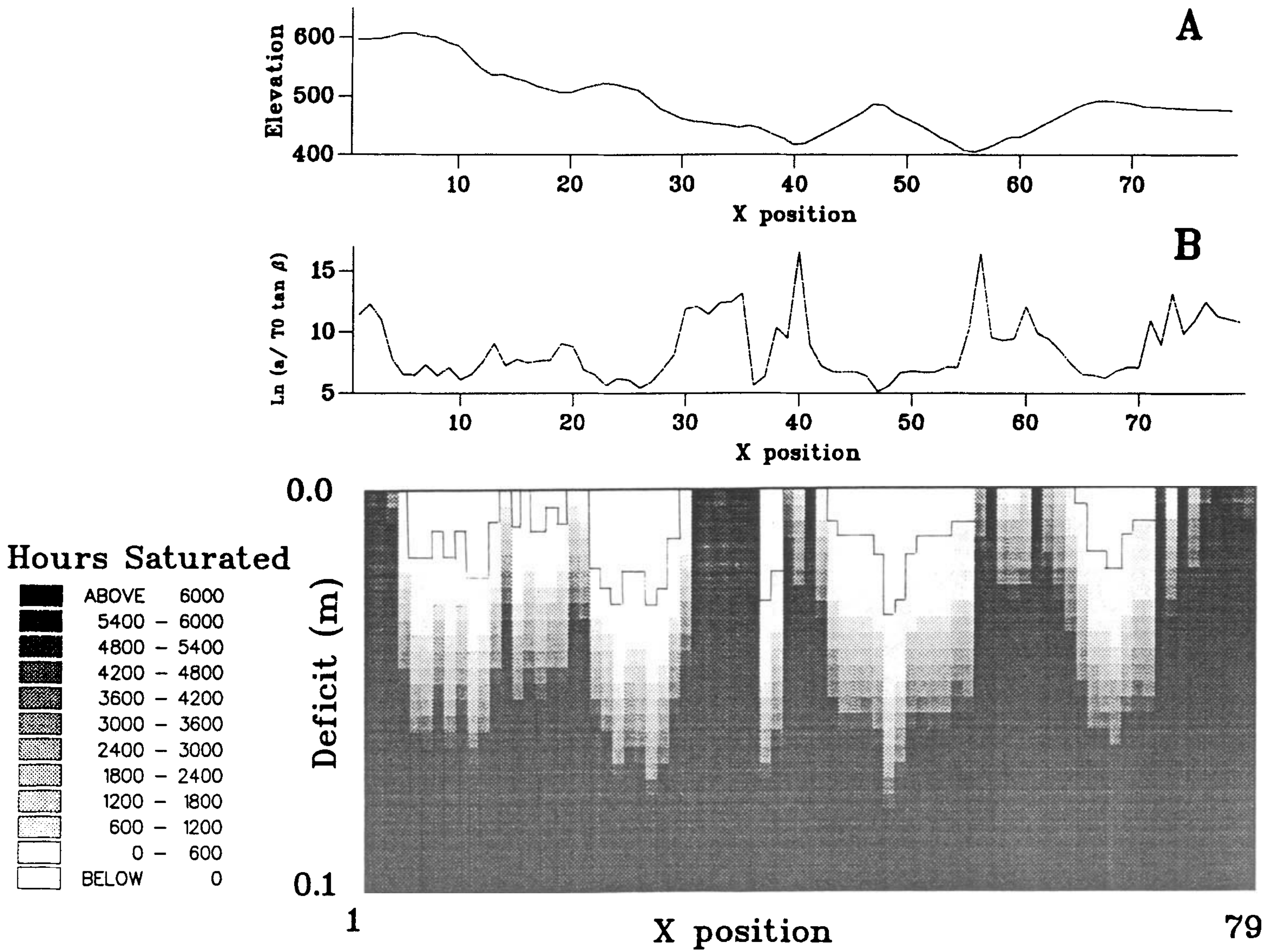

Figure 9. Variable source area dynamics: actual cross-section of terrain. (A) Elevation of the actual cross-section. (B) Soil topographic index values for this cross-section. (C) Cumulative frequency of saturation for each cell across a range of deficits for nine months in 1985

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Table IV. Calibration parameters determined for the wet and dry periods of the 1985 flow record

	T_0 (m^2/hour)	m (m)	SRMAX (m)	Efficiency (%)
Dry (1200 h)	24·79	0·012	0·005	68·0
Wet (1400 h)	16·10	0·0079	0·007	68·2
Wye 1985 (nine months)	8·27	0·0093	0·088	85·2

to the fact that the root zone conceptualization is now being tested by the observed data. This is obvious for the dry period in that the value has become small to fit the peaks for the storm events seen (Figure 10A–C). This is also true in the wet period as the first storms in the sequence happened to follow a drier period (Figure 11A and B). Conversely the SRMAX value for the nine month period is free to drift as it is untested by the efficiency criteria. If a parameter is untested by optimization data then it could be said to be 'redundant' (Sorooshian, 1991), hence its value is physically less meaningful. The points raised by Klemes (1986) that the wet and dry periods should be tested independently is valid if meaningful parameters values are to be found. However, many hydrologists would find the low optimized SRMAX values too small to be realistic.

The question is now whether these parameter values alter the quality of the results significantly when validated. So, firstly, the optimized parameters were cross-validated on the wet and dry periods. The Table V shows the efficiency of these runs.

The dry parameter values yield fairly good results on the wet period (Figure 11D) but the wet parameters do not give a good efficiency for the dry period (Figure 10D). The low efficiency on the dry period when using the wet parameters is largely due to the sensitivity of the efficiency calculation. This sensitivity of our goodness of fit criteria is clearly shown by running both sets of parameters on the complete nine month period of 1986 as validation. Efficiency values of 89·48 and 93·84% for the 'optimized values tells us that both sets work well on a long data set with a good mixture of storms. It is tempting to suggest that the dry period parameters are a better optimum parameter set at these values work the best on all periods. Reservations remain, however, that the efficiency calculation may not be an accurate criteria for the goodness of fit due to its sensitivity to the chosen data period. For example, the efficiency of 4·03% for cross-validation of the wet parameters on the dry period leapt to a 90% efficiency if two later storms were included.

The early part of the wet period simulation shows a number of storms being under-predicted. This is due to an error induced by the antecedent condition not being as dry as that suggested by the model. The effect of over-drying the model before a storm is that the root zone needs to be filled, hence both runoff processes (overland flow and subsurface stormflow) have to wait for filling to be completed before beginning operation. This suggests a weakness in the root zone conceptualization. Also the very low values of SRMAX do not reflect the field measurements of Calder *et al.* (1983), where local deficit values of 50 mm were observed, rising to values of 150 mm for the 1976 drought.

EVAPOTRANSPIRATION DYNAMICS

These calibration exercises have resulted in some very different values of the SRMAX parameter. The sensitivity of evapotranspiration predictions to changing the SRMAX value (Table VI) is thus examined here. The values compared are the 89·9 mm value derived from the nine month calibration of 1985 and the 7·9 mm value found in the wet period simulation (Figure 11B and 11C and Table IV). Alteration in evaporative flux would be of concern to a general circulation modeller who requires an accurate hourly evapotranspiration term for their climate models.

Small SRMAX values are needed to fit storms after dry antecedent conditions and by reducing evapotranspiration, the effect this has on the root zone status is shown in Figure 10C. Figure 10C compared

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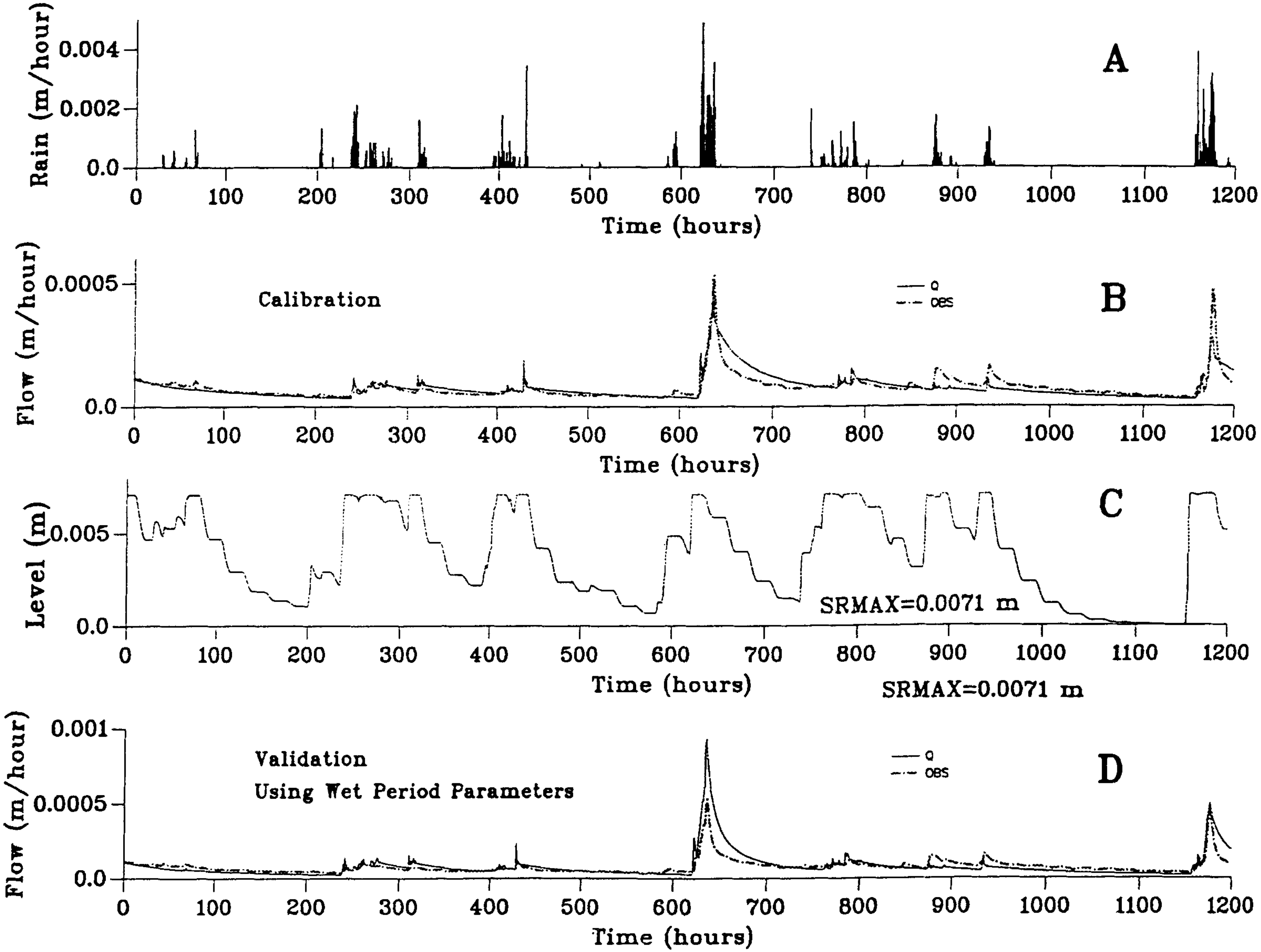

Figure 10. Dry period calibration and validation. (A) Rainfall showing small storms and relatively long interstorm periods. (B) Observed and simulated flow for the dry period. (C) Status of the root zone storage where SRMAX = 7·1 mm. (D) Validation of the dry period simulation using the wet period calibrated parameters

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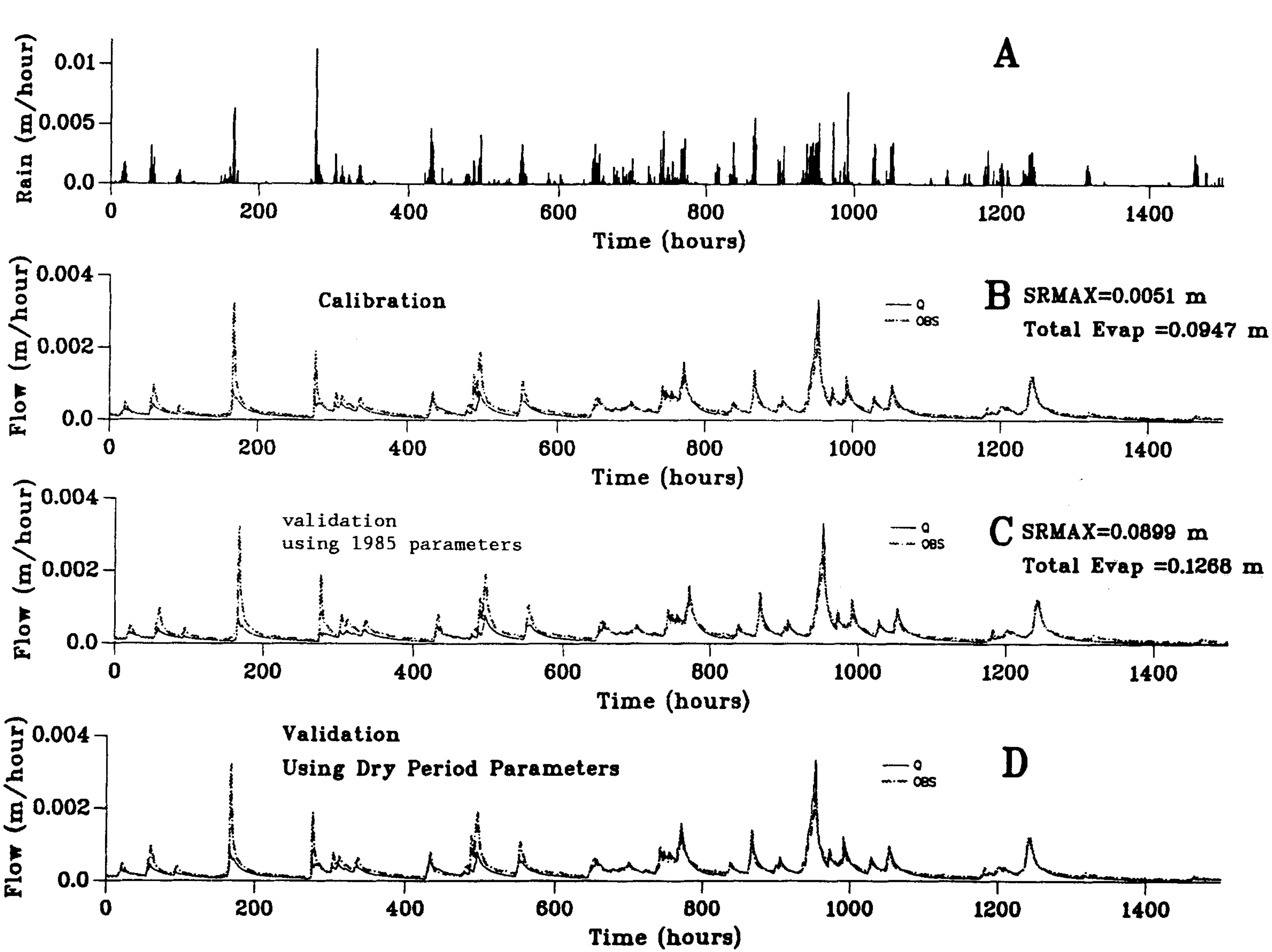

Figure 11. Wet period calibration and validation. (A) Rainfall showing a sequence of repeating storms. (B) Observed and simulated results when SRMAX = 5·1 mm. (C) Observed and simulated results when SRMAX = 88·9 mm. (D) Validation of wet period simulation using the dry period parameters

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Table V. Cross-validation of the wet and dry period parameter sets

	Validation periods (%)		
	Wet	Dry	1986 nine months
Dry	62·62	—	89·48
Wet SRMAX = 0·0079 m	—	4·03	93·84

with Figure 7D shows the difference in evapotranspiration loss from the root zone, with large SRMAX values keeping the evapotranspiration at close to potential for longer as larger amounts of moisture are available. The current conclusion is that the weakest part of the model is in the root zone conceptualization. The model needs to be made more complex to reflect the range of processes operating in the field. The problem is, however, that the number of important model parameters would increase, thus making the optimization process more difficult or unrealistic. This is especially true for the case where the new parameters are not independent of each other.

The current evapotranspiration conceptualization gives reasonable predictions of the hydrographs and the overall water balance for long periods of simulation. Hence further analysis of the evapotranspiration process was made to test the spatial variability of the predictions. High index values produce Ered when the root zone is depleted during dry periods, hence giving a higher total evapotranspiration (Figure 12A). The difference between individual grid cells and the average evaporation total are small due to the dominance of the Esrz process. The total amount of Ered depends on the number of grid cells in the classes that have a high water-table. The small number of cells in the tail of the $\ln(a/T_0 \tan\beta)$ distribution keeps the total volume of the Ered contribution small (Figure 12B). The relative evaporation rate for each $\ln(a/T_0 \tan\beta)$ class was also calculated so that the difference in response caused by the Ered process could be seen (Figure 12C). Three distinct zones can be seen: (1) a background evapotranspiration rate for the whole catchment, derived totally from the Esrz process; (2) a build up of extra contribution from the redistributed water component for cells that are saturated for only some of the period; and (3) an upper maximum evapotranspiration rate from those cells which are saturated for the whole time period.

Although the redistribution of water has had little effect for the river Wye, it may be of significance to catchments during drought periods and may directly influence the vegetation pattern of the catchment.

DISCUSSION

TOPMODEL has been thoroughly described in terms of its spatial and temporal modelling capabilities. These processes require further testing against spatial field data and the model must be tested on catchments in different hydrological regions (which is an ongoing process). As stated earlier, when applied to

Table VI. Sensitivity of evapotranspiration total to the root zone depth for nine months in 1985

	Epot (m)	Esrz (m)	Ered (m)	Ea (m)
SRMAX = 0·0079	0·1268	0·0921	0·0026	0·0947
SRMAX = 0·0889	0·1268	0·1236	0·0032	0·1268

Epot = Institute of Hydrology value used as the upper limit of potential evapotranspiration.
Esrz = Evapotranspiration calculated from Equation (4), the root zone depletion relationship.
Ered = Evapotranspiration generated by water redistributed into the root zone from upslope.
Ea = Esrz + Ered, the total evapotranspiration generated by the model.

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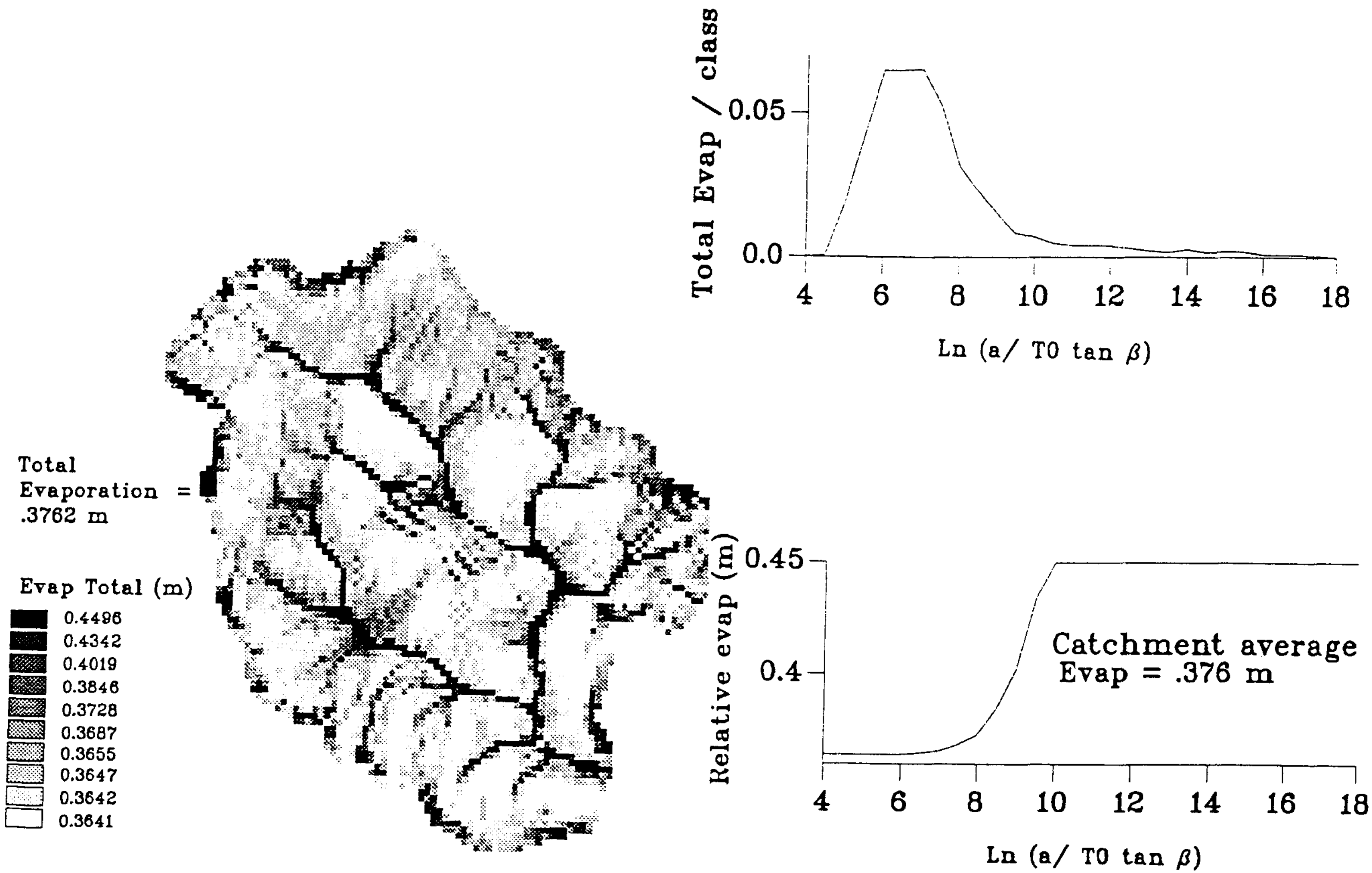

Figure 12. Spatial evapotranspiration dynamics. (A) Actual cumulative evaporation for a cell in that class for nine months in 1985. (B) Total cumulative evaporation for each soil–topographic index increment. (C) Results of 1A, scaled to show the relative evaporation rate per soil–topographic index increment

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a particular catchment the model should be thought of as being a conceptual toolkit for analysis. The model is simple so it can be easily changed to reflect particular local phenomena. The low number of optimization parameters make the calibration of the model possible for a wide range of circumstances.

TOPMODEL has predicted, with mathematical simplicity, processes that are dynamic in time and space thus, demonstrating the use of distribution functions for catchment analysis. The model has predicted soil moisture patterns that show a variable source area that is generally small and restricted to the riparian area, but during larger storm events has the ability to expand to a substantial size.

A major conclusion from this study, reinforcing that of Hornberger *et al.* (1985), was that the efficiency criteria may not be sensitive enough to define a single unique modelling parameter set. It is possible to have reasonable model efficiency and have poor hydrograph simulation and, conversely, to have low efficiencies where the hydrograph simulation is visually adequate. An extra qualitative weighting criteria was needed in this study to define a parameter set applicable over the widest range of circumstances. The use of wet and dry periods has aided in this process; sometimes it is better to trade a small amount of efficiency over long periods to gain efficiency on storms of a shorter, more difficult period. Also the fact that a number of parameter sets were determined for different calibration data sets tends to suggest that there is probably not one unique parameter set for this catchment. It is probably true that there is a zone of the parameter space that is the best area for deriving model parameters such as T_0 and m. This was shown by the interchangeability of the parameters from one period to another with only small changes in the overall modelling quality on validation.

SRMAX as an important model parameter for the evaporation total was shown to have a wider optimization zone. It may be true that the parameter has no effect on the efficiency calculation over long data periods and could be pre-set and lost from optimization. However, the point raised by Klemes (1986) that the model should be appropriate to distinct hydrological conditions (in this instance a drier period) is still valid. Hence, it is necessary to use an appropriate period of calibration data to set SRMAX. It is more likely that the root zone conceptualization and subsequent parameterization needs to be improved. For this hydrological regime, a minimum parameterization has been chosen, with one parameter to be optimized. It is easy to make the evapotranspiration process more complex, as has been attempted in the multiple parameter soil–vegetation–atmosphere–transfer schemes or SVATS [for example, SiB, Sellers *et al.* (1986) or BATs, Dickinson *et al.* (1986)]. There still appears to be scope, however, for an improved simple parameterization with minimal requirements for calibration. Some knowledge of the patterns of soil moisture in space and time would be a considerable help in this respect.

Both fieldwork data sets and the hydrological model must be interactive tools for hydrological analysis and must evolve in unison.

ACKNOWLEDGEMENTS

Thanks are due to NERC for funding this work as part of the PhD of Paul Quinn and to the Institute of Hydrology for their data and advice. Thanks also to our current grant under the TIGER initiative for our work on evaporation dynamics.

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