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Distributed TOPMODEL Approach for Rainfall-Runoff Routing Modeling in Large-Scale Basins

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Synopsis

The objective of this study is to apply a modified distributed TOPMODEL approach to three large-scale basins as a Rainfall Routing Model (RRM): the Amazon, the Brahmaputra and the Yangtze basins. This modified approach uses a multi-velocities parameterization to routing the flow. The modified model also uses spatially distributed rainfall and evapotranspiration data through a cell-to-cell routing method. Monte Carlo method was used to find the best set of parameters. The modified model obtained Nash-Sutcliffe coefficient values of 0.48, 0.82 and 0.78 for the Amazon, the Brahmaputra and the Yangtze basins, respectively. The simulations showed that the modified TOPMODEL approach seems to be a reasonable hydrological model to estimate stream flow discharges in large-scale river basins.

Keywords: Large-scale, TOPMODEL, GCM, RRM, distributed model, multi velocities

1. Introduction

Global scale hydrological modeling is an important issue in order to promote the understanding of the global water cycle in terms of quantity and quality.

General Circulation Models (GCMs) are the current tool to simulate the global climate system. According to Durcharne *et al*. (2003), GCMs need the discharges from the rivers to model the water cycling through oceans, atmosphere and land.

In the past decade, GCMs using large-scale rainfall routing models (RRMs) have received special attention. The use of RRMs has basically three purposes: (1) to study the freshwater flux into the oceans, which may affect ocean convection, ocean salinity and ice formation, (2) to evaluate the GCM performance and (3) to study the impact of climate change on water resources (Arora, 2001).

RRMs need routing the flow through the river channels until the basin outlet.

According to Ngo-Duc *et al*. (2007), in state-of-the-art global routing models, most of the approaches either assume a constant velocity or use simple formulas that use time-independent flow velocities parameterized as a function of the topographic gradient. In general, these approaches are sufficient to model discharges in monthly or longer time scales. Addressing this issue, Ngo-Duc *et al*. (2007) updated their model, Total Runoff Integrating Pathways (TRIP), to take into account variable velocities in the river channels throughout the basin and, therefore, to model the short-term discharge fluctuations. The attempt to consider variable velocities is also noticed in local distributed hydrological models, as can be seen in Ivanovi *et al*. (2004).

In distributed hydrological models the multi-velocities criteria can be applied to every cell in the grid. However, for lumped models this can be applied to the area-distance function (Rodriguez-Iturbe & Rinaldo, 1996).

TOPMODEL (Beven & Kirkby, 1979; Beven *et al.*, 1995) is a hydrological model based on variable source area assumption. The TOPMODEL framework has two components: (1) the storage component, which is represented by three reservoirs and (2) the routing component, which is derived from a distance-area function and two velocities parameters. Its main parameter is the topographic index derived from a digital elevation model. This index represents the propensity of a cell or region to become saturated.

The TOPMODEL is considered as a semi-distributed model, as it uses distributed topographic information to determine the topographic index and to distribute saturation deficits throughout the basin, as well. However, its main limitation is the impossibility to use distributed input data, such as rainfall and evapotranspiration.

The objective of this study is to apply a modified distributed TOPMODEL approach in three large-scale basins as a RRM: the Amazon, the Brahmaputra and the Yangtze basins. This modified approach uses a multi-velocities parameterization for routing the flow and uses spatially distributed rainfall and evapotranspiration data.

2. TOPMODEL background

The TOPMODEL is a rainfall-runoff model that uses the concept of hydrological similarity based on topography. This similarity is defined by the topographic index λ_j:

$$\lambda_j = \ln\left(\frac{a_j}{\tan\beta_j}\right) \quad (1)$$

where a_j [L] is the upslope contributing area per unit contour length for each cell j in the catchment and tan β_j [-] is the slope of this cell measured with respect to plan distance .

Once a number of classes with the same hydrological similarity is defined, the storage deficit S_i [L] for each class i is:

$$S_i = S + m(\lambda - \lambda_i) \quad (2)$$

where S [L]is the lumped or mean storage deficit for the entire catchment; λ is the mean topographic index (approximated by a weighted average over the areas with the same hydrological similarity); λ_i is the local topographic index and m is a parameter associated with the rate of decline of the catchment recession curve.

For each time step the mean storage deficit is updated following the equation:

$$S_t = S_{t-1} + \left\lfloor Q_{b_{t-1}} - Q_{v_{t-1}} \right\rfloor \quad (3)$$

where S_t is the updated value of the storage deficit; S_{t-1} is the storage deficit in the previous time step; Q_{bt-1} [LT^{-1}] is the base flow in the previous time step and Q_{vt-1} [LT^{-1}] is the unsaturated zone recharge in the previous time step. This recharge is defined by:

$$q_{Vi} = \frac{S_{UZ}}{S_i T_D} \quad (4)$$

where S_{UZ} [L] is the unsaturated zone deficit and T_D [TL^{-1}] is residence time in the unsaturated zone.

The baseflow Q_b [LT^{-1}] is defined by:

$$Q_b = Q_S e^{\left(\frac{-S}{m}\right)} \quad (5)$$

where Q_S [LT^{-1}] is the discharge when the catchment is saturated and it is calculated by:

$$Q_S = T_0 e^{-\lambda} \quad (6)$$

where T_0 [LT^{-1}] is the soil saturated transmissivity, which is constant for the entire catchment.

In the first time step the mean storage deficit is estimated by:

$$S_{t=0} = -m\ln\left(\frac{Q_0}{Q_S}\right) \quad (7)$$

where Q_0 [LT^{-1}] is the initial discharge at the first time step.

The TOPMODEL uses the Dunne overflow generation mechanism (Dunne & Black, 1970), *i.e.*, when the storage deficit equals to zero.

In the TOPMODEL approach there is a reservoir intended to represent root and vegetation storage. This is called Root Zone reservoir and can be depleted through the following equation:

$$E_a = E_p\left(1 - \frac{S_{RZ}}{S_{RMAX}}\right) \quad (8)$$

where S_{RZ} [L] is the root zone reservoir deficit, S_{RMAX} [L] is the maximum deficit in the root zone reservoir, E_p [LT^{-1}] is the potential evapotranspiration and E_a is the evapotranspiration rate. Its deficit at the first time step is set through a parameter called S_{RO}.

Flow routing is done through a time-area function. This function is derived from a distance-area function (Rinaldo *et al.*, 1995; Rodriguez-Iturbe & Rinaldo, 1996) using the following equation:

$$tc_k = \sum_{k=1}^{N} \frac{l_k}{V} \tag{9}$$

where tc_k [T] is the time of concentration of a determined class area of the catchment; V [LT^{-1}] is a velocity parameter, as k = 1, V is equal to the input parameter V_{CH} [LT^{-1}] and for $k > 1$, V assumes the value of the input parameter V_R [LT^{-1}]. V_{CH} represents the main channel velocity and V_R represents the average velocity of lower order rivers and hillslopes; l_k is the plan flow path length from a class area k to the basin outlet and N is the total number of classes which the area-distance function is composed.

3. Methodology

Study areas and data series

For this study were selected three large-scale basins, the Amazon, the Brahmaputra and the Yangtze basins (Fig. 1). The Amazon is a forest basin located in South America. Its area is roughly 7.05 million km^2. According to Beighley *et al.* (2009), the annual flow of the Amazon basin accounts for approximately one fifth of the all river discharges to the oceans. The Brahmaputra basin has roughly 1.7 million km^2 and it is the fourth largest river in the world in terms of discharge. The Yangtze basin located in China has 1.72 million km^2. It is the third largest river worldwide in terms of discharge.

For all basins the global topographic data were extracted using ETOPO5 data (Fig. 1), from the National Geophysical Data Center (NGDC), National Environmental Satellite (NOAA). Basins boundaries (Fig. 1) and stream networks were acquired from the Global Runoff Data Center (GRDC) and the Global Drainage Basin Database (GDBD), respectively. Although ETOPO5 grid has a 5-minute resolution, the resolution was interpolated to 0.5 degree, in order to match the climate and GRDC data resolutions. The climate daily data were obtained from the National Centers for Environmental Prediction (NCEP). Observed daily discharges from the GRDC for the Amazon encompass the period from 1990 to 1995 at Obidos station. For the Brahmaputra from 1990 to 1991 at Goalundo station. For the Yangtze basin the period of data used corresponds to the 2004 year at Datong station. The Penman-modified method (Doorenbos & Pruit, 1992) was used to estimate evapotranspiration. Figures 2 and 3 show examples of the spatial distribution of the rainfall and evapotranspiration for a day, respectively.

Model approach modification

This work used a TOPMODEL-GRASS version, implemented in C language (GRASS, 2010).

It is assumed that there is a power law relationship between cumulative area and velocity.

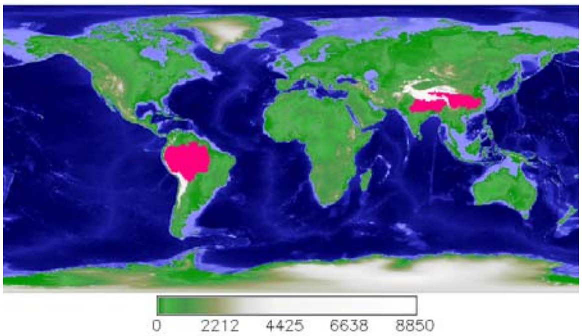

Fig. 1 ETOPO5 data with elevations in meters. For the Amazon basin (South America), the Brahmaputra basin (north of India) and the Yangtze basin (China).

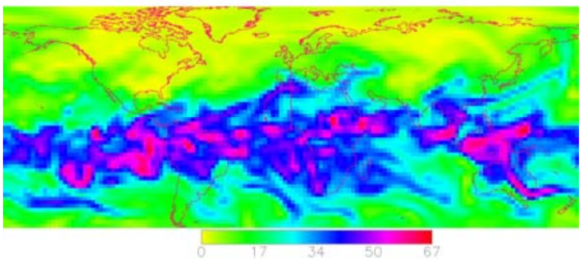

Fig. 2 Spatial distribution of daily rainfall from GCM daily data, 0.5 degree resolution. Units in 0.1 m.

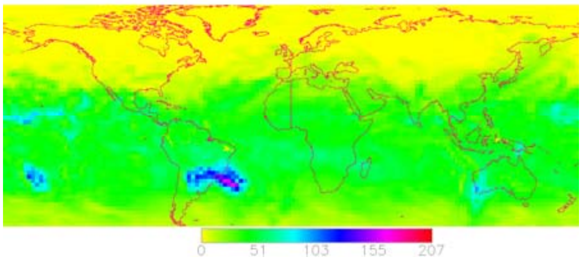

Fig. 3 Spatial distribution of daily evapotranspiration from Penman-modified method and GCM daily data, 0.5 degree resolution. Units in 0.1 mm.

This assumption is supported by the work of Leopold *et al*. (1964). The meaning of the TOPMODEL velocity parameters was modified, instead of representing velocities, the parameters represent coefficients in the following equation:

$$v_K = V_{CH} A_K^{V_R} \tag{10}$$

where v_K [LT^{-1}] is the velocity of the area-distance class K; A_K [L^2]is the cumulative area of the area class K; V_R is a power law exponent [-]; V_{CH} is a proportionality constant [$L^{-1}T^{-1}$].

Introducing Eq. (10) into Eq. (9), a time-area function can be derived from a distance-area function applying the following equation:

$$tc_k = \sum_{k=1}^{N} \frac{l_k}{V_{CH} A_K^{V_R}} \tag{11}$$

Equation (11) tries to take into account the spatial distribution of velocities in a basin. Furthermore, this approach tries to distinguish velocities on the river channels and velocities on the hillslopes. According to Lazzaro (2008), these velocities can differ by orders of magnitude and have been recognized as a primary source of the overall variance of the hydrograph. However, Eq. (11), as well as Eq. (9), is temporally invariant, that is, the velocity never changes over the simulation time.

In order to use the spatial distribution of rainfall and evapotranspiration, the water balance calculation was modified. Instead of carrying out the water balance for a certain number of classes, defined previously using the topographic index (usually 20-30 classes in the original approach), the modified approach carries out the water balance for every cell in the basin. In this way, the overland flow can be routed cell-to-cell. However, the saturation deficits for every cell are updated over the simulation using an average deficit. In other words, there is no sub-surface flow from one cell to another cell.

Hydrograph simulations

A Monte Carlo procedure was used to find the best set of parameters from pre-defined ranges. The parameter values were spread according to a uniform distribution of probability. Nash-Sutcliffe coefficient (Nash & Sutcliffe, 1970) was chosen as an objective function to evaluate the stream-flow efficiency with a threshold of 0.2. In this way, it was carried out 1,000 simulations for each basin and all simulations with Nash-Sutcliffe values above or equal to 0.2 were selected to estimate uncertainty bounds.

A normal probability distribution was assumed to delineate the lower and upper uncertainty bounds. These bounds, in this work, encompass 90 percent of all discharges for each time step. Therefore the distance of 1.645 standard deviation from the mean was delimited.

The uncertainty bounds were delimited as a means to evaluate the model performance. Uncertainty limits are useful to identify errors in model structure or in input data.

4. Results and discussion

For the Amazon basin 81 iterations from 1000 iterations stayed above the threshold of 0.2. The Nash-Sutcliffe coefficient for the best set of parameters was 0.48. For the Brahmaputra basin, 90 iterations with Nash-Sutcliffe equal or above 0.2 threshold. For this basin the best set of parameters obtained Nash-Sutcliffe equal to 0.82. For the Yangtze basin after 1000 iterations, 456 iterations stayed above the threshold limit. The best simulation obtained a 0.78 Nash-Sutcliffe coefficient. Table 1 summarizes the results from the simulations.

Looking at Table 1, it is possible to realized that the saturated hydraulic transmissivity parameter (ln

T_0) were quite different among the basins. Also, the velocity parameters (V_{CH}, V_R). This might be explained as in the Yangtze basin there are many reservoirs, and as the model does not take this fact into account, velocities and transmissity try to represent the delay time imposed by the reservoirs. The Amazon basin is fair flat, and this might explain the lower value in the V_R parameter. Velocities values change slightly as the cumulative area increases.

Brahmaputra basin seems to be a intermediate case between the two another basins.

Using the calibrated velocity parameters, the spatial distribution of delay times can be determined, according to Eq. 11. Figs. 4 to 6 show the travel times for each basin. Observing the travel times, also is possible to see the reservoirs effect in the Yangtze basin, whose total travel time, which might be interpreted as the time of concentration, is higher than the another basins (19295 hours). Taking a look at the Figs. 4 to 6 one can see the effect of Eq. 11. Travel times along the river channels are quite different from those in the hillslopes.

Figures 7 to 9 show the hydrograph simulations for each basin. Through Figs. 7 to 9 is possible to notice that the modified TOPMODEL simulated the discharges for all basins in a satisfactory way.

Table 1 Simulation results.

Parameter	Basin		
	Amazon	Brahmaputra	Yangtze
ln T_0 ($\ln(m^2\ h^{-1})$)	26.53	0.46	-6.78
m (m)	0.06	0.02	0.08
S_{RMAX} (m)	0.0022	0.0015	0.0019
T_D ($h\ m^{-1}$)	109.7	1638.63	1224.52
V_{CH} ($m\ h^{-1}$)	357.66	50.96	5.08
V_R ($m\ h^{-1}$)	0.06	0.19	0.28
Nash-Sutcliffe	0.48	0.82	0.78

This model obtained 0.48, 0.82 and 0.78 Nash-Sutcliffe for the Amazon, Brahmaputra and Yangtze basins, respectively. These values, show a slight improvement compared to the results obtained by Silva *et al.* (2010) (0.39 for the Amazon basin and 0.65 for the Yangtze basin), despite the raising in the computational time.

It was noticed some discrepancies between observed discharge and simulated discharge.

It could be due to data errors and/or model assumptions.

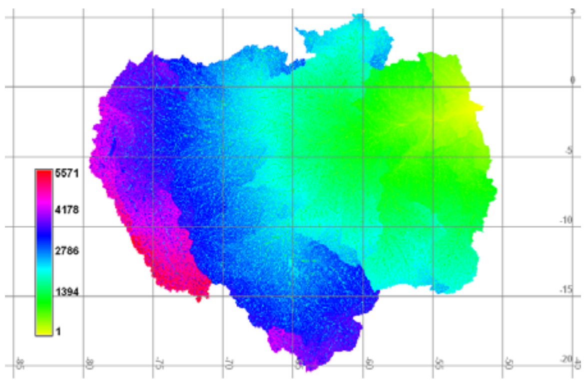

Fig. 4 Travel times in hours in the Amazon basin.

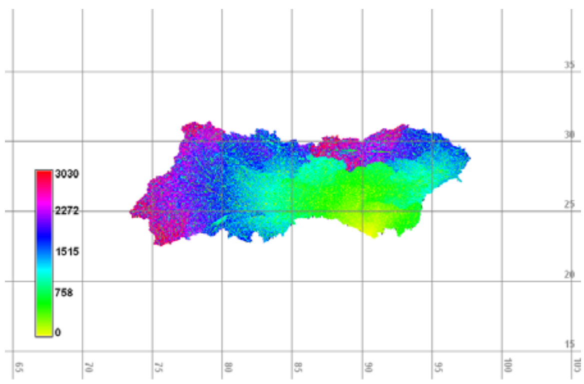

Fig. 5 Travel times in hours in the Brahmaputra basin.

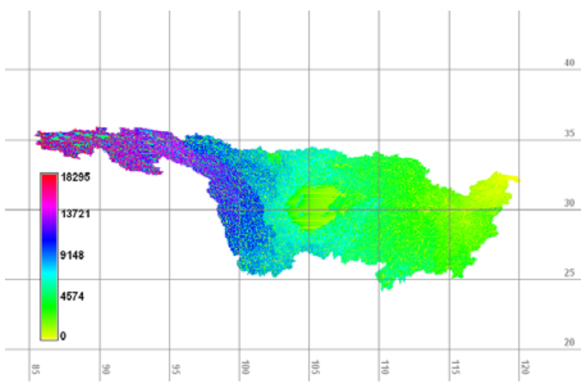

Fig. 6 Travel times in hours in the Yangtze basin.

Observing the uncertainty intervals in the Figs. 7 to 9, it is noticed that the model for the Brahmaputra and the Yangtze basins involved the observed discharges in a better way than in the Amazon river. It is clear that the model could not represent the observed data variance, even using distributed rainfall and evapotranspiration. This problem might be associated to data errors or model limitations.

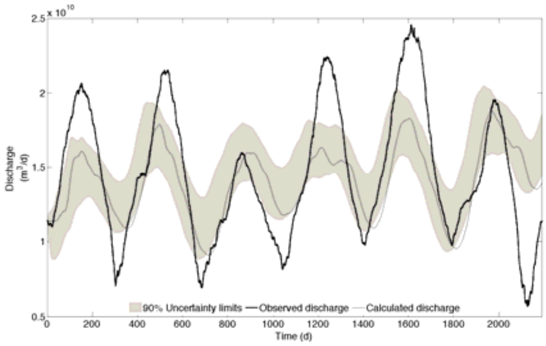

Fig. 7 Hydrograph simulation for the Amazon basin. Calibration period from 1990 to 1995. Observed data at Obidos station.

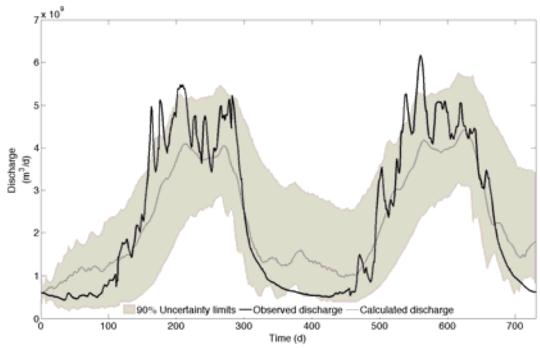

Fig. 8 Hydrograph simulation for the Brahmaputra basin. Calibration period from 1990 to 1991. Observed data at Goalundo station.

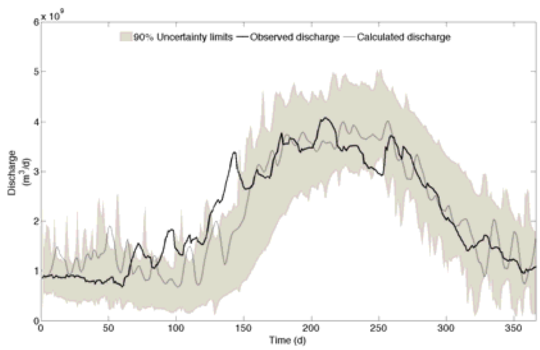

Fig. 9 Hydrograph simulation for the Yangtze basin. Calibration period for 2004. Observed data at Datong station.

For instance, a model limitation would be the model incapacity to model the water storage in the river channels.

5. Conclusions

In this study, three global basins were selected, according to daily data availability, to apply a modified distributed TOPMODEL approach. The velocity parameters in TOPMODEL were modified with the objective to give a more realistic representation of the velocities in the river basins, instead of using only one velocity parameter, which is a usual approach in RRM. In addition, a cell-to-cell approach was implemented to use the spatially distributed data. The method presented here implemented a power law relationship between cumulative area and velocity. Thus, the river velocities were spatially distributed. The global topographic data, basins boundaries and stream network, all them in GIS format, were acquired from the Geophysical National Data Center, the Global Runoff Data Center (GRDC) and the Global Drainage Basin Database (GDBD), respectively. The climate daily data were obtained from the National Centers for Environmental Prediction (NCEP). Monte Carlo simulations were used to find the best set of parameters in terms of the Nash-Sutcliffe coefficient.

The simulations showed that the modified TOPMODEL approach, using distributed rainfall and evapotranspiration seems to be a reasonable hydrological model to estimate stream flow discharges, despite the increasing in computational time compared to the original approach. Further studies should be carried out in order to identify source of errors and/or improve TOPMODEL approach for large-scale rainfall-runoff modeling through the implementation, for example, of a water storage river channel method.

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分布型TOPMODELを用いた大規模流域に対する降雨流出解析

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要　旨

本研究は修正分布型TOPMODELを３つの大流域（アマゾン川，ブラマプトラ川，揚子江)に適用し，その降雨追跡(流出)モデルとしての特性を評価するものである。修正分布型TOPMODELにおいては，従来のTOPMODELで一定値を利用していた流域内の流下流速を小流域の特性を反映して変化させ，セル毎の降水量と蒸発散量の変化を考慮し，最終的にモンテカルロ法で最適化し決定する。修正分布型TOPMODELではGRDCの日観測流量に対してナッシュ係数でそれぞれアマゾン川 0.48，ブラマプトラ川 0.82，揚子江 0.78 を得ており，本手法が信頼に足る方法であることを裏付けている。

キーワード：大規模流域，TOPMODEL，GCM，降雨追跡モデル，分布型流出モデル，分布型流速