A Distributed Spatial Sediment Delivery Model for the Arid Regions

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ABSTRACT

A distributed parameter sediment delivery model has been derived from basic-principles and linked with a personal computer-based, low cost, geographic information system (GIS) to facilitate preparation, examination and analysis of spatially distributed input parameters as well as to link the sediment delivery from a micro-scale to the drainage basin-scale. The heterogeneous and complex land surface within the drainage basin divided into sub-areas. Data on vegetative cover are derived from the digital analysis of satellite images. Spatial sediment delivery in arid drainage basins in Argentina and India is analyzed through preprocessing employing GIS, calculation of sediment delivery, and post processing and display of spatial output in the GIS. The model accurately predicts, even for the same value of flow shear stress, the higher sediment delivery from bare land surfaces as compared to land surfaces protected by vegetative cover. The study further enables the identification of vulnerable regions within a drainage basin, thus facilitating improvements in planning of soil conservation systems.

INTRODUCTION

Arid regions have a potential for generating and transporting large quantities of sediment (Schick, 1970) mainly due to the torrential rainfall (Bell, 1979), weathering (Goudie and Wilkinson, 1977), almost total lack of natural protection against soil erosion due to sparse vegetative cover (Magfed, 1986; Pilgrim *et al.,* 1988), aeolian surficial deposits providing readily available material to be eroded by runoff (Jones, 1981) and biotic interference (FAO, 1973). Models of sediment delivery by water may dynamically route the sediment by solving the continuity equation for sediment transport (Bennett, 1974). The solution of this equation is generally accomplished using numerical methods in association with a water balance model, which provides the required hydrologic inputs. These dynamic models provide estimates of total soil loss from a drainage basin and predict the sediment delivery by considering the processes of soil detachment, transport and deposition.

However, the use of numerical methods is limited because of high cost, stability and convergence problems, uncertainty concerning friction losses, parameterization problems and inherent variability. In data deficit arid regions, a closed-form solution to the governing differential equation under steady state conditions is preferred. The closed-form solution reduces computational time and alleviates instabilities associated with numerical solutions.

Spatial heterogeneities in topography, soil properties, vegetation and land use within a drainage basin (BAHC, 1993) cause area differences in rainwater infiltration, runoff generating processes and sediment delivery. Geographic information systems (GIS) are becoming more widely used in spatial hydrologic modeling. The GIS are being integrated with hydrologic and hydraulic models to quickly assemble model input data and store model output for analysis and display (Vieux, 1991). Jett *et al.* (1979) discussed the use of slope, soil and land use data from a GIS for parameterizing hydrologic models. Logan *et al.* (1982), Pelletier (1985), Hession and Shanholtz (1988) and Gonzalez *et al.* (1990) used GIS for analysis of soil erosion by water employing USLE. De Vantier and Feldman (1993) reported that soil erosion potential prediction was a practical and widely applied GIS operation. In those studies, the GIS provided an important spatial/analytical function performing the time consuming geo-referencing and spatial overlays to develop the model input data. 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 is limited by the requirement of large amounts of high-quality data. In this chapter a distributed parameter sediment delivery model has been developed and is linked with raster-based GIS to predict the spatial sediment delivery in arid regions (Sharma *et al.*, 1996 a, b). These results can also be used with a digital elevation model (DEM) to display distributed parameter model results. Linking a GIS with a distributed parameter sediment delivery model identifies vulnerable areas to use in planning and designing soil conservation systems and allows realistic answers to questions such as "where are high soil erosion hazard zones located in a drainage basin"?

THEORY

Sediment Continuity Equation

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

$$\partial q_s/\partial X = D_f + D_i \quad (1)$$

where q_s (kg s^{-1} m^{-1}) is sediment transport rate per unit of width, X (m) is down slope 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 (Lu *et al.,* 1989), equation (1) can be written as:

$$\partial q_s/\partial X = D_f \quad (2)$$

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, 1982; Jones, 1981) deposition is assumed to occur at a rate:

$$D_f = G\,(T_c - q_s) \quad (3)$$

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

Hydrologic Inputs

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

$$h = (qnS_c^{-0.5})^{0.6} \quad (4)$$

where h (m) is overland flow depth, q (m^3 s^{-1} m^{-1}) is flow discharge per unit of width, n is Manning's roughness coefficient and is equal to 0.046 (for moderate vegetative cover and rough surface / depressions of 10 to 15 cm depth, a moderate value; Foster, 1982) and S_c is mean bed slope. Although the Darcy-Weisbach equation with a varying friction factor for laminar flow might me more accurate for the 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 the 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:

$$\tau_s = \gamma\, h\, S_c \quad (5)$$

where γ (kg m^{-2} s^{-2}) is the 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) after evaluating nine sediment transport equations, concluded that the Yalin equation (Yalin, 1963) provided reliable estimates of transport capacity for shallow overland flow. Foster and Meyer (1972b) also concluded that the Yalin equation was the most appropriate for the shallow flows associated with soil erosion.

The Yalin equation is defined as:

$$T_c/(SG)d\delta_w^{0.5}\tau_s^{0.5} = 0.635\delta\{1-(1/\beta)\ln(1+\beta)\} \quad (6)$$

with β, δ and Y are expressed as:

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

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

$$Y = (\tau_s/\delta_w)/(SG-1)gd \quad (9)$$

Where SG is particle specific gravity (2.65 for fine sand and silt), d (m) is particle diameter, δ_w (kg m^{-3}) is mass density of water, Y is dimensionless shear stress, Y_{cr} is dimensionless critical shear stress read from the Shield's diagram, revised as per Abrahams *et al*., (1988) for the overland flow on desert hill slopes, g (m s^{-2}) is acceleration of gravity and β and δ are dimensionless parameters as defined by equations (7) and (8), respectively. The modified Yalin equation, which considers a mixture of particle of varying size and density (Foster, 1982), was used.

Sediment Delivery Model

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

$$\partial q_s/\partial X + G(q_s - T_c) = 0 \quad (10)$$

The closed-form solution of equation (10) is:

$$\ln((T_c - q_s) = -G\,X + \ln(C) \quad (11)$$

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; Sharma *et al*., 1996c), 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 (Feddes *et al*., 1978; Belmans *et al*., 1983).

In its simplest form the water balance states that in a given volume of soil the difference between the amount of water added and the amount of water withdrawn during a certain period is equal to the change in water content during the same period, i.e.:

$$\Delta W = W_{in} - W_{out} \tag{12}$$

where ΔW (m) is change in water content, W_{in} (m) is amount of water added and W_{out} (m) is amount of water withdrawn. The interception of rainfall by the vegetation canopy, transpiration of water by plant communities and evaporation from the bare soil surface were assumed negligible in comparison with the amount of water being redistributed in the soil-plant-atmosphere system during a rainstorm. Thus, a water balance equation can be expressed for each time step Δt (s) as:

$$\Delta W = (P - Q + q)\Delta t^{j} \tag{13}$$

where P (m) is precipitation, Q (m) is rainfall excess equal to the surface runoff, q (m) is flux through the bottom of soil compartment (positive upwards and negative downwards) and j is the time index. The quantity ΔW can be calculated as:

$$\Delta W = \Sigma^{m} \theta \{^{+1}\Delta Z\}^{+1} - \Sigma^{m} \theta_i^j \Delta Z_i^j \tag{14}$$

where m denotes the total number of compartments in the soil layer (Belmans *et al.*, 1983), Z (m) is the depth below the surface, θ (m m^{-1}) is local moisture content and i is the soil depth index. The superscript +1 refers to the next step in the calculation. For each time step all values, except q are known, hence q can be calculated as:

$$q = (\Delta W/\Delta t^{j}) - P + Q \tag{15}$$

To calculate the unknown terms of the water balance equation, it is essential to describe the process of water movement within the soil. The equation of continuity for vertical flow in a soil column is:

$$\partial\theta/\partial t = -\partial q/\partial Z \tag{16}$$

The equation of motion for vertical movement of soil moisture is:

$$Q = -K(h)\ \{(\partial h/\partial Z) + 1)\} \quad (17)$$

where K (m s^{-1}) is the hydraulic conductivity and h (m) is the soil water pressure head, which is negative in the unsaturated zone. Equations (16) and (17) can be combined in a single prognostic equation:

$$\partial h/\partial t = [1/C(h)](\ \partial/\partial Z)\{K(h)[(\partial h/\partial Z) + 1]\} \quad (18)$$

which is often referred to as Richard's Equation. Here C(h) (m^{-1}) is differential moisture capacity and is equal to $\partial\theta/\partial h$. Equation (18) is a combination of conservation of mass equation and Darcy's law and represents a partial differential equation of unsaturated flow in an isotropic non-swelling soil. As K(h) and C(h) are nonlinear functions of h, equation (18) must be solved numerically. Equation (18) is solved by an implicit finite difference scheme that applies an explicit linearization, as proposed by Haverkamp *et al*., (1977) with a time step of one minute and a maximum simulation period of one day. 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. The SWAMIN model considers up to five distinct soil layers.

The sediment delivery model (equation 11) has been evaluated through two independent case studies each in Argentina and India in order to establish its efficacy.

CASE STUDY – 1

Study Area

The Divisadero Largo (5.47 km^2) drainage basin is located within the piedmont and precordilleran region of the Andes Mountains in western Mendoza (33.0-33.5^{o}S, 68.8-69.1^{o}W), Argentina (Figure 1). The region is intersected by steeply eroded gullies and rock outcrops. The soils are shallow, undeveloped and consist of medium to fine sand and vegetation is composed of low shrubby pastures ranging from 5 to 45 % cover.

The area lies in a subtropical arid climate and is characterized by convective summer thunderstorms that cause flash floods and high sediment delivery rates. The annual average precipitation is 201 mm, 77 % of which is received within the summer months of October to March. The hydrological network consists of four automatic raingauges and one runoff gauging station; the data have been recorded through a telemetry network since 1984 (Fernandez *et al.*, 1984). The sediment delivery for each runoff event was calculated from the sediment concentrations obtained from an automated sampler installed at the drainage basin outlet.

Data Sources and Procedures

Topographic maps, aerial photographs, satellite images, published literature, field observations, commercially available GIS packages such as IDRISI (IDRISI, 1992) and PC-RASTER (van Deursen and Wesseling, 1992) and custom written software were used in the study.

Contours at 5-m intervals were digitized from the topographic map of the Divisadero Largo drainage basin on a Tektronix workstation and interpolated to a 1-m vertical resolution, 1:5000 digital elevation model (DEM) with a 30- X 30-m sub-areas. Importing the DEM in PC-RASTER, and using a 3 X 3 kernel calculated the topographic characteristics such as slope and slope direction. A ratio vegetation index (RVI) map was produced using a Landsat Thematic Mapper (TM) image acquired on 22 February 1986 (path 232 and row 083), when the vegetation density is the maximum. Table 1 gives RVI in relation with fractional vegetative cover and digital numbers for the study drainage basin. The aerial photographs of the series ‘Cuencas Aluvionales Gran Mendoza’ of the

Table 1. Characteristics of vegetation in the Divisadero Largo drainage basin

Characteristics of vegetation	Class			
	1	2	3	4
Ratio vegetation index	< 40	41-45	46-48	> 48
Fractional vegetation cover (%)	0.5	5-20	20-35	> 35
Digital number	20	40	60	80

Direccion Privincial de Geodecia y Catastros Mendoza, Argentina, were used in combination with ground truth to produce a map of the hydrological soil groups (USDA-Soil Conservation Service, 1972). The proportion of the hydrological soil groups B, C and D in the drainage basin is about 30, 40 and 30 %, respectively. The Thiesen polygons were drawn from the coordinates of raingauge stations and digitized on the Tektronix workstation.

The water balance model, SWAMIN, is valid only for hydrologic response units (HRU), which have uniform characteristics of soil, slope, vegetation and rainfall distribution pattern. In the present study, individual map layers representing soil type, slope, vegetation density and rainfall distribution pattern were converted to a common scale, geo-referenced and overlaid in IDRISI to formulate 30 X 30 m sub-areas. Attributes were then averaged over the sub-areas, and the results of 26 HRU delineated within the Divisadero Largo drainage basin were stored in a data file to be read prior to the actual simulation. Figure 2 shows the flow diagram of GIS application in the study.

While analyzing the drainage basin in PC-RASTER, the subroutine WATERSHED keeps track of the flow path traveling through each sub-area using the down-slope direction of the steepest gradient. The flow path connects one sub-area with its downstream sub-area up to the drainage basin outlet thereby generating the local drainage direction map. On each sub-area, the sediment available for delivery was the soil detached on that sub-area plus the sediment carried to it from the upstream sub-area. This sum was compared with the transport capacity at that sub-area. If the total sediment available for transport was less than the transport capacity, the sediment delivery to the downstream sub-area equaled the amount of available sediment. However, if the transport capacity was less than the sediment available for delivery, the sediment delivery equaled the transport capacity. The sediment delivery was linked through the channel sub-area using the similar procedure as in the overland flow sub-area. This procedure continued to the drainage basin outlet, and the sediment load at the outlet equaled the sediment delivery from the drainage basin.

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). Vich *et al.*, (1983) assessed the sediment delivery from 10 smaller plots, each 2 X 10 m, maintained under different stages of plant cover in the Divisadero Largo drainage basin. Fitting a least squares technique to the sediment delivery data, the parameters G and C were found to be 0.036 m^{-1} and 0.73 kg s^{-1} m^{-1}, respectively. These values are in agreement with the similar values obtained by Singh and Regl (1983) and Sharma *et al.*, (1993) for the arid regions.

Model Validation

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

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

$$Q_s' = 1.21\ Q_s - 0.02 \qquad (19)$$

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 for the predicted sediment delivery (E_s) was calculated by the relationship:

$$E_s = (Q_s' - Q_s)/Q_s \qquad (20)$$

The average E_s was found to be 6.1 %, the maximum was 16.5 %, and the minimum was only 2.2 %.

An exponential relationship (Figure 3) was found between the actual sediment delivery and shear stress acting on the soil, calculated using equation (5). The relationship in Figure 3 indicates the occurrence of two kinds of surfaces within the drainage basin – one, a highly erodible surface such as bare medium to fine sand; and another, a surface resistant to soil erosion such as a soil surface protected by vegetation. As an example, at 1.0 kg s^{-2} m^{-1} shear stress, the former loses 0.22 kg m^{-2} sediment whereas the latter loses only 0.01 kg m^{-2} sediment.

CASE STUDY – 2

Study Area

The Luni River and its tributaries form the only integrated drainage system (34,866 km^2) in the arid northwest India. It originates in the Arawalli hill ranges near Ajmer (20.45°N, 74.65°E) and flows into a swamp near the Gulf of Kutchch. The elevations in the Luni

Basin ranges from 886 m at the source to 10 m at the outlet; the gradient of the channels vary from 0.0006 to 0.0034 m m^{-1}. Fifty-two percent of the basin is rugged mountainous terrain comprised of hilly and rocky piedmont of igneous and metamorphic rocks of Precambrian and Paleozoic age; the rest of the basin comprised of Pleistocene alluvium and Holocene sand ranging from a 1 to 40 m depth.

Annual precipitation in the Luni Basin ranges from 600 mm in the southeast to 300 mm in the northwest; 80 % of rainfall occurring during the summer months of July and August. Typical of the desert climate, the rainfall is characterized by a rapid onset and short duration; it is infrequent, localized and variable within the basin. Mean pan evaporation is 2640 mm $year^{-1}$, about five times the precipitation, thus, stream flow is near zero for much of the year.

Data Sources and Procedures

Study on the spatial variation in runoff and sediment delivery is based on 34 gauging stations, which are located at various tributaries in the Luni Basin. Stage heights were measured hourly at each gauging station during the flow, and discharge was calculated by the slope-area method. The first water samples for sediment concentrations were collected when flow began, with subsequent samples collected at times whenever there were significant changes in the flow discharge. The measurements of flow rate and sediment concentration allowed computation of sediment delivery rates for each runoff event. This study was conducted for 16 years from 1979 to 1994.

In the present case study, the arid upland basins are considered as those areas of the drainage basin where the runoff is only related to rainfall at the drainage basin surface, i.e., hilly, mountainous and shallow rocky/gravelly wastes.

Model Calibration

The outlet of a drainage basin may control the amount of sediment leaving the basin. Therefore, the conditions at the outlet of a drainage basin can be used to calibrate a sediment delivery model with the expectation that the model would give the best estimates of drainage basin sediment yield using parameters based on the characteristics of the outlet (Finkner *et al*., 1989).

The first calibration option determined τ_s based on the sediment discharge at the drainage basin outlet and the reference slope (mean slope of the drainage basin), S_o. The dual slope method considered the average of two τ_s values – the first value of τ_s based on the reference slope while the second τ_s value was based on the actual slope at the outlet. The third option for calibrating the sediment delivery model, average shear method, would take into account the combination of slope and discharge along the flow path. The average shear stress could be calculated as:

$$\tau_s = \int \tau_s(X)dx/L \tag{21}$$

where L (m) is the length of flow path.

The sediment delivery model was tested for 10 arid upland drainage basins forming a part of the Luni Basin in the Indian arid zone. To account for the basin complexity, each drainage basin was segmented into three zones, upper, middle and lower, based on the degree of steepness and stream order (Sharma, 1992). One such segmented basin is depicted in Figure 4 as an example. The characteristics of these segments were used to calculate the values of the calibration options. The values of coefficients G and C, determined by the least squares technique at each stage of the flow hydrograph viz. rising, peak and recession, varied between 0.0022 and 0.0072 m^{-1}, and 0.66 and 89.23 kg s^{-1} m^{-1}, respectively.

Model Validation

The reference slope, dual slope and average shear methods of calibrating the coefficients G and C were evaluated through independent runoff events at each stage of the flow hydrograph (Table 2). At the rising stage of the hydrograph, the root mean squared difference was consistently the lowest with the reference slope method. This is because at the rising stage the desert streams convey the highest sediment concentrations, which is attributed to the existence of a thin loose surface layer produced by weathering, drying and biotic interference within the drainage basin during the dry season (Sharma *et al*., 1984). The splash erosion process may provide additional amount of material during the time interval between the initiation of rainfall and that of runoff. The presence of this abundant loose erodible material within the drainage basin rendered only the reference slope as a controlling factor for the sediment delivery, i.e., the average conditions within the drainage basin affected the sediment transport rate at the rising stage of the hydrograph.

Table 2. Summary of statistical analysis of three calibration methods for the sediment delivery model

Hydrograph stage	Calibration method	Sum of squares	Root mean squared difference	Maximum deviation	Number of observations
Rising	Reference slope	3.46	0.20	6.1	84
	Dual slope	4.73	0.24	6.4	
	Average shear	5.75	0.26	15.0	
Peak	Reference slope	114.51	1.33	25.5	65
	Dual slope	41.68	0.80	6.4	
	Average shear	43.95	0.82	6.7	
Recession	Reference slope	3.73	0.23	31.2	70
	Dual slope	1.21	0.13	4.5	
	Average shear	1.03	0.12	3.9	

At peak flow the root mean squared difference was the lowest with the dual slope method. This is because at peak flow the flow conditions were at quasi-steady state and the flatter slopes at the drainage basin outlet resulted in the deposition of the large sediment load, which was picked up from the upstream areas within the drainage basin.

During the recession stage of the hydrograph the receding flow deposited the sediments on its flow path throughout the drainage basin. The flow velocity dropped below the critical value, thereby resulting in a rapid decrease in the sediment concentration towards the end of flow. The average shear stress showed the least root mean squared difference (Table 2) since it represented not only the slope of each segment within the drainage basin, but also the combination of slope and discharge along the length of each segment, i.e., flow path and their cumulative effect at the outlet.

A comparison of observed and predicted sediment delivery rate (Figure 5) shows a good agreement. Further, the maximum deviation between the observed and predicted sediment delivery rates were less than 10 % (ranges between 3.9 and 6.4 %) for the best-fit calibration methods. Thus, a deposition-based sediment delivery model is a suitable approximation to sediment transport rates in the arid zone drainage basins.

EPILOGUE

A closed-form solution of the continuity equation of sediment transport is preferred since it reduces the number of computations and alleviates instabilities associated with the numerical solutions. In this study, we have evaluated a model based on such a solution for arid zone drainage basins, where initial potential sediment load is greater than the sediment transport capacity of overland flow and large amounts of sediment delivery are of concern. A model of this kind based on the Manning's turbulent flow equation and the Yalin sediment transport capacity equation predicted the sediment delivery rates at the rising, peak and recession stages of the hydrograph within ±10 % accuracy. The dynamic forces active at each hydrograph stage can be accounted through various calibration methods involving slope and flow length at the drainage basin outlet. Integration of sediment delivery from sub-area to sub-area up to the outlet using the local drainage direction and principle of conservation of mass was also successful. Preliminary results indicate that the model described herein can simulate the spatial sediment delivery.

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 (Sharma *et al*., 1994). Such information is useful in the identification of vulnerable areas within a drainage basin. Localized storms and high temporal and spatial variability of sporadic rainfall in the arid regions (Jones, 1981; Sharma, 1992) cause considerably greater variability of runoff within the drainage basin. Also, the role of transitory hydrological and hydraulic regimes is not fully understood in the arid regions (Pilgrim *et al.,* 1988). 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). Further, the validation of spatially distributed hydrologic response data such as sub-area overland flow depth, average sub-area infiltration, sub-area flow transport capacity, etc. is extremely difficult, expensive and may not be realistic. Nevertheless, the study demonstrates that the application of GIS to sediment delivery modeling can be accommodated in a low-cost personal computer-based computing environment.

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