Sediment yield modelling at micro-basin and basin scales in semi-arid regions of Brazil

E. E. DE FIGUEIREDO

Water Resources Research Engineering Area, Department of Civil Engineering, Federal University of Campina Grande, PO Box 505, 58100-970 Campina Grande/PB, Brazil

eduardo@dec.ufcg.edu.br

Abstract Four distributed models, two event-oriented, KINEROS2 and WESP, and two continuous models, SHETRAN and MOSESS_D, have been utilized to investigate their capability in simulating hortonian flows, typical of semi-arid environments, and soil erosion in northeast semi-arid areas of Brazil. Parameters of the models were first determined for vegetated and non-vegetated plots (100 m^2) of the experimental basin of Sumé (EBS) located within the representative basin of Sumé (RBS) in the state of Paraíba. For the event oriented models they were calibrated by comparing models results with observed values, while for the continuous models they were fixed, based on field data and used unchanged to simulate micro-basins (0.5–1.0 ha) and basin responses (10–140 km^2) of the RBS. The MOSESS_D model was also applied to one sub-basin (19 km^2) of the representative basin of Tauá (RBT), in the semi-arid region of the state of Ceará, as a validation test. In general, models results approached the observed values, which indicate that they would be useful in the semi-arid northeast Brazil. Good runoff results were obtained with the continuous models implying their suitability for simulating the dynamics of ephemeral rivers in the investigated areas, but the observed sediments were better approached by the event models. The sediment yields simulated with the continuous models decreased as basin area increased, at rates (1000–10 t km^{-2} $year^{-1}$) comparable with those in other river systems worldwide.

Key words flow dynamics; distributed model; runoff-erosion; semi-arid; scale effects

INTRODUCTION

Soil erosion is a complex process that depends on the dynamics of surface runoff. Modelling the runoff–erosion processes at different scales with physically-based models enables us to simulate semi-arid landscape dynamics (Zhang *et al*., 2007) from plots to catchment scales (Srinivasan *et al*., 2004; Figueiredo & Bathurst, 2005), and the related impacts of rainfall variability and land use on basin responses (Figueiredo & Bathurst, 2006, 2007). Large scales are heterogeneous and may affect simulated processes and model parameters, principally those that do not take into account antecedent conditions and need calibration, e.g. the event-oriented ones. Conversely, continuous models have the advantage of simulating processes, taking into account antecedent conditions (with parameters assessed from field data). These have recently been pointed out as important aspects in the evaluation of erosion processes through different scales (Gupta, 1986; Julien & Frenette, 1986; Walling, 1994). In this research, the physically-based event-oriented rainfall–runoff–erosion models KINEROS2 (Woolhiser *et al*., 1990) and WESP (Lopes, 1987), and the continuous models SHETRAN (Bathurst *et al*., 1995) and MOSESS_D (Sousa & Figueiredo, 2007) have been applied at various basin scales (plots, micro-basins and catchments) in the northeast semi-arid region of Brazil where the flow is of the Horton type. In this article, the modelling approach and the results are presented and discussed.

THE SIMULATION MODELS

The event-oriented models KINEROS and WESP

KINEROS and WESP are event-oriented physically-based models. Both of them consider the watershed as made up of planes and channels. Evapotranspiration is not modelled. Infiltration is calculated with the Smith and Parlange, and Green and Ampt equations. Runoff is routed using the kinetic wave approximation. The runoff lamina and load of sediment at the outlet of the watershed are determined for each event of rainfall. The soil erosion results from the rainfall impact (negligible in the channels) and shear stress of the surface runoff. Sediment transport in the WESP model is determined from a mass balance, while the KINEROS2 model takes into account the concept of transport capacity. The sediment flux is similar for both models.

The Smith and Parlange equation takes into account the saturated hydraulic conductivity of the soil K_s, a soil-type parameter α, and a factor that combines the effect of the capillary potential G, the depth of flow h, and the deficit of moisture content of the soil $\Delta\theta = (\theta_s - \theta_i)$ where θ_s is the volumetric water content at saturation, and θ_i is the initial water content of the soil. The equation of Green and Ampt accounts for the saturated hydraulic conductivity of the soil, the cumulative infiltration I, and the capillary potential of the soil $N_s = \Delta\theta$.G.

In both models, the flow component (planes and channels) is described by similar dynamic equations. Erosion on planes and channels in the KINEROS model is calculated as:

$$\frac{\partial(AC_s)}{\partial t} + \frac{\partial(QC_s)}{\partial x} - e(x,t) = q_s(x,t) \tag{1}$$

where A is the cross sectional area of the flow, C_s is the volumetric concentration of the sediment in the flow, Q is the discharge, e is the volumetric soil erosion rate per unit width, q_s is the lateral influx of the sediments per unit length of the channel, x denotes position, and t time.

The erosion rate e is the sum of the erosion resulting from the raindrop impact $e_s = c_f\, e^{-c_h h}\, i^2$ and the erosion or deposition $e_h = c_g(C_m - C_s)$ resulting from the effects of the flow shear stress and gravity, where c_f is a coefficient related to the characteristics of the surface soil (determined by calibration), i is the intensity of rainfall, $e^{-c_h h}$ is a function that represents reduction in the impact of rainfall by the surface runoff, c_h is a fixed parameter in the model, c_g is a sediment transfer coefficient, C_m is the maximum concentration at the equilibrium capacity of transport, $C_s = C_s(x, t)$ is the actual concentration of sediments in the flow that depends on the position x and time t, and $c_g = c_o(v_s/h)$ if $C_s \le C_m$ (erosion) or $c_g = v_s/h$ if $C_s > C_m$ (deposition), c_o represents the soil cohesion, h is water depth, and v_s is the fall velocity. The total load transport capacity equation of Engelund & Hansen (1967) is used to estimate C_m.

In the WESP model, the erosion due to raindrop impact is $e_s = K_i\, i\, r_e$, where K_i is a parameter representing the susceptibility of the soil for raindrop erosion by impact, i is the rainfall intensity, and r_e is the rainfall excess over infiltration. The erosion due to shear is $e_h = K_r \tau^{1.5}$, where K_r represents the erodibility parameter due to flow shear and τ is the effective shear stress on the soil surface varying with flow depth, time and space. Soil deposition in the planes is $d = \varepsilon_p v_s C_s$, where ε_p is a coefficient, v_s is the fall velocity of the sediment particles and C_s is the mean concentration of the sediments. The erosion rate in the channels e_h is determined according to one of the following conditions: $e_h = a(\tau - \tau_c)^{1.5}$ for $\tau \ge \tau_c$ or $e_h = 0$ for $\tau \le \tau_c$, where a is the erodibility parameter for channels, τ is the shear stress of the surface runoff, and $\tau_c = \delta(\gamma_s - \gamma)D$ (δ is a numerical factor, γ_s and γ are specific weights of the sediments and water, respectively, and D is the sediment size) is the shear stress of the sediment. Deposition of the sediments in channels is $d = \varepsilon_c T_W v_s C_s$, where ε_c is a coefficient, T_W is the flow top width, and C_s is the mean concentration of the sediments.

The continuous models SHETRAN and MOSESS_D

SHETRAN and MOSESS_D are continuous spatially distributed models. The former is physically-based in that it integrates surface and subsurface processes at the basin scale, which is divided horizontally into grid cells and vertically into nodes within each soil layer. Evapotranspiration is based on the Penman-Monteith equation (also pan evaporation data). Interception is according to Rutter (1971/1972). Infiltration is represented by the variation of soil moisture content over time, determined with the equation of Richards (1931). Runoff is based on the Saint Vennant equation (1871). The Boussinesq equation is used for the saturated zone (not discussed in this article). Runoff and soil erosion processes are simulated using a fully implicit finite difference solution of the governing equations. MOSESS_D is not fully physically-based. It was developed based on its version for small scales, namely MOSESS (Figueiredo & Davi, 2006). The watershed is made up of elements and channels. Soil layers represent the soils vertically. Interception is based on a proportion of vegetated areas. Evapotranspiration can be either estimated with the methods of Penman, Thornthwaite or Blaney-Criddle, or with pan evaporation data. The model brings the moisture content up to date over time for the soil layers by considering the actual evapotranspiration, surface infiltration capacity, rainfall intensity, and hydraulic conductivities of the soils. Groundwater is not modelled. Runoff is generated by excess rainfall over infiltration capacity, by saturation of the surface layer (due to infiltration) or both, and routed alternatively by the Convex method of the SCS (McCuen, 1982) or by the method of Muskingum, both with the translation time expressed by the Kirpich equation (Chow *et al*., 1988) or estimated with the Manning equation. In both models soil erosion is due to rainfall impact and runoff. Sediment transport depends on the flow transport capacity.

Interception in the SHETRAN model depends on the net rainfall, canopy water depth, storage capacity S_c, and drainage parameters k_c and b_c. The net rainfall varies with the total rainfall, maximum and actual surface cover fractions p_1, p_2, and potential evapotranspiration E_p. Actual evapotranspiration is a function of the type $E_a/E_p = f(\psi)$, determined based on Feddes (1976), where E_a/E_p is the ratio of actual to potential evapotranspiration and ψ is the soil tension. The total actual evapotranspiration is from the interception, roots (based on a root function R_t), and bare soil.

Infiltration in the unsaturated zone is represented by the variation of soil moisture content over time t in the vertical axis z, that is $\partial\theta/\partial t = f[K, \psi, S](\theta)$, where θ is the volumetric moisture content, S is the source/sink term, $K(\theta) = K_S\ [(\theta - \theta_r)/(\theta_S - \theta_r)]^{\eta}$ is the Brooks & Corey (1964) hydraulic conductivity, θ_s and θ_r are the saturated and residual moisture contents, and η is the Averjanov exponent (Mualem, 1978).

The Saint Venant equation (for overland and channel flows) considers the water depth h, and the u (x axis) and v (y axis) flow velocities calculated with Manning's equation.

$$\frac{\partial h}{\partial t} + \frac{\partial (uh)}{\partial x} + \frac{\partial vh}{\partial y} = q \qquad \text{Overland flow} \tag{2}$$

$$\frac{\partial A}{\partial t} + \frac{\partial (Au)}{\partial x} = q_L \qquad \text{Channel flow} \tag{3}$$

Erosion by rainfall is based on the momentum squared of raindrops, and initiation of sediment motion by runoff (e.g. Shields), corrected by k_r and k_f, respectively, which are the rainfall and runoff erosivity coefficients, and by a factor for reducing rainfall detachment by the flow. The transport capacity can be calculated either with the total load equation of Engelund & Hansen (1967), or with the bed-load formulae of Yalin (1963), which account for the representative sediment diameter D_{50}.

Sediment transport is based on the mass conservation equations (overland and channel flows) given by:

$$\frac{\partial (hc)}{\partial t} + (1-\lambda)\frac{\partial z}{\partial t} + \frac{\partial g_x}{\partial x} + \frac{\partial gy}{\partial y} = 0 \qquad \text{Overland flow} \tag{4}$$

$$\frac{\partial (Ac)}{\partial t} + (1-\phi)B\frac{\partial z}{\partial t} + \frac{\partial (AcV_s)}{\partial x} = q_s \qquad \text{Channel flow} \tag{5}$$

where c is sediment concentration, λ is the soil porosity, g_x and g_y are the x, y transport rates, t is time, z is depth of loose soil, A is the flow area, ϕ and B are channel bed porosity and width, and V_s is the sediment velocity.

The MOSESS_D model takes the rainfall intensity i based on input data of rainfall heights P and corresponding rainfall time intervals Δt. Actual evapotranspiration rate E_a is zero when $P > 0$ and/or $\Psi \geq 1500$ kPa (wilting point); $E_a = E_p$ (potential) when $\Psi \leq 33$ kPa (field capacity); and when 1500 kPa $\leq \Psi \leq 33$ kPa, E_a is determined by $E_a = E_p.exp[-\kappa.(\psi(\theta)-33)/(1500 - 33)]$, where

κ is a local factor, and $\Psi(\theta) = A.\theta^B$ is the equation of Saxton *et al.* (1986), with A and B expressed in terms of sand and clay percentages; if $\psi \leq 33$ kPa then $\theta = \theta_s$ with θ_s taken as porosity ϕ (also expressed in terms of soil texture with the Rawls & Brakensiek (1989) equation) corrected by the air entrapment factor f_a ($0.8 < f_a < 0.9$), that is $\theta_s = f_a\ \phi$.

The actual soil moisture content $\theta_{t+\Delta t}$ is determined for the surface and sub-surface layers, with depths h_S and h_{SS}, for when $P = 0$ and $P > 0$. When $P = 0$, the moisture content of the surface layer θ_S decreases by $\Delta\theta = [E_a + K_S(\theta)].\Delta t/h_S$ due to evapotranspiration, and by $\Delta\theta = [E_a + K\ (\theta)].\Delta t/h_S$, where $K(\theta)$ is taken as the lower hydraulic conductivity of the layers, that is $K(\theta) = K_S(\theta)$ if $K_S(\theta) < K_{SS}(\theta)$ or $K(\theta) = K_{SS}(\theta)$ otherwise. The moisture content of the sub-surface layer θ_{SS} will decrease by $\Delta\theta = E_a.\Delta t/h_{SS}$ owing to evapotranspiration if $\theta_S \leq \theta_{Smin}$ (a minimum value up to which the water cannot percolate to the sub-surface layer). Otherwise θ_{SS} increases by $\Delta\theta = K(\theta).\Delta t/h_{SS}$ with $K(\theta)$ taken as the lower hydraulic conductivity of the layers. When $P > 0$ the actual evapotranspiration rate $E_a = 0$. Therefore, the moisture content of the surface layer increases according to the infiltration capacity f by $\Delta\theta = f.\Delta t/h_{SS}$ if $i > f$, or to the rainfall intensity i by $\Delta\theta = i.\Delta t/h_{SS}$ if $i \leq f$, and the moisture content of the sub-surface layer increases by $\Delta\theta = K(\theta).\Delta t/h_{SS}$ where $K(\theta)$ is taken as the lower hydraulic conductivities of the layers.

The surface infiltration capacity is calculated with the empirical equation $f = C.\theta^{-D}$, where C and D are soil factors determined through infiltration experiments. It is limited to f_{max} when $\theta = \theta_{wp}$ (moisture content at wilting point) and f_{min} when $\theta = \theta_{fc}$ (moisture content at field capacity). The hydraulic conductivity is based on the Brooks & Corey (1964) equation with the Averjanov exponent $\eta = 3 + 2/\lambda$, where $\lambda = -1/B$ is the pore size factor, B is the Saxton *et al* (1986) parameter, K_s is the saturated hydraulic conductivity, and θ_r the residual moisture content (determined with the Rawls & Brakensiek (1989) equation), all in terms of soil texture.

The overland flow depth is determined for the following conditions $h_{t+\Delta t} = (i - f).\Delta t$ if $i > f$, $h_{t+\Delta t} = (\theta - \theta_s).h_s$ (h_s is the depth of the surface layer) if $i > \theta > \theta_s$, $h_{t+\Delta t} = (i - f).\Delta t + (\theta - \theta_s).h_s$ if $i > f$ and $\theta > \theta_s$. The water discharge Q for each element is determined by $A.h_{t+\Delta t}/\Delta t$, where A is the area of the element, $h_{t+\Delta t}$ is the runoff lamina and Δt is the time interval, and placed at the neighbour upstream channel link. Two procedures are used for the flow routing: the Muskingum method, and the Convex method of the SCS (McCuen, 1982). The actual output flow in the Muskingum method is $Q_{t+\Delta t} = C_1 I_{t+\Delta t} + C_2 I_t + C_3 Q_t$, where $I_{t+\Delta t}$ and I_t are the actual and antecedent input flows, Q_t is the antecedent output flow (unknown), and C_1, C_2 and C_3 are the Muskingum propagation coefficients given in terms of the water surface form X, the hydrograph translation time K, and the time interval Δt. The convex method of the SCS gives the actual output discharge as $Q_{t+\Delta t} = CI_t + (1 - C)\ Q_t$, where I_t and Q_t (unknown) are the antecedent input and output flows, $C = \Delta t/K$ is the propagation coefficient and K is the hydrograph translation time. For ephemeral rivers Q_t can be set to zero.

The hydrograph translation time can be estimated with the equation of Kirpich, $K_t = 0.0078L^{0.77}/S^{0.385}$, where L (ft) and S (–) are the length and slope of the reach, or with the equation $K_t = (L/V)$ where L is the length of the reach and V is mean water flow velocity that can be estimated with Manning's equation.

The soil erosion detached by rainfall and runoff form the total load of sediment available for transport, which are simulated with the same equations as for the SHETRAN model. The sediment available (by rainfall and runoff) will be totally transported if it does not exceed the flow transport capacity. Otherwise, the difference will be deposited. The flow transport capacity can be calculated with either the equation of Yalin (1963), Engelund & Hansen (1967), or with the equation of Laursen (1958) for bed and/or total load that accounts for the soil size distribution.

CATCHMENT MODELLING AND DATA USED

Data of catchments in the semi-arid regions of Sumé and Tauá (Fig. 1), in the Northeast region of Brazil, were utilized, which are: runoff–erosion plots (100 m^2) and micro-basins (0.5–1.0 ha) in

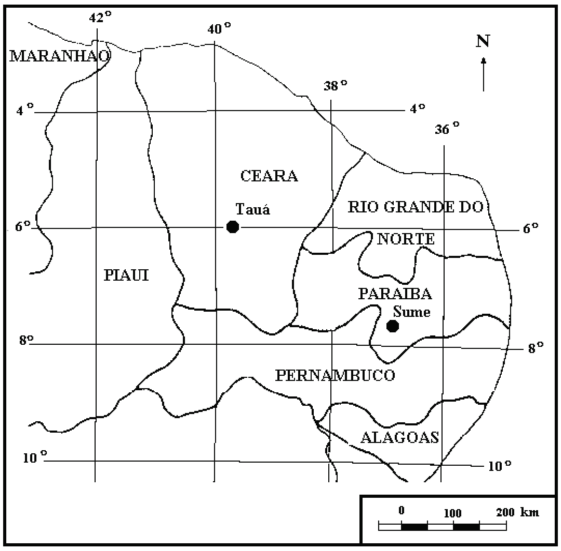

Fig. 1 Location of the study areas.

the experimental basin of Sumé (EBS), bare cleared and with native vegetation (Caatinga), within the Umburana (11 km^2) sub-basin of the representative basin of Sumé-RBS (140 km^2 at Gangorra) and its sub-basins Jatobá (27 km^2) and Umburana (11 km^2), in the state of Paraíba, and the sub-basin Mundo Novo (19 km^2) in the representative basin of Tauá (RBT) in the state of Ceará.

Micro-basins and catchments were modelled according to the rivers network and basins relief (Figueiredo, 1988; Lopes, 2003). Several rainfall–runoff–erosion events observed at the plots and micro-basins were utilized to calibrate the event models KINEROS2 and WESP. Data from automatic stations, in the EBS (1982–1986) as well as in the RBS (1977), were utilized to run the SHETRAN model. The MOSESS_D applications were carried out for the plots, micro-basins and Umburana with data from the EBS (1986) and RBS (1977), and some non-continuous rainfall–runoff events observed at the sub-basins of Umburana and Mundo Novo. Model performances were investigated based on comparisons of model results with observed runoff and erosion values, hydrographs, and also with sediment yields observed worldwide.

MODEL PARAMETERIZATION

Models parameters were first tested at the plots. Some parameters of the event models were fixed according to the soils, such as $\alpha = 0.85$ that characterizes the soil type, and the soil porosity $\phi = 0.398$. The capillary potential was adjusted to G = 260 mm for the plots, but it varied for the RBS basins from 10 to 960 mm. The other parameters were calibrated and varied with the scale. The Manning roughness coefficient for the planes ranged from 0.02 to 0.1 (for the channels it was fixed to 0.035), and the hydraulic conductivity from 3.5 to 15 mm h^{-1} depending on the plot surface condition. The initial relative saturation S_i varied from 0.42 to 0.76 (plots and micro-basins), and from 0.19 to 0.94 (catchments of the RBS). The erosion parameter c_f varied from 10^3 to 10^8 (plots and micro-basins) and N_s from 0 to 220 mm (plots and micro-basins), K_r from 0 to 10 kg.m/N$^{1.5}$ s^{-1} for the plots (a mean of 1.76 was set to the micro-basins), and a from 1.10^{-5} to 0.15 kg.m/N$^{1.5}$ s^{-1} for the channels.

The SHETRAN and MOSESS_D model parameters (Table 1) were evaluated based on the texture (Saxton *et al.*, 1986; Rawls & Brakensiek, 1989) of the soils of the two soil layers (*S* and *SS*) of the RBS: the loam (50.2% sand; 15.8% clay) surface layer (10 cm) and the sandy clay loam (50.2%, 32.5% clay) sub-surface layer (40–200 cm). The interception parameters were set to S_c = 0.5 mm, k_c = 1.7 and b_c = 7.77 according to Jetten (1996). Cover fractions were set to 0.1, 0 and 0.5 to represent ground cover by stones, bare and vegetated areas, respectively. The parameter κ

=10, C = 17 and D = 0.197 were determined via infiltration experiments. The Muskingum routing parameter *X* was set to 0.1. The erosivity parameters were k_r = 18 s^2 kg^{-1} m^{-2} and k_f = 6 mg m^{-2} s^{-1} (k_f was set to 0 in the SHETRAN and MOSESS_D models for comparisons of sediment yields). The Manning-Strickler roughness coefficient, the inverse of Manning's coefficient (Chow, 1959), was set to 29.5 for the channel reaches, and to 50 for the elements. The sediment diameter (D_{50}) was fixed to 0.5 mm in the methods of Yalin and Engelung & Hansen, while the sediment size distribution in Table 2 was utilized in the method of Laursen. The convex method was used for runoff propagation and Kirpich for the time of concentration. Based on the water temperature of 25 °C, γ = 9779 N m^{-3}, ν (kinematic viscosity) = 8.94 × 10^{-7} m^2 s^{-1} and γ_s/γ = 2.65.

Table 1 Soil parameters used in the SHETRAN and MOSESS_D models.

Soil layer	A	B	K_s	θ_s	θ_r	θ_{wp}	θ_{fc}	θ_{smin}	ϕ
S	0.021257	–5.077	5.850	0.4081	0.0747	0.1109	0.2348	0.2976	0.448
SS	0.001055	–8.338	0.216	0.3908	0.1118	0.1829	0.2887	0.3335	0.488

A (kPa); K_s (mm.day^{-1}); θ, ϕ (m^3 m^{-3}).

Table 2 Percentages P_i of diameters D_i from the eroded soil.

	D_1	D_2	D_3	D_4	D_5	D_6	D_7	D_8	D_9
D_i (mm)	0.002	0.063	0.100	0.200	0.630	1.000	2.000	6.300	10
P_i (mm)	5.34	5.06	13.56	40.71	14.44	13.81	6.15	0.58	0.35

RESULTS AND DISCUSSION

In general, models results approached the observed values. Peak runoffs and sediments were overestimated with the continuous models, most likely because of the modelling of the river systems. Some results are presented in Figs 2–6 and Table 3. It is seen that the simulated runoff laminas and hydrograph (Figs 2 and 3) approached the observed values at micro-basin M3 (5200 m^2). Hydrographs and sediment observed and simulated (Figs 4 and 5) with the model MOSESS_D (Umburana and Mundo Novo) suggest that the hydrological propagation procedures simulated the outputs with consistency. Table 3 shows laminas and erosion simulated with the

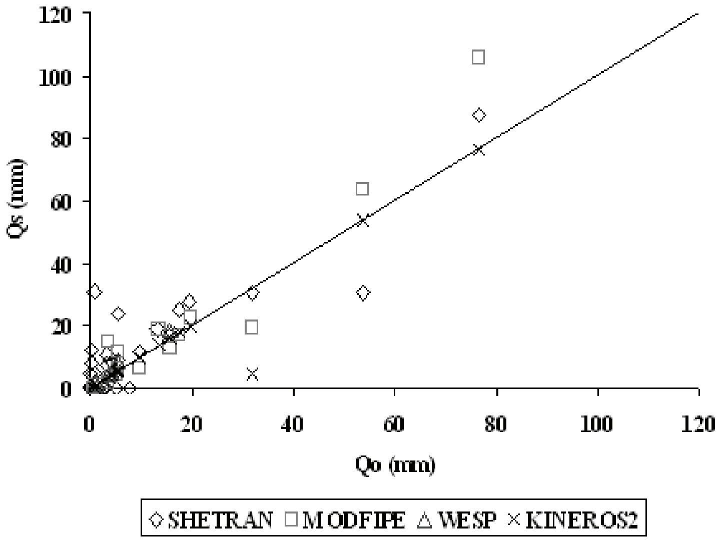

Fig. 2 Observed (Qo) and simulated (Qs) laminas in 1986 (micro-basin M3).

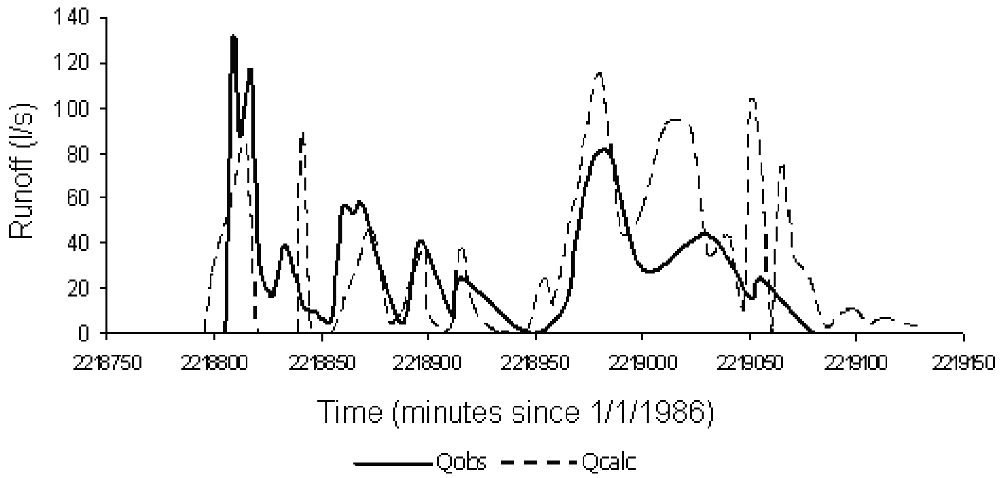

Fig. 3 Observed and simulated hydrographs with the MOSESS_D model (micro-basin M3).

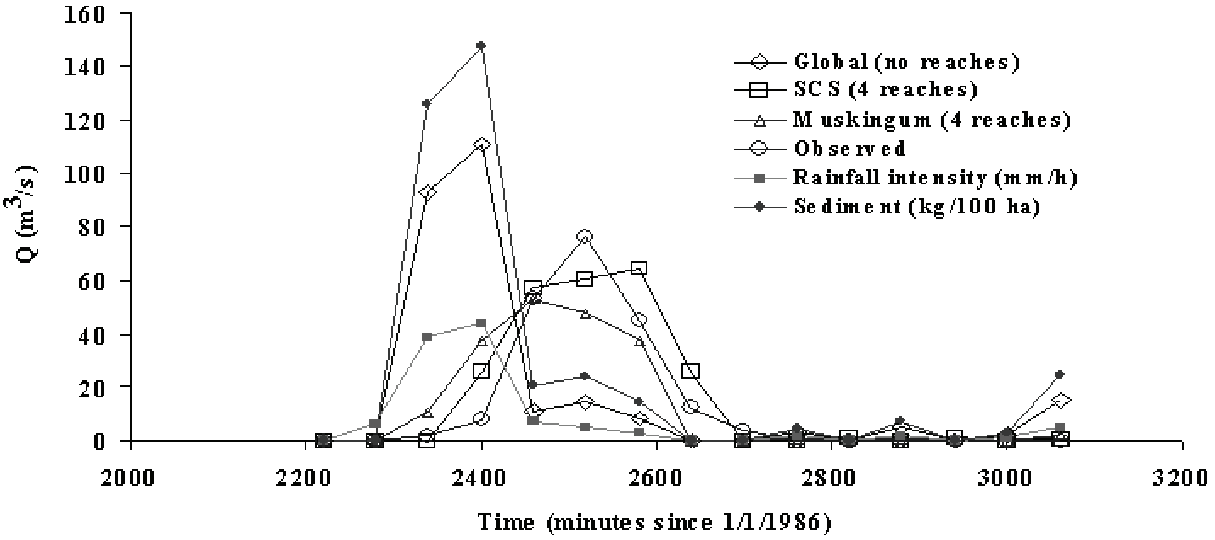

Fig. 4 Observed and simulated hydrographs and sediment at Umburana (EBS).

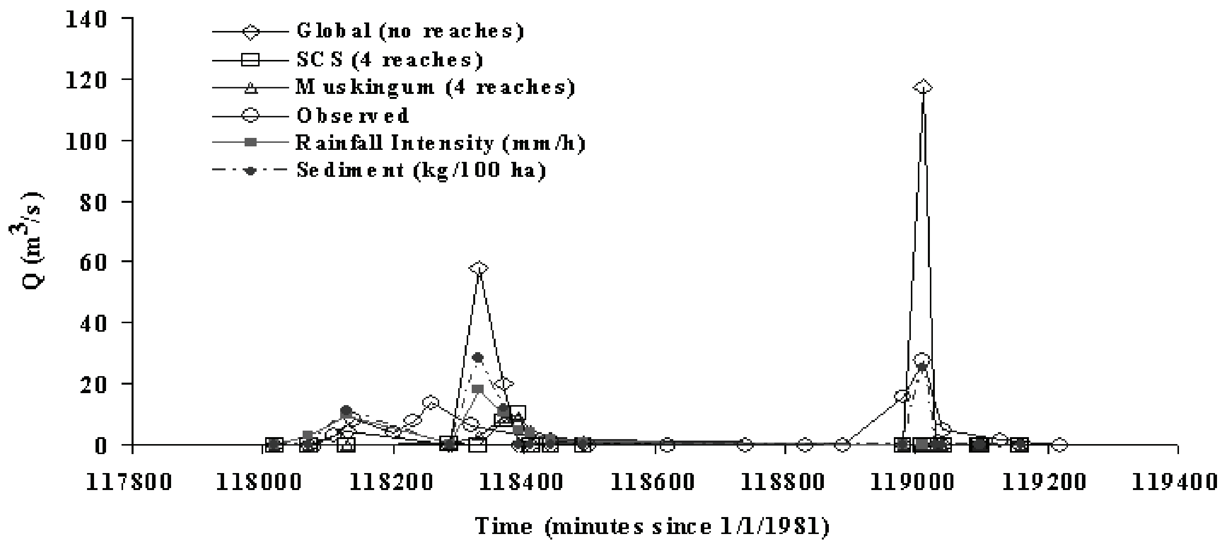

Fig. 5 Observed and simulated hydrographs and sediment at Mundo Novo (RBT).

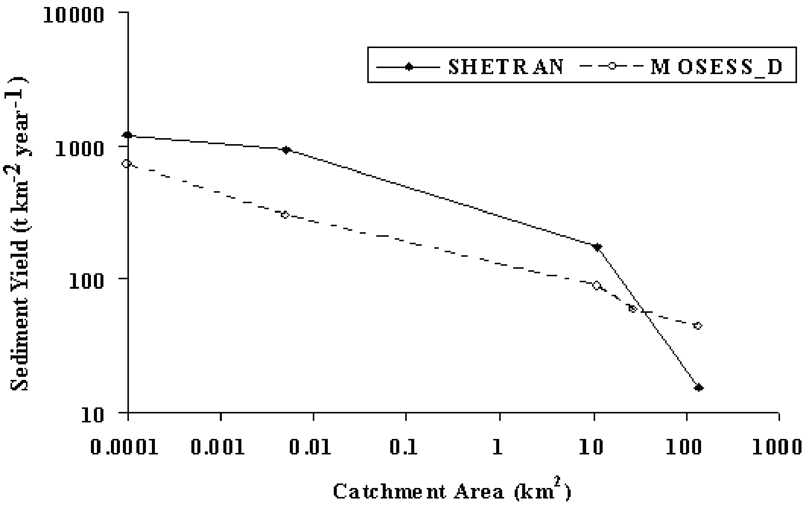

Fig. 6 Simulated annual sediment yields for the EBS and RBS catchments.

Table 3 Observed and simulated laminas and erosions with the models (micro-basin M4).

Month/Year	$Q_{observed}$ (mm)	$Q_{shetran}$ (mm)	Q_{mosess_D} (mm)	$Q_{kineros2}$ (mm)	$E_{observed}$ (kg/ha)	$E_{shetran}$ (kg/ha)	E_{mosess_D} (kg/ha)	$E_{kineros2}$ (kg/ha)
Feb/86	83	46	64	64	9733	11027	9604	8794
Mar/86	206	298	134	133	8979	24551	17392	9864
Apr/86	3.4	32	71	43	4434	14769	6370	3817
Total	298	376	308	240	23146	50500	33366	22475

models (except WESP due to the few calibrated events) and observed at the micro-basin M4 (4800 m^2) in the rainy months of 1986. The continuous models represented the runoffs better, but KINEROS2 simulated the best results for the sediments. Simulated sediment yields with the SHETRAN and MOSESS_D models, for the scales of the EBS and RBS, range from 1000 to 10 t km^{-2} $year^{-1}$ (Fig. 6), compared well with figures observed in semi-arid continents (see Walling, 1994) and in river systems located in areas of low precipitation all over the world (Fig. 7). These yields are high for semi-arid landscapes as a result of land misuse. Results of land use modelled for the investigated areas with the MOSESS_D model showed no major impact on sediment yields, owing to the simplicity with which basin cover is accounted for in the model. In contrast, the sediment yields simulated with SHETRAN showed a great impact by deforestation (Figueiredo & Bathurst, 2007).

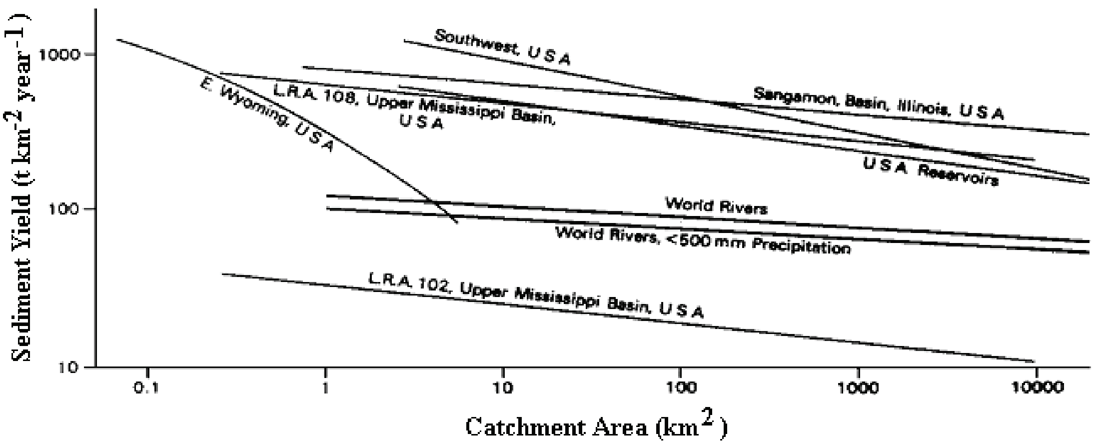

Fig. 7 Observed annual sediment yield in areas of low precipitation (Walling & Kleo, 1979).

CONCLUSIONS

From the results in this research it can be concluded that the investigated physically-based models were able to represent the dynamics of the hortonian flows in semi-arid environments, in the states of Paraíba (Sumé) and Ceará (Tauá). The observed runoffs were approached by the continuous models simulations, but overestimated peaks and sediments, while the event model simulations come close to the sediments. Sediment yields simulated with the continuous models, in the range of 1000 to 10 t km^{-2} year^{-1}, are high for the semi-arid environment, but compared well with observed values in semi-arid continents and in areas of low precipitation all over the world.

Acknowledgements The CAPES and CNPq, Brazil, supported the work described in this paper.

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