A METHOD OF MODELING SOURCE AREA RESPONSE TO CLIMATE VARIABILITY

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ABSTRACT: A modeling framework for understanding spatially-explicit relationships between soil moisture dynamics and streamflow generation in upland humid forested watersheds is described. The framework consists of a dynamic canopy interception module and a 2D finite element hillslope hydrology model (IHDM4) having hillslope planes objectively delineated using contour-based terrain analysis (TAPES-C). This approach is fine-scaled both in space and time allowing for the inclusion of topographic and soil heterogeneities necessary for mapping oscillations in the variable source areas of streamflow generation. The modeling framework is implemented for a small control watershed (WS2) at the Coweeta Hydrologic Laboratory. Simulation results to be presented at the conference include the climate-scale response of variable source areas for hillslope cross-sections to hourly climate data spanning years in which total precipitation was: (a) >20% above average, (b) near average, (c) >20% below average.
KEY TERMS: distributed hydrologic modeling; contour-based terrain analysis; variable source areas; climate timescales; watershed hydrology.

INTRODUCTION

In humid upland forested watersheds, the extent of the saturated near-stream areas, or variable source areas, at the time of a storm primarily determines the timing and volume of stormflow (Hewlett and Hibbert 1963, 1966, Hewlett and Nutter 1970). Variable source areas are spatially constrained by the distribution of topographic and soil characteristics as well as temporally constrained by the frequency and amount of antecedent precipitation. For a given watershed morphology, the variable source area can be depicted as a nonlinear oscillator responding to climatic input functions.

As summarized by Hibbert and Troendle (1988), various models have been constructed to account for the effect of variable source areas on runoff hydrographs. These range from lumped parameter expressions (Hewlett and Troendle 1975) to semi-distributed parameter models such as TOPMODEL (Beven and Kirkby, 1979) to more complex fully-distributed parameter watershed models (Bernier 1982, Thomas and Beasley 1986, Beven et al. 1987). A central objective of these studies has been to account

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for watershed hydrologic processes at the time scale of a storm. Much less investigation, however, has been conducted on variation of saturated near-stream areas at seasonal or annual time scales. Climate scale source area variation was first described by Dunne et al. (1975) and has since been simulated for surface features only using a semi-distributed parameter approach (O'Loughlin 1986).

In order to understand the source area oscillator for a given watershed morphology, a fully distributed model offers the best resolution to capture the fine spatial scale dynamics of near-stream moisture states and fluxes. The increased complexity of these models, however, incurs costly increases in parameter estimation with sometimes only marginal or no gain in predictive capability (as has been discussed by Stephenson and Freeze 1974, Beven 1989, Loague 1990). Also, complexity in the equations of water flow can make solutions intractable, with respect to either the mathematical structures or the computational requirements, a complication particularly severe when attempting to incorporate topographical and soil heterogeneities found in any upland forested watershed. Due to these problems, implementation of a distributed model necessitates simplifying assumptions regarding the nature of watershed morphology, for example specifying control volumes in the models as raster-based grid cells or wedge-shaped hillslope sections. While such assumptions reduce complexity, they result in increased errors. For this reason, Beven (1989) has pointed out that many distributed models are just lumped parameter models with a finer mesh, and called for models with closer correspondence between model equations and field processes.

The spatial and temporal framework of any modeling effort should be constructed to match the scale of the process(es) in question (Mark 1978, Yeakley and Cale 1991). For variable source area modeling in complex terrain, the process in question is the hydraulics of fluid flow. As discussed by Moore and Grayson (1991), there are three primary ways of structuring a network of topographic data: (1) triangulated irregular networks (TINs); (2) raster or grid networks; (3) vector or contour line based networks. Of the three, contour based networks provide more physical realism than grid based networks which restrict water flow from a given node to only one of eight possible directions. TINs provide physical realism, but require interpretive alignment of the elements, many times based on vector digital elevation maps (DEMs). Moore and Grayson (1991) provide an automated contour-based method (TAPES-C) for partitioning watersheds into natural units bounded by irregularly shaped polygons. These polygons are bounded by equipotential (or contour) lines on two sides and by streamlines, orthogonal to the contours, on the other two sides. The streamlines are assumed to be no-flow boundaries, thus groundwater flow is constrained to flow through a series of elements positioned along a natural gradient. Such a series of cells is termed a "stream tube." By orienting the flow equations of a distributed parameter model along stream tubes, spatial complexity in the equations may be reduced from three dimensions to two, while accomplishing a terrain-based model structure.

Here we describe the implementation of a distributed parameter model (IHDM4) having hillslope sections constrained naturally by the terrain using an automated contour-based partitioning package (TAPES-C) for a small forested watershed (WS2) at the Coweeta Hydrologic Laboratory in western North Carolina. The advantage of this framework is that 2-D transient moisture dynamics can be simulated within a terrain-sensitive framework which can ultimately be implemented for ungauged watersheds. This modeling framework will be used to simulate

longterm dynamics of the variable source areas of WS2, using hourly climate data for three years of differing rainfall regimes: (1) near average; (2) >20% above average; (3) <20% below average. The objective of this exercise is to gain a spatially explicit understanding of the near-stream moisture conditions of the watershed, in terms of both nominal oscillations and overall climate response range.

MODEL DESCRIPTION

Above-ground Processing

The distributed model used here (IHDM4: Institute of Hydrology Distributed Model, v.4) receives inputs from an above-ground model which accounts for canopy and litter fluxes. The above-ground model used here was specified by Rutter et al. (1971, 1975). The Rutter model follows a dynamic canopy storage (C) with input of a constant fraction of rainfall determined by leaf area index and vegetation type, and output as evaporation and drainage. The equations of the model are:

$$dC/dt = Q - K(\exp(bC)-1) \quad (1)$$

$$Q = (1-p)R - E_p * f(C) \quad (2)$$

where K and b are drainage parameters, R is the total rainfall, E_p is potential evaporation (determined by a Penman-Monteith equation with stomatal resistance set to zero), p is the canopy throughfall fraction and if C>S, f(C)=1, else if C<S, f(C)=C/S, where S corresponds to a completely wet canopy. The model allows for simultaneous evaporation and transpiration from a partially wet canopy (C<S), a phenomenon particularly important during longterm low-intensity winter storms at Coweeta.

The model is regulated by a water balance given as: rainfall = throughfall + change in C + evaporation loss. Transpiration demand is calculated as E_p for that fraction of the canopy which is dry. An effective precipitation is then calculated which is the throughfall amount (which includes direct throughfall as well as drainage) minus the transpiration demand. In the absence of throughfall then, effective precipitation at the soil surface is negative, which is input to hillslope hydrology model (IHDM4) as a sink term at the surface. The sink is regulated by soil moisture availability times the fractional root distribution in a given layer in the hydrology model as given by Feddes (1976). If positive, i.e. if rainfall is occurring, then input to the surface is as a source term.

Terrain Analysis

Contour based terrain analysis as developed by Moore et al. (1988) and Moore and Grayson (1991) requires three general steps. First, a contour map of the watershed is digitized, creating a vector DEM. Here we used the Arc/INFO geographic information system (GIS) to accomplish this task for WS2 (see Figure 1a). Then a preprocessing program (PREPROC) is used to transform the vector DEM into a north-south oriented coordinate system. Finally, the program TAPES-C partitions a watershed into "stream tube" subwatershed units using a constant offset between trajectory (i.e. stream tube boundary) starting points. Figure

1b shows results from a TAPES-C computation for WS2 using a 50 meter offset. Further processing is then required to transform the streamtube output of TAPES-C into a structure suitable for IHDM4.

Each hillslope plane in IHDM4 is represented by a two-dimensional vertical cross-section of finite-element nodes running longitudinally from watershed divide (or interior high point) to stream. At each vertical set of nodes in the cross-section, a constant width is assumed. So from map view, a hillslope plane in IHDM4 is constrained to a series of adjacent trapezoids beginning at the stream and continuing to the divide. To fit the TAPES-C output to IHDM4, we extended the no-flow boundaries shown in Figure 1b to permanent stream locations using Arc/INFO to derive fifteen hillslope planes (see Figure 1c). Arcs bounding streamtube cells were then selected to maximize the criteria that (a) the arcs be positioned parallel to the contours and (b) the streamside arc be parallel to the stream.

Area and slope for the surface of each cell (i.e. each 4-sided polygon) were calculated using Arc/INFO. For each streamtube, widths of the cells were allowed to vary in order to transform the cells into trapezoids while maintaining area and slope for each cell. The transformation proceeds iteratively from the streamside arc up the streamtube using the relation:

$$Y_{N+1} = Area_{N,N+1}/X_{N,N+1} \qquad (3)$$

where Y_N is the effective width of arc N, $Area_{N,N+1}$ is the original area of the polygon bounded by streamtube orthogonals and arcs N and N+1, and $X_{N,N+1}$ is the average distance between arcs N and N+1.

Hillslope Hydrology Model

Within a streamtube, a cell is bounded by two vertically-layered sets of finite element nodes. The top surface of the soil (i.e. highest set of nodes) is treated as a flux boundary with fluxes controlled by the applied input rates of effective precipitation unless the surface becomes saturated and overland flow develops. The surface boundary then changes to a fixed head boundary while saturation persists, with the potentials fixed at atmospheric pressure. The change of boundary conditions at the soil surface can occur locally on the slope to enable simulation of an oscillating variable source area (Beven et al. 1987, p.14).

For a given streamtube, subsurface flow is given by the Richards equation expressed as

$$BC(\psi)\frac{\partial \phi}{\partial t} - \frac{\partial}{\partial x}\{BK_x(\psi)\frac{\partial \phi}{\partial t}\} - \frac{\partial}{\partial z}\{BK_z(\psi)\frac{\partial \phi}{\partial z}\} = Q_s \qquad (4)$$

where B is a gradually varying streamtube width, ψ is capillary potential, x is horizontal distance downslope, z is gravity potential (measured vertically from some arbitrary datum), ϕ ($= \psi + z$) is total hydraulic potential, $C(\psi)$ is the specific moisture capacity of the soil (slope of the relation between θ and ψ), θ is soil moisture content by volume, K_x, K_z are saturated hydraulic conductivities in the x, z directions (functions of ψ), Q_s is a source/sink term, and t is time. Implementation of (4) requires several assumptions, including: (a) water is of constant viscosity and unit density; (b) flow is laminar and occurs in an isothermal medium; (c) Darcy's law applies with time-invariant parameters; (d) only single phase water flow in response to

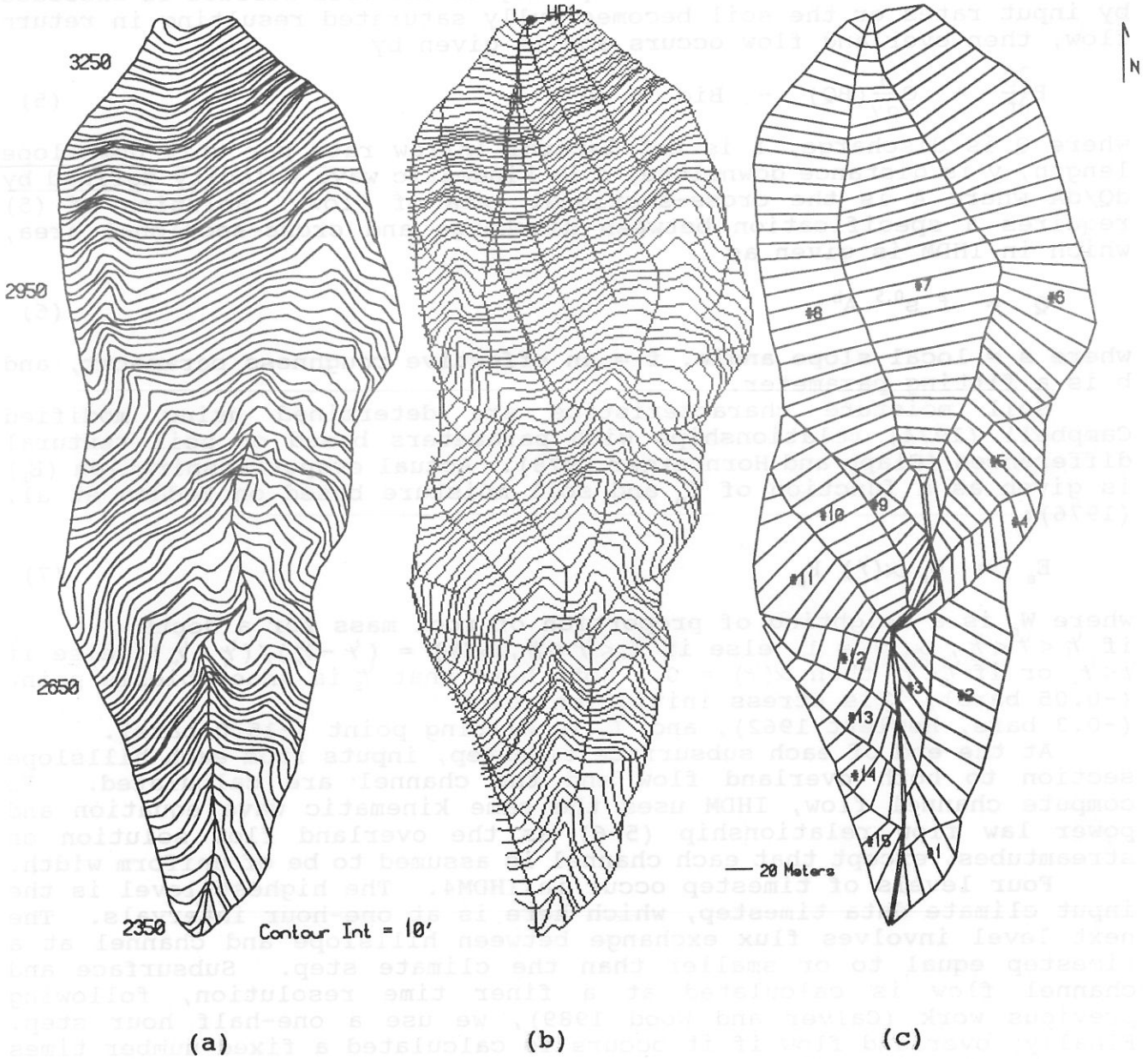

Figure 1. Terrain analysis for WS2 at the Coweeta Hydrologic Laboratory. Shown are: (a) original contour map with stream location and approximate watershed boundaries, (b) "stream tube" delineation using TAPES-C software, (c) resulting hillslope planes fitted (but untransformed) for IHDM4.

hydraulic gradients is considered; (e) the relationship between ψ and θ is locally differentiable (Beven et al. 1987).

If either the infiltration capacity of the soil surface is exceeded by input rates or the soil becomes fully saturated resulting in return flow, then overland flow occurs and is given by

$$B\frac{\partial Q}{\partial t} + C\frac{\partial}{\partial y}\{BQ\} - Bic = 0 \qquad (5)$$

where Q is discharge, i is net lateral inflow rate per unit downslope length, y is distance downslope, c is kinematic wave velocity defined by dQ/dA where A is the cross-sectional area of flow. Solution of (5) requires a specification between discharge and cross-sectional area, which in IHDM is given as

$$Q = f\, s^{0.5}\, A^{b} \qquad (6)$$

where s = local slope angle, f = an effective roughness parameter, and b is a fitting parameter.

Soil moisture characteristics are determined using modified Campbell (1974) relationships with parameters based on soil textural differences (Clapp and Hornberger 1978). Actual evapotranspiration (E_a) is given as a function of E_p and soil moisture based on Feddes et al. (1976):

$$E_a = W_r\, \alpha(\psi)\, E_p \qquad (7)$$

where W_r is a weighting of proportion of root mass for a depth, if $\psi_i < \psi < \psi_s$, $\alpha(\psi) = 1$, else if $\psi_w < \psi < \psi_i$, $\alpha(\psi) = (\psi - \psi_w)/(\psi_i - \psi_w)$, else if $\psi < \psi_w$ or if $\psi_s < \psi$ then $\alpha(\psi) = 0$. Note here that ψ_s is anaerobiosis point (-0.05 bars), ψ_i is stress initiation point (-0.3 bars, Hewlett 1962), and ψ_w is wilting point (-15.0 bars).

At the end of each subsurface timestep, inputs from each hillslope section to both overland flow and the channel are calculated. To compute channel flow, IHDM uses the same kinematic wave equation and power law flow relationship (5-6) as the overland flow solution on streamtubes, except that each channel is assumed to be of uniform width.

Four levels of timestep occur in IHDM4. The highest level is the input climate data timestep, which here is at one-hour intervals. The next level involves flux exchange between hillslope and channel at a timestep equal to or smaller than the climate step. Subsurface and channel flow is calculated at a finer time resolution, following previous work (Calver and Wood 1989), we use a one-half hour step. Finally, overland flow if it occurs is calculated a fixed number times in each subsurface flow timestep (Beven et al. 1987).

Calibration of the model is first performed at the storm scale for both wet and dry years using an objective function of SSE of observed vs. predicted streamflow (Hornberger et al. 1985). This calibration is then extended to climate timescales using soil moisture data measured along a WS2 hillslope transect from a drought period through precipitation recharge (Yeakley 1992) as a behavioral constraint (Hornberger and Cosby 1985). Parameters tuned primarily include those identified as sensitive in IHDM (Calver 1988): saturated hydraulic conductivity; porosity; initial soil moisture potential; surface roughness. All possible parameter ranges are based on measurements conducted at similar low elevation watersheds at Coweeta (Swank and Crossley 1988, Gaskin et al. 1989, Vose and Swank 1991).

Simulations and Analyses

After calibration, simulations are conducted for three years of five hourly climate variables: precipitation, temperature, windspeed, relative humidity, solar radiation. All data were collected from climatic stations within a kilometer of WS2. From the recent record, we selected three consecutive years (1979-1981) in which annual precipitation steadily decreased from 28% above the mean (1979), to within 5% of the mean (1980), and finally to 20% below the mean (1981).

Model output includes streamflow values for the simulation period as well as soil moisture response range profiles for both a lower and an upper hillslope in WS2, as determined by the contour-based partitioning. Further output includes time series of saturated area percentage of each hillslope cross-section for the three year period simulated.

Analysis includes cross-correlations performed between precipitation and saturated area percentage, as well as between saturated area percentage and streamflow. Trend analysis is performed to determine if the amplitudes of variable source oscillations change significantly as the precipitation regime becomes more droughty.

EXPECTED RESULTS AND FUTURE DIRECTIONS

We expect that source area variation will be greater during drier periods and so the amplitude of the variable source area oscillation should increase as precipitation drops from 20% above the annual mean in 1979 to 20% below the annual mean in 1981. This amplitude increase, however, should be affected by the degree of serial correlation of storm occurrence and intensity at subannual timescales. Further, we expect that convergent hillslope planes (e.g. plane #11 in Figure 1c) will show a greater response range for the near-saturated source area than divergent planes (e.g. plane #9).

Our analyses, while explicitly of hydrologic responses to climate variability based on a 3 year climate record, point implicitly to hydrologic response to climate change by projecting response ranges under differing precipitation regimes. Moreover, we explicitly simulate responses to a significant multi-annual decrease in storm frequency. Using an approach following that taken by Wolock and Hornberger (1991), it is possible to represent the input climate variables stochastically using fitted distributions and then adjusting the moments to simulate climate change incorporating variability, as was done recently using a lumped parameter evapotranspiration model (PROSPER) on an annual timescale for WS2 at Coweeta (Yeakley et al. 1991). We foresee such simulations for the more fine-scaled modeling framework presented here in the near future.

An additional future direction is to expand the scale of the simulations spatially from the 12 ha watershed level up to the 1600 ha level of the Coweeta Basin. Once parameterized and calibrated at that scale, it can be implemented for similar large scale ungauged basins in the same region if elevation, vegetation and soils information are available. Such a high resolution approach could be used to identify hydrologically-sensitive areas within basins to changing climatic conditions.

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