Sediment Dynamics and the Hydromorphology of Fluvial Systems (Proceedings of a symposium held in Dundee, UK, July 2006). IAHS Publ. 306, 2006.

MOSESS: a model for soil erosion prediction at small scales

EDUARDO E. DE FIGUEIREDO[1] & HÉRBETE H. R. C. DAVI[2]

1 *Water Resources Research Engineering Area, Department of Civil Engineering, Federal University of Campina Grande, PO Box 505 – 58109-970, Campina Grande – PB, Brazil*
eduardo@dec.ufcg.edu.br

2 *Superintendence of Infra-Structure, Federal University of Rio Grande do Norte, PO Box 1524 – 59072-970, Natal – RN, Brazil*

Abstract In contrast to fully distributed physically-based models, event-oriented models do not consider antecedent conditions. Both types of models are in general distributed and applicable at small and large scales. However, at small scales the effects of storage within the hydrological system is negligible and both types of model can perform similarly. This motivated the design of MOSESS, a simple soil erosion model, which simulates runoff and erosion processes at small scales. The model takes into account antecedent conditions, with the processes and basin features represented by physical parameters. Runoff is generated either by an excess of rainfall over infiltration capacity or by saturation of the soil top layer over time, and soil erosion by rainfall and runoff. Applications of the model to small areas in Sumé, located in the semi-arid Northeast of Brazil showed that the processes were reasonably well simulated. MOSESS demonstrated its suitability as a simple soil erosion model with results comparable to those obtained with other calibrated event-oriented models.

Key words event-oriented model; overland flow; small-scale model; soil erosion

INTRODUCTION

In this paper, a soil erosion model is coupled with and driven by a hydrological model, which is generally designed for determining the process of overland flow that can be generated either by infiltration excess overland flow (Horton, 1933) or by saturation excess overland flow (Dunne, 1978). Soil erosion incorporates the processes of detachment, transportation and deposition of soil particles, which are governed by erosive agents such as rainfall, runoff, wind and gravity (Vanoni, 1975; Lal, 1994). Detachment is governed by the energy of raindrop impact and by forces of the water flow. Flow transport capacity and sediment resistance to motion governs soil transportation and deposition. For modelling these processes, important properties depend on the basin scale. By contrast to large-scale models, small-scale models assume a homogeneous basin, sensitive to high intensity and short-duration rainfalls, and runoff, in which overland flow predominates, and channel storage is negligible (Singh, 1998). Also, the land processes (e.g. infiltration, runoff) are represented by methods valid for small scales (Gupta, 1986). In general, either event-oriented models or fully distributed physically-based models are used to represent these processes. The former models do not take into account antecedent conditions (e.g. WESP, Lopes, 1987) while the latter models are continuous and incorporate basin heterogeneities (e.g. SHETRAN, Ewen *et al.*, 2000).

These aspects motivated the design of MOSESS, which is presented and examined in this paper

DESCRIPTION OF MOSESS

MOSESS is the English translation for MOFIPE, the Portuguese acronym for *Modelo de base física para simulação hidrossedimentologica em pequena escala* developed by Davi & Figueiredo (2003).

The processes involved (rainfall, evapotranspiration, infiltration, overland flow and erosion) are modelled based on three assumptions, which operate at small scales: (a) the basin system is homogeneous; (b) rainfall is space-invariant, and (c) the channel system is negligible. Figure 1 shows the model flowchart. During rainfall events the actual evapotranspiration is zero because the air is already saturated. Between rainfall events, it is calculated in terms of the soil tension that varies with the soil moisture content and potential evapotranspiration. The moisture content depends on the rainfall intensity, infiltration capacity and hydraulic conductivity. Runoff is generated either if the rainfall intensity exceeds the infiltration capacity or if the soil top layer becomes saturated. Soil erosion is generated by the impact of raindrops and runoff, with the rate

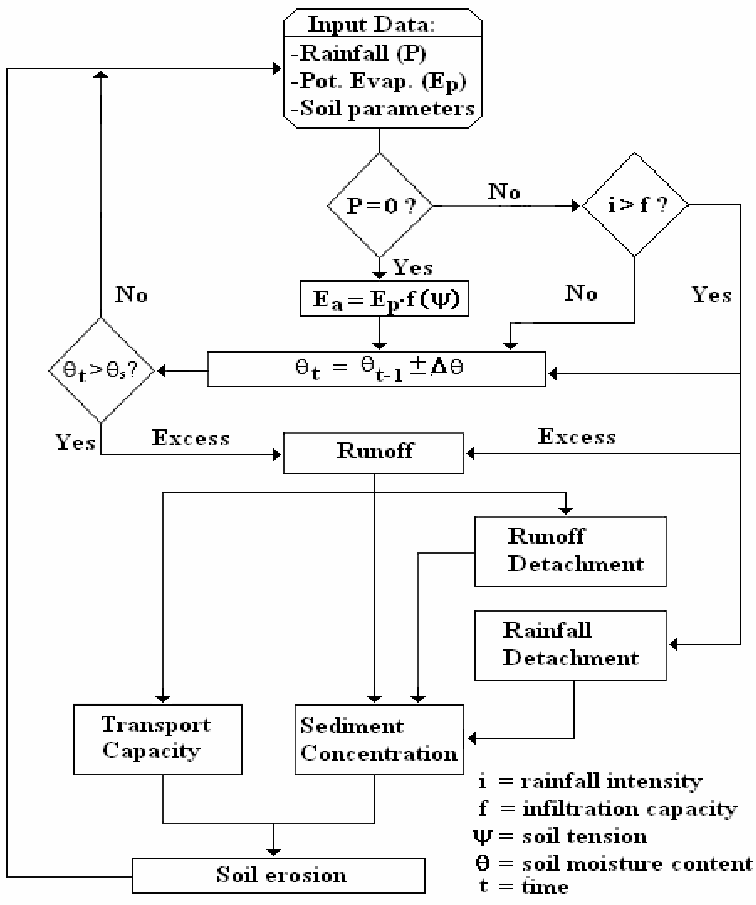

Fig. 1 Flowchart of MOSESS (adapted from Davi, 2004).

of transport depending on the transport capacity of the flow. If the sediment available for transport exceeds the flow transport capacity, the difference is deposited. The model is not complex and does not require numerical solutions. Its components are presented as follows.

Flow component

The input data of rainfall P (mm) and corresponding time intervals Δt (h) of occurrence are used to calculate the rainfall intensities ($i = P/\Delta t$ (mm h^{-1})). Between rainfall events and for soil tensions (ψ) between the field capacity (33 kPa) and wilting point (1500 kPa), the actual evapotranspiration rate E_a (mm h^{-1}) is calculated by:

$$E_a = E_p \cdot \exp[-\delta.(\psi(\theta) - 33)/(1500 - 33)] \qquad \text{if } 1500 < \psi(\theta) < 33 \text{ kPa} \tag{1}$$

where E_p is the potential evapotranspiration rate (mm h^{-1}), θ is the soil moisture content (m^3 m^{-3}), δ (–) is a local factor and $\psi(\theta)$ (kPa) is given by the equation of Saxton *et al*. (1986):

$$\psi(\theta) = A \cdot \theta^B \tag{2}$$

where A and B are expressed in terms of sand and clay percentages (%S, %C).

In equation (2) if $\psi \geq 1500$ kPa then $\theta \leq \theta_{wp}$ (moisture content at wilting point) and $E_a = 0$, and if $\psi \leq 33$ kPa then $\theta \geq \theta_{fc}$ (moisture content at field capacity) and $E_a = E_p$.

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} (mm), for when $P = 0$ and $P > 0$. When $P = 0$, the moisture content of the surface layer (θ_S) decreases due to evapotranspiration to the atmosphere and percolation to the sub-surface layer. Percolation is a function of the lower hydraulic conductivity of the layers ($K_S(\theta)$ or $K_{SS}(\theta)$ mm h^{-1}). The moisture content of the sub-surface layer (θ_{SS}) will decrease 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 the water cannot evaporate from the sub-surface and θ_{SS} may increase. These conditions are expressed as:

$$\theta_{S(t+\Delta t)} = \theta_{S(t)} - \Delta\theta = \begin{cases} \theta_{S(t)} - \dfrac{[E_a + K_S(\theta)] \cdot \Delta t}{h_S} & \text{if } [K_S(\theta) < K_{SS}(\theta)] \text{ and } \theta_S > \theta_{S\min} \\ \theta_{S(t)} - \dfrac{[E_a + K_{SS}(\theta)] \cdot \Delta t}{h_S} & \text{if } [K_S(\theta) \geq K_{SS}(\theta)] \text{ and } \theta_S \leq \theta_{S\min} \end{cases} \tag{3}$$

$$\theta_{SS(t+\Delta t)} = \theta_{SS(t)} \pm \Delta\theta = \begin{cases} \theta_{SS(t)} - \dfrac{(E_a)_t \cdot \Delta t}{h_{SS}} & \text{if } \theta_S \leq \theta_{S\min} \\ \theta_{SS(t)} + \dfrac{K_{S(t)} \cdot \Delta t}{h_{SS}} & \text{if } \theta_S > \theta_{S\min} \text{ and } K_S < K_{SS} \end{cases} \tag{4}$$

When $P > 0$ ($E_a = 0$) the moisture content of the surface layer increases according to the infiltration capacity (f), if $i > f$, or to the rainfall intensity if $i \leq f$ (mm h^{-1}), and the moisture content of the sub-surface layer is a function of the lower hydraulic conductivity of the layers. These conditions are expressed as:

$$\theta_{S(t+\Delta t)} = \theta_{S(t)} + \Delta\theta = \begin{cases} \theta_{S(t)} + \dfrac{f \cdot \Delta t}{h_S} & \text{if} \quad i > f \\ \theta_{S(t)} + \dfrac{i \cdot \Delta t}{h_S} & \text{if} \quad i \le f \end{cases} \tag{5}$$

$$\theta_{SS(t+\Delta t)} = \theta_{SS(t)} + \Delta\theta = \begin{cases} \theta_{SS(t)} + \dfrac{K_S \cdot \Delta t}{h_{SS}} & \text{if } K_S < K_{SS} \\ \theta_{SS(t)} + \dfrac{K_{SS} \cdot \Delta t}{h_{SS}} & \text{if } K_S \ge K_{SS} \end{cases} \tag{6}$$

The surface infiltration capacity is calculated in terms of θ ($f = C\theta^D$ mm h^{-1}, where C and D are soil factors, which can be determined through infiltration experiments). It is limited to f_{max} when $\theta = \theta_{wp}$ and f_{min} when $\theta = \theta_{fc}$. The hydraulic conductivity is calculated with the equation of Brooks & Corey (1964) given by:

$$K(\theta) = K_{sat}[(\theta - \theta_r)/(\theta_{sat} - \theta_r)]^{3+2/\lambda} \tag{7}$$

where $\lambda = -1/B$ is the pore size factor, with B and K_{sat} (saturated hydraulic conductivity (mm h^{-1}) determined by the Saxton *et al.* (1986) equations, and θ_r (the residual moisture content) determined by Rawls & Brakensiek (1989) equation, all in terms of %S and %C.

The overland flow depth (mm) is determined by the following equation:

$$h_{t+\Delta t} = \begin{cases} (i - f) \cdot \Delta t & \text{if} \quad i > f \\ (\theta - \theta_{sat}) & \text{if} \quad \theta > \theta_{sat} \quad \text{and} \quad i < f \end{cases} \tag{8}$$

where θ_{sat} is the moisture content at saturation, taken as the soil porosity (ϕ) corrected by a factor for the air entrapment (f_a), with ϕ given in terms of texture (%S, %C) by Rawls & Brakensiek (1989).

Soil erosion component

Erosion by rainfall detachment is based on the momentum squared of raindrops:

$$D_r = k_r F_w (1 - C_g) M_r \tag{9}$$

where k_r = rainfall and runoff erosivity coefficients (s^2 kg^{-1} m^{-2}); $F_w = \exp(1 - h/d_m)$ is the factor for reducing rainfall detachment ($F_w = 1$ if $h < d_m$), h = water depth (m), $d_m = 0.00124 i^{0.182}$ is the raindrop diameter (m); C_g = ground cover fraction (–); $M_r = \alpha i^\beta$ (kg m^{-2} s^{-1}) is the moment squared for rainfall; α and β are factors that depend on the rainfall intensity (Wicks, 1988).

Erosion by runoff is based on the sediment initiation of motion:

$$D_f = k_f (1 - C_g)(\tau / \tau_c - 1) \quad \text{if } \tau > \tau_c; \text{ otherwise } D_f = 0 \tag{10}$$

where k_f = runoff erosivity coefficient (kg m^{-2} s^{-1}), $\tau = \gamma h S_w$ (N m^{-2}) is the flow shear stress, γ = specific weight of water (N m^{-3}), h = flow depth (m), S_w = water surface slope (–), $\tau_c = (\gamma_s/\gamma - 1)\gamma D_{50}\, a.R_*^{\,b}$ (N m^{-2}), γ_s = sediment specific weight (N m^{-3}), D_{50} = representative diameter (m), $a,b = f(R_*)$; R_* = max (0.03; $(D_{50}/\nu)(g\ \tau/\gamma)^{1/2}$) is the particulate Reynolds number (–), ν = kinematic viscosity (m^2 s^{-1}) and g = 9.81 m s^{-2}.

The sediment available ($D_r + D_f$) for transport 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 from the equations of Laursen (1958), Yalin (1963) or Engelund-Hansen (1967). The equation of Engelund-Hansen is for total load, while Yalin proposed a formula for bed load. The equation of Laursen is for bed and/or total load with the soil size distribution taken into account, while the other methods use the representative diameter D_{50}.

PARAMETERS EVALUATION

Parameters for the applications of MOSESS were evaluated based on information from the experimental area in Sumé (Fig. 2), which consists of nine plots of 100 m^2 (P1, P3 and P4, bare cleared, P5 with native vegetation and the others under different surface treatment), and four micro-basins (0.48–1.0 ha), M1 and M2 with native vegetation and M3 and M4 bare cleared. At these small areas runoff and soil erosion were observed during the period 1982–1988. The soil surface is a shallow loam (50.2% sand; 15.8% clay), underneath which there is a sandy clay loam (50.2% sand, 32.5% clay) (Cadier & Freitas, 1982; Cadier *et al.*, 1983).

Data from a raingauge station in the area were used to determine the rainfall intensities. Equation (1) was used to determine E_a (for 33 kPa < ψ < 1500 kPa) hourly, with the potential evaporation rate (E_p) based on daily pan evaporation data and δ set to 10 (Davi, 2004). The surface and sub-surface parameters were calculated from the texture of the layers, which were: A = 0.021257 and 0.001055, B = –5.077 and –8.338, K_{sat} = 5.85 and 0.216 mm day^{-1}, θ_r = 0.0747 and 0.1118 m^3 m^{-3} and ϕ = 0.448 and 0.488 m^3 m^{-3}; the values of θ_{sat} were set to 0.4081 m^3 m^{-3} and 0.3908 m^3 m^{-3} based on

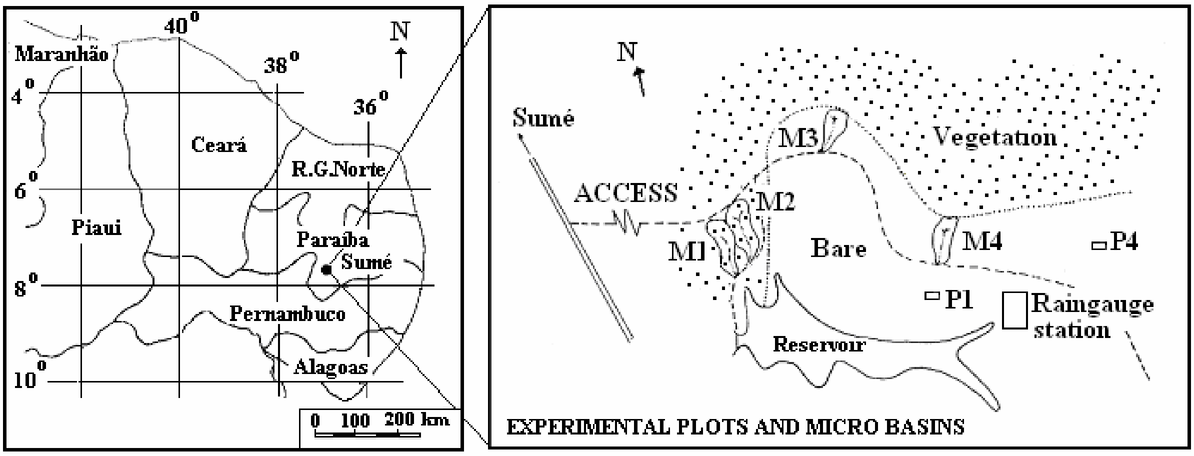

Fig. 2 The Northeast region of Brazil (left) and experimental areas (right) in Sumé.

the values of ϕ and f_a = 0.91 and 0.8. For the infiltration capacity, C and D were set to 17 and 0.197 (Davi, 2004). The values of θ_{Smin} = 0.2976 m^3 m^{-3} (surface), θ_{fc} = 0.2348 and 0.2887 m^3 m^{-3} and θ_{wp} = 0.1109 and 0.1829 m^3 m^{-3} were determined from equation (2) making ψ = 10, 33 and 1500 kPa respectively.

For determining the sediment detached by rainfall and runoff (equations (9) and (10)), C_g was set to 0.1 to represent the ground cover fraction by stones, the raindrop diameter d_m to 0.005 m (Lal, 1990), k_r to 18 s^2 kg^{-1} m^{-2}, k_f to 6 mg m^{-2} s^{-1}, α, β, a, b and R_* determined according to data in Wicks (1988). Based upon the water temperature of 25 °C, γ was set to 9779 N m^{-3}, ν to 8.94 × 10^{-7} m^2 s^{-1} and γ_s/γ to 2.65. The value of D_{50} varied from 0.4 to 1.0 mm depending on the site to model. The transport capacity was investigated with all methods for comparisons. For the Laursen´s method, just one sediment diameter (D_{50}) was considered to keep consistency for comparisons with the methods of Yalin and Engelung-Hansen.

MODEL SIMULATIONS

Simulations (1985–1987) were carried out for the plots P1 (3.8% slope) and P4 (7% slope), and micro-basin M3 (5200 m^2; 7% slope) and M4 (4800 m^2; 6.8% slope), all under bare soil surface condition, with the initial soil moisture contents of the surface (h_S = 0.1 m) and subsurface layers (h_{SS} = 0.4 m) corresponding to the wilting point. This was because the simulations started prior to 1985, during a dry period in the region. In general, the simulations compared well with the observed values (r^2 > 0.67 for runoff and r^2 > 0.41 for soil erosion). Comparison of MOSESS simulated runoff with the results generated by the calibrated event-oriented models WESP (Lopes, 1987; Aragão, 2000) and KINEROS (Smith *et al.*, 1994; Lopes, 2003) for sites P4 and M4 showed reasonable agreement (Fig. 3). Hydrographs and sedigraphs (Fig. 4) for the event 126 (M3) showed that the lowest sediment discharges were obtained with Yalin's equation, with the best results from the Engelund-Hansen equation followed by the Laursen equation. Figure 5 shows the soil moisture content with depth (M3), prior and during the event 126. Before the event, the soil moisture decreases owing to evapotranspiration, and it increases during the rainfall.

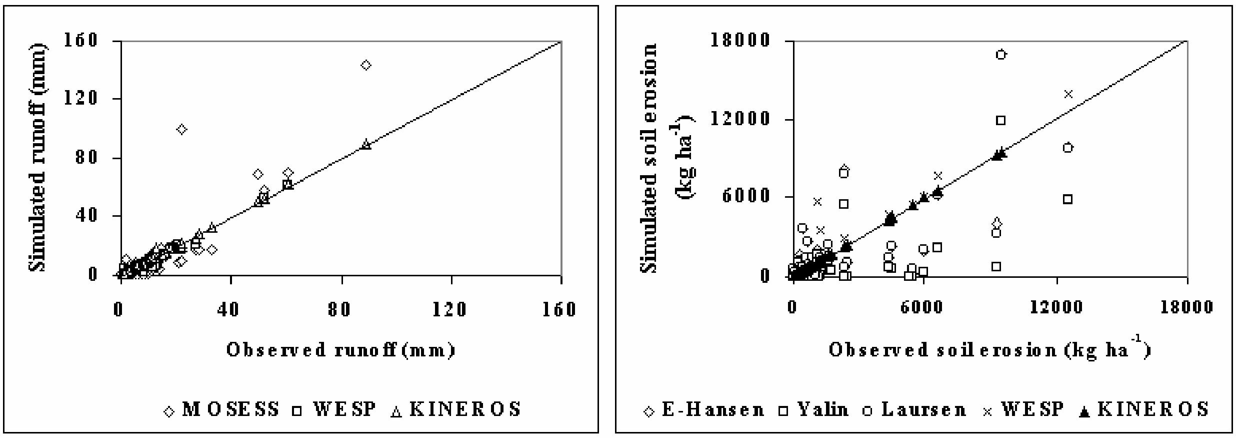

Fig. 3 Observed and simulated runoffs at plot P4 (left) and micro-basin M4 (right)

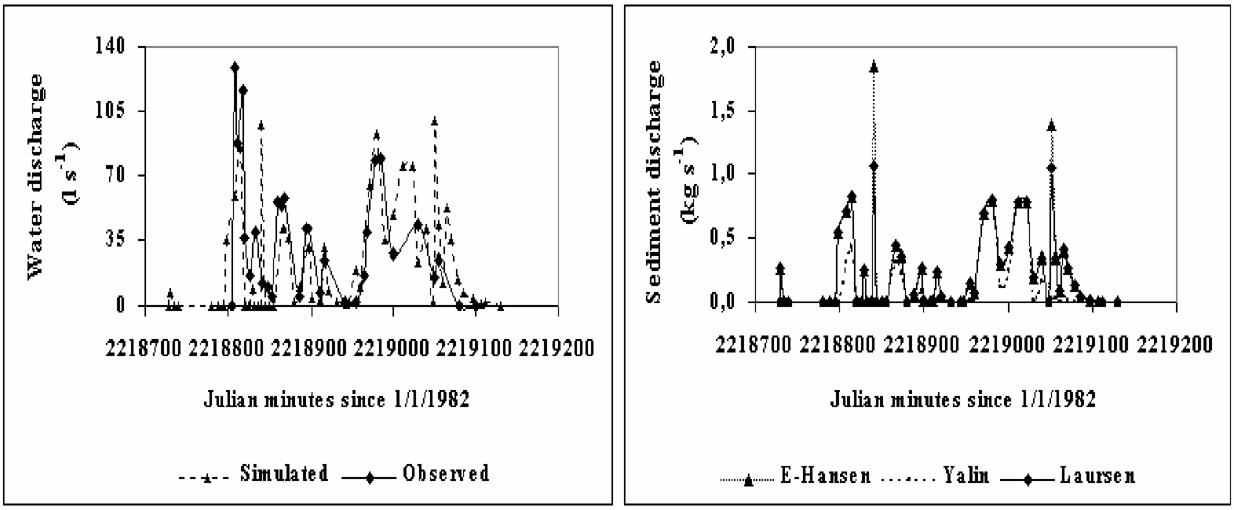

Fig. 4 Hydrographs (left) and sedigraphs (right) at micro-basin M3 (event 126).

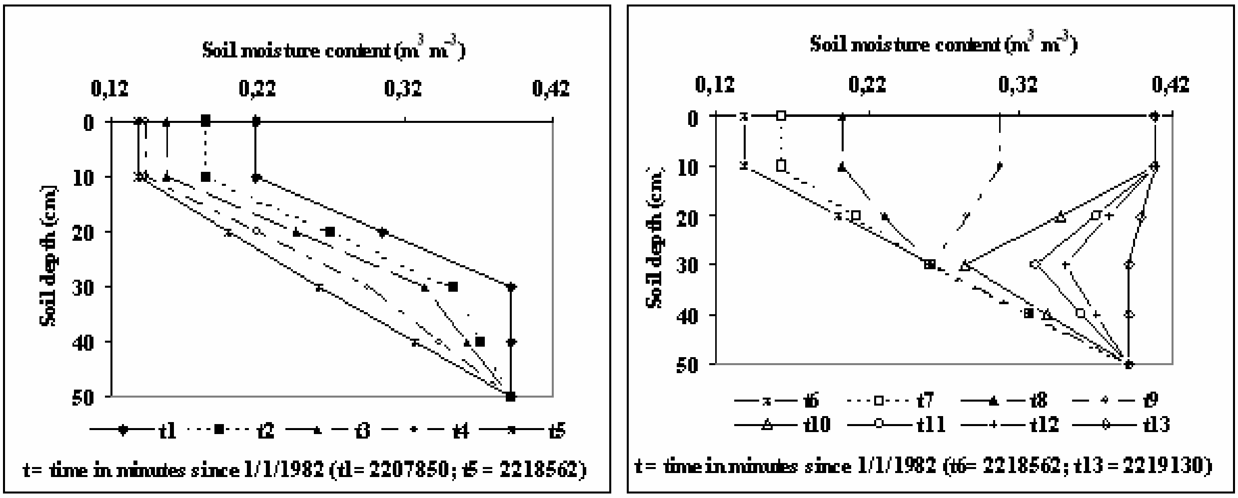

Fig. 5 Soil moisture content before (left) and during (right) event 126 at micro-basin M3.

CONCLUSIONS

MOSESS was designed to represent the processes of runoff and soil erosion at small scales taking into account antecedent conditions. The model was tested at plots and micro-basins in Sumé, in Northeast Brazil from which the following conclusions can be drawn: (a) the model was capable of simulating the processes reasonably well with parameters based on soil texture, and it explained 67 % and 41 % of the observed runoff and soil erosion respectively; (b) the evolution of the soil moisture content with the soil depth and time was consistent as a result of the methods used for calculating actual evapotranspiration and infiltration; (c) the model results compared well with those generated from calibrated event-oriented models, and its further development is encouraged up to the inclusion of the channel system.

Acknowledgements The CT-HIDRO (Brazil) program of scholarship and UFCG are gratefully acknowledged.

REFERENCES

Aragão, R. (2000) Estimativa dos Parâmetros do Modelo Distribuído WESP com os Dados da Bacia Experimental de Sumé, PB. MSc Thesis, Dept of Civil Engineering, Federal University of Campina Grande, PB, Brazil.

Brooks, R. H. & Corey, A T. (1964) *Hydraulic Properties of Porous Media*. Hydrol. Paper 3, Colorado State Univ., Fort Collins, USA.

Cadier, E. & Freitas, B. J. (1982) Bacia representativa de Sumé, primeira estimativa dos recursos de água. *Série Hidrologia* **14**, SUDENE, Recife-PE, Brazil.

Cadier, E., Freitas, B. J. & Leprun, J. C. (1983) Bacia experimental de Sumé – instalação e primeiros resultados. *Série Hidrologia* **16**, SUDENE, Recife-PE, Brazil.

Davi, H. H. R. C. & Figueiredo, E.E. (2003) Um modelo hidrossedimentológico para simulação em pequena escala. In: *Anais of the XV Symposium on Water Res*., Curitiba, PR, Brazil.

Davi, H. H. R. C. (2004) Um modelo para simulação hidrossedimentológica em pequena escala. Msc Thesis, Dept of Civil Engineering, Federal University of Campina Grande, PB, Brazil

Dunne, T. & Leopold, L. B (1978) *Water in Environmental Planning*. W. H. Freeman, San Francisco, USA.

Engelund, F. & Hansen, E. (1967) *A Monograph on Sediment Transport in Alluvial Streams*. Teknisk Vorlag, Copenhagen, Denmark.

Ewen, J., Parkin, G. & O'Connell, P. E. (2000) SHETRAN: Distributed river basin flow and transport modeling system. *J. Hydraul. Engng* ASCE **5**(3), 250–258.

Gupta, V. K., Iturbe, I. R. & Wood, E. F. (1986) *Scale Problems in Hydrology*. Water Science and Technology Library. D. Reidel Publishing Company, Dordrecht, The Netherlands.

Horton, R. E. (1933) The role of infiltration in the hydrologic cycle. *Trans. Am. Geophys. Union* **14**, 446–460.

Lal, R. (1990) *Soil Erosion in the Tropics – Principles and Management*. McGraw-Hill, Inc, London, UK.

Lal, R. (1994) Soil erosion by wind and water: problems and prospects. In: *Soil Erosion – Research Methods* (ed. by R. Lal), 1–9. Soil and Water Conservation Society, St Lucie Press, USA.

Laursen, E. M. (1958). The total sediment load of streams. *J. Hydraul. Div.* ASCE **54**, 1–36.

Lopes, V. L. (1987) A numerical model of watershed and sediment yield. PhD Thesis, University of Arizona, USA.

Lopes, W. T. A. (2003) Efeitos de Escala na modelagem hidrossedimentológica na região semi-árida da Paraíba. MSc Thesis, Dept of Civil Engineering, Federal University of Campina Grande, PB, Brazil.

Rawls, W. J. & Brakensiek, D. L. (1989) Estimation of soil water retention and hydraulic properties. In: *Unsaturated Flow in Hydrologic Modeling – Theory and Practice* (ed. by H. J. Morel-Seytoux), 275–300. Kluwer Academic Publishers, Dordrecht, The Netherlands.

Saxton, K. E., Rawls, W. J., Rosemberger, J. S. & Papendick, R. I. (1986) Estimating generalized soil-water characteristics from texture. *Soil Sci. Soc. Am. J.* **50**, 1031–1036.

Singh, V. P. (1995) Watershed Modelling. In: *Computer Models of Watershed Hydrology* (ed. by V. P. Singh), 1–22. Water Resources Publications, Highlands Ranch, Colorado, USA.

Smith, R. E., Goodrich, D. C., Woolhiser, D. A., & Unkrich, C. L. (1995) KINEROS – A KINematic Runoff and EROSion Model. In: *Computer Models of Watershed Hydrology* (ed. by V. Singh) , 697–732. Water Resources Publications, Highlands Ranch, Colorado, USA.

Vanoni, V. A. (1975) *Sedimentation Engineering*. ASCE-Manuals and Reports on Engineering Practice no. 54. New York, USA.

Wicks, J. M. (1988) Physically-based mathematical modelling of catchment sediment yield. PhD Thesis, Dept of Civil Engineering, Univ. of Newcastle upon Tyne, England, UK.

Yalin, M. S. (1963) An expression for bed-load transportation. *J. Hyraul. Div.* ASCE **89**(HY3), 221–250.