Modelling Soil Erosion, Sediment Transport and Closely Related Hydrological Processes (Proceedings of a symposium held at Vienna, July 1998). IAHS Publ. no. 249, 1998.

A physically-based sediment delivery model for arid regions

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Abstract A sediment delivery model has been developed for estimating the sediment delivery rates in an arid upland basin. The model uses a steady-state sediment continuity equation and a first-order reaction model for deposition, since the initial potential sediment load is greater than the overland flow transport capacity—calculated by the Yalin method—in the arid regions. Spatial sediment delivery is analysed through preprocessing by GIS, calculation of sediment delivery, postprocessing and display of spatial output in the GIS. This study enables the identification of vulnerable regions within a drainage basin, and thus facilitates improvement in planning soil conservation systems.

INTRODUCTION

This paper summarizes the outcome of a cooperative project, and the main contributors are listed in the Acknowledgements section. The geographic information systems (GIS) are being integrated with hydrological models to quickly assemble model input data and store model output for analysis and display (Vieux, 1991). De Vantier & Feldman (1993) reported that erosion potential prediction was a practical and widely applied GIS operation. However, the study was based on the USLE and therefore gave general information on the potential soil erosion in a drainage basin. The GIS technology has also been integrated with ANSWERS (De Roo *et al.*, 1989), and AGNPS (Engel *et al.*, 1993), but the applicability in arid regions was limited by the requirement for large amounts of high-quality data. In this study, a distributed parameter sediment delivery model is derived and linked with raster-based GIS to predict the spatial sediment delivery within an arid zone drainage basin for the identification of vulnerable areas to use in planning and designing soil conservation systems. These results also can be used with a digital elevation model (DEM) to display distributed parameter model results.

GOVERNING EQUATIONS

Sediment continuity equation

Sediment movement downslope obeys the principle of continuity of mass expressed by (Nearing *et al.*, 1989):

$$\frac{\mathrm{d}q_s}{\mathrm{d}X} = D_f - D_i \tag{1}$$

where q_s (kg s^{-1} m^{-1}) is sediment transport rate per unit width, X (m) is downslope distance, D_f (kg s^{-1} m^{-2}) is net flow detachment rate and D_i (kg s^{-1} m^{-2}) is net rainfall detachment rate. The assumption of quasi-steady state allows equation (1) to be written without an explicit time parameter. If D_i is assumed to be small (Lü *et al.*, 1989), then equation (1) can be written as:

$$\frac{\mathrm{d}q_s}{\mathrm{d}X} = D_f \tag{2}$$

The net flow detachment rate D_f is positive for detachment and negative for deposition. In the arid regions, since the initial potential sediment load is greater than the sediment transport capacity (Foster *et al.*, 1980; Jones, 1981) deposition is assumed to occur at a rate of:

$$D_f = G(T_c - q_s) \tag{3}$$

This relationship is a diffusion type equation (Foster & Meyer, 1972) where G (m^{-1}) is a first-order reaction coefficient and T_c (kg s^{-1} m^{-1}) is flow transport capacity.

Hydrological inputs

The flow depth is estimated by Manning's equation as:

$$h = (qns_c^{-0.5})^{0.6} \tag{4}$$

where h (m) is overland flow depth, q (m^3 s^{-1} m^{-1}) is flow discharge per unit width, n is Manning's roughness and is equal to 0.046 (for moderate vegetative cover and rough surface/depressions of 10 to 15 cm depth, a moderate value; Foster *et al.*, 1980) and s_c is mean bed slope.

Although the Darcy-Weisbach equation, with a varying friction factor for laminar flow, might be more accurate for calculation of depth in some cases, most users are better acquainted with estimating Manning's n. The error in estimating a value for n is probably greater than the error in using Manning's equation for laminar flow.

Flow shear stress

Shear stress action on the channel bed, τ_s (kg m^{-1} s^{-2}), is calculated using the equation:

$$\tau_s = \gamma h s_c \tag{5}$$

where γ (kg m^{-2} s^{-2}) is specific weight of water.

Sediment transport capacity

Several generalized formulae have been developed for computing the sediment transport capacity, T_c. Many of the equations were developed for streams, and were later applied to shallow overland and channel flows. However, Alonso *et al.* (1981), who evaluated nine sediment transport equations, concluded that the Yalin equation (Yalin, 1963) provided reliable estimates of transport capacity for shallow overland flow and streamflow. Foster & Meyer (1972) also concluded that the Yalin equation was the most appropriate for the shallow flows associated with upland erosion.

The Yalin equation is defined as:

$$\frac{T_c}{SG \cdot d_w^{0.5} \tau_s^{0.5}} = 0.635\delta\left[1 - \frac{1}{\beta}\ln(1+\beta)\right] \tag{6}$$

where β, δ and Y are expressed as:

$$\beta = 2.45(SG)^{-0.4}(Y_{cr})^{0.5} \tag{7}$$

$$\delta = (Y/Y_{cr}) - 1 \text{ (when } Y < Y_{cr},\ \delta = 0) \tag{8}$$

$$Y = (\tau_s/\rho_w)/(SG - 1)gd \tag{9}$$

and where SG is particle specific gravity (2.65 for fine sand and silt), d (m) is particle diameter, ρ_w (kg m^{-3}) is mass density of water, Y is dimensionless shear stress, Y_{cr} is dimensionless critical shear stress from Shield's diagram (revised as per Abrahams *et al.* (1988) for the overland flow on desert hillslopes) and g (m s^{-2}) is acceleration of gravity. The modified Yalin equation which considers a mixture of particles of varying size and density (Foster, 1982) was used.

Sediment delivery model

Combining equations (2) and (3), the sediment delivery model was written as:

$$\frac{\mathrm{d}q_s}{\mathrm{d}X} + Gq_s - GT_c = 0 \tag{10}$$

The solution of equation (10) is:

$$\ln(T_c - q_s) = -GX + \ln C \tag{11}$$

where C (kg s^{-1} m^{-1}) is a constant of integration and is equal to $T_c - q_s$ at $X = 0$. Thus, C is the difference between sediment transport capacity and the actual sediment transport at the point of initiation of runoff within the drainage basin.

Water balance model

The water balance model, SWAMIN (Huygen, 1993), is a vertically one-dimensional model used to describe the time sequential distribution of precipitation among surface runoff, soil moisture storage and deep percolation. The SWAMIN model is a variant of the water balance simulation model, SWATRE (Belmans *et al.*, 1983).

Feddes *et al.* (1993) described the one-dimensional vertical flow of water, in a homogeneous and isotropic soil profile, in terms of soil water pressure head (negative in unsaturated soil), as:

$$\frac{\delta h}{\delta t}=\left[\frac{1}{C(h)}\right]\frac{\delta}{\delta z}\left\{K(h)\left[\frac{\delta h}{\delta z}+1\right]\right\} \tag{12}$$

where h (mm) is soil water pressure head, t (s) is time, $C(h)$ is differential moisture capacity ($\delta\theta/\delta h$), θ (mm^3 mm^{-3}) is volumetric soil water content, z (mm) is vertical coordinate, with the origin at the soil surface, directed positive upwards and K (mm s^{-1}) is hydraulic conductivity. Equation (12) is solved by an implicit finite difference scheme that applies an explicit linearization. The upper boundary condition is defined by the saturated hydraulic conductivity at the soil surface and the bottom boundary condition is free drainage to the underground. The model output consists of instantaneous rainfall excess, flux at the bottom of soil profile and change in soil moisture storage.

STUDY DRAINAGE BASIN

The study drainage basin, Divisadero Largo (5.47 km^2), is located within the piedmont and precordilleran areas of the Andes mountains to the west of Mendoza (33.0–33.5°S, 68.8–69.1°W), Argentina (Fig. 1). It has an average width of 600 m, a flow length of about 9 km, and a time of concentration of about 3 h. The altitude ranges from 950 m in the east to 1450 m in the west. The basement of the area is formed by Triassic and Tertiary sediments and these are covered discordantly by

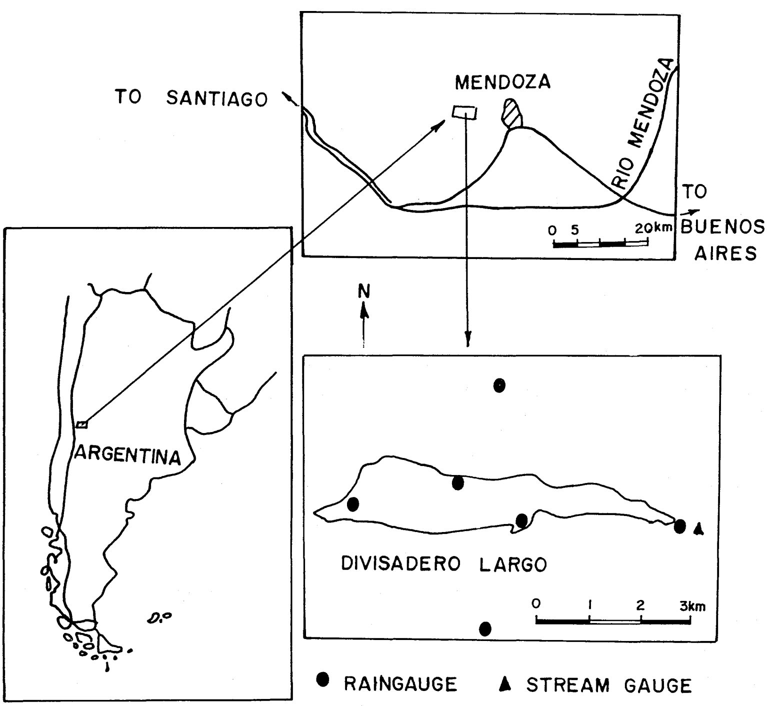

Fig. 1 Location of the Divisadero Largo basin and the gauging stations.

alluvial fan deposits of Quaternary age. The region is intersected by steeply eroded gullies and rocky outcrops. The soils are shallow, undeveloped and consist of medium to fine sand. The vegetation consists of low shrubby pastures ranging from 5 to 45% cover, with less cover on steep slopes.

The area lies in a subtropical arid climate and is characterized by convective summer thunderstorms that cause flash floods and heavy soil erosion. The annual average precipitation is 201 mm, 77% of which is received in the summer months of October–March. Typical of the desert climate, the rainfall is characterized by relatively high intensity and extreme variability, both from year to year and within the rainy season. The average annual temperature is 13°C. The hydrological network consists of four automatic raingauges and one runoff gauging station and the rainfall and runoff data have been recorded through a telemetry network since 1983.

MODEL CALIBRATION

The soil characteristics, such as representative particle diameter (0.22 mm) and specific gravity of the soil particle (2.65) were taken from a detailed soil analysis (Ligtenberg *et al.*, 1992). The model parameters G and C are associated with sediment delivery. According to equation (11), the sediment delivery is proportional to the difference between the transport capacity and the actual transport. Fitting a least squares technique to the sediment delivery data recorded in micro-plots within the Divisadero Largo basin, the parameter G was found to be 0.036 m^{-1}. Singh & Regl (1983) suggested a reasonable value of G as 0.030 m^{-1}. The parameter C was found to be equal to 0.73 kg s^{-1} m^{-1}, which was well within the range of 0.34–2.42 kg s^{-1} m^{-1} as reported by Sharma *et al.* (1996) for arid regions.

HRU DELINEATION USING GIS OVERLAY ANALYSIS

The hydrological response units (HRU) are distributed, heterogeneously structured areas with common land use and pedo-topo-geological conditions controlling their unique hydrological dynamics (Beven, 1989). The water balance model, SWAMIN, and the sediment delivery model (equation (11)) are valid only for HRU which have uniform characteristics of soil, slope, vegetation and rainfall distribution pattern.

The GIS database was generated by digitizing maps of soil types and land use, and by importing a digital elevation model (DEM) into the GIS. The HRU were delineated by GIS analysis as follows:

(a) Using topographic map of the study area at 1:5000 scale, every 5 m contour was digitized and interpolated to a continuous elevation map or DEM with 30 * 30 m grid size. Slopes and aspects were derived from the DEM.

(b) Aerial photographs on 1:25 000 scale were used in combination with the field survey to produce a soil map of the study drainage basin. This soil map was digitized to a 30 * 30 m grid and converted from vector format to the raster format.

(c) A ratio vegetation index map was produced using a Landsat Thematic Mapper image acquired on 22 February 1986 (path 232 and row 083). The maximum vegetative cover occurs during February, which is in the middle of the rainy season.

(d) Area mean daily precipitation was calculated from Thiessen polygons and was applied uniformly to all HRU.

During the GIS overlay analysis, generated subclasses with small areas were merged with similar larger classes based on the insight gained by the hydrological system analysis. Using the entire GIS database, 26 HRU were delineated within the study drainage basin.

INTEGRATION FROM POINT TO DRAINAGE BASIN SCALE

Since pixels in this study were small, only 30 * 30 m, the point estimates of rainfall excess and sediment transport capacity were used as area representative values over each pixel. While analysing the drainage basin using GIS package PC-RASTER (van Deursen & Wesseling, 1992), the subroutine WATERSHED keeps track of the flow path travelling through each of the pixels using the downslope direction of the steepest gradient. The flow path connects one pixel with its downstream pixel up to the drainage basin outlet. On each pixel, the soil available for delivery was the material detached on that pixel plus the material carried to it from the pixel upstream. This sum was compared with the transport capacity at that pixel. If the total soil available for transport was less than the transport capacity, the sediment load carried to the downstream pixel equalled the amount of available material. However, if the transport capacity was less than the soil available for delivery, the sediment load equalled the transport capacity. The sediment delivery was linked through the channel pixel using the same method as in the overland flow pixel. This procedure continued up to the outlet, and the sediment load at the outlet equalled the total sediment delivery from the drainage basin.

RESULTS AND DISCUSSION

The sediment delivery model (equation (11)) was validated on 26 discrete rainfall events for which the soil loss data were recorded at the drainage basin outlet. The rainfall amount varied from 6 to 53 mm in 35 min to about 10 h in duration. The rainfall intensity for a 5-min period ranged between 12 and 168 mm h^{-1}.

A comparison of measured and predicted sediment delivery shows a good agreement (Fig. 2). With a coefficient of determination of 0.98 ($p > 0.01$), the predicted and observed sediment delivery relationship was:

$$Q_s' = 1.21Q_s - 0.02 \qquad (13)$$

where Q_s' (kg m^{-2}) is predicted sediment delivery and Q_s (kg m^{-2}) is observed sediment delivery. For the model verification, the relative error in the predicted sediment delivery (E_s) was calculated by the relation:

$$E_s = (Q_s' - Q_s)/Q_s \qquad (14)$$

The average E_s was found to be 6.1%, the maximum was 16.5% and the minimum

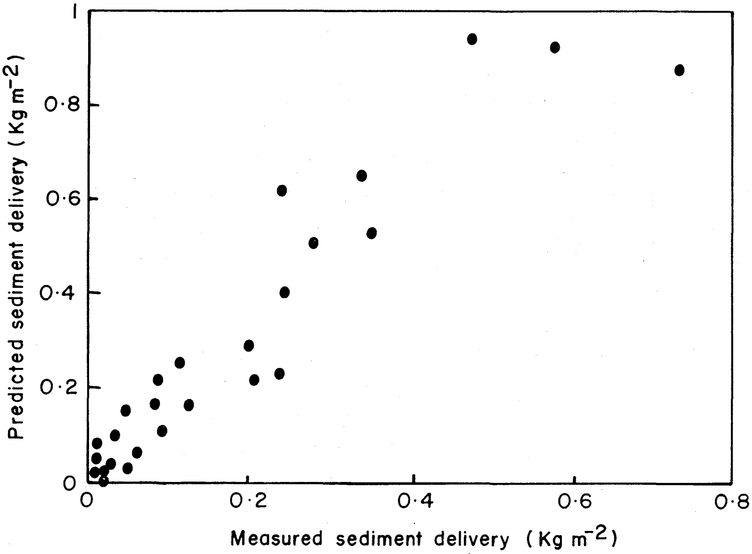

Fig. 2 Measured *vs* predicted sediment delivery in the Divisadero Largo basin.

was only 2.2%.

The sediment delivery model (equation (11)) in conjunction with GIS has a capability to predict the spatial variability of sediment delivery within a drainage basin. Such information is useful in the identification of vulnerable areas within a drainage basin. Apart from the rainfall, the practical impossibility of knowing in sufficient detail the surface characteristics of the drainage basin adds to the spatial variability of sediment delivery in the arid regions (Sharma *et al.*, 1994).

CONCLUSION

Interfacing GIS technology and a distributed parameter sediment delivery model for identification of vulnerable areas within a drainage basin is successful. Preliminary results indicate that the model described herein can simulate the spatial sediment delivery, although additional validation must be conducted. Further research focuses on the validation of spatially distributed hydrological response data such as pixel overland flow depth, average pixel infiltration, or pixel flow transport capacity.

Acknowledgements The work summarized here was undertaken by a team including K. D. Sharma and F. Cazorzi, N. C. Ciancaglini, G. d'Urso, G. Dalla Fontana, P. C. Fernandez, L. Fornero, J. Huygen, M. A. Manzanera, J. Maza, M. Menenti, J. A. Morabito, J. A. Pazos, E. P. Querner, A. Vich and J. M. Juluaga.

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