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

Effects of rainfall variability and land-use change on sediment yield simulated by SHETRAN

EDUARDO E. DE FIGUEIREDO[1] & JAMES C. BATHURST[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 *Water Resource Systems Research Laboratory, Department of Civil Engineering, University of Newcastle, Newcastle upon Tyne NE1 7RU, UK*

Abstract In the Cariri region in the semiarid Northeast of Brazil, precipitation varies considerably in space and time, and land use changes as a consequence of deforestation. The effects of these factors on basin responses are qualitatively known, but an assessment of the magnitude of the effects is still needed. In this paper, the physically based and spatially distributed model SHETRAN was parameterized in order to simulate the processes of runoff and sediment yield in various catchments with changes in the annual rainfall totals varying from 400 mm (dry), through 600 mm (normal) to 800 mm (wet), and percentage of basin area deforested from 10 to 90%. The results showed that, overall, peak discharges and volumes at basin areas from 100 m^2 to 137 km^2, and sediment yields for areas up to 11 km^2, increased by 1.5 to 12-fold as annual rainfall changed from dry to wet and deforestation from 10 to 90%, while annual sediment yields for the largest area (137 km^2) increased by 38 times as annual rainfall increased from dry to normal, then to wet.

Key words deforestation; rainfall variability; soil erosion

INTRODUCTION

It is well known that climate characteristics and land use have an effect on the process of soil erosion, which is governed by rainfall and runoff. In semiarid regions runoff varies considerably as a consequence of rainfall variability (Nemec & Rodier, 1979). In the semiarid region of northeast Brazil for instance, the climate is a complex phenomenon that greatly influences rainfall characteristics (Kousky, 1985; Hastenrath, 1990). Rainfall is generally high, with annual totals of 400–800 mm (Cadier, 1996). Land use is characterized by deforestation (Figueiredo, 1998), and runoff and sediment yield vary depending on basin scale (Figueiredo & Bathurst, 2005), channel network and morphology. In order to evaluate the effects of rainfall variability and land-use change on basin responses, the model SHETRAN (Ewen *et al*., 2000) was used to simulate the processes of runoff and sediment yield at various catchments in the Cariri region with annual rainfall totals varying from 400 mm (dry), through 600 mm (normal) to 800 mm (wet), and the percentage of basin area deforested from 10 to 90%.

THE CATCHMENTS IN THE CARIRI REGION

The representative basin of Sumé (Fig. 1) lies within the Cariri region. It contains nine plots (100 m^2), four micro-basins (0.48–1.0 ha) and three basins (10.7–137.4 km^2).

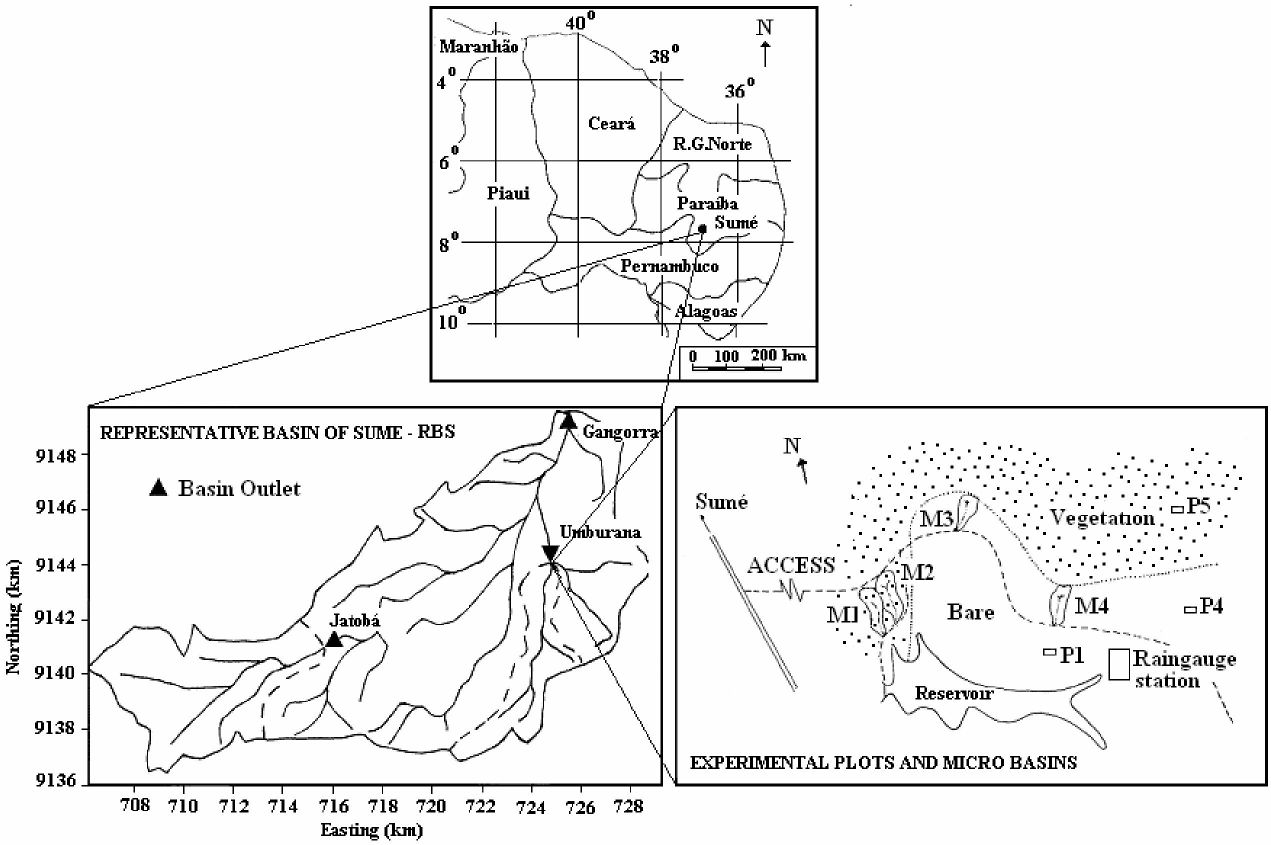

Fig. 1 The northeast region of Brazil (top), RBS (bottom left) and experimental areas (bottom right).

Deforestation is common in the region; rainfall intensities are high and runoff is flashy (Cadier & Freitas, 1982; Cadier *et al.*, 1983). As a result, sediment is increasingly eroded from the surface of a loam soil (50.2% sand; 15.8% clay), which is shallow (0.1 m) and has low permeability (0.31 m day^{-1}). Underneath, lies a sandy clay loam soil (50.2% sand, 32.5% clay) with depth <0.7 m and low permeability (0.03 m day^{-1}). Occasionally alluvium can also be found, which is deeper and permeable (~2 m; 8 m day^{-1}). The vegetation is Caatinga (bushes and typical trees).

SHETRAN

SHETRAN is a physically-based spatially-distributed model that integrates surface and subsurface processes at the catchment scale; it is divided horizontally into grid cells and vertically into nodes within each soil layer. It simulates the runoff and soil erosion processes using a fully implicit finite difference solution of the governing equations. The model equations of interest are given in brief here (for details, see Bathurst *et al.*, 1995).

Interception is based on the equation of Rutter (1971/1972) which depends on the net rainfall (Q), canopy water depth (C), storage capacity (S_c), and drainage parameters (k_c and b_c):

$$\frac{\partial C}{\partial t} = Q - k_c e^{b_c (C - S_c)} \tag{1}$$

The net rainfall varies with the total rainfall (P), maximum and actual ground cover fractions (p_1, p_2), and potential evapotranspiration (E_p) that can be determined with observed data and a relationship $E_a/E_p = f(\psi)$, where E_a is the actual evapotranspiration and ψ the soil water tension. The total actual evapotranspiration is that from the interception, roots (based on a root function R_t) and bare soil.

Infiltration in the unsaturated zone (UZ) is modelled according to Richards (1931):

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}[K \frac{\partial \psi}{\partial z}] + \frac{\partial K}{\partial z} - S(z) \tag{2}$$

where θ = volumetric moisture content, t = time, z = vertical axis, S = source/sink term, $K = K_S\ [(\theta - \theta_r)/(\theta_S - \theta_r)]^{\eta}$ is the hydraulic conductivity (Brooks & Corey, 1964), $K_S = K(\theta_S)$, θ_S and θ_r are the saturated and residual moisture contents, and η is the Averjanov (1950) exponent (Mualem, 1978).

Overland and channel flows are modelled based on the Saint Venant (1871) equations given in terms of the water depth (h), Manning's (1895) flow velocities (u, v) that depend on the Manning-Strickler roughness coefficient (MSRC), the inverse of Manning's coefficient (Chow, 1959), discharges q (overland flow) and q_L (channel flow), and flow area (A):

$$\frac{\partial h}{\partial t} + \frac{\partial (uh)}{\partial x} + \frac{\partial vh}{\partial y} = q \quad \text{Overland flow} \tag{3}$$

$$\frac{\partial A}{\partial t} + \frac{\partial (Au)}{\partial x} = q_L \quad \text{Channel flow} \tag{4}$$

Erosion by rainfall (D_r) is based on the momentum squared of raindrops (equation (5)) and by runoff (D_f) on the sediment initiation of motion (equation (6)):

$$D_r = k_r F_w (1 - C_g)[(1 - C_c) M_r + M_d] \tag{5}$$

$$D_f = \begin{cases} k_f (1 - C_g)[\frac{\tau}{\tau_c} - 1] & \text{if } \tau > \tau_c \\ 0 & \text{if } \tau \le \tau_c \end{cases} \tag{6}$$

where $k_{r,f}$ is rainfall and runoff erosivity coefficients, F_w is a factor for reducing detachment based on the ratio water depth/raindrop diameter; $M_{r,d}$ is momentum squared for rainfall and leaf drip, $C_{g,c}$ are ground and canopy cover fractions, and τ, τ_c are flow and critical shear stresses.

Sediment transport is based on the mass conservation equations (overland and channel flows), and on the flow transport capacity (e.g. Yalin, 1963; Engelund-Hansen, 1967):

$$\frac{\partial (hc)}{\partial t} + (1 - \lambda)\frac{\partial z}{\partial t} + \frac{\partial g_x}{\partial x} + \frac{\partial gy}{\partial y} = 0 \quad \text{Overland flow} \tag{7}$$

$$\frac{\partial(Ac)}{\partial t}+(1-\phi)B\frac{\partial z}{\partial t}+\frac{\partial(AcV_s)}{\partial x}=q_s \quad \text{Channel flow} \qquad (8)$$

where c is sediment concentration, λ is soil porosity, $g_{x,y} = x, y$ transport rates, t is time, z is depth of loose soil, A is flow area, ϕ, B is bed porosity and width and V_s is sediment velocity.

TEST OF THE MODEL

This phase consisted of generating the processes of runoff and sediment yields by modelling the catchments and fixing the input data and model parameters taking into account the main characteristic of land se in the region (i.e. deforestation).

The five units chosen to test the model were the plots P1, P4 (bare cleared) and P5 (with native vegetation), the micro-basins M1 and M2 (0.6–1.0 ha, bare cleared), and M3 and M4 (0.5 ha, with native vegetation), and the sub-basins of Umburana (11 km^2) and the representative basin of Sumé at Gangorra (137 km^2). As the catchments have to be divided horizontally (grid network) and vertically (nodes), different grid resolutions were used for the micro-basins (5 × 5 m to 10 × 20 m) and sub-basins (250 × 250 m to 2 × 2 km), while for the plots, one single active grid cell was set (4.5 × 22.2 m). Soil depths for each grid were set to 0.1 m (A horizon) and 0.7 (B horizon). For the division of the soil column, 1 cm between nodes was set around the root zone and impermeable layer, and 5 cm in between. The root function was set to 0.2 for soil depths from 1 to 4 cm, and to 0.1 for depths from 5 to 10 cm (Weischet & Caviedes, 1993).

The input climate data were break-point precipitations and daily pan evaporation (1982 to 1988 for the plots and micro-basins, and 1977 for the representative basin of Sumé) observed in the study area.

For interception, the ground cover parameters were set to $p_1 = 0.9$ ($p_1 = 0$ for bare soil) and $p_2 = 1.0$, based on information in Cadier *et al.* (1983), and the canopy parameters set to $k_c = 1.7 \times 10^{-5}$ mm s^{-1} and $b_c = 7.77$, based on Jetten (1996) considering the storage capacity $S_c = 0.5$ mm. The actual evapotranspiration was determined using a relationship between the ratio of actual to potential evapotranspiration and soil tension (see Figueiredo & Bathurst, 2005) following Denmead & Shaw (1962), with the potential evapotranspiration based on observed pan evaporation data. The soil tension was modelled (see Figueiredo & Bathurst, 2004) based on texture (Saxton *et al.*, 1986).

The infiltration parameters were set for the soils of the A and B horizons based on texture (Saxton *et al.*, 1986; Rawls & Brakensiek, 1989; Mualem, 1978): $K_s = 0.306$ and 0.057 m day^{-1}, $\theta_r = 0.075$ and 0.112 m^3 m^{-3}, $\theta_s = 0.448$ and 0.488 m^3 m^{-3}, and $\eta = 15$. For the overland and channel flows the MSRC values were set to 50 $m^{1/3}$ s^{-1} (bare surface), 1.0 $m^{1/3}$ s^{-1} (vegetated surface) and 30 $m^{1/3}$ s^{-1} (channels), all based on data in the literature (e.g. Chow, 1959; Woolhiser, 1975; Engman, 1986; Wicks *et al.*, 1992).

The soil erosion parameters (k_r and k_f) were set to 11 s^{-1} kg^{-1} m^2 (Wicks *et al.*, 1992) and 0 because the process of soil erosion is dominated by rainfall in the region, which is generally intense. The soil size distributions were set from sieve analysis of the eroded sediment at places. The soil and channel bed porosities for the sediment transport were set to 0.448. The equations of Yalin (1963) and Engelund-Hansen (1967) were chosen for the surface and channel flow transport capacities.

Parameters of bare and vegetated surfaces were tested at the plots P1, P4 (bare) and P5 (with vegetation), respectively, and used unchanged to generate the processes at similar surfaces, M3 and M4 (bare) and M1 and M2 (with vegetation). For Umburana and the representative Sumé basin, a distribution of vegetation was fixed and the parameter values set accordingly. The results showed that volumes, peak discharges and sediment yields were not well simulated in dry years, probably because of a poor representation of the soil-water functions and the characteristics of rainfall. The peaks were better simulated at the largest basins because of a good representation of the river characteristics and morphology (network, relief, sediment, roughness, etc.), which affect the hydrodynamics of the basins. At the micro-basin scale small rivulets characterize the drainage system, which it was not possible to model. Overall, the model simulated well the observed values (R^2 > 70%), which were well contained by output bounds (>70% of the time), but varied markedly depending on the basin area (Figueiredo & Bathurst, 2001, 2004, 2005).

MODEL SIMULATIONS

Simulations were realized, using the parameter values fixed during the test of the model, to investigate the effects of rainfall variability on catchments responses by simulating the most common annual rainfall (*P*) in the study region, varying from dry (400 mm), through normal (600 mm), to wet (800 mm). In combination, five levels of deforestation were fixed, 10%, 30%, 50%, 70% and 90% of basin area, and the model applied to the plot P4 (100 m^2), micro-basin M4 (4800 m^2), Umburana (10.7 km^2) and the representative Sumé basin at Gangorra (137.4 km^2).

Table 1(a), (b) and (c) shows the results. It is seen that, in general, the simulated annual peak discharges, volumes and sediment yields increased as annual rainfall and deforestation (*D*) increased. Annual peak discharges and volumes at areas from 100 m^2 to 137 km^2, and sediment yields at areas up to 11 km^2, increased by 1.5 to 12-fold as annual rainfall changed from dry to wet and deforestation from 10 to 90%, while annual sediment yields at Gangorra (137 km^2) increased by 38 times as annual rainfall changed from dry to wet. This huge increase in sediment yield at the representative Sumé basin is a result of the great influence of land-use, channel characteristics and morphology, represented by the vegetation distribution, channel and sediment sizes, which affect the flow resistance included in the Saint Venant equations through the MSRC. Faster runoff is associated with high MSRC (smooth surface), and the opposite effect is produced by an increase in surface roughness. For the representative Sumé basin, the channels were considered, with the MSRC set to 30 $m^{1/3} s^{-1}$ for every channel link. Except for the plot scale, the roughness flow resistance was fixed according to the assumed vegetation distribution. Small rivulets characterize the channels of the micro-basin scale and were not modelled. The plot scale was represented as a single grid cell, with the value of the MSRC weighted since it is not possible to represent vegetation distribution in such cases. Channels route the water more rapidly than shallow overland flow does, and an increase in the area covered by channels also decreases the amount of infiltration (Grayson *et al.*, 1992). These aspects may explain the large amount of sediment yield at the larger basin scales.

Table 1 (a) Simulated annual peak discharges (in L s^{-1} P4 and M4, and in 10^3 m^3 s^{-1} Umburana (UMB) and the representative basin of Sumé (RBS).

D (%)	*P* = 400 mm				*P* = 600 mm				*P* = 800 mm			
	P4	M4	UMB	RBS	P4	M4	UMB	RBS	P4	M4	UMB	RBS
90	0.028	1.6	3.3	14.1	0.084	4.1	7.3	32.1	0.13	6.3	11.1	46.2
70	0.027	1.3	2.3	15.1	0.079	3.7	4.9	30.2	0.13	6.3	7.6	48.3
50	0.022	1.2	2.5	9.6	0.071	3.4	5.5	18.6	0.13	6.3	8.8	47.9
30	0.018	1.1	2.1	7.7	0.071	3.1	4.7	16.3	0.13	6.3	7.6	53.6
10	0.019	1.0	1.7	5.8	0.064	2.7	4.0	11.8	0.13	6.3	6.6	39.1
(b) Simulated annual volumes (in m^3 P4 and M4 and in 10^3 m^3 Umburana and RBS)												
90	5.3	278	899	6 400	13.5	706	1 980	14 800	26.2	1 307	3 400	24 200
70	4.4	245	890	6 400	12.2	641	1 850	13 900	26.3	1 320	3 200	23 200
50	2.9	218	831	3 900	11.6	583	1 740	10 800	25.1	1 280	3 100	23 700
30	3.1	195	887	4 800	11.2	540	1 790	10 100	24.7	1 250	3 150	23 600
10	3.2	169	854	3 900	10.8	480	1 730	8 400	24.6	1 200	2 800	17 300
(c) Simulated sediment yield (in t $year^{-1}$)												
90	0.10	3.72	781	330	0.21	8.25	2 643	3 298	0.36	15.76	4 901	7 694
70	0.08	2.90	599	220	0.16	6.31	1 915	2 473	0.28	12.03	3 360	5 633
50	0.06	2.14	556	124	0.11	4.26	1 851	1 786	0.20	7.66	3 360	4 672
30	0.03	1.70	503	110	0.07	3.11	1 541	1 511	0.13	1.70	2 696	5 359
10	0.01	1.15	385	82	0.02	1.92	1 081	1 099	0.10	2.72	2 536	3 985

CONCLUSIONS

The SHETRAN model was used to evaluate the sediment dynamics of several catchments in the semiarid region in Northeast Brazil for given changes in annual rainfall and land use. From the results it can be concluded that: (a) the model explained well (>70%) the observed runoff and sediment yields, which were contained more than 70% of the time by simulated output bounds; (b) the processes increased consistently as annual rainfall and percentage of deforestation increased; (c) the runoff and sediment dynamics at the catchments scales were affected by the channel network and morphology; (d) on the whole, peak discharges and volumes at all basin areas studied, and sediment yields of areas up to 11 km^2, increased similarly (1.5- to 12-fold) while for the representative basin of Sumé, areal annual sediment yields increased by 38 times as annual rainfall changed from dry to normal, then to wet and deforestation increased from 10–90%.

Acknowledgements The funding agency CNPQ (Brazil) and UFCG supported this research and is gratefully acknowledged.

REFERENCES

Averjanov, S. F. (1950) About permeability of subsurface soils in case of incomplete saturation. *Engng Collect.* **7**, 1950.

Bathurst, J. C., Wicks, J. M. & O'Connell, P. E. (1995) The SHE/SHESED basin scale water flow and sediment transport modelling system. In: *Computer Models of Watershed Hydrology* (ed. by V. P. Singh), 563–594.Water Resources Publications, Highlands Ranch, Colorado, USA.

Brooks, R. H. & Corey, A T. (1964). Hydraulic properties of porous media. Hydrol. Paper 3, Colorado State Univ., Fort Collins, USA.

Cadier, E. (1996) Hydrologie des petits bassins du Nordeste Brésilien semi-aride: typologie des bassins et transposition écoulements annuels. *J. Hydrol*. **182**, 117–141.

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.

Chow, V. T. (1959) *Open-Channel Hydraulics*. McGraw-Hill Int., New York, USA.

Denmead, O. T. & Shaw, R. H. (1962) Availability of soil water to plants as affected by soil moisture content and meteorological conditions. *Agronomy J.* **54**, 385–390.

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

Engman, E. T. (1986) Roughness coefficients for routing surface runoff. *J. Irrig. Drain. Engng ASCE* **112**(1), 39–53.

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.

Figueiredo, E. E. (1998) Scale effects and land-use change impacts in sediment yield modelling in a semi-arid region of Brazil. PhD Thesis, Dept. of Civil Engineering, University of Newcastle upon Tyne, UK.

Figueiredo, E. E. & Bathurst, J. C. (2002) Sediment yield modelling at various basin scales in a semiarid region of Brazil using SHETRAN. In: *Fifth Int. Conf. on Hydroinfomatics* (ed. by R. A. Falconer, E. L. Harris, B. Lin, C. Wilson, I. D. Cluckie, D. Han, J. P. Davis & S. Heslop), 316–321. IWA.

Figueiredo, E. E. & Bathurst, J. C. (2004) Uncertainty analysis in up-scaling the SHETRAN model parameters. In: *Sixth Int. Conf. on Hydroinfomatics* (ed. by Shie-Yui Liong, Kok-Kwang Phoon &Vladan Babovic), (2), 1737–1744. World Scientific Publishing Company.

Figueiredo, E. E. & Bathurst, J. C. (2005) Relationship between simulated sediment yield and scale in a semiarid region of Brazil. In: *Sediment Budgets 2* (ed. by A. J. Horowitz & D. E. Walling) (Proc. Symp. S1 at Foz do Iguaçu, Brazil, April 2005), 110–118. IAHS Publ. 292. IAHS Press, Wallingford, UK.

Grayson, R. B., Moore, I. D. & McMahon, T. A. (1992) Physically based hydrologic modelling. 1: A terrain-based model for investigative purposes. *Water Resour. Res*. **28**(10), 2639–2658.

Hastenrath S. (1990) Prediction of northeast Brazil rainfall anomalies. *J. Climate* **3**(8), 893–904.

Jetten, V. G. (1996) Interception of tropical rain forest: performance of a canopy water balance model. *Hydrol. Processes* **10**, 671–685.

Kousky, V. E. (1985) Atmospheric circulation changes associated with rainfall anomalies over tropical Brazil. *Mon. Weather Rev*. **113**(11), 1951–1957.

Manning, R. (1895) On the flow of water in open channels and pipes. *Trans. Inst. Civil Engineers of Ireland* **20**, 161–207. Dublin, 1891; supplement 24, 179–207, 1895.

Mualem, Y. (1978) Hydraulic conductivity of unsaturated porous media: generalized macroscopic approach. *Water Resour. Res*. **14**(2), 325–334.

Nemec, J. & Rodier, J. A. (1979). Streamflow characteristics in areas of low precipitation (with special reference to low and high flows). In: *The Hydrology of Areas of Low Precipitation* (Proc. Camberra Symposium), 125–140. IAHS Publ. 128. IAHS Press, Wallingford, UK.

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.

Richards, L. A. (1931) Capillary Conduction of liquids through porous mediums. *Physics* **1**, 318–333.

Rutter, A. J., Kershaw, K. A., Robins, P. C. & Morton, A. J. (1971/1972) A predictive model of rainfall interception in forests. 1. Derivation of the model from observations in a plantation of Corsican pine. *Agric. Meteorol*. **9**, 367–384.

Saint-Venant, A. J. C. Barre de (1871) Théorie du mouvement non permanent des eaux, avec application aux crues des rivières et à l'introduction des marées dans leur lits. *Comptes rendus des séances de l'Académie des Sciences* **73**, 147–154 & 237–240. Paris.

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.

Weischet, W. & Caviedes, C. N. (1993) *The Persisting Ecological Constraints of Tropical Agriculture*. Longman Scientific & Technical, New York, USA.

Wicks, J. M., Bathurst, J. C. & Johnson, C. W. (1992) Calibrating SHE soil-erosion model for different land covers. *J. Irrig. Drain. Engng* **118**(5), 708–723.

Woolhiser, D. A. (1975) Simulation of unsteady overland flow. In: *Unsteady Flow in Open Channels* (ed. by K. Mahmood & V. Yevjevich). Water Research Publications, Fort Collins, USA.

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