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Application of TOPMODEL in Buliu River Basin and comparison with Xin'anjiang model

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Abstract: Along with the rapid development of computer and GIS technology, hydrological models have progressed from lumped to distributed models. TOPMODEL, a bridge between lumped and distributed models, is a semi-distributed model in which the predominant factors determining the formation of runoff are derived from the topography of the basin. A test application of TOPMODEL in the Buliu River Basin is presented. For the sake of comprehensively evaluating the TOPMODEL, the Xin'anjiang model, a classic lumped hydrological model, was also applied in the basin. The structural differences and the simulation results of the two models are compared and analyzed.

Key words: *TOPMODEL*; *the Buliu River Basin*; *topographic index*; *the Xin'anjiang model*

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1 Introduction

In the 1970s and 1980s, due to the restrictions of available data observation and collection methods, distributed hydrological models developed more slowly than lumped hydrological models. With the rapid development of computer and GIS technology, more hydrological models have progressed from lumped to distributed models. A distinct characteristic of these distributed models is their use in conjunction with digital elevation models (DEMs). Distributed models based on DEMs have combined the research results of traditional hydrology, information technology and other related fields, and become both a hydrological simulation technology research hotspot and a developing aspect of existing hydrological models. In recent years, many advanced distributed models have been developed. Popular distributed hydrological models include SHE, TOPMODEL and SWAT (Reed et al. 2004). The first can be used only in small watersheds, and the latter two can be applied in large or semi-large basins (Liu et al. 2006).

TOPMODEL is a rainfall-runoff model that uses a simple theory of the relationship between different points in a catchment, their hydrological similarity, to predict the distributed responses of the catchment to rainfall. In the original form of the model (Beven and Kirkby 1979), the index of hydrological similarity is based on the topographic index $\ln(\alpha / \tan\beta)$. This form of the index assumes that the water table is, in a quasi steady state, parallel to the ground surface slope and that transmissivity varies locally as an exponential function of depth.

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Other forms of local transmissivity lead to different forms of the topographic index, as Beven (1986) demonstrates for spatially variable transmissivity at saturation, Ambroise et al. (1996) for linear and parabolic transmissivity, and Iorgulescu and Musy (1997) and Duan and Miller (1997) for a general power law function. The Xin'anjiang model has been described by Zhao et al. (1984) and is now widely and successfully used in China. TOPMODEL is always described as a bridge between the lumped and the distributed models, which succeeds in simulating the watershed runoff in many areas, as shown by Guo et al. (2000) in the Baishui River Basin in China and by Xiong et al. (2002) in the King Watershed in Ireland.

2 Fundamentals of TOPMODEL

In transforming the formal conceptualization of hydrological processes into functional mathematical structures, TOPMODEL makes three basic assumptions by Beven et al. (Singh 1995): (1) that successive steady state representations can come close to representing saturated zone dynamics; (2) that the slope of the local ground surface is an acceptable approximation of the hydraulic gradient; and (3) that distributed downslope transmissivity can be calculated as an exponential function of the storage deficit or water table depth.

These assumptions lead to simple relationships between catchment storage (or storage deficit) and the local water table level (or storage deficit due to drainage), in which the main factor is the topographic index $\ln\left(\alpha / \tan\beta\right)$ (Beven 1997). The index represents the propensity of any point in the catchment to develop saturated conditions (Wolock 1993).

3 Application of TOPMODEL

3.1 Buliu River Basin

The Buliu River Basin, which is located in Southeast China (Figure 1), covers an area of approximately $3\,310\ \mathrm{km}^2$ with primarily loamy soil, as well as clay near its outlet. The length of the Buliu River is 132 km. The Buliu River Basin lies in the subtropical monsoon region and its mean annual precipitation is 1 193 mm. The basin is represented by a DEM, which can be accessed on the internet (Figure 2(a)), with a grid cell side length of 1 000 m (USGS 2003). The drainage network was extracted automatically from the DEM by setting the threshold for the drained area at 1% of the total area (Figure 2(b)) (Fairfield and Leymarie 1991). The drainage area can be created by identifying the DEM grid cells that drain to the network (Figure 2(c)). Hourly values for seven rain gauges were used in this case study, as well as available discharge data from Pingla. The regional rainfall distribution was estimated using the Thiessen polygon method.

Figure 1 Main tributaries of the Buliu River Basin

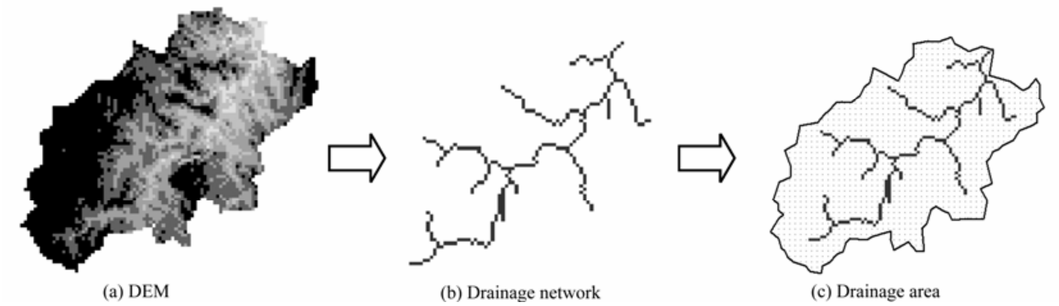

Figure 2 Drainage network and area extracted from DEM

3.2 Topographic index

The multiple flow direction algorithm (Quinn et al. 1991, 1994) is used to calculate the topographic index, which is based on the distribution of contributing area to all downslope grid elements. The upslope contributing area can be assigned eight connected neighbors (Figure 3). The contour length L_1, L_2, L_3 and L_4, perpendicular to the flow directions in Figure 3(a), is determined by the geometric frame in Figure 3(b). Four adjacent cells have a weight of 0.5 and the adjacent cells at the corner have a weight of 0.35. Furthermore, assuming that the contributing area of each downslope direction is proportioned with the runoff gradient in that direction, the direction with a higher gradient would have a greater upslope contributing area (Kong and Rui 2003; Rui and Shi 2004).

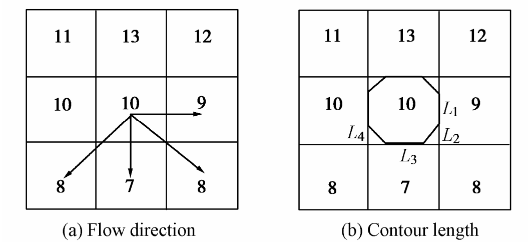

Figure 3 Runoff distribution of multiple flow direction algorithm

The contributing area of each downslope direction can be denoted as

$$\Delta A_i = A\left(\tan\beta_i L_i\right) / \sum_{j=1}^{n}\left(\tan\beta_j L_j\right) \tag{1}$$

where n is the number of downslope directions, ΔA_i is the contributing area of downslope direction i, A is the upslope contributing area, $\tan\beta_i$ is the slope gradient of downslope direction i, and L_i is the contour length of downslope direction i. The function of representative slope gradients is shown in Eq. (2):

$$\tan\beta = \sum_{j=1}^{n}\left(\tan\beta_j L_j\right) / \sum_{j=1}^{n} L_j \tag{2}$$

As there is a relationship:

$$\alpha / \tan\beta = A / \left(\sum_{j=1}^{n} L_j \tan\beta\right) \tag{3}$$

Substituting the definition of $\tan\beta$ in Eq. (2) into Eq. (3) leads to Eq. (4):

$$\alpha / \tan \beta = A / \sum_{j=1}^{n} \left(\tan \beta_j L_j \right) \tag{4}$$

Eq. (4) can be used to calculate the topographic index of each DEM grid cell (Figure 4).

3.3 Routing method based on DEM

The flow path of each DEM grid can be described based on the maximum elevation difference (Wigmosta and Lettenmaier 1999). Time-area routing can be achieved by classifying DEM grid cells according to the flow paths. Here we apply the time-area routing method based on the DEM to deal with runoff concentration. This method considers the effect of slope gradient and can generate the general distance-area accumulation curve conveniently. The distance-area curves for the river basin are shown in Figure 5.

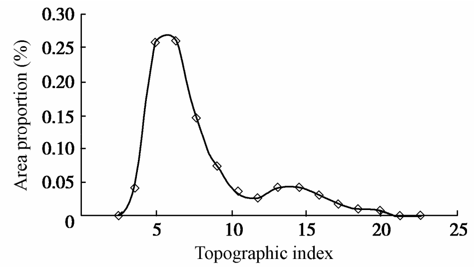

Figure 4 Distribution curve of the topographic index

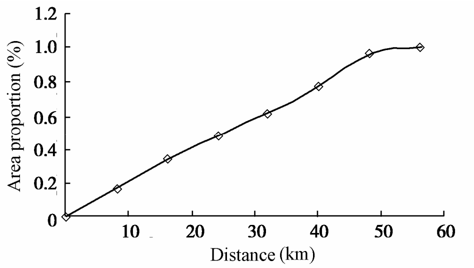

Figure 5 Distance-area accumulation curve of the routing method based on the DEM

3.4 Flood simulation results

The calibration of the model was performed at a 1-hour time step using available hydrological data from 2002 to 2004. Data from 2005 to 2007 were used for validation. The flood simulation results of TOPMODEL are shown in Table 1.

Table 1 Flood simulation results of TOPMODEL

Flood starting date	Rainfall (mm)	Runoff			Peak discharge			T_d (h)	Deterministic coefficient
		Observation (mm)	Calculation (mm)	Error (%)	Observation (m^3/s)	Calculation (m^3/s)	Error (%)		
2002-07-18	134.7	59.9	58.9	-1.67	602	493.2	-18.07	2	0.70
2002-08-12	157.7	96.8	104.5	7.95	867	879.4	1.43	1	0.61
2003-05-28	39.6	25.4	25.0	-1.57	457	488.0	6.78	0	0.83
2003-06-04	165.0	91.5	101.4	10.82	1280	1208.1	-5.62	-1	0.92
2003-06-19	120.6	67.0	77.3	15.37	1110	929.2	-16.29	0	0.67
2003-07-06	61.5	26.3	27.2	3.42	359	324.5	-9.61	1	0.90
2003-07-24	87.4	45.9	51.2	11.55	415	454.3	9.47	-1	0.58
2003-09-08	34.4	29.1	28.6	-1.72	262	260.0	-0.76	-1	0.85
2004-07-08	76.2	27.4	28.3	3.28	573	561.7	-1.97	-2	0.68
2004-07-18	128.2	52.8	57.4	8.71	401	406.2	1.30	2	0.60
2004-08-18	25.9	23.7	25.9	9.28	204	209.3	2.60	1	0.90
2004-09-01	60.2	25.4	27.5	8.27	205	231.7	13.02	0	0.82
2005-06-06	49.9	17.6	17.2	-2.27	332	317.8	-4.28	-4	0.57
2005-06-10	46.4	18.2	17.4	-4.40	492	489.2	-0.57	1	0.79
2006-06-16	54.8	19.1	17.3	-9.42	430	493.6	14.79	1	0.75
2006-07-16	128.8	71.5	73.9	3.36	782	659.3	-15.69	0	0.96
2006-08-01	211.8	118.3	129.8	9.72	1140	1163.4	2.05	0	0.96

T_d: time difference of peak appearance

4 Comparison with the Xin'anjiang model

4.1 Structure of the models

In the Xin'anjiang model, the actual evapotranspiration of the basin is related to both potential evapotranspiration and soil moisture conditions: evapotranspiration is calculated using a three-layer soil moisture model. The concept of runoff formation on repletion of storage is applied to calculate runoff production, which means that runoff is not produced until the soil moisture content of the aeration zone reaches field capacity, and thereafter runoff equals additonal rainfall without further loss. The concepts of free water storage and free water distribution are developed to separate runoff into three components: surface runoff, groundwater flow and interflow (Zhao 1984). Flow concentration in sub-basins is calculated according to a linear reservoir algorithm. The Muskingum successive reaches model is applied to route flow along main rivers (Lin 2001).

TOPMODEL follows the generally adopted practice of calculating actual evapotranspiration as a function of potential evapotranspiration and root zone moisture storage, in which actual evapotranspiration cannot be specified directly. In the Xin'anjiang model, the volume of evapotranspiration is controlled by tension water storage, the ratio of potential evapotranspiration to pan evaporation, and the coefficient of deep evapotranspiration (Xie et al. 2007). Because it considers evapotranspiration calculation from the vertical aspect, the Xin'anjiang model is more realistic and has more advantages in controlling the water balance.

TOPMODEL confirms source areas in accordance with the spatial distribution of the water table. It first calculates subsurface and saturated surface runoff separately, then sums them up for the total runoff. The Xin'anjiang model, on the other hand, calculates the total runoff first, then divides total runoff into runoff components through a free water storage framework for subsequent calculation of flow concentration. TOPMODEL describes the phenomenon of spatial heterogeneity of moisture using the distribution curve of the topographic index, while the Xin'anjiang model uses the tension water capacity curve to describe it.

In TOPMODEL, a distance-area curve from the time-area routing method embodies the lag decay function of the hydrological system. Saturated slope runoff and subsurface runoff are computed together for flow concentration. The Xin'anjiang model calculates flow concentration through its three components separately, in accordance with each decay law (as surface runoff, subsurface runoff and groundwater flow can all be described by a linear reservoir algorithm).

4.2 Parameters and results

The Xin'anjiang model has 17 primary parameters. The parameters of a single flood that have been calibrated using the same calibration and validation data used with TOPMODEL are listed in Table 2. The comparison of simulation results is shown in Table 3. The data in Table 3 are mean values of all individual floods.

Table 2 Parameters of the Xin'anjiang model

Parameter symbol	Meaning of the parameter	Value
K	Ratio of potential evapotranspiration to pan evaporation	0.93
B	Exponent of the tension water capacity distribution curve	0.3
C	Coefficient of deep evapotranspiration	0.15
WM	Areal mean tension water capacity	120
UM	Tension water capacity in the upper layer	20
LM	Tension water capacity in the lower layer	60
IM	Ratio of the impervious area to the total area of the basin	0.01
SM	Maximum possible deficit of free water storage	40
EX	Exponent of the free water capacity curve	1.5
KG	Outflow coefficient of free water storage to groundwater	0.35
KI	Outflow coefficient of free water storage to interflow	0.35
CG	Recession constant of groundwater storage	0.98
CI	Recession constant of the lower interflow storage	0.55
X	Parameter of the Muskingum method	0.1
CS	Recession constant in the lag and route method	0.2
L	Lag in the lag and route method	1

Table 3 Comparison of flood simulation results of two models

Type	Runoff error (%)		Peak discharge error (%)		Time difference of peak appearance (h)		Deterministic coefficient	
	(1)	(2)	(1)	(2)	(1)	(2)	(1)	(2)
Calibration	6.14	-5.76	-1.48	7.21	0.17	0.38	0.76	0.85
Validation	-0.60	-14.73	-0.74	-5.23	-0.20	-0.33	0.81	0.87

(1): TOPMODEL; (2): the Xin'anjiang model

TOPMODEL and the Xin'anjiang model are comparably efficient. TOPMODEL has much fewer parameters that need to be calibrated than the Xin'anjiang model, and many of its parameters have a physical basis. For flood simulation, parameter calibration is more simple and convenient in TOPMODEL than in the Xin'anjiang model. On the whole, both models provide good simulation results for the study catchments. The discharge hydrographs for the results of the two models are shown in Figure 6.

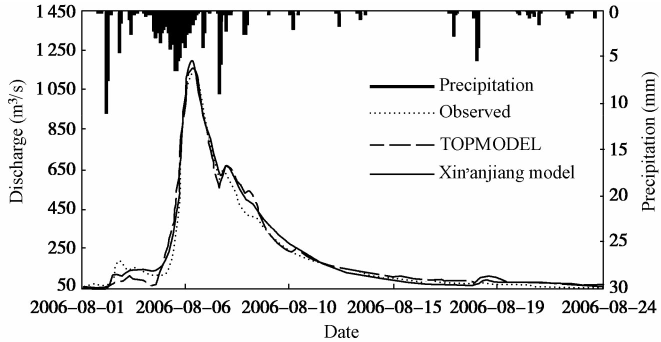

Figure 6 Buliu River Basin discharge hydrograph

5 Conclusions

TOPMODEL, based on the variable source area concept, uses a topographic index to reflect the influence of the spatial variability of the earth's surface on watershed hydrological cycle processes. It provides a good method for solving the spatial problems of a hydrological model (Liu et al. 2003). Compared with the Xin'anjiang model, TOPMODEL has three important characteristics: (1) its dynamic simulation of water movement in the unsaturated zone and shallow underground layers is more appropriate for practical hydrological application; (2) it incorporates the influence of the earth's surface topography on hydrological response and visualizes the simulation results of the spatial variation of runoff production; and (3) it is used in conjunction with GIS technology.

Compared with distributed hydrological models, the important characteristics of TOPMODEL are its simple structure and its low number of parameters that need to be calibrated. The model only considers the surface topography. A set of other hydrological factors are assumed to be uniform over space. Furthermore, except for topographic index calculation, watershed grids have no practical significance. The distinguishing characteristic of distributed hydrological models is the consideration of the spatial distribution of hydrological factors and the relationships between them. By this definition, TOPMODEL is a semi-distributed hydrological model. TOPMODEL's advantage is that, due to its simple structure, it only requires a DEM and basic hydrological data. Its disadvantage is that it gives less consideration to the spatial variation of some hydrological factors and relationships between hydrological cells. At present, TOPMODEL is undergoing much improvement, due to further research on basic assumptions made by the model, the effects of precipitation distribution and vegetation in model input. TOPMODEL illustrates the idea of building a distributed model on a DEM. Further exploration of the model, whether in runoff and water quality simulation, climate and land cover variation research, or in water resources management, has promising application potential.

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