Implementation of subgrid-scale spatial variability of parameters in a regional climate–hydrology coupled model

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Abstract Climate variability has an important impact on various hydrological processes. In order to represent small-scale hydrological processes on the climate and hydrology interactions, it is necessary to couple the regional climate model with a fine-scale hydrological model. In this paper, the parameterization of subgrid spatial variability is implemented in a coupled climate–hydrology model. The subgrid-scale spatial variability in hydraulic conductivity and precipitation is considered in the simulation to represent small-scale hydrological processes in the regional model. The results show that the simulated streamflow hydrograph at the basin outlet is consistent with the observed one. The incorporation of subgrid-scale spatial variability in hydraulic conductivity and precipitation could greatly improve the simulation. It is necessary to incorporate these subgrid hydrological processes to obtain realistic and accurate simulations of hydrological responses.

Key words climate–hydrology coupled model; subgrid; spatial variability; runoff generation mechanism

INTRODUCTION

Hydrological processes such as soil water movement, evaporation and runoff generation are directly influenced by the spatiotemporal distribution of rainfall, the variation of air temperature, humidity and wind velocity, as well as of solar radiation intensity induced by the climate change. In the form of evaporation, precipitation and runoff, the hydrological cycle is the major carrier for the energy and substance exchange in climate and hydrological systems (Liu, 1997). There is a close interrelationship between climate and hydrological systems. The integrated simulation of climatic and hydrological responses and parameterization of hydrological processes have become focal points of recent research (Xue *et al.*, 2005).

However, coupling hydrological and climate models poses new challenges for researchers (Liu, 2003). The first problem is the complex mechanism of rainfall–runoff generation. It is known that the infiltration excess (Horton) runoff and saturation excess (Dunne) runoff are two main mechanisms for rainfall–runoff generation. Generally, Horton runoff is the control mechanism in dry areas, while Dunne runoff occurs in wet areas. However, the Horton runoff can happen in wet areas during a high intensity rainstorm. Precipitation with a long duration can also produce saturation excess runoff in dry areas. The rainfall–runoff generation mechanism is controlled by the characteristics of landform, soil, geology and climate. There is still a lack of an effective method to describe hydrological processes such as soil water movement and evapotranspiration due to the high heterogeneity of land cover and soil texture in hydrological models. Traditional conceptual models, with features of strong reliability on data and indirect relationship with geographical features, are not suitable for coupling climate–hydrology models (Zeng *et al.*, 2003). Distributed hydrological models have become an important tool for the climatic and hydrological study of various processes (Beven & Kirkby, 1979; Paniconi & Wood, 1993; Ren & Liu, 2000; Walko *et al.*, 2000), but most distributed models still have deficiencies in characterizing non-homogeneity of rainfall–runoff generation and considering the flow concentration process in a grid or between grids (Yu & Schwartz, 1998). Secondly, the scale discrepancy between climate models of large scale and hydrological models of relatively smaller scale is still an issue for coupling two systems (Wood *et al.*, 1990). Hydrological processes such as rainfall, runoff, evapotranspiration and infiltration all show subgrid variability. So it is difficult to develop a parameterization scheme which can describe actual physical processes in large spatial grids. The previous studies show that streamflow would be underestimated without considering such a subgrid variability (Yu *et al.*, 2001). Further, research on the data decomposition in smaller time and space intervals is also an effective way to solve the problem of scale discrepancy between climate and hydrological models.

In this study, the Hydrologic Model System (HMS) (Yu *et al.*, 2002), which is physically based and linked with a regional climate model (RCM), is used for the simulation of various hydrological processes (Fig. 1). The method to describe the spatial variation of precipitation and hydraulic parameters is also implemented in the model (Genuchten, 1980; Thomas & Henderson, 1991).

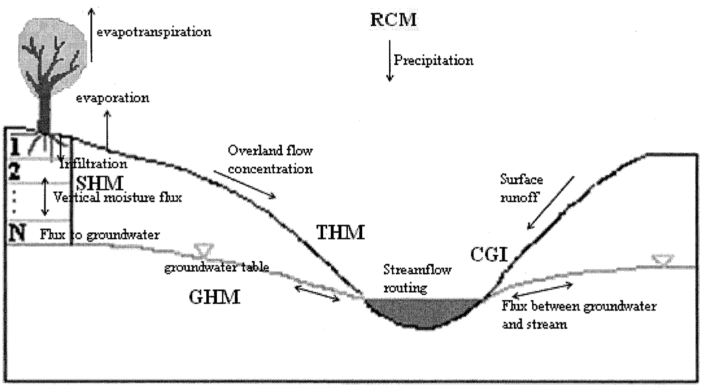

Fig. 1 Schematic map of climate and hydrology models.

METHODS

Climate and hydrology models

Regional climate change is greatly influenced by local characteristics, which are difficult to represent in the large-scale grid of global climate models. The regional climate model (RCM) used in the study has a finer resolution, which is adapted from MM5. RCM has the multi-level Blackadar-type planetary boundary layer parameterization and Grell cumulus parameterization (Grell, 1993). RCM models 20 layers distributed unevenly in a vertical direction from the surface to a height of 100 mb. The distance between two layers is smaller near the surface (Yu *et al.*, 1999, 2000). In this study, RCM is used in the non-hydrostatic mode and simulate precipitation per hour in the resolution of 4 km.

The Hydrologic Model System (HMS) is a physically-based, parameter distributed model system. It includes four submodels, a Soil Hydrologic Model (SHM), a Terrestrial Hydrologic Model (THM), a Ground-Water Hydrologic Model (GHM), and a Channel Ground-Water Interaction Model (CGI) (Yu & Schwartz, 1998). In SHM, based on the water storage including precipitation and flow from neighboring grids, the Richards equation is used to simulate the vertical soil water flow in the unsaturated zone, which is solved with the Crank-Nicholson method numerically. The evapotranspiration is calculated using the Penman-Monteith method. The Green-Ampt method is implemented to partition rainfall into the infiltration and surface runoff. The spatial distribution of soil moisture content simulated with SHM is used for the calculation of other hydrological components in every time step. In THM, the kinematic wave algorithm is used to model the overland flow based on the flow direction derived from DEM. The Muskingum-Cunge procedure is used to perform the channel routing from grid cell to grid cell. Through these flow simulations, the flow through each grid is routed to the basin outlet. The second order partial differential equation is applied to describe groundwater flow in the saturated zone, which is solved with a finite difference method in GHM. Darcy's law is used in CGI to calculate the interactive flux between stream and groundwater based on the hydraulic gradient between groundwater level and channel water level in the stream cell.

HMS uses the remotely sensed data and digital data as the parameters in the model system and they have physical meaning. The four submodels in HMS are interrelated to describe actual physical processes. For example, the interaction of groundwater and streamflow are considered and the loss and acquisition of water in a grid cell or between grid cells can also be described in the model.

Characterizing the subgrid scale spatial variability

As stated in the previous section, it is important to apply subgrid-scale spatial variability in climate and distributed hydrological models, not only for better coupling two systems, but also for improving the forecasting precision of hydrological processes (i.e. soil moisture, runoff). The hydrological responses strongly rely on the spatial variation of precipitation and hydraulic characteristics of soil (Pitman *et al.*, 1990). The precipitation simulated with the RCM is the spatially-averaged value in a large-scale grid cell, and this averaging may lower the actual rainfall intensity. In the Horton method, the rainfall intensity plays an important role in runoff generation and leads to the underestimation of runoff. In this study, the probability density distribution is implemented to represent the first-order approximation of the spatial variation of precipitation in each RCM grid cell. Based on the observed data, the exponential distribution is used to downscale the simulated precipitation in each RCM grid cell (at a resolution of 4 km) into hydrological grid cells (at a resolution of 1 km), which fall into each RCM grid cell (Yu *et al.*, 2000):

$$f(p_i) = \frac{c}{P}\exp(-\frac{cp_i}{P}), \quad \int_0^\infty f(p_i)dp_i = 1 \tag{1}$$

where f is the area fraction of a RCM grid cell with precipitation p_i, P is the precipitation simulated by RCM, p_i is the downscaled precipitation, c is the conversion coefficient which range from 0.3 for convective rainfall to 1.0 for large scale rainfall (Thomas & Henderson, 1991). The downscaled precipitation for a RCM grid cell is assigned to every corresponding hydrological grid cell randomly.

With the same method, the exponential distribution is also used to downscale the hydraulic conductivity into subdivisions within each hydrological grid cell (Yu *et al.*, 2000):

$$f(k_i) = \frac{1}{K}\exp(-\frac{k_i}{K}), \quad \int_0^\infty f(k_i)dk_i = 1 \tag{2}$$

where f is the area fraction of a HMS grid cell with hydraulic conductivity k_i, K is the hydraulic conductivity in a hydrological grid cell, k_i is the downscaled hydraulic conductivity.

With this method, the precipitation in each RCM grid cell is disaggregated to hydrological grid cells and the hydraulic conductivity in hydrological grid cells is downscaled into more detailed hydrological subdivisions. The hydrological simulation is conducted in each subdivision. The runoff and infiltration in subdivisions within a hydrological grid cell are summed for the calculation of water flow and water balance.

APPLICATION

Study area

The upper West Branch watershed with a drainage area of 14 710 km^2 is a subbasin of the Susquehanna River basin in Pennsylvania, USA (Fig. 2). Most of the upper West Branch watershed lies within the Appalachian Plateau physiographic province, which is of broad upland plain separated by steep-walled, narrow valleys. The southern portion of the upper West Branch watershed lies within the Valley Ridge province of the Appalachian, which has long, low, even-crested ridges with elevation of approx. 400–500 m a.m.s.l (Fig. 3). The land cover of the upper West Branch watershed is predominately forest, especially in the northern and western of the basin. Agricultural and urban lands are primarily distributed in the eastern and southern parts of the watershed.

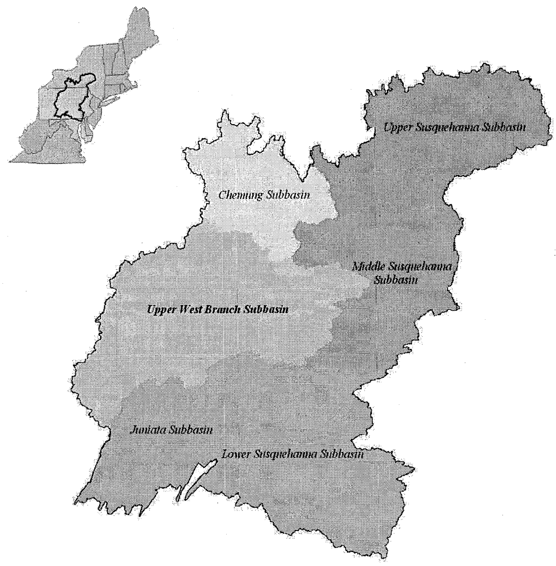

Fig. 2 Location map of the study area.

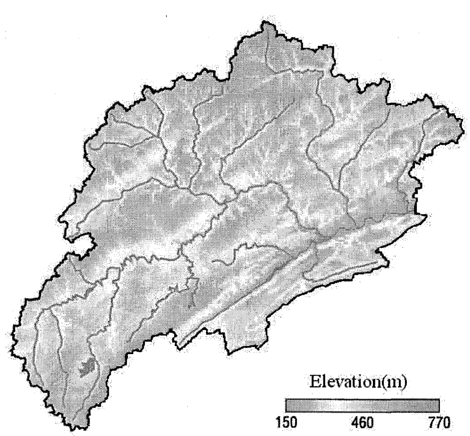

Fig. 3 DEM of the upper West Branch watershed.

Data processing

As a distributed hydrological model system, HMS uses remote sensed and other data sets as input data source. These image data have to be processed into digital ones which can be recognized by the model. After the correction of atmospheric attenuation, radiance and temperature values can be extracted from AVHRRR and Landsat TM images according to the grid resolution. The fractional vegetation cover, impervious surface area, evapotranspiration and surface soil moisture content can also be derived through processing these images. The land cover information is drawn based on basic land cover types. The spatial distribution of hydraulic parameters is obtained based on the soil data from the State Soil Geographic (STATSGO) and land cover information (Yu *et al.*, 2001). Assuming that hydraulic parameters are unchanged in a hydrological grid cell, values of two hydraulic parameters used in SHM, hydraulic conductivity and average capillary suction, are assigned into each hydrological grid cell with a resolution of 1 km.

A digital elevation model (DEM) is the basis for the distributed hydrological simulation and coupling of climate and hydrological systems. The DEM is generated from USGS 3-arc second data in the study. In order to match other data in HMS, the DEM is reprocessed into 1-km^2. The DEM is used to derive basin features such as grid slope, drainage area and flow direction, and to delineate the basin boundary and channel network in HMS.

Model calibration

Trial-and-error is still the most commonly used way to calibrate the model. In general, the calibration can then be classified into direct algorithm and indirect algorithm. The former uses the inverse operator while the latter adopts the iterative method. The calibration method for a distributed watershed model (Yu & Schwartz, 1999) was implemented in this study, which belongs to the indirect algorithm. Values of parameters in their own ranges are adjusted to optimize the model performance based on the multi-objectives such as soil moisture, groundwater level and discharge in the basin outlet. The model calibration has been conducted in previous studies (Yu *et al.*, 1999) for the upper West Branch watershed. The calibrated model can be applied to simulate hydrological responses in this watershed and describe the mechanisms and interrelationship of various hydrological processes.

RESULTS AND ANALYSIS

Influenced by the cyclone on 14 April 1986, the storm event began on 15 April and precipitation continued until 18 April. RCM is used to simulate the precipitation for this event. Using observed and simulated precipitation, HMS is run from 14 to 22 April with a time step of 10 minutes. The results are shown in Figs 4, 5 and 6. S1 and S2 represent the streamflow simulation with or without considering subgrid-scale variability of hydraulic conductivity with observed precipitation. S3 represents the simulation considering subgrid-scale variability of hydraulic conductivity with simulated precipitation. S4 represents the simulation considering subgrid-scale variability of hydraulic conductivity and precipitation with simulated precipitation.

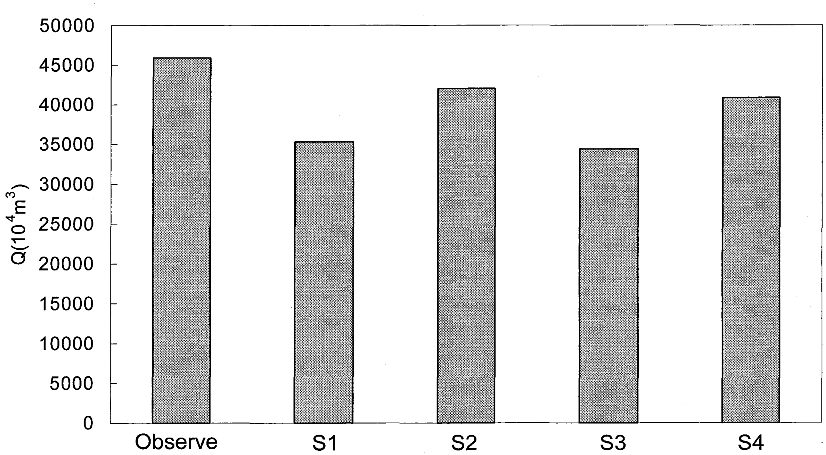

Fig. 4 Observed and simulated flood water volume.

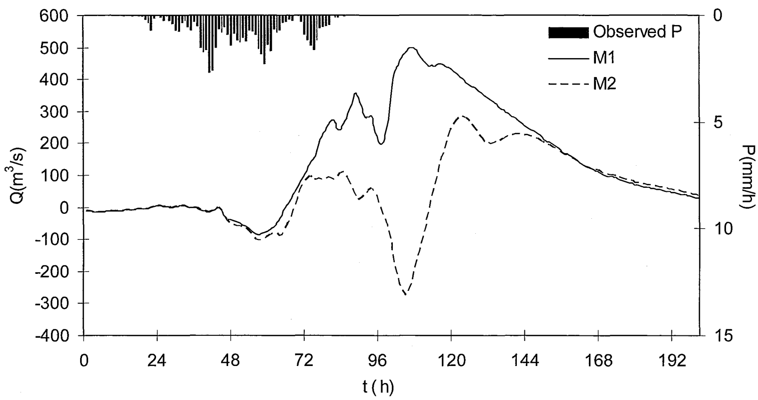

Fig. 5 Hydrographs of discharge differences between observed and simulated streamflow with observed precipitation.

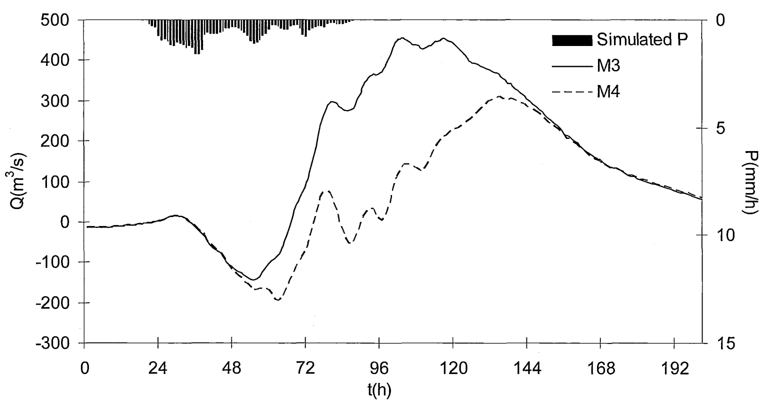

Fig. 6 Hydrographs of discharge differences between observed and simulated streamflow with simulated precipitation.

Comparing Figs 5 and 6, the precipitation simulated with RCM begins at the same time, with the observed one and the histogram shape for the simulated and observed precipitation being basically identical. However, the histogram of simulated rainfall is flatter than the observed, and the highest rainfall intensity generated with RCM is smaller than the observed. The reason for this phenomenon is that the grid cell size of RCM is so large that the simulated precipitation is the averaged value over grid cell and the spatial variability of rainfall is neglected. The spatial variation of rainfall could be large and should not be ignored, especially for the convective rainfall event. Therefore, this spatial averaging also causes the underestimation of precipitation.

In Fig. 4 it is obvious that the flood water volumes of S3 and S4 are less than those of S1 and S2, respectively, which is mainly caused by the underestimation of precipitation. With the observed precipitation and considering subgrid-scale variability of hydraulic conductivity, the discharge amount of S2 is still less than that observed. This is because this rainfall event is of long duration and low intensity so that the saturation excess runoff is the dominant runoff generation mechanism. According to previous studies (Dunne & Black, 1970), the expansion of the near stream area is the dominating mechanism of runoff generation in the study area. Through considering subgrid-scale variability of hydraulic conductivity, the HMS can describe the runoff generated with the expansion of saturated area implicitly, but because there is no explicit spatial description of this process in a grid cell, the streamflow is still underestimated.

For a better comparison, the hydrographs in Figs 5 and 6 show the differences between observed and simulated discharges. M1, M2, M3 and M4 represent the differences between observed discharge and discharge of S1, S2, S3 and S4, respectively. It can be seen that M1 shows poor performance because the spatial variability of hydraulic conductivity is not considered. With this variability, the performance of M2 improves a lot. Comparing hydrographs M3 and M4, it can be concluded that considering the subgrid-scale variability of precipitation can also enhance the performance for the streamflow simulations.

In Figs 5 and 6, there are relatively large differences between observed and simulated discharges in some sections of hydrographs, which may be primarily due to the DEM effect. DEM in a resolution of 1 km may ignore many terrain details, which causes the underestimation of discharge. It is also worth noticing that part of simulated values during the small and large discharges may be larger than observed ones, although most simulated discharge is less than the observed. On one hand, it is caused by the smoothing effect of flow calculations. On the other hand, not considering human activities (i.e. reservoirs) may also contribute to this problem.

CONCLUSION

Climate variability is the driving force for changes on various hydrological processes. Integrated study of climate and hydrological processes is necessary to further understand the interrelationship among climate, hydrology, ecology and human activities so that we can improve water resources planning and water quality management.

This paper describes how to downscale the precipitation generated with RCM and hydraulic parameters, which is not only necessary for better coupling of climate models with large-scale grids and hydrological models with small-scale grids, but also improves the description of runoff generation processes. Results show that the simulated precipitation is smaller than the observed in general due to the averaging effect of the large-scale grid in RCM. The spatial variability of precipitation and hydraulic parameters is important for the runoff generation process. When this spatial variability is described in the model, the simulated discharge hydrographs are improved.

Results also show that such small-scale processes as expansion of saturation area still can not be explicitly described in the model, although the discharge at the outlet has been greatly improved. So how to represent such processes in regional models while keeping their natural temporal and spatial scales is the next research point. Human activities should also be included in the model for more realistic and comprehensive simulations.

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