An Investigation of the Effect of Drainage Density on Hydrologic Response

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Abstract

The sensitivity of streamflow simulations to the drainage density of river basins was investigated. A physically based spatially distributed hydrologic model was used in the model experiments. The hydrologic model was applied to the Monongahela river basin in the United States of America for the simulation of 1988 and 1993 hydrologic regimes for selected periods between April and July. Model simulations of streamflows for 3 different drainage density scenarios (0.2, 0.24 and 0.38 km^{-1}) were compared against the observations. Evaluation of the model results indicated that the hydrologic model response changes significantly for the prescribed drainage densities. In general, the hydrologic model overestimated the stream discharges in response to an increase in drainage density. This outcome was attributed to an increase in the number of channel pixels, and thus an increase in the subsurface flow contribution to the total streamflow.

Key words: Drainage density, Hydrologic model, Streamflow, River basin.

Introduction

Drainage density, a fundamental concept in hydrologic analysis, is defined as the length of drainage per unit area. The term was first introduced by Horton (1932) and is determined by dividing the total length of streams within a drainage basin by the drainage area. A high drainage density reflects a highly dissected drainage basin with a relatively rapid hydrologic response to rainfall events, while a low drainage density means a poorly drained basin with a slow hydrologic response (Melton, 1957).

The objective of this study was to investigate the sensitivity of streamflow simulations to the drainage density of river basins. A physically based spatially distributed hydrologic model developed by Yildiz (2001) was applied to the Monongahela river basin in the USA for the simulation of 1988 (a dry year) and 1993 (a wet year) hydrologic regimes for a selected period between April and July. Three different stream network configurations, which actually yield 3 different drainage densities, were used for the model simulations of streamflow hydrographs. The effect of drainage density was evaluated through comparison of the model simulated streamflows against the observations on a daily basis at the outlet of the river basin.

The 1988 and 1993 extreme hydrologic regimes

The drought of 1988 and the flood of 1993 are among the most severe occurrences of climatic extremes in the continental United States during recent decades. The occurrence of these extremes has been linked to modifications in the general circulation induced by pronounced sea surface temperature (SST) anomalies in the tropical Pacific (Trenberth and Guillemot, 1996).

During the summer of 1988, a strong La Niña was underway, with below normal SSTs in the eastern tropical Pacific, while the summer of 1993 was characterized by conditions of a mature El Niño with pos-

itive SST anomalies over the same region. As noted by Trenberth and Guillemot (1996), these in turn affected the distribution of the extratropical jet stream and mid-latitude storm track, thus causing anomalous circulations over the continental United States. Severe drought conditions during the summer of 1988 afflicted much of the continental United States, especially the Great Plains and the Midwest. In the lower Mississippi Valley, rainfalls were at record lows from April through June 1988. The most intense period of drought and above normal atmospheric temperatures occurred in June 1988, but the conditions leading up to the spring-summer drought were in place as early as March. The heat waves that accompanied the dryness extended throughout the summer, although the weather patterns and rainfall returned to nearly normal in July.

The 1993 summer flooding in the Mississippi river basin was produced by one of the largest rainfall anomalies of the century. Heavy rainfall that persisted through June and July caused record high river levels in the central United States. The total rainfall over the summer period was twice as large as the normal value. During the spring of 1993, rainfall in the central United States was already above normal, and the soil moisture levels were near saturation. Therefore, this region was poised for potentially severe flooding prior to the onset of excessive and localized rainstorms at the beginning of June.

Hydrologic model description

Physically based spatially distributed hydrologic models have become an important tool for simulating the effects of spatial heterogeneities in watersheds by utilizing physical parameters that have physical significance and represent spatial variability. They can easily incorporate detailed information on topography, soil, vegetation, and climate from digital and remotely sensed data resources. During recent decades, several physically based distributed hydrologic models (Abbott *et al.*, 1986; Grayson *et al.*, 1992; Johnson and Miller, 1997; Biftu and Gan, 2001, among others) have been developed for various hydrologic applications in watersheds.

Due to the physical basis of the approach and the increasing availability of digital and remotely sensed spatial data, physically based distributed models have some advantages over simple lumped conceptual models. Historically, conventional lumped models generally have not incorporated spatially variable data including topography, soil and vegetation.

Further, their physical parameterizations are valid in small-scale homogeneous media, and thus they only can be an approximate representation of the hydrologic processes of a real landscape. Consequently, such models can not reproduce spatial heterogeneities in hydrologic system responses by using basin-averaged parameters (Abbott *et al.*, 1986; Beven, 1989).

Incorporating detailed information on climate, soil, vegetation, and digital elevation a physically based spatially distributed hydrologic model developed by Yildiz (2001) was used in the model experiments. With a simple, yet physically realistic representation of surface-subsurface flow interactions, the model couples an existing land surface model (Devonec and Barros, 2002) with a surface flow routing model and a lateral subsurface flow routing model (Figure 1). At the land-atmosphere interface water and energy fluxes in the vertical direction are calculated by the land surface model through the use of simplified conceptual descriptions of the physics, the so-called parameterization schemes. A vertical soil column is discretized into a number of layers with a thin superficial layer at the top to function as the interface between the ground and the atmosphere, and other deeper layers to store water and energy. The surface of the soil is subdivided into vegetation and bare soil areas.
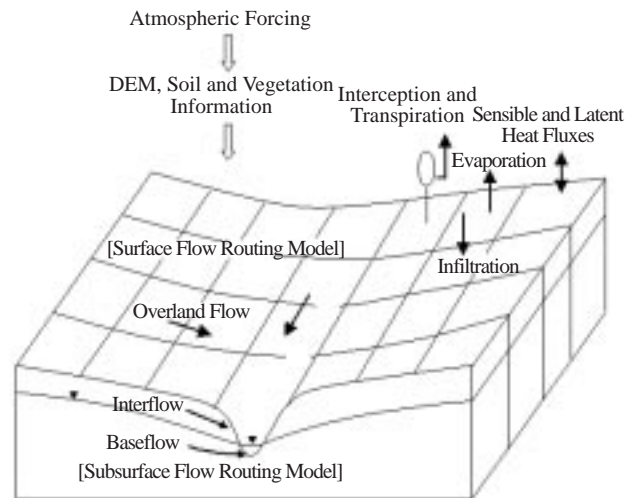

Figure 1. Structure of the hydrologic model.

Excess rainfall on the land surface is routed by the surface flow routing model, which relies on an algorithm using a single down-slope flow direction to facilitate the simulation of surface flow. Assuming a

linear flow surface across grid cells a one-dimensional kinematic wave approach is employed in overland routing to simulate the inflow and outflow discharges for each grid cell. A modified Muskingum-Cunge method of variable parameters developed by Ponce and Yevjevich (1978) is applied to route the water through the channel network to the basin outlet. Finite difference approximations were used in numerical solutions of routing equations and the time-step is adoptive, changing with hydraulic conditions on the hillslopes and in structures.

Subsurface flow (i.e. interflow and baseflow) is routed in the lateral directions by the subsurface flow routing model. A multicell approach proposed by Bear (1979) for aquifer systems was adopted for subsurface flow routing. Therefore, water balance equations are written for every grid cell and the system of equations is solved simultaneously for the entire aquifer system by finite difference approximations. Given the river stage in the channel, the flux between the channel and the ground water system is determined at the end of each time step. In the model, groundwater divides are assumed to correspond with the digital elevation model (DEM)-derived basin boundaries, and thus there is no interaction between the local and regional groundwater system. In addition, the water table is assumed to follow the topographic surface slope.

The current version of the model does not have a dynamic vegetation component but vegetation can be dynamically introduced into the model simulations through the adaptive assimilation of remotely sensed or digital data.

The stream network of the watershed is constructed from DEM using a threshold value of the flow contributing area and is optimized through the visual comparison with the actual stream network. Specifically, a pixel with a flow contributing value lower than the threshold value is treated as a plane pixel; otherwise it is treated as a channel pixel. The choice of threshold value is important in approximating the actual shape of the stream network as well as in obtaining accurate streamflow hydrographs. Several techniques are presented in the literature for simulating stream networks from DEMs (Montgomery and Foufoula-Georgiou, 1993). The most common technique is to choose an arbitrary threshold value on the basis of the visual similarity between the extracted network and topographic maps. The reader is referred to Yildiz (2001) for further details on the model's structure.

Model study area

The Monongahela river basin is located on the western slopes of the Appalachian Mountains (38.56N-40.47N, 79.07W-80.76W) with portions in Pennsylvania and West Virginia. The basin is a tributary of the Ohio river basin and has a drainage area of approximately 13,875 km^2 with outlet at Elizabeth, PA. The actual stream network includes the West Fork, Tygart Valley, Cheat, and Monongahela rivers and their tributaries.

As part of the Appalachian Plateau, the basin is characterized by strong spatial variability in the soil-terrain-hydrogeology system. Elevations in the basin range from about 400 to 1200 m, being greatest in the southern mountainous areas and lowest in the northern areas. At elevations above 400-500 m, the bedrock is highly dissected, and consists of sandstone with almost flat-lying layers of shale, clay, stone, and dense limestone. The soil layers above the bedrock are very thin, and thus most of the rainfall runs off the slope. The small amounts of water that infiltrate move vertically through fractures, and then move horizontally through sandstone or coal layers over large distances until they find another region of fractures, or an unconfined flow region such as colluvium and alluvium deposits. Accordingly, the base flow and interflow is very small during non-rainy periods in the warm season. At low elevations, productive unconsolidated alluvial aquifers ensure significant and sustained baseflow and interflow contributions during summer months (Trapp and Horn, 1997).

The vegetation cover in the watershed area also presents significant spatial variability with a predominance of deciduous trees at high altitudes and short grass and crops at low altitudes. A small fraction of the southeastern part is covered by coniferous trees, while a narrow band of bare ground can be found along the northeast-southwest direction.

The regional climate is humid to temperate, with topographic difference influences leading to local anomalies. The average annual temperature is about 9 °C. Mean monthly temperatures range from -2 to 22 °C. Average annual precipitation is 1067 mm and ranges from 940 mm in northern areas to 1524 mm in the southern mountainous areas. Precipitation during the winter is cyclonic in origin, whereas thunderstorms are responsible for most of the summer rainfall. The average annual runoff (1951-1980) ranges from 635 to 1016 mm in the mountainous southeastern areas and from 458 to 660 mm elsewhere. The

average annual recharge is estimated to range from 200 to 378 mm. The remainder of the average annual precipitation is estimated as evapotranspiration ranging from 90 to 410 mm across the north-south direction (McAuley, 1995).

Data description

Using 3-arc second DEM data (approximately 100 m) from the United States Geological Survey (USGS) watershed boundary delineation and stream network construction were performed at 1-km spatial resolution. Therefore, the original DEM data (i.e. 3-arc second) were aggregated into 1-km spatial scale.

The hydrologic model was driven by atmospheric forcing data including air temperature, pressure, humidity, wind velocity, and shortwave and longwave radiations obtained from regional climate forecasts. The data were produced by the National Center for Atmospheric Research (NCAR) Regional Climate Model (RegCM2) for spring and summer 1988 and 1993 periods over the Midwest United States. The climate model was driven at the lateral boundaries by European Center for Medium Range Forecast (ECMWF) data analyses and model outputs were produced at a temporal resolution of 6 h for the pressure and 3 h for the remaining data sets at 25-km spatial scale (Jenkins and Barron, 1997). The climate forecast data were downscaled from 25-km to 1-km spatial resolution with a bilinear interpolation scheme. The downscaled data were further linearly interpolated into a 1-h temporal scale.

Although the RegCM2 precipitation exhibited a close temporal correlation with the basin averaged observed precipitation, the climate model simulated excessive precipitation during the entire simulation period. Therefore, observed precipitation of 14 point measurements within the basin for two 5-month periods between April and August at an hourly time step were used in model simulations. Spatially distributed precipitation over the entire river basin was obtained by interpolating techniques using a modified Thiessen polygon approach in which each Thiessen polygon is represented by a raingauge, and thus, at a given time step, rainfall is uniform over a Thiessen polygon but spatially variable over the entire river basin. The standard Thiessen polygon method was modified in order to include orographic precipitation effects, especially during the spring months.

The physically based model parameters were de-rived from the ancillary data using digital and remotely sensed data resources. Specifically, soil parameters including hydraulic conductivity, porosity, field capacity and wilting point were obtained from the STATSGO data base, which was designed primarily for regional, multi-county, river basin, state, and multi-state resource planning, management, and monitoring (USDA, 1995). The dominant soil texture in the basin was silt loam, while loam and sandy loam were found scattered across the river basin, especially in the south.

Vegetation was included dynamically in the hydrologic model utilizing time-series of remotely sensed data. Vegetation characteristics including leaf area index (LAI) and fractional vegetation coverage (F_r) were estimated by parameterizations (LAI: Choudhury et al., 1994; F_r: Carlson and Ripley, 1997) using normalized vegetation difference index (NDVI) data from the Advanced Very High Resolution Radiometer (AVHRR). Given the soil and vegetation information, the other model parameters were selected from the literature (albedo: Dingman, 1994; roughness length and minimum stomatal resistance: Dickinson et al., 1993; Manning's roughness coefficients: Chow, 1959).

Streamflow simulations of the 1988 and 1993 hydrologic regimes

Hydrologic model simulations of streamflow hydrographs in the basin were performed at 1-km spatial scale at an hourly time step for selected spring and summer periods. As depicted in Figure 2, 3 different stream network configurations were developed for both 1988 and 1993 simulations using flow contributing threshold values of 25, 15 and 5 km^2, which produced drainage densities of 0.2, 0.24 and 0.38 km^{-1}, respectively. The simulated streamflows were compared against the observations at the outlet of the river basin on a daily basis.

For both 1988 and 1993 simulations, the soil column was assumed fully saturated at the beginning of the simulation period. Soil moisture can be updated as new soil moisture information becomes available. The hydrologic model was initialized for a period of 1 month (spin-up period) at the beginning of the simulation in order to allow the state variables to reach equilibrium conditions. The model was not calibrated; that is, the physically based model parameters extracted from the ancillary data were not submitted to optimization, because the simulation

years of 1988 (a dry year) and 1993 (a wet year) represented 2 extreme hydrologic regimes. Bindlish and Barros (2000) showed that the calibration of model parameters is particularly sensitive to the underlying climate regime, and thus calibration does not lead to an improved model response.

Using the fractional factorial design method (Box *et al.*, 1978) sensitivity testing of the hydrologic model to selected model parameters listed in Table 1 showed that the impact of vegetation is significant on the hydrology of the Monongahela river basin to different hydroclimatological extremes. Evaluation of the model sensitivity analysis in 1988 and 1993 showed that the model's sensitivity to model parameters changes as the climate regime changes. It also showed that spatial and temporal variability can affect sensitivity significantly. During the dry year of 1988, vegetation properties of fractional vegetation coverage and leaf area index had significant effects on model results, suggesting that hydrologic processes of evaporation and transpiration are expected to play an important role in such a dry climate regime. During the wet year of 1993, in addition to the vegetation parameters of Fr and LAI, soil hydraulic conductivity was of primary importance due to higher soil water availability (Yildiz, 2001).
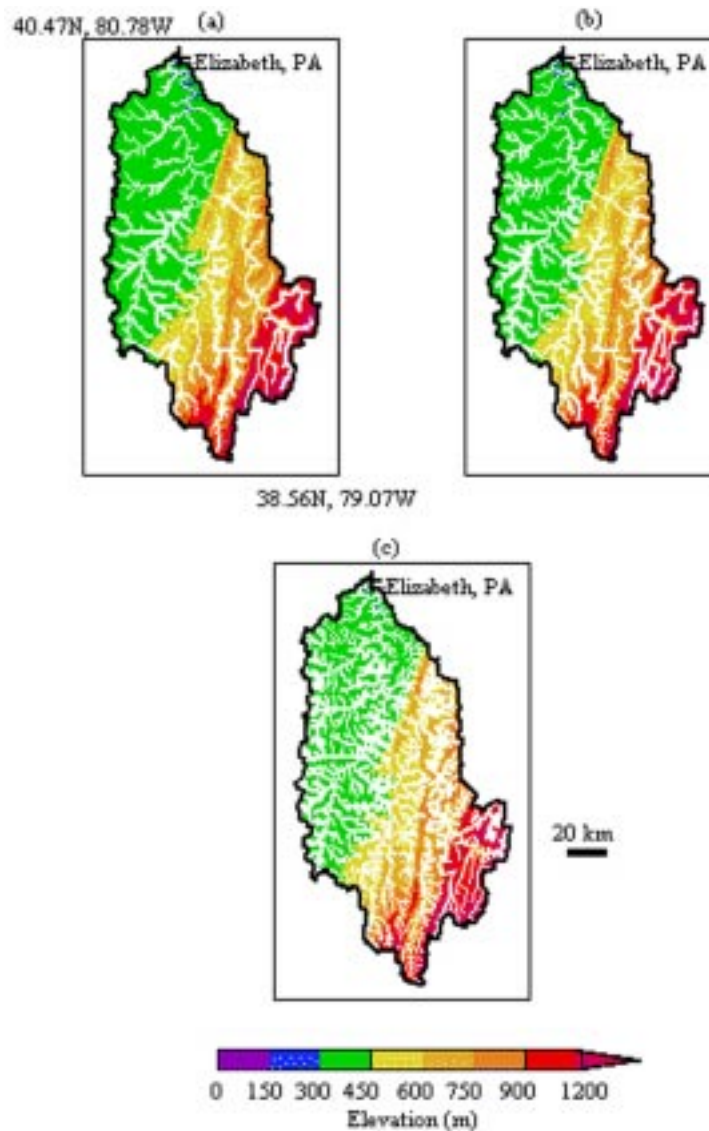

Figure 2. Comparison of the stream networks for the threshold values of the flow contributing area used in the delineation of the stream network: (a) 25 km^2, (b) 15 km^2, and (c) 5 km^2.
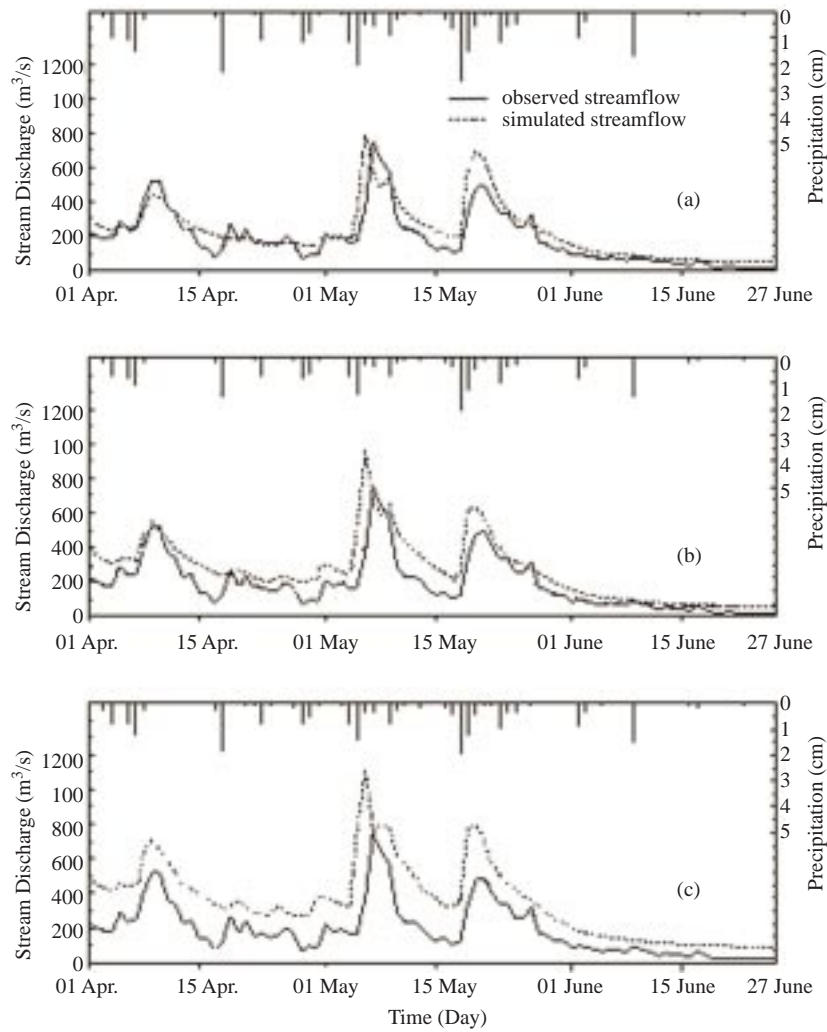
Figure 3. Observed and simulated streamflow hydrographs at Elizabeth in 1988 for the threshold values of the flow contributing area used in the delineation of the stream network: (a) 25 km^2, (b) 15 km^2, and (c) 5 km^2.

Table 1. Selected model parameters for the model sensitivity analysis.

Parameter Name	Classification
Leaf area index	Land use/Land cover
Fractional vegetation coverage	Land use/Land cover
Root depth	Vegetation
Minimum stomatal resistance	Vegetation
Albedo	Land use/Land cover
Roughness length	Land use/Land cover
Soil field capacity	Soil hydraulics
Soil wilting point	Soil hydraulics
Hydraulic conductivity	Soil hydraulics

Discussion of Results and Conclusions

The 1988 and 1993 model simulations of streamflow hydrographs performed for the prescribed stream network configurations along with the observations are shown in Figures 3 and 4, respectively. Comparison of the simulated and observed streamflow hydrographs clearly reveals that the hydrologic model response changes significantly for drainage densities of 0.2, 0.24 and 0.38 km^{-1}. In both years, the simulated streamflows steadily increased as the drainage density of the river basin increased. The hydrologic model generally overestimated the stream discharges in response to an increase in drainage density, producing relatively higher streamflow statistics

of mean, standard deviation, root mean square error and bias, but relatively lower coefficients of variation (Table 2). As shown in the figures, substantial differences between the observed and simulated peak flows, especially during the spring season, were obtained as a result of a drainage density increase.

This outcome can be attributed to an increase in the number of channel pixels, and thus an increase in the subsurface flow contribution (i.e. interflow and baseflow combined) to the total streamflow. In fact, comparison of the ratio of the model simulated subsurface flow to the total streamflow for the prescribed stream network configurations indicated that an increase in drainage density resulted in an increase in subsurface flow response (Figure 5).
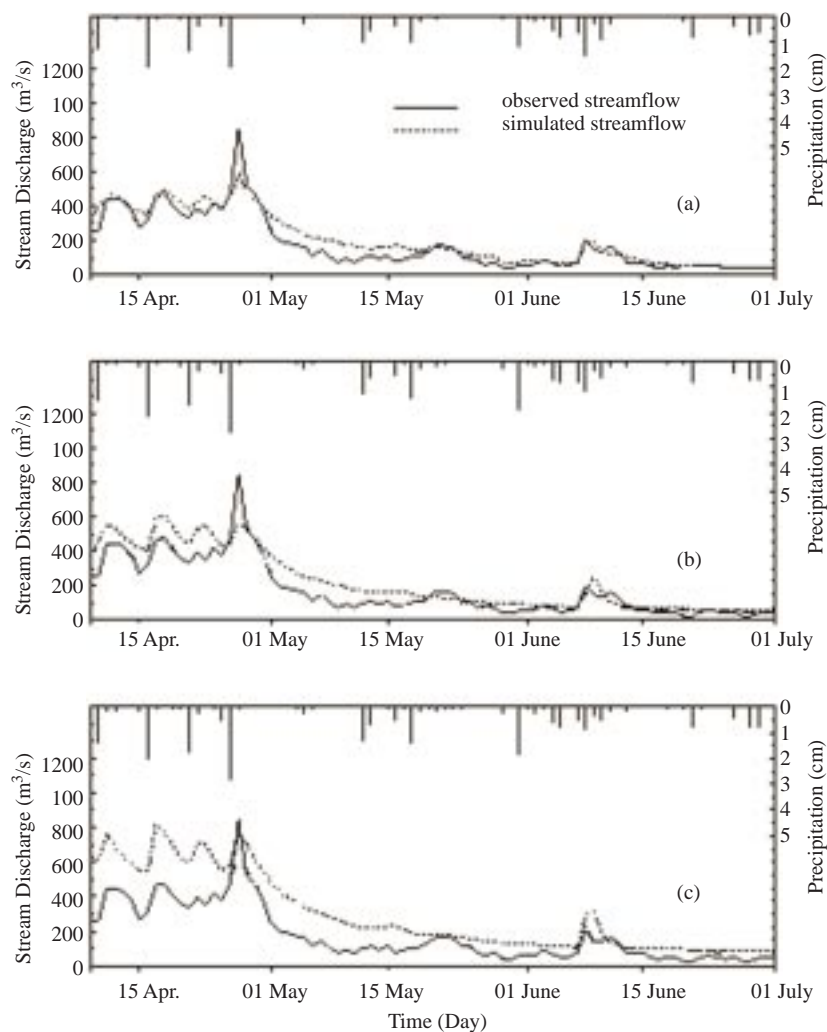

Figure 4. Observed and the simulated streamflow hydrographs at Elizabeth in 1993 for the threshold values of the flow contributing area used in the delineation of the stream network: (a) 25 km^2, (b) 15 km^2, and (c) 5 km^2.

(01 April-27 June 1988)
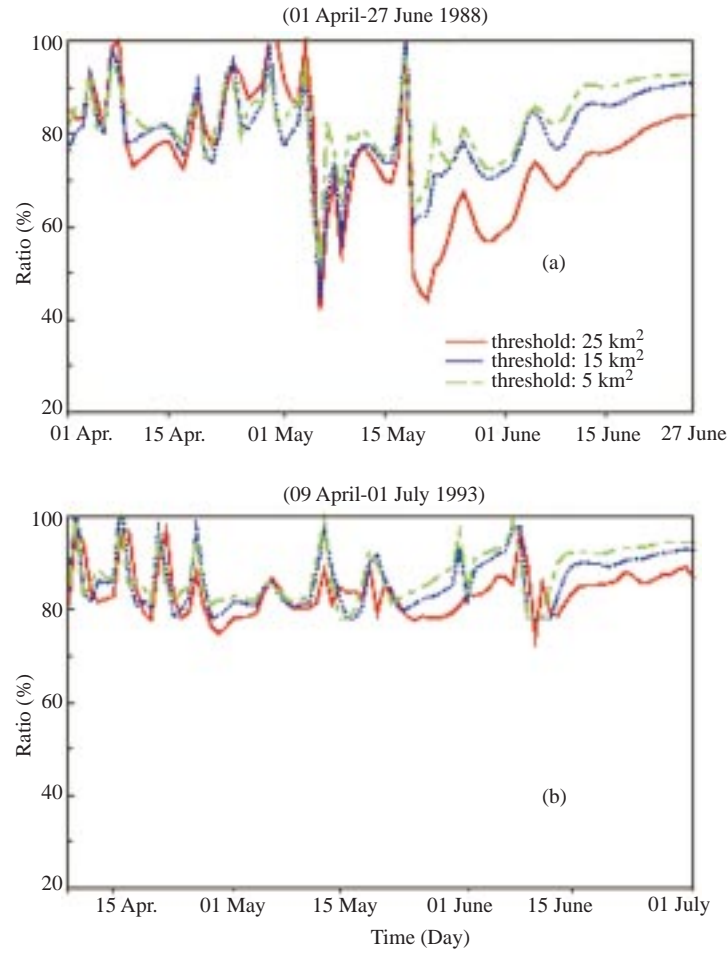
(09 April-01 July 1993)

Figure 5. Comparison of the ratio of subsurface flow to streamflow for the threshold values of the flow contributing area of 25, 15, and 5 km^2 in (a) 1988 and (b) 1993.

Table 2. Streamflow statistics of the observed and simulated streamflow hydrographs for flow contributing threshold values of 25, 15, and 5 km^2.

	1988				**1993**			
	Obs.	25 km^2	15 km^2	5 km^2	**Obs.**	25 km^2	15 km^2	5 km^2
Mean[1]	**143.5**	174.2	205.4	278.3	**140.5**	172.5	196.6	252.3
Std. Dev.[2]	**82.4**	85.3	86.7	99.5	**100.5**	91.5	102.5	129.7
CV[3]	**0.57**	0.49	0.42	0.36	**0.72**	0.53	0.52	0.51
RMSE[4]		67.5	84.5	151.5		57.5	68.4	127.4
Bias[5]		30.5	62.5	135.5		33.1	44.5	110.5

[1] Arithmetic average

[2] Standard deviation

[3] Coefficient of variation (Std. Dev./Mean)

[4] Root mean square error defined by $RMSE = \left(\frac{1}{n} \sum\limits_{i=1}^{n} [Q_s(i) - Q_o(i)]^2 \right)^{1/2}$

[5] Bias defined by $Bias = \left(\frac{1}{n} \sum\limits_{i=1}^{n} Q_s(i) - \frac{1}{n} \sum\limits_{i=1}^{n} Q_o(i) \right)$ in which $Q_s(i)$ and $Q_o(i)$ are the simulated and observed streamflow rates respectively, and n is the number of items of data.

This simple test of drainage density effect on hydrologic response indicated the importance of stream network construction for use in hydrologic model simulations of river basins. The evaluation of the model results suggested that drainage density can significantly affect model response, and that it can be a source of ambiguity in model calibration.

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