 Predictions in Ungauged Basins: PUB Kick-off (Proceedings of the PUB Kick-off meeting held in Brasilia, 20–22 November 2002). IAHS Publ. 309, 2007.

Runoff and sediment yield predictions in a semiarid region of Brazil using SHETRAN

EDUARDO E. DE FIGUEIREDO[1] **& J. C. BATHURST**[2]

1 *Área de Engenharia de Recursos Hídricos, DEC/CCT/UFCG, PO Box 505, 58109-970 Campina Grande, PB-Brazil*
eduardo@dec.ufpb.br

2 *Water Resource Systems Research Laboratory, Department of Civil Engineering, University of Newcastle, Newcastle upon Tyne NE1 7RU, UK*

Abstract Physically-based models are, in principle, capable of simulating ungauged basins using data of short length. This paper discusses runoff and sediment yield modelling using the physically-based and distributed model SHETRAN. The model was utilized to predict the runoff and sediment yields in the semiarid region of the Northeast of Brazil (NeB). The study was carried out using data observed at various basin scales within the Representative Basin of Sumé (RBS; 137.4 km^2) in the semiarid region of Paraíba state, Brazil, ranging from plots (100 m^2), through micro basins (0.5–1.0 ha), to sub-basins (10.0–140.0 km^2). Model parameters were evaluated using field data and techniques based on soil texture. The evaluated parameters were sufficient to represent the characteristics of the region. The results achieved at every basin scale and grid size, and at different time resolutions (daily, monthly and annual) showed that observed runoff and sediment yields were simulated with physically meaningful results. Scale effects on model parameters were not significant, although the Manning-Strickler coefficient changed with basin size. Land-use change effects on simulated runoff and sediment yields were considerable. For the homogeneous area of the RBS, the parameters tested for the plot scale could be used to simulate larger areas with different grid sizes, opening a possibility of simulating relevant processes for ungauged basins, an important requirement in vast areas such as the NeB.

Keywords micobasins; plots; runoff; sediment yield; SHETRAN

INTRODUCTION

In the last two decades, physically-based distributed models have been used increasingly to investigate land-use change and climate change impacts on water and sediment yields (e.g. Grayson *et al.*, 1992a; Bathurst *et al.*, 1996). However, several studies have already expressed concern with regard to the reality of the concept involved in designing these models (e.g. Grayson *et al.*, 1992b), principally the prediction of ungauged basins (Storm & Refsgaard, 1996), the effects of scale on model parameters (Beven, 1995) and uncertainty in model results (Melching, 1995). Thus, applications of these types of models (e.g. Connoly & Silburn, 1995) are still necessary for investigations on land-use and climate change impacts, parameter scale dependency and uncertainty so as to form a sound background. In this study, SHETRAN (Ewen, 2000), an enhanced version of the SHE/SHESED model (Abbott *et al.*, 1986a,b; Bathurst *et al.*, 1995), was applied for various watersheds (plot, micro-basin, sub-basin and basin) in the semiarid Northeast of Brazil (NeB), in order to investigate the model's capability of simulating runoff and sediment yields using estimated parameters based on field data and soil texture, the effects of model grid size on parameter values, the transfer of parameter values across scales, which are important requirements to bridge confidence in simulating ungauged areas, and the effects of land-use changes on runoff and sediment yields.

SHORT DESCRIPTION OF SHETRAN

SHETRAN is a physically-based spatially distributed model that simulates the major processes of the land phase of the hydrological cycle. An orthogonal grid network represents spatial distribution. Interception is calculated using the Rutter (1971/1972) equation which depends on ground cover areas, canopy drainage and storage capacity. Actual evapotranspiration can be calculated using either the Penman-Monteith equation or observed evaporation data (e.g. pan evaporation) as a function of soil tension (water potential; e.g. Feddes *et al*., 1976). Transpiration is accounted for by a root density function. Infiltration in the unsaturated zone (UZ) is calculated using the Richards equation, which is governed by the soil-hydraulic parameters (soil matrix potential and the hydraulic conductivity that is calculated using the equation of Brooks & Corey, 1964). The process governs the generation of runoff and is coupled to the saturated zone (SZ). Overland flow (runoff) can be generated by excess of rainfall over infiltration capacity (Horton flow) or by excess of saturation (Dunne flow). The Saint Venant equations are used for routing overland flow, with flow velocity determined by the Manning equation. Soil erosion consists of soil detachment (by rainfall and runoff) and transport. Erosion by rainfall is calculated based on the momentum squared of raindrops, and by overland and channel flow based on the equation of Shields. Transport capacity can be calculated using the equations of Yalin (1963) or Engelund-Hansen (1967). The model equations are not presented herein, but they can be found elsewhere (e.g. Bathurst *et al*., 1995; Figueiredo, 1998).

STUDY AREA

The representative basin of Sumé (RBS, 137.4 km^2) is the study area, located in the semiarid NeB. It is situated between 7°40′–7°50′S and 37°10′–36°55′W and has two internal sub basins, Umburana (10.7 km^2) and Jatobá (26.8 km^2). The basins were instrumented for determining the water discharge, and were monitored from 1975 to 1980. Break-point precipitation for 1977 and daily data for 1975–1980 are available. The climate in the RBS is semiarid. Droughts are common and precipitation is concentrated in three to four months. Mean annual precipitation and evaporation are 600 mm and 2500 mm respectively. The dominant vegetation cover (known as caatinga) consists of grass, small sized trees having small leaves, bushes with a shallow root system and open spaces.

Within the Umburana sub basin, four micro basins (0.5–1.0 ha, ~7% slope) and nine plots (100 m^2, 4–9% slope), with different surface conditions, were equipped for measuring the water-sediment mixture. They were operated for 10 years (1982–1991).

Break-point precipitation and daily precipitation, pan evaporation, flow and sediment yield data are available for 1982–1988. Soils are predominantly brown non-calcic (80%) and red-yellow podzol (20%). The brown non-calcic soil is shallow and has two horizons of low permeability, A (0.1 m; 0.31 m/day; 50.2% sand; 15.8% clay) and B (0.7 m; 0.03 m day^{-1}; 50.2% sand, 32.5% clay). The red yellow podzol is deeper and permeable (~2 m; 8 m day^{-1}). Details on the study area can be found in Cadier & Freitas (1982), Cadier *et al.* (1983), Cadier (1996) and Figueiredo (1998).

MODELLING APPROACH

The spatially distributed basis of the model required definition for the catchments grid resolutions (grid sizes of 5 m × 5 m to 2 km × 2 km were used, except at the plot scale for which a single rectangle of 22.22 m × 4.5 m was set) and parameter values (baseline) for the processes of interception, evapotranspiration, flow in the unsaturated (UZ) and saturated (SZ) zones, overland and channel flow, soil erosion and sediment transport. The model parameters were first tested for the plots, and then validated with the procedure proposed by Ewen & Parkin (1996). Simulations were realized first for the plots P1 and P4 (bare cleared), and P5 (vegetated), for testing the baseline values. Then, they were used to simulate the overland flow and sediment yields at the micro basins M1 and M2 (bare cleared) and M3 and M4 (vegetated). Parameters for the larger basins were defined for vegetated and non-vegetated areas based on an assumed vegetation distribution. The effects of land-use change on runoff and sediment yields were simulated based on assumed deforestation scenarios.

For the interception, ground proportions covered by the canopy were set to match 90% of the total area for plot P5 and micro basins M1 and M2, and 30–50% for the larger basins. The Rutter (1971/1972) parameters (canopy storage and drainage) were set based on information in Jetten (1996). Jetten defined values of between 0.3 and 1.4 mm for the canopy storage and proposed equations to estimate the drainage parameters. The values were set to 0.5 mm for the storage capacity, and 0.000017 mm s^{-1} and 7.77 for the drainage parameters. The height of vegetation and proportion of drainage as leaf drip were set to 1.0 m and 0.5. Pan evaporation data were used to estimate the potential evapotranspiration. The function between the ratio of actual to potential evapotranspiration (E_a/E_p) and soil tension (ψ) was fixed considering that, E_a = 0 when ψ is at its wilting point, and $E_a = E_p$ when ψ is at its field capacity. In between, values were set following Denmead & Shaw (1962). The root density function was fixed based on data reported in Cadier *et al*. (1983).

Parameters for the flow in the UZ such as the soil matrix potential (soil moisture, θ, at different soil tensions ψ) and the hydraulic conductivity, $K(\theta)$, were determined using the equations of Saxton *et al*. (1986), Rawls & Brakensiek (1989), Brooks & Corey (1964) and field data reported by Cadier & Freitas (1982). The estimated saturated hydraulic conductivity (K_s) and soil moisture (θ_s), adopted here, were 0.306 m day^{-1} and 0.448 m^3 m^{-3} (A horizon), 0.057 m/day and 0.488 m^3 m^{-3} (B horizon) and 8.0 m day^{-1} and 0.5 m^3 m^{-3} (red-yellow podzol). The exponent of the equation of Brooks & Corey (1964), known as the Averjanov (1950) exponent, for estimating $K(\theta)$, was determined based on soil characteristics (see Mualem, 1978).

Values of 15 (A and B horizons) and 3 (red-yellow podzol) were used. For the SZ flow, boundary conditions were set to avoid the groundwater discharging to the soil surface because in the study region overland flow is mainly Hortonian.

The Manning-Strickler coefficient (the reciprocal of the Mannings' roughness) is used for the overland and channel flow. For the overland flow it was set according to information in Engman (1986), Wicks *et al.* (1992) and Chow (1959). For the vegetated areas the coefficient was set to 1 $m^{1/3}$ s^{-1}. For a bare clay loam soil type, similar to those in the study region (bare plots), a value of 50 $m^{1/3}$ s^{-1} was used. For the bare micro basins, however, this value decreased to 40 $m^{1/3}s^{-1}$. For the larger basins the value decreased even more, changing to 25 $m^{1/3}$ s^{-1} for Umburana, and to 15 $m^{1/3}$ s^{-1} for the Jatobá and RBS basins. These changes were necessary to best fit the observed hydrographs. For the channels, a value of 30 $m^{1/3}$ s^{-1} was adopted.

Soil detachment by rainfall was modelled following prior experience reported in Wicks *et al.* (1992). Soil detachment by runoff was not allowed, because in the study region soil detachment is mainly provided by rainfall, which is generally intense, and overland flow is relatively shallow. The equations of Yalin (1963) and Engelund-Hansen (1967) were used for the sediment transport capacity in the surface and channel respectively. A soil size distribution was fixed based on data from the eroded material at sites. Soil porosity and bulk density were set to 0.448 and 1.46 kg m^{-3}.

RESULTS AND DISCUSSION

Parameter estimation results

Comparisons of estimates of moisture contents (θ), at different soil tensions (ψ), and hydraulic conductivity (K_s) using the equations of Saxton *et al.* (1986) and Rawls & Brakensiek (1989), with averaged observed values, using data reported in the literature (see Figueiredo, 1998), were generally reasonable, except for the residual moisture content (θ_r). Mean percentage errors, excluding θ_r, were in the range of –74% to –29%. The estimates for the residual moisture content were not as good as for the other moisture contents. However, of the parameters of the Brooks & Corey equation, θ_r is the least important and it was set to zero (see for example Saxton *et al.*, 1986). Parameters and functions were examined too via the model simulation results. Most of them were used unchanged for all basin scales, except the Manning-Strickler coefficient that changed as basin size increased.

Model simulation results

Figure 1 shows graphical plots. Table 1 shows r^2 (coefficient of determination) values.

In general, the results were fairly reasonable, except for the dry years. This may suggest an improved model response for wetter antecedent conditions (see Wicks *et al.*, 1992). For the water discharge, the simulations were quite good. For the sediment yield, they were, in many cases, underestimated, principally in dry years.

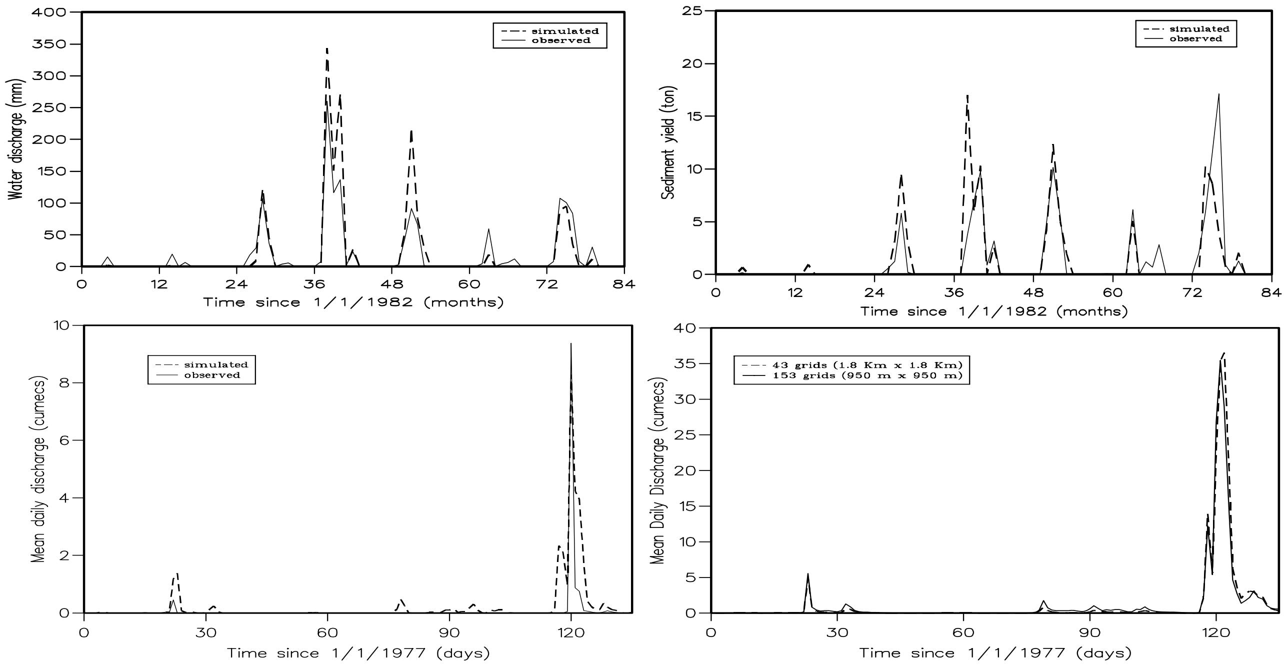

Fig. 1 Comparisons: Top: Pl (left), M3 (right); Bottom: Umburana (left), RBS (right).

Table 1 Coefficient of determination (r^2).

Site	Flow Daily	Monthly	Annual	Sediment Daily	Monthly	Annual
Plot P1[1]	0.63	0.86	0.86	0.19	0.43	0.85
Plot P4[1]	0.69	0.92	0.98	0.23	0.48	0.61
Plot P5[2]	0.58	–	–	–		
Micro basin M1[2]	0.49	–	–	0.44		
Micro basin M2[2]	0.34	–	–	0.10		
Micro basin M3[1]	0.60	0.89	0.94	0.35	0.42	0.69
Micro basin M4[1]	0.52	0.83	0.92	0.28	0.31	0.71
Umburana[3]	0.67	–	–	–	–	–
Jatobá[3]	0.76	–	–	–	–	–
RBS[3]	0.80	–	–	–	–	–

[1] Using data from 1982–1988; [2] using data of 1986; [3] using data of 1977.

Model validation

Model parameters were verified through using the containments (% of time observed values fall within output bounds) as a measure of reliability. The procedure by Ewen & Parkin (1996) was used to determine output bounds based on a set of simulations with parameter bound values (defined in this study according to field data and variations in soil texture). The containments were better for the plots and micro-basins (80%) than for the larger basins (10%). For the larger basins, however, the observed peak discharges were 100% contained and, additionally, some sub-hourly events at Umburana had more than 50% of its observed discharges contained by the output bounds. Figure 2 shows some bar diagrams showing the contained features.

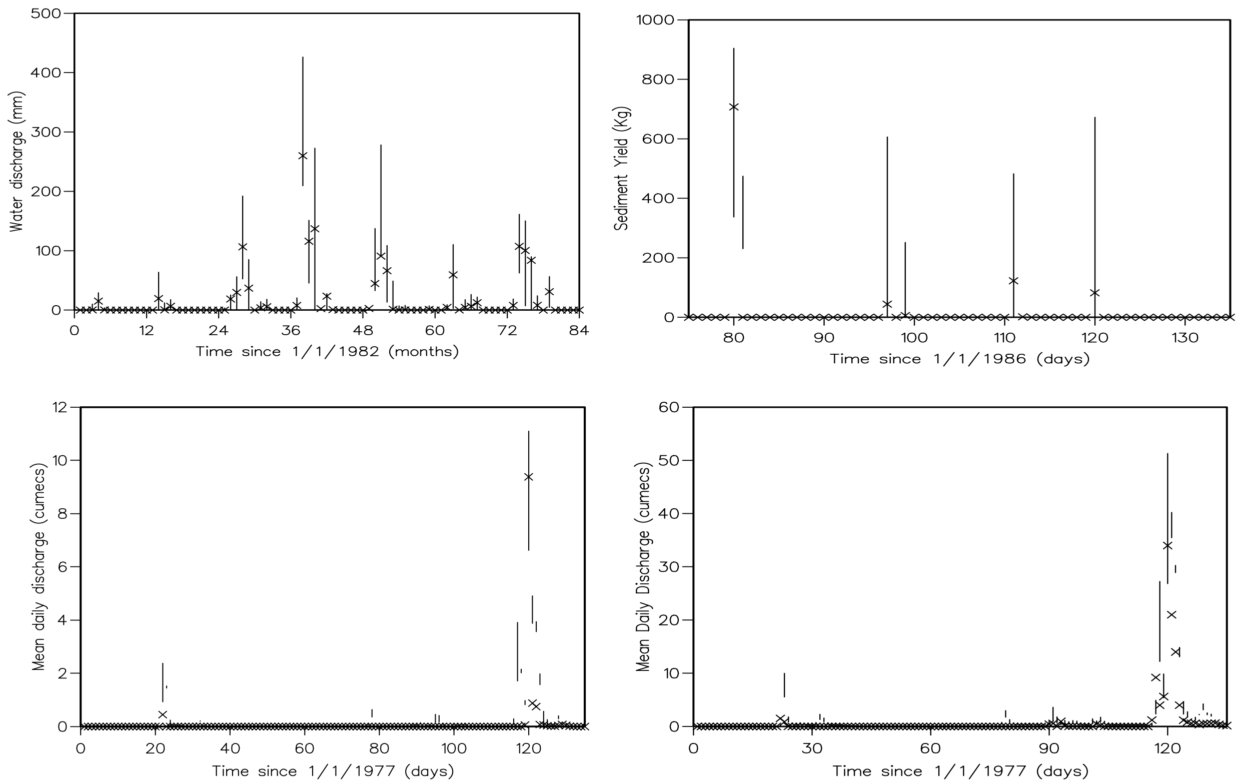

Fig. 2 Output bounds (bars) containing observed characteristics (X). Top: P1 (left), M2 (right); Bottom: Umburana (left), RBS (right)

Effects of scale on model parameters

Similar results were obtained with the various grid sizes used for a particular basin and, therefore, little parameter dependency on grid size could be observed (see the result for the RBS in Fig. 1). However, in transferring the model parameters across scales, from the plot, through the micro-basins, to the basin scales, the Manning-Strickler coefficient changed. This was necessary to best fit the observed discharges, and likely because of changes in surface cover conditions (relief, river network, vegetation cover, etc.), as discussed in the literature (see for example Chow, 1959; Shih & Rahi, 1982).

Effects of land-use change on runoff and sediment yield

Runoff and sediment yields were simulated for five scenarios of deforestation (10, 30, 50, 70 and 90% of basin area) and common annual rainfalls namely dry (400 mm), normal (600 mm) and wet (800 mm) in the study region. The simulated runoff (peak discharges and volumes) and sediment yields (see plots in Fig. 3) were both affected by the rainfall regime, land use, and basin area. Runoff increased while sediment yield (in t km^{-2} per year) decreased as the catchment area increased, suggesting the effects of spatial scale on these processes as reported in the literature (see Fournier, 1960; William *et al*., 1966; Walling & Kleo, 1979; Lal, 1993; Sahin & Hall, 1996).

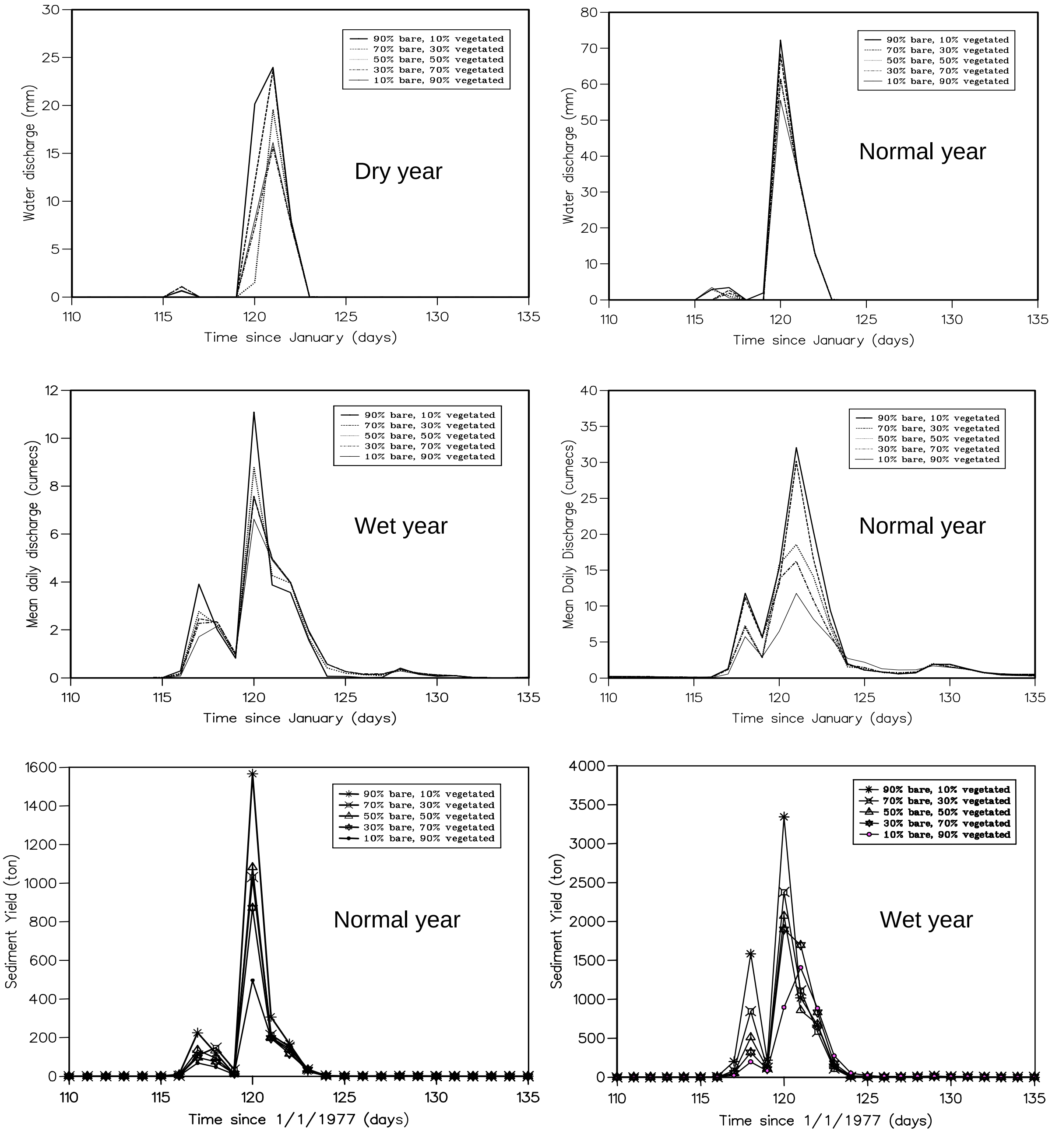

Fig. 3 Simulated discharge and sediment yield for different land uses and rainfall. Top: P4 (left), M4 (right); Middle and Bottom: Umburana (left), RBS (right).

CONCLUSIONS

The SHETRAN model was parameterized, using field data and functions that require little and cheaply obtained information, and applied to areas in the semiarid NeB. The model simulated the overland flow and sediment yields at different scales, without significant parameter change, with meaningful results.

The equations of Saxton *et al.* (1986) and Rawls & Brakensiek (1989) estimated the soil-moisture parameters (the soil matrix potential, hydraulic conductivity, etc.) reasonably well (mean error varied from –74 to –29%). Using the estimated parameters, the simulations results were quite good for the observed runoff (mean r^2 = 0.8) and sediment yield (mean r^2 = 0.46). The containments (overall mean 76%) of observed runoff (86%) and sediment yields (67%) suggest that the model parameters and their bound values can be used for simulating similar ungauged basins, an important requirement for the vast semiarid region of the NeB.

Effects of model grid size on model parameters were not significant. The Manning-Strickler roughness coefficient changed throughout the scales, decreasing as basin area increased (50–15 $m^{1/3} s^{-1}$), because of changes in the soil surface conditions, as discussed in the literature (see Chow, 1959; Shih & Rahi, 1982). The sediment yields decreased from 8–2053 t km^{-2} per year, for a year with normal rainfall, depending on the land use, as basin area increased, suggesting the effects of spatial scale on the soil erosion process, with results comparable with those reported in the literature (see Fournier, 1960; Walling & Kleo, 1979; Lal, 1993).

Peak discharges, volumes and sediment yields were markedly affected by land use and precipitation, increasing, for a year with normal rainfall, by 7–30% (peaks), 3–16% (volumes) and 26–84% (sediment), depending on the basin size, as deforestation increased from 10–90%.

Acknowledgements This research was funded by CNPq-Brazil.

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