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Forest ecosystem processes at the watershed scale: incorporating hillslope hydrology

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ABSTRACT

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An approach to distributed modeling of watershed hydro-ecological processes over large spatial scales is described. A data and simulation system, RHESSys (Regional HydroEcological Simulation System), combines a set of remote sensing/GIS techniques with integrated hydrological and ecological models in order to automate the parameterization and simulation of a suite of ecological and hydrological flux and storage processes through the watershed. Specifically, we simulate forest canopy net photosynthesis (*PSN*) and total evapotranspiration (*ET*) through the year with a modeling package that integrates FOREST-BGC, a stand level model of forest carbon, water and nitrogen budgets, with TOPMODEL, a quasi-distributed hydrological model. The latter model introduces the effects of hillslope hydrological processes, incorporating surface redistribution of soil water by saturated throughflow processes. The maintainance of regular patterns of soil water by throughflow processes cause forest ecosystem activity to vary dependent on hillslope position. The distributed framework is based on a terrain partition in which each terrain object (hillslopes and stream reaches) comprising the watershed are separately parameterized and simulated. The location of each terrain object within the watershed is explicitly represented while the internal variability of each object is represented as a joint parameter distribution. Generalization of the surface into different numbers of terrain objects (by growing or shrinking the extent of the stream network) is automatically accomplished using digital terrain data. This allows us to flexibly alter the spatial representation of the watershed by shifting surface information either into the internal hillslope parameter distributions or into greater numbers of hillslopes (and stream reaches). Limited simulations of a mountainous watershed in western Montana indicate that incorporation of within hillslope throughflow and the soil, topography and vegetation distribution has the effect of significantly altering the seasonal *PSN* and *ET* trends in comparison with lumped surface representations without lateral water flux.

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INTRODUCTION

In mountainous and hilly terrain the spatial patterns of forest ecological and hydrological storages and fluxes are dependent on the topography through the variability of the radiation environment, precipitation and temperature conditions, and soil water drainage. These factors can show significant variations on scales ranging from individual hillslopes owing to soil drainage and exposure, to larger regions affected by temperature and precipitation gradients. The close coupling and mutual evolution of the forest canopy and local soil properties with the energy and water environment contributes to a complex pattern of landform, soils and vegetation over the landscape. Analysis or modeling of any one of these components requires the simultaneous treatment of the other components, incorporating significant feedbacks and variations that may occur over the time and space scales of interest. If either the hydrological or ecological components are prescribed, the lack of dynamic feedbacks could seriously alter the modeled systems' behavior. Because of this, a major challenge to their effective integration is to find efficient methods to represent the processes at a level of conceptual and computational detail at which their spatial variability and mutual interaction are adequately expressed.

In this paper, we present an integration of a forest ecosystem process model, FOREST-BGC (Running and Coughlan, 1988), with a spatially distributed hydrologic model, TOPMODEL (Beven and Kirkby, 1979). Both models have been previously validated in a number of forest stands and catchments. The integration of the models is carried out in order to investigate the distributed feedbacks between ecological and hydrological processes at the watershed scale. In particular, we are interested in the impact of lateral water redistribution processes (surface and subsurface water flow) and the resulting patterns of available soil water on the distribution and magnitude of forest canopy evapotranspiration (*ET*) and net canopy photosynthesis (*PSN*). An important feature of the model integration is the incorporation of all significant water pathways which may drain or modify the patterns of soil water. We demonstrate the model integration and the distributed hydro–ecological interactions with a set of numerical experiments carried out for the area of Soup Creek, a 15 km^2 mountainous watershed in western Montana. While results presented here are to some extent unique to the specific climate and terrain of this area, the advantages and limitations of the present approach are clearly illustrated, and the impact of the canopy, water interactions are considered representative for northern Rocky Mountain ecosystems.

We have embedded both FOREST-BGC and TOPMODEL into a data and simulation framework, RHESSys (Regional HydroEcological Simulation System), designed to automate the distributed parameterization and execution of the integrated models. RHESSys incorporates a set of geographic information processing techniques to combine information drawn from digital terrain data, remote sensing and digitized soil maps into a distributed parameter set necessary to operate the hydro–ecological models described here. The GIS/remote sensing approach to parameterization is required for the simulation of large watersheds as manual methods are inefficient, inaccurate and do not offer a methodology that can be implemented on even larger scales of interest in the future (Shuttleworth, 1988). FOREST-BGC (Running and Coughlan, 1988) is a stand level simulation model of carbon, water and nitrogen budgets that calculates, among other things, forest evapotranspiration (*ET*) and *PSN*. The carbon and water budgets for a forest stand are computed from daily meteorological data, basic canopy information (e.g. LAI, and stomatal physiology) and a simplified soil water budget. This budget is based on the soil 'bucket' model, and includes calculations of rainfall input, interception, snowmelt, capacity overflow and evapotranspiration.

TOPMODEL (Beven and Kirkby, 1979) uses surface topographic and soils information to simulate the lateral redistribution of soil water over a catchment by saturated throughflow. The model is based on demonstrated principles of hillslope hydrology in which locations with large upslope contributing areas and low surface gradients maintain higher soil moisture levels than locations that are steep (rapid throughflow drainage) or have low upslope contributing areas. The patterns of soil transmissivity, T (m^2 day^{-1}), defined as the saturated hydraulic conductivity integrated over depth are combined with these topographic properties to represent the effects of lateral soil moisture flux, such that the patterns of available soil water over the landscape can be predicted. In our modeling framework the moisture patterns provide spatial variability to the available soil water required by FOREST-BGC, leading to differential rates of forest evapotranspiration and productivity over the terrain. O'Loughlin (1990) has recently reviewed and compared the simulation approach taken by TOPMODEL and TOPOG, a modeling system with a similar conceptual genesis, with emphases on land management and forestry applications.

We have previously presented RHESSys in a form to parameterize FOREST-BGC (without TOPMODEL) over complex terrain (Band, 1991; Band et al., 1991). The parameterization method for mountainous landscapes is based on the extraction of the stream network of the watersheds from a digital elevation model (DEM), and the segmentation of the topographic surface into their component hillslopes (Band, 1989). Each hillslope is then separately parameterized with information drawn from geographically registered soil and remote sensing data sets. The advantage to this approach is the relatively conservative variation of the important model variables of net radiation, air temperature and leaf area index within the hillslope facets, and the higher variation of these variables between the hillslope facets. Simulation of *ET* and *PSN* over the population of hillslopes defining the topography is then able to capture a large proportion of the landscape variability in these processes.

The substitution of TOPMODEL as the surface hydrological component for the landscape replaces the functions carried out by the bucket model previously used in FOREST-BGC. This addition results in the following major modifications in the simulation behaviour. (1) It provides for lateral subsurface drainage of soil water from the hillslopes and the production of saturation runoff from partial contributing areas of the watershed. This produces more realistic outflow hydrographs from a catchment, and represents important pathways for depletion of soil water that are not incorporated into the bucket model. (2) A distribution of available soil moisture is developed over the watershed on the basis of hillslope position. This allows the representation of important topographic controls on canopy processes and forest growth, particularly through the zonation of soil moisture deficits. (3) The necessary distributions of soil transmissivity and topographic variables requires a more distributed framework for surface representation in which the observed assemblages of terrain, soil and vegetation parameters are maintained. The higher resolution information describing these patterns includes their spatial covariation. Aside from distributing the computed flux processes, integration of the nonlinear water and carbon flux processes over these parameter distributions may lead to significant shifts in the timing and magnitude of the total areal exchange of carbon and water with the atmosphere.

The more distributed treatment of the watershed hydrology accomplished by incorporating TOPMODEL into RHESSys includes modifications in the geographic information processing for model parameterization and in the integrated simulation and visualization techniques. In the following sections the current version of RHESSys is described in terms of its structure and the interaction of its parameterization, simulation and visualization components. The simulation components, FOREST-BGC and TOPMODEL, are described in sufficient detail to evaluate their integration within RHESSys as presented here. The demonstration of RHESSys parameterization and simulation results for Soup Creek includes an evaluation with limited field measurements of leaf water potential and basin runoff and comparison with the previous generation of the model run with the 'bucket' soil water representation. The model comparisons are used to investigate the impact of distributed vs. lumped hydrology on the spatial and temporal patterns of computed carbon and water flux. Note that the lack of distributed snowpack and soil moisture observations for model initialization for the simulation year of 1988 makes difficult the direct comparison of model results with observed discharge and canopy variables for validation purposes. Instead, the limited field data are used to assess the overall behavior of the model in comparison to alternative model forms. Finally, ongoing and planned improvements in the model structure, parameterization schemes and validation strategies are discussed.

DESCRIPTION OF RHESSys

The RHESSys parameterization and simulation components (RHESSys pieces) are specifically designed to represent the surface soil, topographic and vegetation patterns along with certain ecological and hydrological processes, respectively, at a level of detail commensurate with 'landscape level' modeling. By landscape level, we refer both to a geographical area (order of magnitude areas of 10^1–$10^3\,km^2$) over which there is significant variation in the major factors controlling both storage and flux of ecological and hydrological variables, and a conceptual level of treating and representing that variability. The conceptual approach seeks to represent the land surface characteristics and the physical processes at a level of detail for which the necessary parameters can be realistically estimated to reproduce the dominant patterns of hydro–ecological dynamics (such as runoff generation or canopy productivity) over the landscape. The requirement of adequate parameterization over large heterogeneous areas is generally very difficult for those models that seek to cast the physical processes as a set of differential equations to be solved at fine space and time steps over the nodes of a finite element of difference mesh. While these approaches model the processes with a high degree of physical realism, they require a large precise set of parameters to give solutions commensurate with that level of process specification. While current computing technology may be adequate to solve the equations, it is not feasible to collect all the physical parameters necessary to adequately describe the actual field condition over extensive, diverse terrain. Beven (1989) has proposed that in many circumstances we must consider such detailed descriptions of the surface and subsurface characteristics of a watershed to be essentially unknowable from a modeling perspective. Therefore, in the presence of this uncertainty and 'unknowability' hydrological and ecological models must be constructed to represent as much of the significant pathways and processes as possible with the restriction of being realistically parameterized and operated.

It is not our purpose or goal to model these processes with high precision at any given point, but to distinguish between process rates over landscape units, such as between ridges, hillslopes and valleys or the regular variations that will occur within a hillslope as a result of present trends of insolation or soil moisture. Therefore, our approach is to reduce the complexity of the physical representation of the processes to the level where we can distinguish the approximate spatial pattern of responses over the landscape. Emphasis is therefore placed on a better description of the dominant spatial patterns of surface properties and hydrologic pathways at the landscape level, rather than a more precise description of surface properties or processes at any given point or small set of points. In effect, the uncertainty introduced by this restriction of the process representation may be no greater, and is less misleading, than uncertainty introduced by greater numbers of less well known parameters.

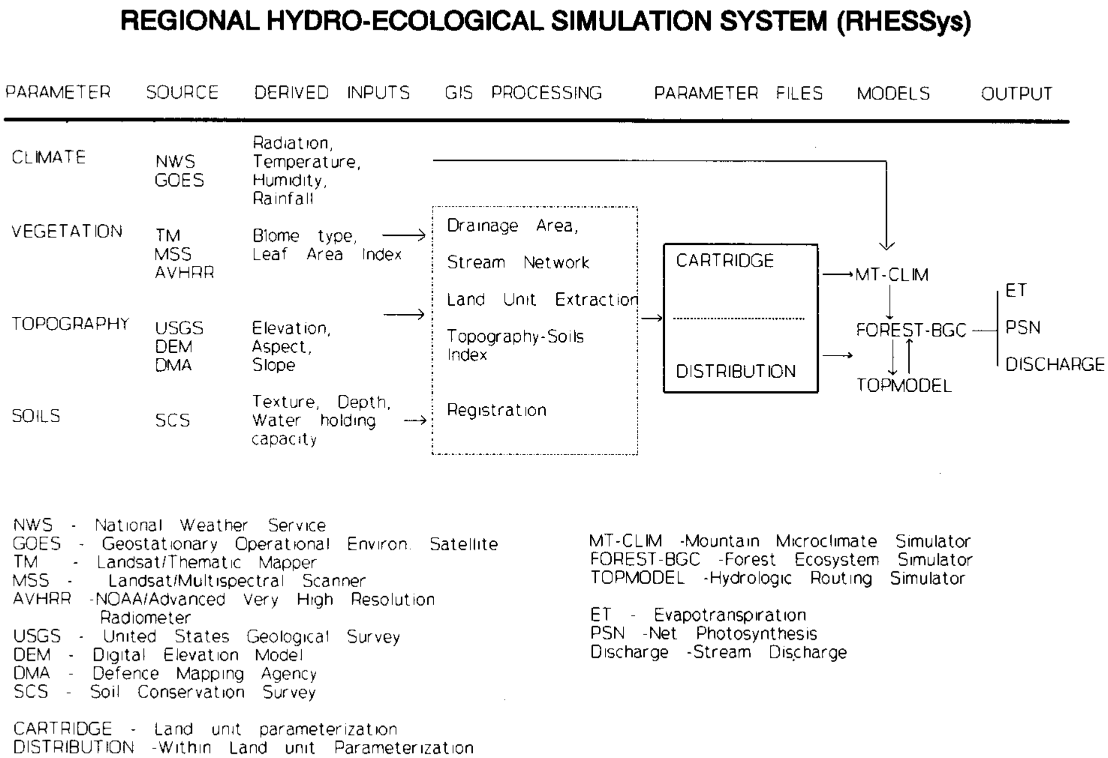


Fig. 1. Flow diagram of information from raw image data through model parameter sets and simulation results for RHESSys. Surface parameter fields are computed from remote sensing imagery, digital elevation data and digitized maps, and aggregated to landform levels in the cartridge files. Within landform variability is represented in the distribution files.

Parameterization strategy

The spatial resolution of process and surface representation is given at two different levels. The hillslopes and stream reaches are explicity located within the watershed with a topological description of stream network and hillslope organization following Band (1989). Within hillslope variability is represented by aspatial frequency distributions computed from high resolution remote sensing imagery and digital terrain data. Figure 1 shows the data flow in the current RHESSys implementation, routing landscape and meteorological information through a set of geographic information processing modules for parameter extraction and organization, and then into the integrated FOREST-BGC/TOPMODEL simulations. Raw input data include remote sensing data for determination of canopy variables (specifically biome type and LAI), grid digital elevation models (DEM) for calculation of surface slope, aspect, elevation, and the contributing drainage area for each pixel as described by Band (1989), Band and Wood (1988) and others. Soil maps of the area are used for the calculation of expected soil depth, texture and transmissivity. The DEM is also used for the extraction of the watershed structure and its component landforms (stream lines, ridge lines, hillslopes).

The parameters required for the simulations are aggregated to the landform level (e.g. individual hillslopes), and maintained either as mean values in the cartridge parameter files or as the range or pattern of parameters in the distribution files (see Fig. 1). The hydro–ecological simulations are then carried out on the landform (land unit) rather than the pixel level. In keeping with the conceptual level of simulation, this represents the landscape patterns to the level of observable variation between hillslopes (given in the cartridge file) and the trends in canopy, soil and topographic variables within the hillslopes (given in the distribution files), but not to the high frequency pixel-by-pixel variations.

A strength of the landscape representation and parameterization is the ability to easily reform the watershed with greater or fewer numbers of landscape units. This is done at the level of parameter field aggregation by substitution of a new watershed partition model, with a greater or lesser extent of the stream network (Fig. 2). As each hillslope is referenced to an individual stream link, the extent of the stream network determines the detail of the watershed partition into elementary hillslope and stream units. At the parameter file level each cartridge file (e.g. Table 1) represents a different scale of surface parameterization, such that the simulations are easily reparameterized and executed over a range of scales by replacing the cartridge. As the watershed is partitioned into fewer land units, less surface information is represented as the variance of parameter means within the cartridge file (which shows explicit spatial patterns of parameters between the land units) and more of the variance is implicitly incorporated into the distribution files which contains landform internal variance. This provides the ability to perform the simulations at any desired level of surface aggregation, or over a range of scale representations while preserving the watershed structure in terms of the arrangement of runoff producing regions and the drainage network.

FOREST-BGC

FOREST-BGC is a compromise between detailed models of canopy biophysics and more general, functional or statistical approaches to climate–vegetation interactions (Running and Coughlan, 1988). The model is designed to simulate carbon, water and nitrogen cycles of forested ecosystems at landscape levels, allowing the use of remote sensing and geographic information systems (GIS) for spatially distributed parameterization. Only canopy *ET* and *PSN* results are addressed in this paper. The input data required consists of daily meteorological conditions for a base station, including minimum and maximum temperature, relative humidity or dewpoint, precipitation and solar insolation, if available. The base station data is then extrapolated to different

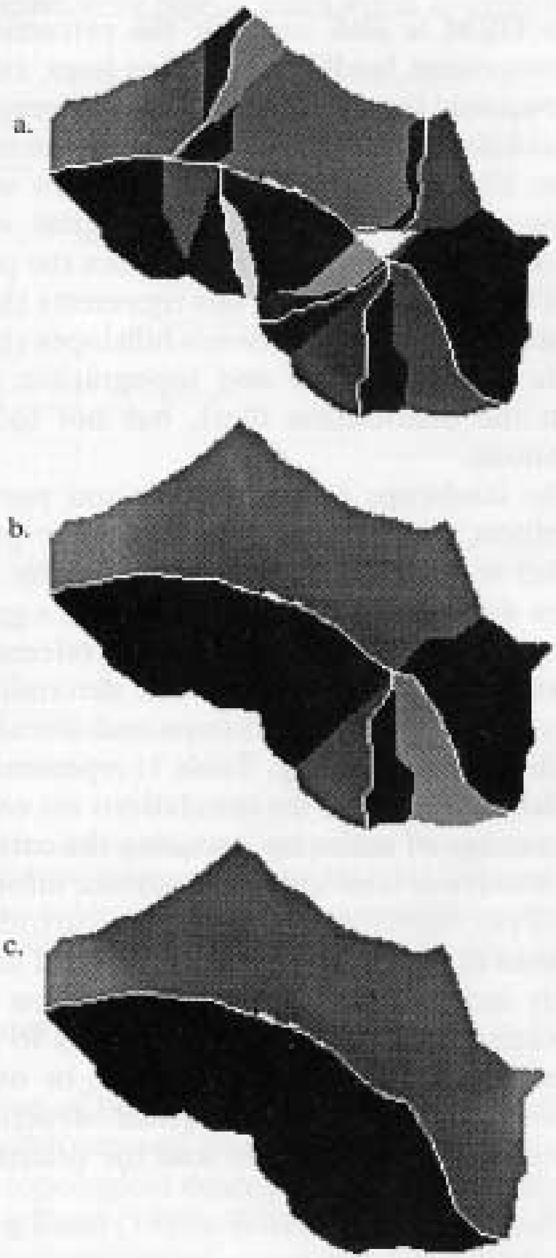


Fig. 2. Representation of Soup Creek by the segmentation of the DEM into: (a) 11 stream links and 22 hillslopes; (b) 3 stream links and 6 hillslopes; (c) one stream link and two hillslopes. The partition strategy can easily produce different levels of surface detail with greater or lesser numbers of stream links and hillslopes (two hillslopes draining into each stream link) for use as the basic landform information and simulation units.

portions of the topography with the use of MT-CLIM (Running et al., 1987), a subroutine that adjusts the meteorological data for typical lapse rates, atmospheric optical depth and illumination angles using site elevation, gradient and exposure.

FOREST-BGC takes the daily precipitation, subtracts canopy and litter

TABLE 1

Between hillslope variations in topographic, canopy and soils information

Slope	Aspect degrees	Elevation (m)	Gradient (degrees)	LAI	SWC (m)	Area (ha)	λ	m	Number of intervals
1	177.0	1600.0	30.4	10.6	0.108	129.3	5.7	0.02	10
2	330.0	1627.0	30.9	7.7	0.122	101.3	5.4	0.02	9
3	167.0	1847.0	30.7	9.1	0.101	23.6	5.3	0.02	6
4	236.0	1885.0	33.6	5.9	0.102	35.2	5.2	0.02	7
5	206.0	1519.0	17.2	10.1	0.162	18.1	5.4	0.02	7
6	7.0	1640.0	28.5	7.9	0.134	38.4	5.8	0.02	7
7	209.0	1771.0	26.8	6.5	0.131	238.3	5.9	0.02	9
8	23.0	1692.0	27.2	7.1	0.145	101.4	5.6	0.02	9
9	184.0	1834.0	22.2	8.1	0.116	14.6	5.9	0.02	6
10	245.0	1923.0	29.1	6.3	0.105	57.6	5.9	0.02	8
11	242.0	1597.0	10.8	4.0	0.142	8.7	5.7	0.02	6
12	45.0	1711.0	22.4	6.6	0.163	25.5	5.9	0.02	7
13	225.0	1623.0	13.6	6.1	0.150	3.2	5.9	0.02	6
15	259.0	1968.0	28.4	4.9	0.107	172.1	6.1	0.02	9
16	13.0	1842.0	23.9	6.0	0.124	83.1	6.0	0.02	9
17	334.0	1883.0	24.9	6.1	0.115	43.4	5.9	0.02	9
18	58.0	1925.0	26.9	5.3	0.108	65.0	6.0	0.02	8
19	25.0	1941.0	30.6	4.9	0.110	25.1	5.5	0.02	6
20	94.0	1932.0	27.4	5.7	0.113	16.5	5.7	0.02	6
21	308.0	1838.0	33.7	6.9	0.105	16.3	4.8	0.02	5
22	31.0	1847.0	34.7	8.03	0.107	38.2	5.43	0.02	8

interception and routes the remaining amount into soil water or snowpack for later release by snowmelt (Fig. 3). Canopy variables, the most important of which is LAI, are estimated with remotely sensed data (Peterson et al., 1988; Spanner et al., 1990). At this point, soil information is only available from existing soil maps and requires field inspection to determine any significant trends in soil properties with topographic position (e.g. the existence of a catena). This is a time consuming and very approximate portion of the parameterization, although we are exploring methods of improving and automating the soil parameterization process. Canopy transpiration, calculated by a Penman–Monteith equation based on daily micrometeorologic information, the predawn leaf water potential (ψ_{leaf}) and LAI, drives the uptake and conductance of soil water as modified by physiologic conductance and soil water availability. Canopy photosynthesis is computed with the CO_2 diffusion gradient, radiation and temperature controlled mesophyll CO_2 conductance, the canopy water vapor conductance, LAI and daylength. Beer's Law is used to approximate extinction of canopy shortwave radiation, which is then integrated over the canopy to produce average radiation levels.

A simplified bucket model is used for the soil water budget, with a capacity set at the available soil water capacity (AWC, mm). The net input to the soil

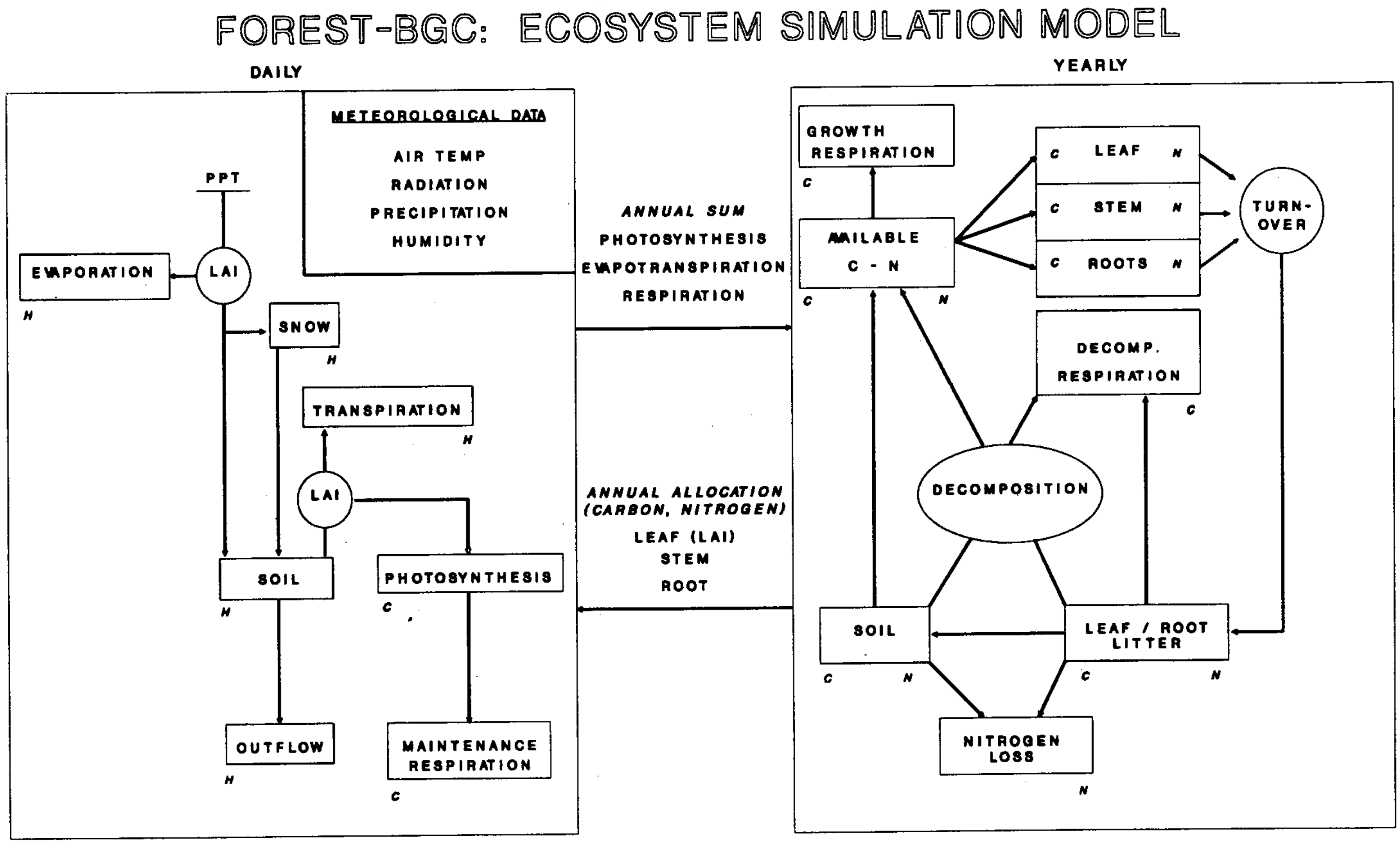


Fig. 3. Flow diagram of FOREST-BGC, after Running and Coughlan (1988).

water storage is given by

$$\partial S_w = P - (I_c + I_l) + M - ET - D \qquad (1)$$

where D is deep percolation, dS_w is the change in soil water, M is snowmelt, P is precipitation and I_c and I_l are canopy and litter interception respectively, all measured as mm day^{-1}. When S_w(mm) exceeds AWC, the difference is immediately released as runoff. In this respect, the hydrology is restricted to vertical exchange. No lateral redistribution of water is considered, and runoff occurs instantaneously when $S_w > AWC$ and is zero if $S_w < = AWC$. This results in unrealistically flashy outflow hydrographs and does not adequately represent soil water drainage in hilly to mountainous watersheds. More importantly in terms of the ecological processes it is largely independent of topographic position as there are no lateral discharge or recharge exchange processes with downslope or upslope terrain, respectively.

TOPMODEL

While the bucket representation may be adequate for landscapes with negligible gradients (and therefore, negligible lateral soil water flux) it is well known that in steeper topography saturated matric throughflow and macropore flow cause significant redistribution of soil water over the landscape. Typically, areas with larger upslope drainage areas and lower gradients have higher values of S_w while areas with low upslope drainage areas (near ridges) and higher gradients are better drained and have lower moisture contents. Localized differences in S_w over the terrain within a single catchment of hillslope can cause significant variations in transpiration and productivity of the canopy. Spatial covariation of S_w with soil hydraulic properties and vegetation canopy properties over the terrain are necessary to incorporate in distributed hydroecological framework, as they result in both short and long term patterns in carbon and water flux rates. Once again, emphasis is on capturing that portion of the landscape that is regular and measurable (in a parameterization sense). High frequency random variations in the model parameters are not represented over the space and time scales we wish to model, and may be considered a more purely stochastic characteristic of the landscape.

TOPMODEL (Beven and Kirkby, 1979) takes an approach to hydrologic simulation that is complementary to the conceptual level of FOREST-BGC. TOPMODEL emphasizes the regular variations in soil water storage and flux that are expected under certain hydrologic conditions. A central assumption of TOPMODEL is that of a quasi-steady subsurface throughflow system in which recharge to the saturated zone in the contributing area above a unit length of contour on the hillslope is approximately equal to the saturated throughflow across that contour length. These conditions are generally restricted to areas of moderate to steep topography with permeable soils in which rapid subsurface throughflow occurs, and local hydraulic gradients can be approximated by the ground surface gradient.

The approach taken in TOPMODEL is considered to be 'quasi-distributed' as each point or region on the surface is not explicitly simulated. Instead, the watershed is modeled as a distribution of points or elements that are parameterized by a hydrologic similarity index

$$ln(aT_e/(T_i \tan\beta))$$

where a is the upslope contributing area (m^2), β is the local gradient, T_i is the local soil transmissivity (m^2s^{-1}) and T_e is a catchment or hillslope wide transmissivity index (defined in the Appendix). This parameter is used to compute local soil water deficit, S, measured as a depth below saturation, as

$$S = S' + m\lambda - m\,ln\left(\frac{aT_e}{T_i \tan\beta}\right) \qquad (2)$$

where S' is the hillslope mean soil water deficit, λ is the mean value of ln $(a/\tan\beta)$, and m is a soil specific parameter associated with the vertical profile of the soil saturated hydraulic conductivity (Beven and Kirkby, 1979). The derivation of the similarity index and eqn. (2) are given in the Appendix. Note that the variability of S from S' is proportional to m.

The hillslope frequency distributions of the similarity index are maintained in the hillslope distribution files of RHESSys (Fig. 1). Any two points that share the same similarity index are considered hydrologically similar. Therefore, the model is integrated over the distribution function of the similarity index for each hillslope. Elements in the catchment may be updated after each time step on the basis of their index value and hillslope S'. Simulation progresses by updating the hillslope S' each time step by water balance computation, and then calculating S_i for each index interval using eqn. (2). For each index interval, infiltration from precipitation and snowmelt is routed through canopy and litter interception stores with defined interception capacities (dependent on index interval LAI and litter depth). Water in excess of these interception stores is added to an unsaturated zone storage, SUZ (mm). Water is routed from SUZ to S at a rate proportional to the value of SUZ/S, the vertical saturated hydraulic conductivity, K_s, and a soil drainage rate parameter, α. In the present application, ET is then computed for each interval calculated by components of FOREST-BGC with the interval available soil water in the rooting zone. The interval specific flux values are areally weighted to produce hillslope level flux values to update S'. Baseflow is approximated for the full hillslope as

$$q_b = q_0 \exp(-S'/m) \qquad (3)$$

where

$$q_0 = T_e \exp(-\lambda) \quad (4)$$

after Sivapalan et al. (1987) and Famiglietti and Wood (1990). Values of the topography–soils index can be computed from digital elevation data (Band and Wood, 1988) and digitized soil maps. At each time step, local soil water content is computed for discrete intervals of $In(aT_e/T\tan\beta)$ and may be mapped back onto the catchment surface (Band and Wood, 1988).

Note that the above method calculates interval specific values for all flux processes except net throughflow drainage, for which the effect is approximated using the implicit solution given by eqn. (2). Explicit routing of throughflow drainage over each hillslope would be very computationally consumptive, and specific parameterization of each node on the hillslopes would be extremely difficult. The price paid for the implicit solution given by TOPMODEL is some ambiguity in the coupling of the canopy ET and local soil moisture updating. If the rate of throughflow divergence is small compared with ET rates, canopy ET may deplete soil moisture faster than throughflow can recharge it. This indicates that the coupling proposed here works best in steep catchments with high transmissivity soils. While our current application basins fit this requirement, as we move into flatter areas or finer grained soils, this coupling may become problematic. We are currently exploring the range of environmental conditions for which the integrated model is applicable.

TOPMODEL has been parameterized and validated on a number of steep upland catchments (e.g. Beven et al., 1984; Hornberger et al., 1985), with validation largely done on the basis of catchment discharge hydrographs and only limited comparison of soil water or saturation patterns. Validation has largely been restricted to storm events as interstorm potential evapotranspiration has only been handled as simple seasonal sinusoidal functions (Beven, 1986), spatially constant prescriptions or functions of base station meteorological observations (Famiglietti and Wood, 1990). In mountainous to hilly terrain, this ignores topographically induced variations in potential ET induced by the variable radiation environment.

Previous use of TOPMODEL has been applied to either entire watersheds or to a system of subcatchments. Familiagetti and Wood (1990) incorporated a spatially constant potential evapotranspiration (ET_p) term for a full watershed, then used the spatially variable soil water produced by TOPMODEL to estimate the distribution of actual ET across the Konza site in Kansas. For flat to gently rolling topography like the Konza site, variation in insolation and surface conditions (other than soil water content) may be conservative enough to allow the assumption of spatially constant atmospheric 'evaporative demand.' In the mountainous topography we are working in, exposure of the surface is quite variable, with the two hillslopes draining into opposite sides of each stream link potentially having very different net radiation environments included in the same subcatchment area. The topography–soils index expresses similarity in terms of the runoff and throughflow drainage system, but does not include effects of variable ET_p, particularly that induced by microclimate within a subcatchment. If different locations with the same value of $In(aT_e/T_i \tan\beta)$ have variable ET rates over time, they are not hydrologically similar, will not respond similarly to a storm event and will have different soil water curves through the year. In this respect, just as FOREST-BGC has been limited in applications in moderate to steep topography by the lack of distributed soil water computation, TOPMODEL has been limited for continuous simulations by the lack of distributed ET computations. While this may not be noticed when validation is done for single storm events on the basis of hydrograph matching or in gentle terrain where ET_p has low variance, it would be more apparent in steep terrain and for continuous simulations with soil water or other distributed variables used for validation.

In order to preserve hydrological similarity, all significant hydrological pathways should be considered. For the present circumstance of variable ET_p induced by exposure, each subcatchment region should be stratified by net radiation. In mountainous terrain this is effectively done by our method of partitioning the surface into hillslope units. Therefore, the conceptual and spatial framework adopted for integration of FOREST-BGC and TOPMODEL is to treat the watershed as a population of hillslopes, each of which is separately parameterized and simulated. Hillslope scale parameters (e.g. elevation, slope, aspect, area) are represented in the cartridge file (Table 1). Within hillslope patterns are represented with a distribution of the topography–soils index and index interval mean values of key variables, such as soil depth and LAI, according to observed soil–vegetation–topography catenae. In the semi-arid area where we have worked, there is a regular association of canopy LAI and soil properties with the similarity index (in the absence of recent disturbance). It is therefore useful to distribute these additional parameters with the similarity index by computing the mean values of LAI for each index interval as determined from remote sensing imagery, and assigning interval values of soil properties as well. Therefore, the replacement of the bucket model with TOPMODEL requires a more detailed parameterization of within hillslope patterns, but allows the computation and discrimination of a soil moisture gradient within each hillslope grading from the more xeric ridge environment, through the wetter riparian zones.

ILLUSTRATION OF THE INTEGRATED SYSTEM

System operation

RHESSys is designed to produce the landscape structure, compute the distributed parameter set and integrate and run the hydro-ecological processes of interest. It is therefore necessary that RHESSys be modular so that the individual components can be modified or replaced to suit particular hydrological or ecological environments. As an example, TOPMODEL is appropriate only to environments where throughflow is a dominant hydrologic process and is sufficiently rapid to justify the use of the steady throughflow assumption given above. In much flatter terrain, a simple bucket or multiple store model emphasizing vertical infiltration and soil water movement would be appropriate. In addition, we have found that the soils in the mountainous areas of western Montana we have been working in have extremely high infiltration capacities, such that use of an infiltration model was not necessary and was deactivated. As we move into other environments with less permeable surface soils, activation of the infiltration model would be appropriate.

Field site

The current implementation of RHESSys is described and illustrated in this paper for Soup Creek, a 15 km^2 watershed located on the west slope of the Swan Mountain Range in northwestern Montana. This alpine basin is a remnant glacial trough with an east–west orientation in the lower portions of the trough, grading into a northwest-southeast trend in the headwaters (see Fig. 4). The Swan Range forms a major barrier for Pacific Ocean air masses. The resultant orographic effects form a strong annual precipitation gradient ranging from 850 mm at the base of the mountains to over 2000 mm at the highest elevations. Deep snowpacks prevent soils from freezing during the winter, so that high infiltration capacity and throughflow rates can be maintained. Soil textures typically range from gravelly silty loam at the surface to extra gravelly silty loam or extra gravelly coarse sandy loam toward the soil column base. Coarse textures are found throughout the basin on alluvial fans, glacial tills and colluvial deposits. Limited field observations show the soils to increase in depth from 1 m or less near the ridges to over 2 m in the valley bottom. Soil moisture appears to follow a similar trend. In the summer of 1988, a year of extreme drought in western Montana, soil pits on the hillslopes and near the ridges were well drained while soils approached saturation near the valley bottom. The watershed supports a coniferous canopy, with a mix of Douglas fir (*Pseudotsuga menzeisii*), red cedar (*Thuja plicata*), larch (*Larix occidentalis*), lodgepole pine (*Pinus contorta*), subalpine fir (*Abies lasiocarpa*) and Engelmann spruce (*Picea englemannii*). Crown closure ranges from 30 to greater than 70% with an understory ranging from broadleaved shrubs in the more open canopies to forest litter in the closed canopies. The area was chosen for simulation work as it had been used for remote sensing determination of LAI (Peterson et al., 1988, Spanner et al., 1990) and ground measured stand LAI were available for the calibration of Thematic Mapper imagery. As noted

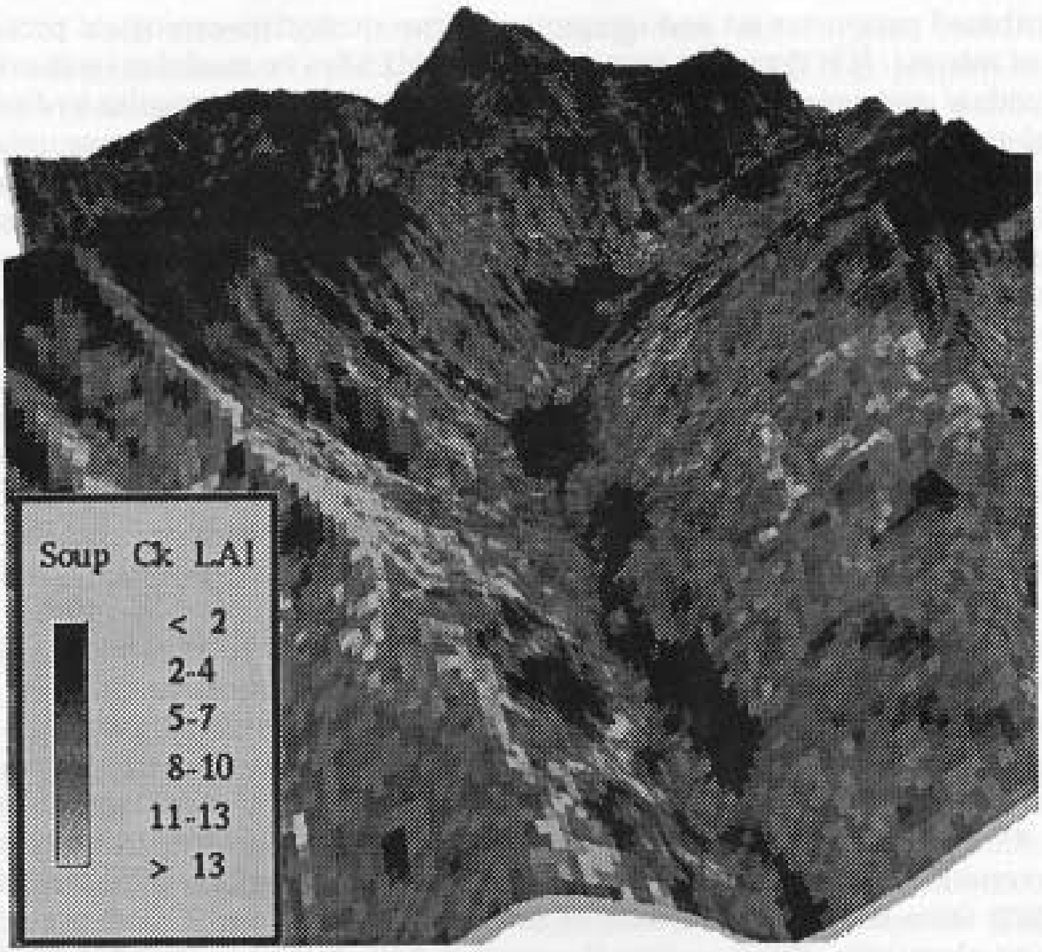
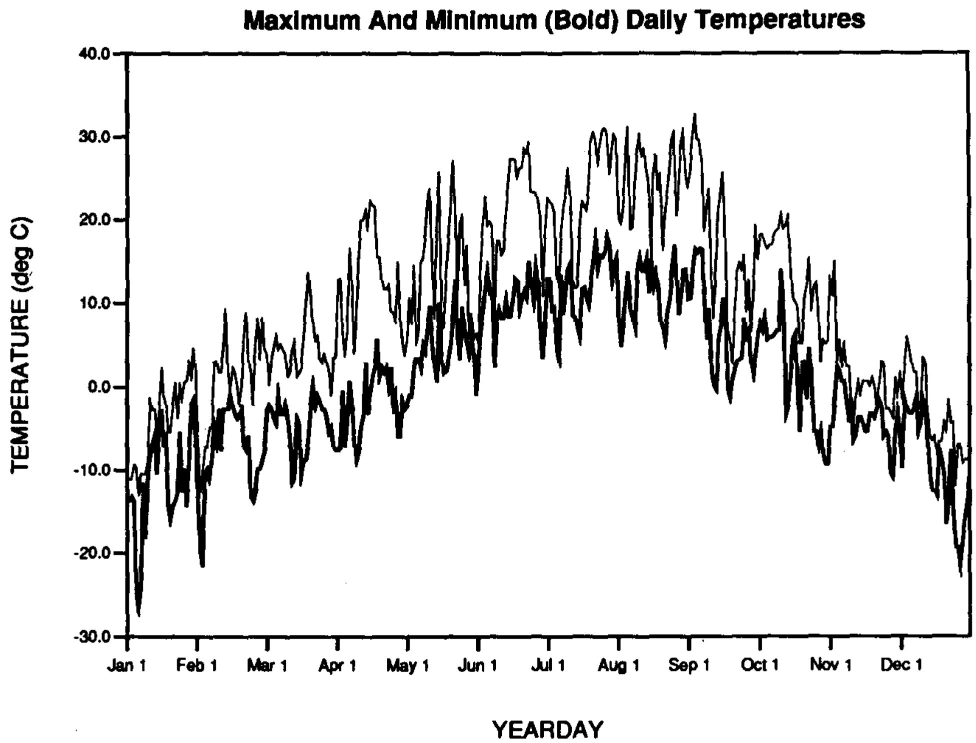


Fig. 4. Distribution of LAI as estimated from Thematic Mapper Bands 3, 4 and 5, draped over Thunderbolt Mountain 7.5 min DEM (30 m resolution). The trunk stream flows from east to west out of the Swan Mountains of western Montana.

above, detailed initialization data for the distribution of snowcover were not available for the 1988 simulation year. As the hydrology of the area is very dependent on the accumulated distribution of the snowpack in the watershed, the current simulations cannot be considered adequately initialized, and this leads to some deviations between observed and simulated results. In the present context, we have used the 1988 meteorological data to self-initialize the snowpack and soil moisture by running the simulation for an initialization year. Heavy precipitation in December (see Fig. 5) probably leads to an overprediction of the initial snowpack, and consequent runoff during the melt season.

Soup Creek parameterization and simulation

A DEM with a horizontal resolution of 30 m interpolated from the digitized contours of the Thunderbolt Mountain 7.5 min USGS topographic quadrangle



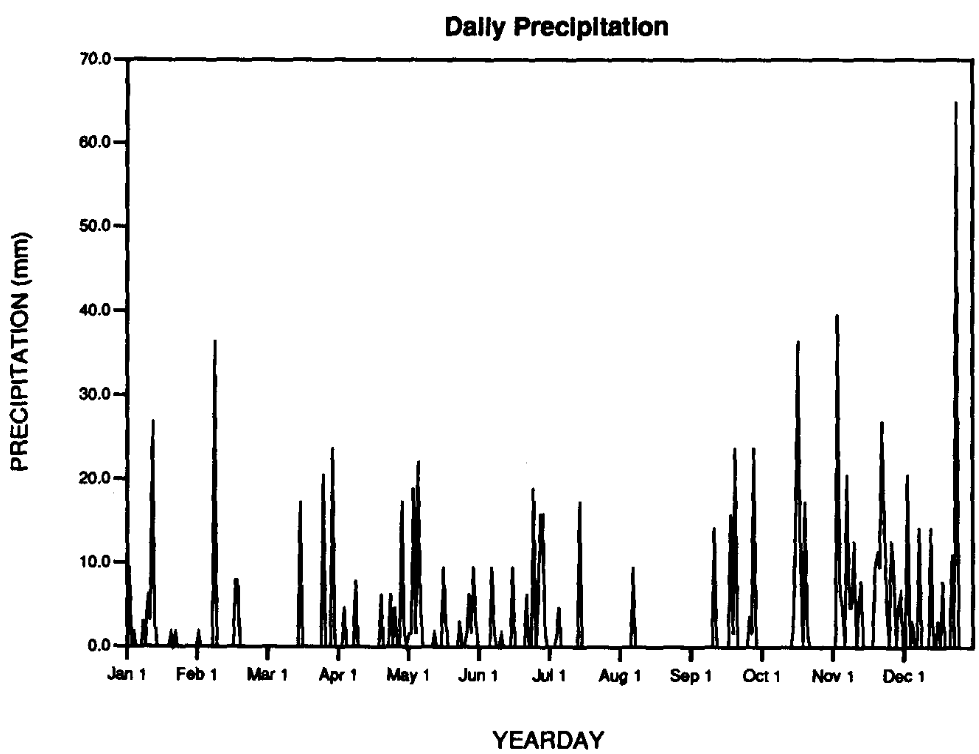


Fig. 5. Daily series of precipitation and maximum and minimum temperatures for the Swan Lake Ranger station for the year 1988.

TABLE 2

Sample distribution file for a low elevation, south facing slope in Soup Creek. Note that soil properties are simplified for this example, but can incorporate greater within hillslope trends if information is available

Index value	Frequency	LAI	K_s ($m\,day^{-1}$)	Root zone capacity (mm)
2.500000	13.00000	7.0	0.93	400.0
3.500000	95.00000	11.3	0.93	450.0
4.500000	266.0000	11.5	0.93	500.0
5.500000	549.0000	10.3	0.93	550.0
6.500000	359.0000	10.4	0.93	600.0
7.500000	92.00000	10.6	0.93	650.0
8.500000	33.00000	11.5	0.93	700.0
9.500000	15.00000	10.4	0.93	750.0
10.50000	14.00000	9.3	0.93	800.0
11.50000	1.000000	7.0	0.93	850.0

was produced by the US Forest Service Centronics Laboratory in Salt Lake City (Fig. 4). The DEM was used to partition the watershed into eleven stream channels and twenty-two hillslopes (Figure 2(a)) and to compute the necessary topographic information including exposure, gradient (β) and upstream drainage area (a). These information fields were combined with information from digitized soil maps to produce a field of the topography–soils index, $ln(aT_e/T_i\tan\beta)$, using the techniques of Band and Wood (1988) and Band (1989). LAI was estimated using an empirical correlation of Thematic Mapper Bands 3, 4 and 5 (630–690 nm, 760–900 nm, and 1550–1750 nm, respectively) with field estimated LAI for a number of sampled forest stands in the region (Fig. 4). The estimation techniques were modified from the simple function of the B4/B3 (infrared to red) ratio used by Peterson et al. (1988) and Spanner et al. (1990) to include a normalization by B5 to account for variable canopy cover. This modification is the subject of another paper (Nemani et al., 1992) and is not discussed further here.

An important part of the parameterization scheme is the availability of water in the soil column to roots. In the mountainous watersheds we have worked in, soils are typically shallow such that the rooting zone can be considered the entire soil profile. Where soils are locally deeper (up to 2 m), primarily the valley bottoms, tap roots are often found extending below the shallower root crown. Therefore, we have considered that the root zone is equivalent to the soil depth (to a maximum of 2 m). In the present context we simply compute the total soil water storage in the root zone for prediction of soil water potential, ψ_{soil}, and ψ_{leaf}. Table 2 shows the distribution file of parameter values within one of the Soup Creek hillslopes. A simplified soil catena is represented, showing increasing soil depth and root zone water capacity with increasing values of the topography–soils index (moving downslope). K_s is approximated for the uniformly coarse soil textures on the hillslopes, and is treated as a hillslope constant for this application, while LAI is computed as mean values for each topography–soils index interval. Simulation results should be interpreted with this parameterization in mind as alternate parameterizations would yield varying results.

We also have not incorporated shadows cast by adjacent terrain (although slope–aspect corrections are made for irradiance) as our sensitivity analyses have shown only small variance in computed daily *ET* with horizon angles ranging up to 40° while daily *PSN* is not significantly lowered until horizon angles exceed 30°. This may occur as the integrated solar radiation over the early morning and late afternoon is much smaller than the integrated midday insolation and the absorbed insolation assymptotes with increasing incident insolation at higher LAI.

The watershed partition with 22 hillslopes was used for the present illustrations to aggregate and compute mean values of elevation, aspect, gradient, LAI and other parameters following Lammers and Band (1990). The cartridge file represented in Table 1 contains the appropriate model parameters for each hillslope. From this point, the system loops through a full year's worth of simulation for each hillslope unit, reading in and extrapolating meteorological base station information to the given hillslope. The distribution of the topography-soils index distribution and index specific parameters computed for the hillslope are then read in and the integrated TOPMODEL/FOREST-BGC model is run over the index values. Results can be reported for each index interval in each hillslope, as aggregate hillslope results computed with areally weighted index values, or as aggregated full watershed values. Time series of intermediate and final results can be produced for any areal unit (index interval, hillslope or watershed) and as images for any time step. More details on the model integration and parameterization process for this area can be found in Patterson (1990). Different values of m were used to simulate scenarios with greater and lower spatial variance of soil water. The parameter α was set to a value of 0.04 in all cases as results were not found to be as sensitive to its variation as the parameter m.

For comparative purposes, RHESSys was also run using the FOREST-BGC soil bucket model to simulate soil water dynamics. For the case of the bucket model, simulations were carried out for each hillslope, treating the bucket as representative of soil water for the full slope. Two sets of bucket model simulations were run. The first used an *AWC* of 15 cm which is generally used as a representative value for a loamy soil, assuming a soil depth of one meter. These can be considered 'default' parameterizations in the absence of specific soil depth and texture data. A second set of simulations used the available estimates of soil depth through the watersheds to compute an areally weighted mean *AWC* of 22 cm. These model runs differ from the

15 cm bucket runs only in the 'depth of the bucket' but illustrate the effect of having more knowledge of the local soil depths through the watershed. The TOPMODEL runs then use the available soil information with the modeled throughflow processes for water redistribution, giving the third level of parameterization complexity.

Annual simulations were run for each model version for the year 1988. Trunk stream discharge was measured for most of 1988 with a stage recorder. Base station meteorological data was recorded at the Swan Lake Ranger Station of the US Forest Service. The daily series of precipitation and maximum and minimum temperatures for this year (Fig. 5) shows a period of relatively high temperatures and a lack of rain over the middle to late summer (approximately 15 July to 10 September). This drought period resulted in massive forest fires in the northern Rocky Mountains as soils and litter dried, and caused significant moisture stress in the forest ecosystem. This is a key period of time for evaluating and comparing the methods accounting for soil water dynamics in the simulations. It should be noted, however, that information was not available for the areal distribution of snowpack depth or soil moisture for model initialization. Therefore, the current simulations cannot be considered valid field testing of the different model forms and we do not attempt to 'tune' the models to match the seasonal discharge curve. Therefore, the observed seasonal hydrograph should only be used for comparison of the 'form' of the hydrographs and other flux trends rather than the specific flux magnitudes.

Simulation results

Seasonal hydrological trends for the full watershed

All versions of the simulations show nearly identical patterns and depths of watershed mean *ET* through the early part of the growing season until moisture stress conditions are reached (Fig. 6). These conditions are generated at different times for the different model runs depending on the soil water capacity and the degree of soil water redistribution. The 15 cm bucket shows more severe moisture limitations than the 22 cm bucket version. In the bucket models, there is no spatial variance of soil moisture within the hillslopes. The timing and severity of moisture stress is fully determined by the drought extent and duration, along with the elevation and exposure of the hillslopes. The moisture limited evapotranspiration continues through 10 September, when rainfall adequately recharges the soil.

The effects of redistributing soil water over the topography can be seen in the variance of watershed *ET* for the TOPMODEL simulations with m set as 0.02 and 0.12 (Fig. 6). With m set to 0.12, a large variation in soil moisture develops across the hydrological similarity indices according to eqn. (2). The lower interval values drain very rapidly while the higher values remain sat-

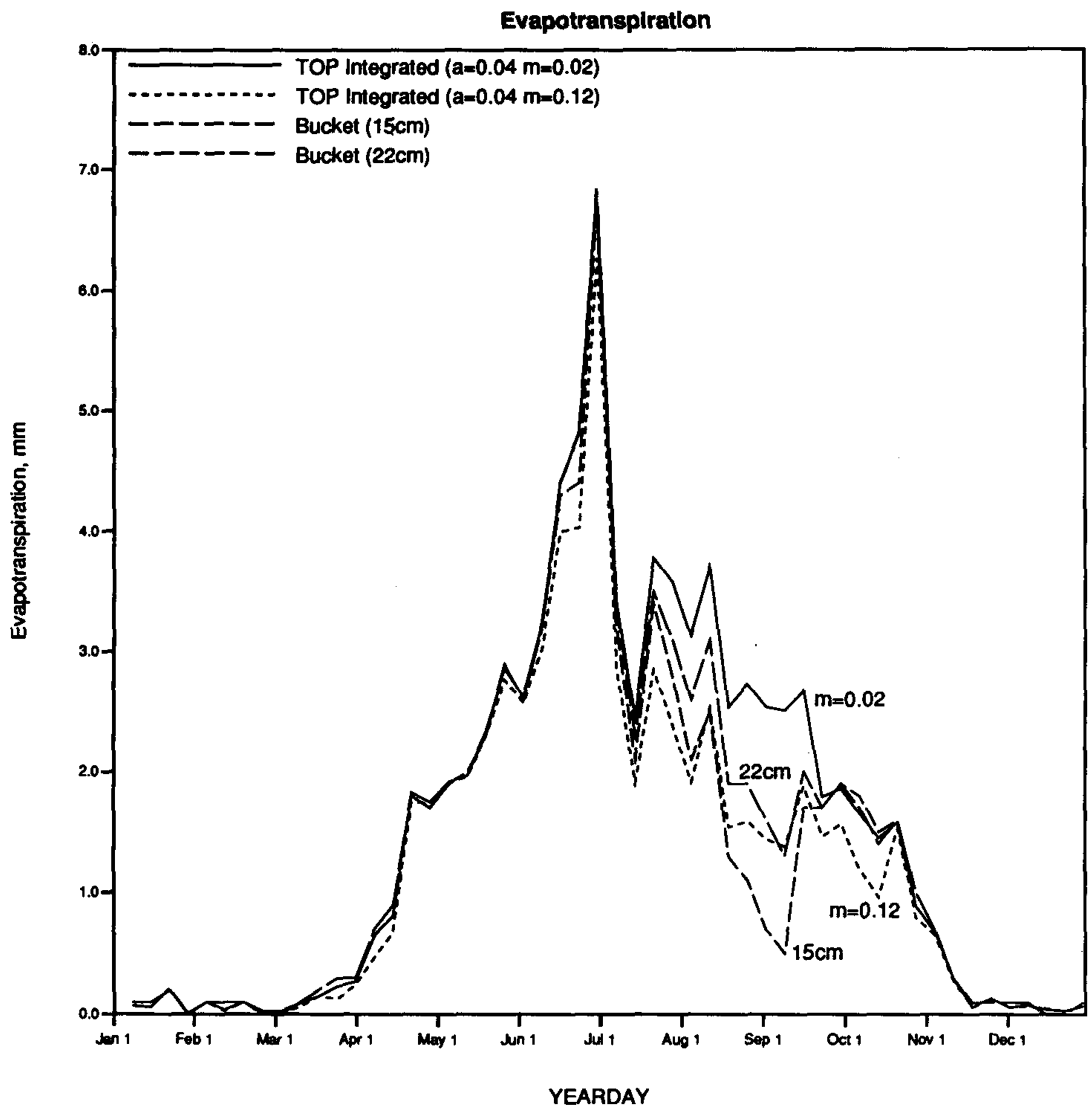


Fig. 6. Seasonal trajectories for areally weighted mean *ET* for the two bucket versions (*AWC* set to 15 and 22 cm) and two TOPMODEL versions (*m* set to 0.02 and 0.12) of RHESSys.

urated. The persistant saturation in the higher intervals maintains saturation overland flow as a consistent outflow of soil moisture. This active pathway drains the overall hillslope soil water, raising S' which expands the moisture stress in the lower intervals. The effect of saturation runoff production as a hydrologic pathway is to cause a small reduction in basin *ET* compared with the bucket models prior to the drought through the augmented hillslope drainage, but to partially buffer the impact of the drought by maintaining soil moisture and high levels of *ET* in a portion of the basin through the driest period of the summer. Therefore, a crossover of *ET* rates with the 15 cm bucket and, for a very short period, the 22 cm bucket can be seen during the drought.

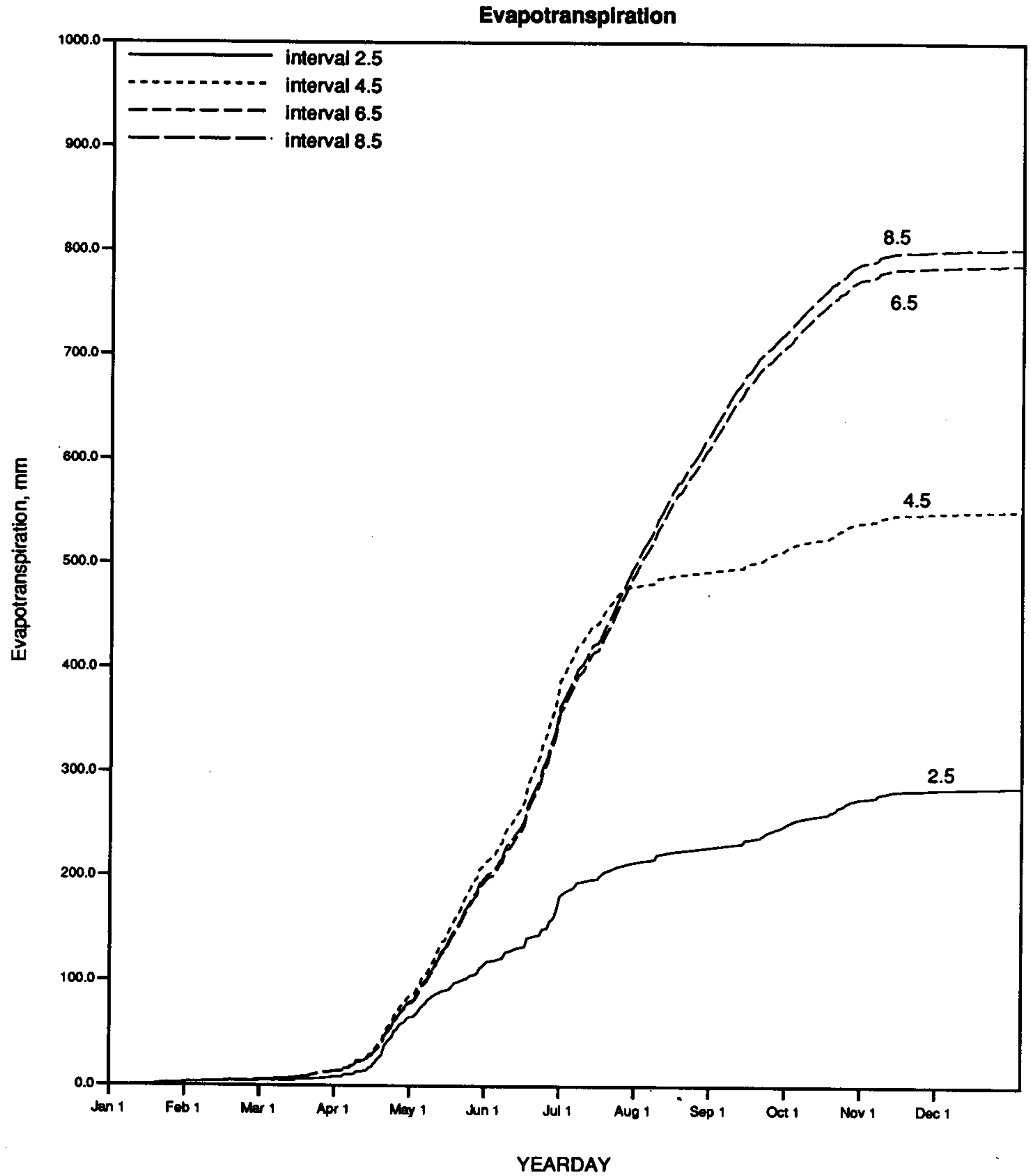


Fig. 7. Cumulative *ET* for different topography–soils indices simulated for a low elevation, south facing slope of Soup Creek. The lower index values show earlier decline of *ET* rates owing to moisture stress, while the higher index values show continuous, unstressed *ET* through the drought period.

Figure 7 shows the distribution of annual *ET* over the topographic–soils indices for a midelevation, south facing slope in Soup Creek. The higher index value zones maintain a high level of *ET* through the drought and contribute an annual *ET* as much as three times the *ET* of the lower index values. The spread of *ET* values among the intervals is largely determined by a combination of soil water recharge and site LAI. Note that at an index value of 6.5 and above, soil water does not appear to be limiting.

By comparison, simulation with m set to 0.02 does not provide the degree of moisture stress in the lowest hydrologic intervals or the degree of saturation and runoff production in the highest intervals. Basin ET rates do not drop as rapidly during the drought as the bucket models as a larger proportion of the basin is maintained with adequate moisture. The presence of the higher soil water zones in combination with the higher VPD during the drought maintains higher average watershed ET. In this case, soil water redistribution acts to buffer the impact of the drought without significant expansion of the saturation runoff pathway that would drain the hillslope soil water. At the driest portion of the drought (early September), daily watershed ET for the TOPMODEL simulation with lower m is nearly twice the daily ET of the higher m simulation and the 22 cm bucket, and nearly six times the 15 cm bucket.

Varying the degree of soil water distribution, and the consequent variations in ET and runoff production is illustrated over a range of parameter m values (Fig. 8). As the distribution of soil moisture increases with m, the increase in runoff production is balanced by the decrease in average ET. The increasing sensitivity of these processes as m rises results from the expansion of saturated or very dry conditions into the larger areas with moderate, rather than extreme topography-soils index values (see Table 2). Therefore, the partition of precipitation into ET and runoff production can be very sensitive to the representation of soil properties and drainage pathways. Note that mean basin precipitation is about 970 mm for 1988.

Figures 9(a), 9(b) show computed daily ET for 20 June, 1988 and 8 September, 1988 respectively, over the Soup Creek watershed. On 20 June radiation levels are near the highest levels of the year, but much lower on the steep north facing hillslopes, limiting ET. ET is probably temperature limited at higher elevations. The lower elevation south facing slopes have uniformly high ET rates with little spatial variance over their surface areas. On 8 September north facing and upper elevation slopes are probably radiation and temperature limited, respectively, during this time as well. The lower elevation south facing hillslopes now show much greater variance in ET rates, reflecting the much lower soil moisture contents near the ridges and on topographic knobs, while the lower portions of the hillslopes and surface hollows (high index values) have higher soil water content owing to recharge from above.

By comparison with the within hillslope patterns discernible in Figs. 7 and 9 the changing patterns of soil water and consequent canopy ET within the hillslopes is lost with the bucket versions of the model (Fig. 10). The ability to discriminate the relative patterns of ecological and hydrological flux contributed by different hillslopes and portions of hillslopes can have significant implications for watershed management and for our ability to validate model

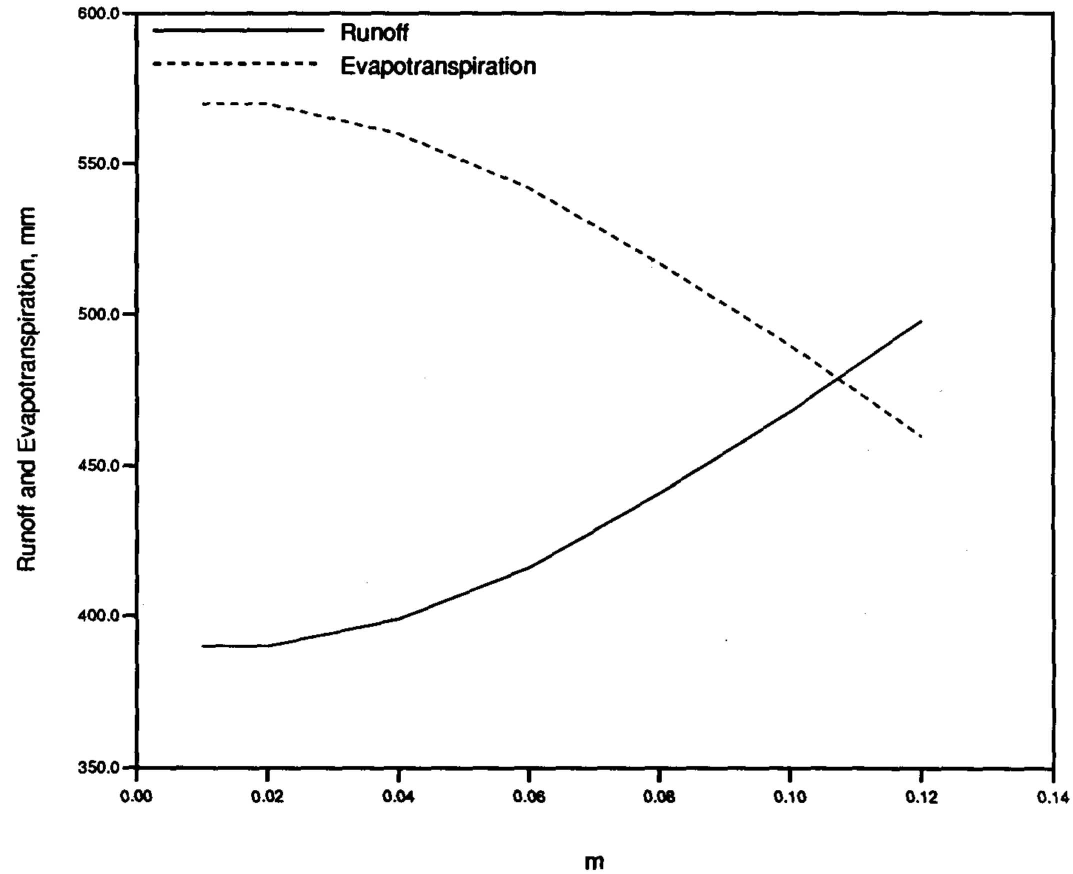


Fig. 8. Sensitivity of annual *ET* and watershed runoff to the parameter *m*, which controls the distribution of soil moisture over the topography–soils index values.

behavior. If both between hillslope and within hillslope patterns of state and flux variables (e.g. soil water storage, predawn ψ_{leaf}, *ET*,*LE*) could be independently estimated on the ground or with remote sensing techniques, model assessment could be carried out using measures of both integrated watershed response in the form of discharge curves and in the progression of surface state and the spatial patterns of water, energy and carbon flux over the season.

The timing of the impact of the drought on the ecosystem causing this response can be clearly seen in the plot of predawn ψ_{leaf} for three model versions (Fig. 11) simulated for the low elevation south facing slope. ψ_{leaf} for the TOPMODEL version is computed as mean values over all index intervals and with m set at 0.02. When the bucket soil water is depleted, ψ_{soil} and ψ_{leaf} drop very rapidly, increasing stomatal resistance. In the TOPMODEL version, the drop in water potential is largely restricted to the lower value (drier) index intervals, which eventually close their stomatae, restricting *ET* to cuticular levels, while the higher index intervals are not stressed and are still freely transpiring. Some comparison may be made with mean values of predawn ψ_{leaf} measured at a few locations on the hillslope, although it should

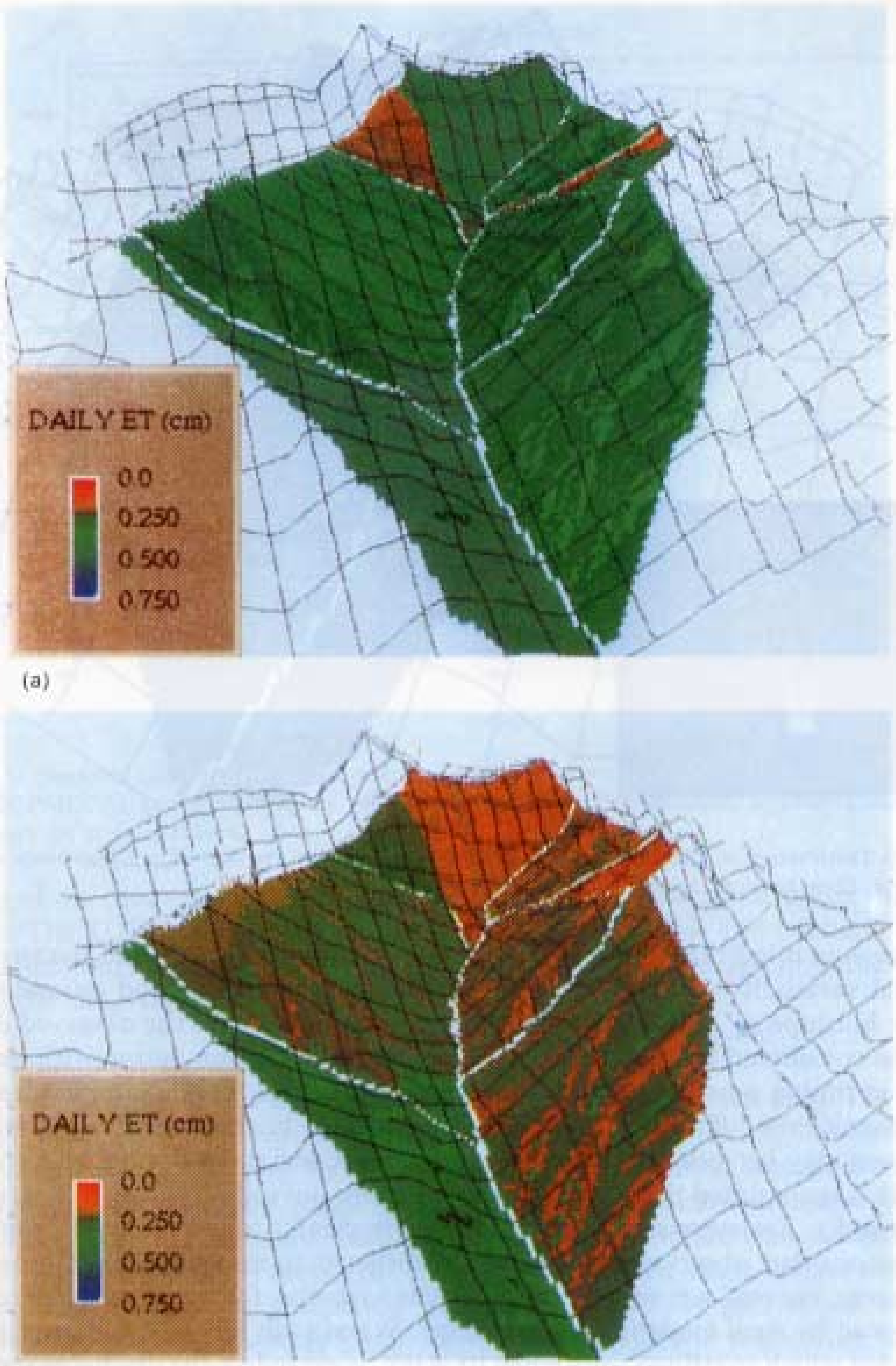


Fig. 9. Distribution of daily *ET* on: (a) 20 June, 1988 shortly after snowmelt; (b) 8 September, 1988 at the end of the summer drought period. Simulations run with $m = 0.085$.

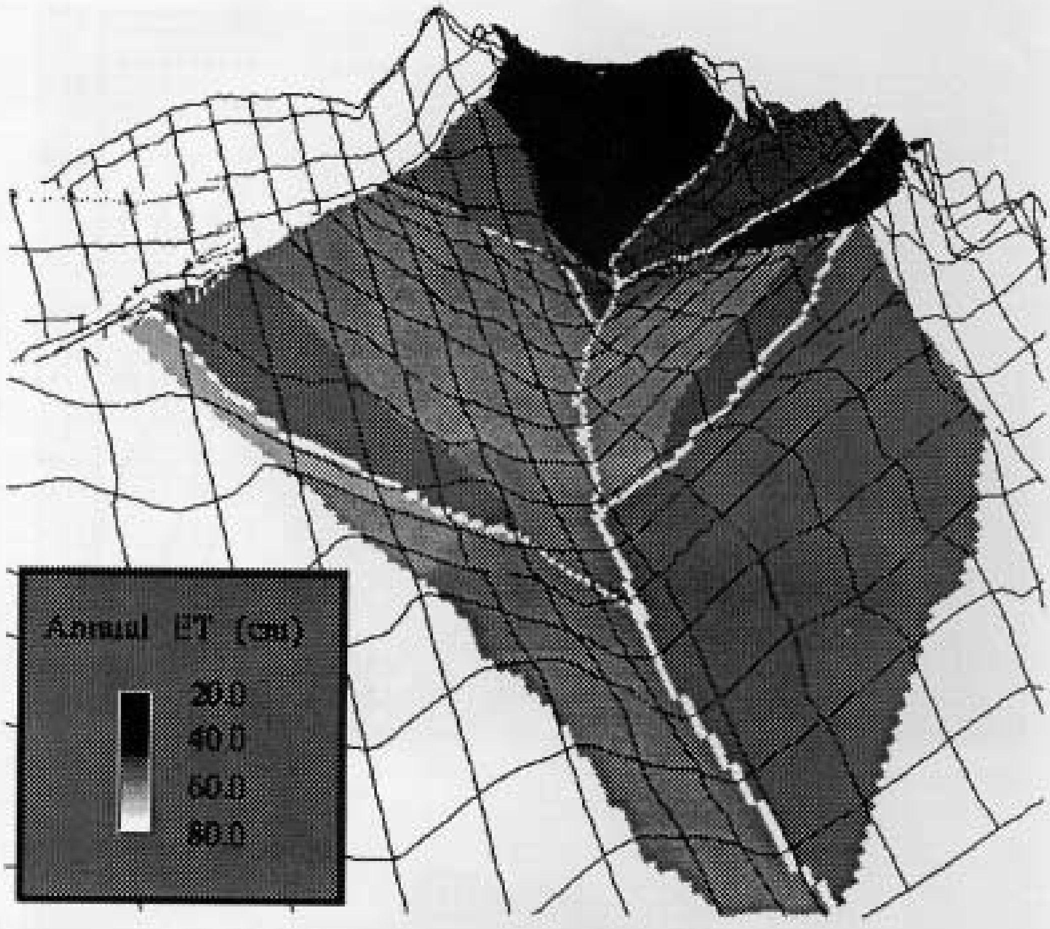


Fig. 10. Distribution of cumulative *ET* computed with a bucket version of soil hydrology with *AWC* = 22 cm towards the end of the growing season.

be borne in mind that the small number of sampling locations and the strong spatial variability of these values probably do not yield a good estimate of mean hillslope ψ_{leaf}. However, the trajectories and range of the observed ψ_{leaf} can be considered broadly similar to the TOPMODEL simulation with some shift in timing possibly owing to the specific locations of the sampled trees.

A significant difference in the performance of the bucket and TOPMODEL versions may be seen in the simulated discharge series for Soup Creek (Fig. 12). It is again noted that these simulations were not initialized with observed snowpacks, nor were model parameters specifically tuned to match runoff curves or other observations. Therefore, rather than matching magnitude of discharge curves, we simply seek to compare the forms of hydrographs produced by each modeling methodology. When each hillslope is considered a bucket which overflows only when its *AWC* is exceeded, the discharge series is strongly spiked with all discharge occuring on the day that *AWC* is exceeded, and none occuring on all other days. This record is quite unrealistic in comparison with the observed record, indicating that the poor representa-

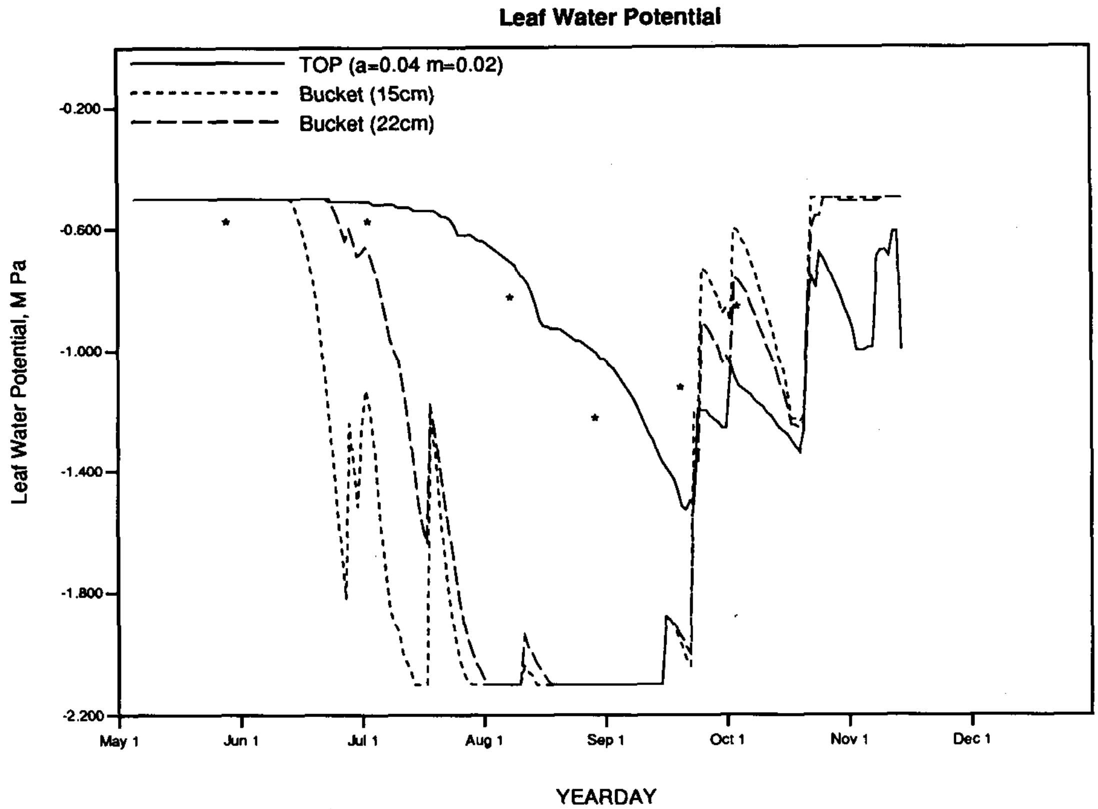


Fig. 11. Seasonal trajectories for areally weighted mean predawn ψ_{leaf} for the two bucket versions and the TOPMODEL version ($m = 0.02$), with the mean of several field measured ψ_{leaf} values observed through the growing season.

tion of the processes invalidates runoff records as a method of model validation. The TOPMODEL version releases water to baseflow and overland flow as a function of local soil water storage levels and yields a much smoother and realistic discharge curve.

Seasonal carbon balance trends

Average watershed *PSN* resembles *ET* through the season (Fig. 13) with some differences in the magnitude and timing of the crossover between the TOPMODEL and bucket versions that is dependent on the balance between primary production and respiration losses as distributed over the range of LAI and soil moisture encountered. *PSN* for the TOPMODEL simulation with $m = 0.02$ greatly exceeds the 15 cm bucket and the 22 cm bucket simulations through the drought period. Setting $m = 0.12$ for greater soil water distribution shows generally lower *PSN* rates through the year owing to the saturation overland flow drainage path, with the exception of a crossover with the 15 cm bucket during drought, similar to the *ET* trends.

Figures 14(a), (b) show daily *PSN* computed for 20 June and 8 September. The changes in the relative surface patterns of *PSN* appear similar but more

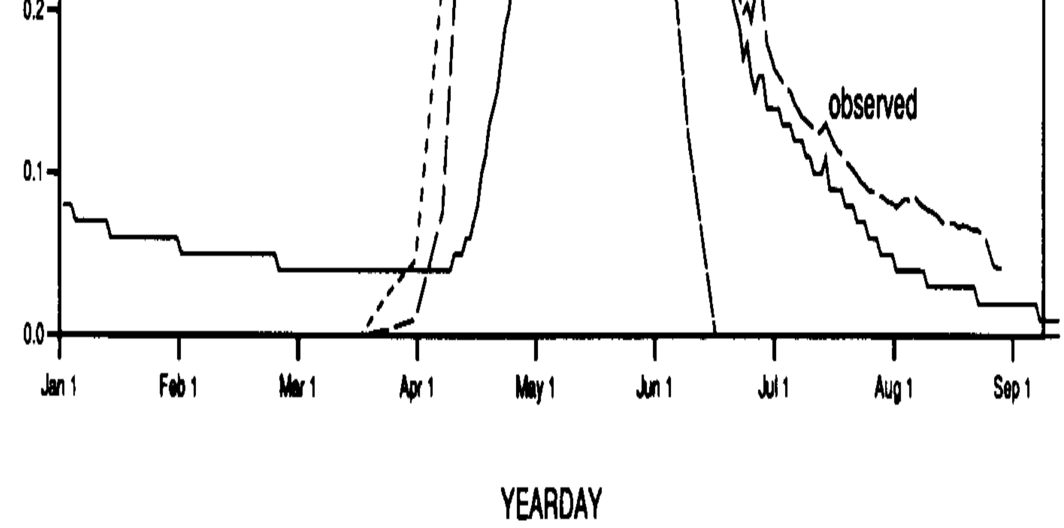


Fig. 12. Annual discharge hydrographs for the two bucket versions and a TOPMODEL version (m = 0.085) with observed discharge values.

pronounced than those followed by *ET*. 20 June shows uniformly high *PSN* rates with peak values on the low elevation, south facing hillslopes. The only areas with low *PSN* are hillslope sections characterized by talus (rapid drainage) and low LAI. By mid-September, night-time minimum temperatures are beginning to drop below freezing on a more regular basis and daylength is reduced. In combination with the water stress in selected topographic positions, a larger overall reduction in the magnitude of *PSN* and increase in *PSN* variance results. Much of the low elevation south facing hillslope that was formally the peak producing region now shows near zero or negative carbon balance with the exception of the wetter hillslope hollows. In contrast, the north facing slope is only at near zero or negative carbon balance

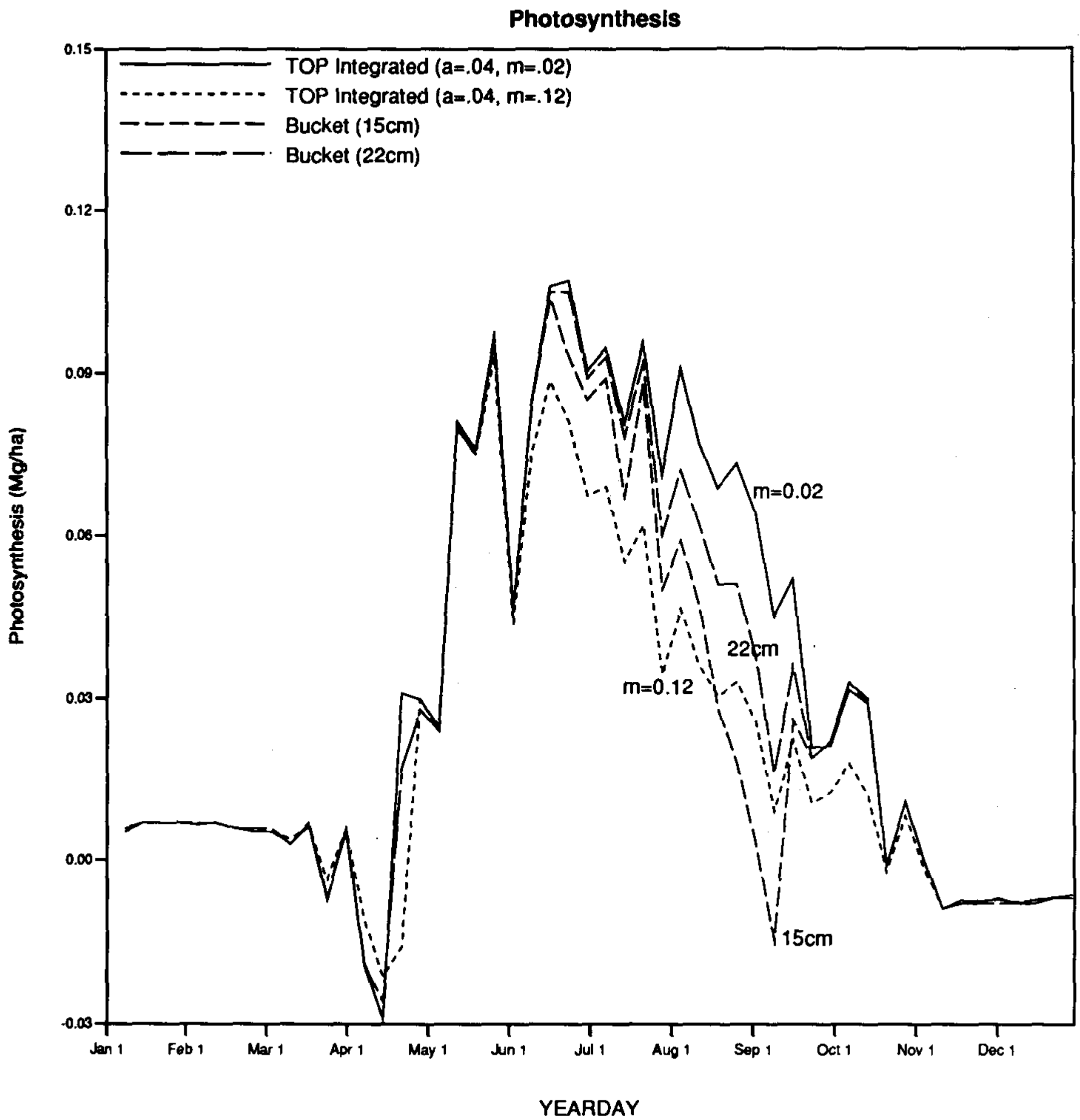


Fig. 13. Seasonal trajectories for areally weighted mean *PSN* for the two bucket versions (*AWC* set to 15 and 22 cm) and the two TOPMODEL versions (m = 0.02 and 0.12).

on topographic ridges and knobs (low index values). Higher elevations appear to have more spatially uniform, moderate *PSN* rates, partially owing to greater orographic precipitation input (computed with MTCLIM). Note that some areas marked as 0.0 Mg ha^{-1} actually have negative carbon balances on this day as the low gross carbon assimilation is offset by high respiration due to high surface temperatures.

DISCUSSION

The integration of the distributed soil water strategy with FOREST-BGC appears to yield a set of important observations in comparison with limited

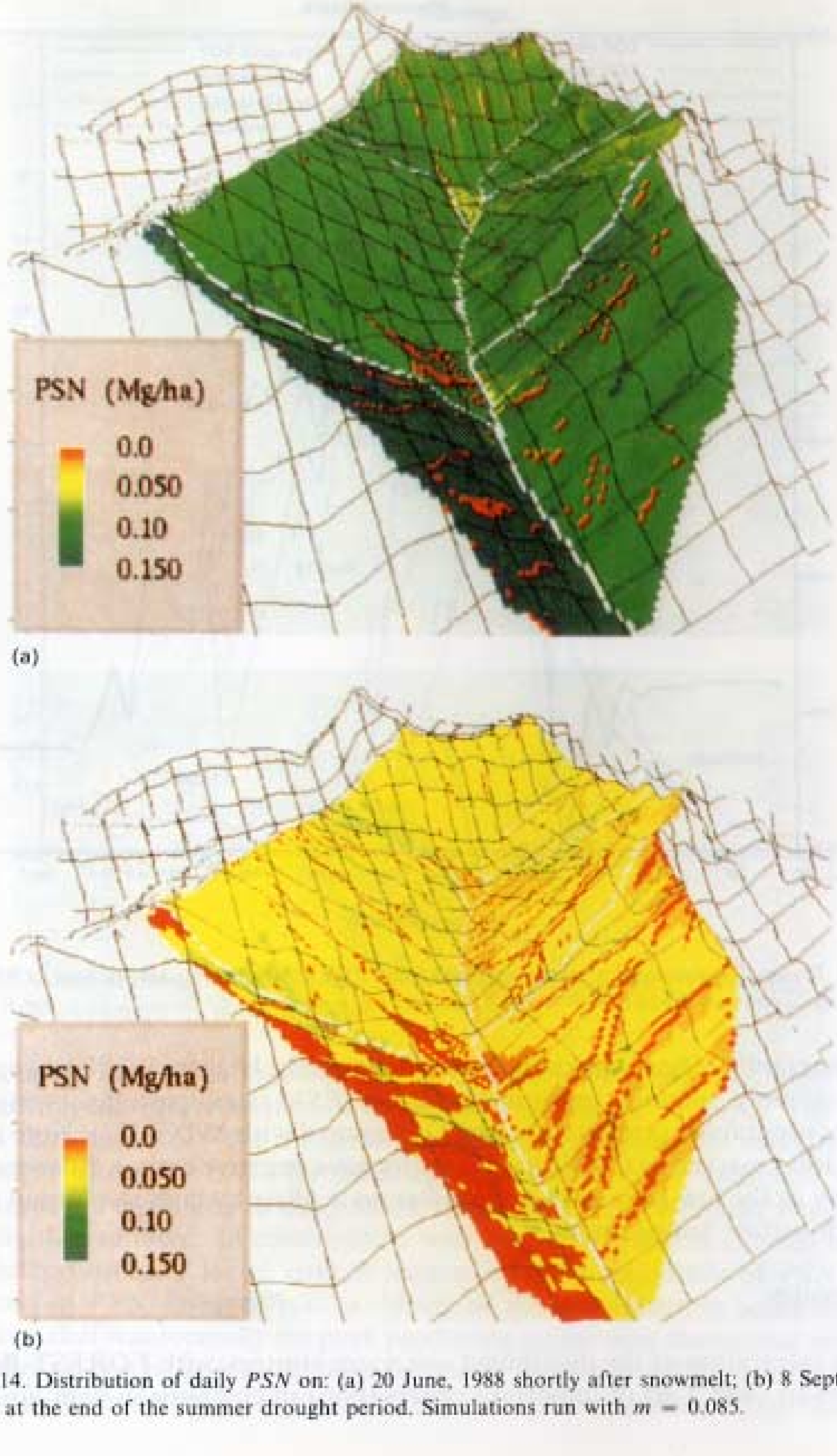


Fig. 14. Distribution of daily *PSN* on: (a) 20 June, 1988 shortly after snowmelt; (b) 8 September, 1988 at the end of the summer drought period. Simulations run with $m = 0.085$.

simulations and 'bucket' versions. First, there are significant differences in evapotranspiration rates and net productivity rates which result from the different model strategies and the degree of simulated soil water distribution (although the specific parameterization of soil rooting depth and shading should be borne in mind). These differences are due to a combination of the variable soil moisture patterns that develop within the hillslopes with the TOPMODEL version, along with the magnitude of both throughflow and runoff pathways. Distributed parameterization of LAI and soil attributes such as depth, texture and hydraulic variables are represented by stratification with the drainage area and slope computed from the DEM. The general trends represented by the parameter covariance are important to represent as they yield distinct microenvironments which behave very differently over the growing season.

One significant trend that has been illustrated by this study is the sensitivity of watershed *ET* and *PSN* to the spatial distribution of soil moisture and canopy properties. The combination of continued soil water recharge by throughflow processes in limited areas of the watershed that generally have higher LAI can buffer the drought impact on areal *ET* and *PSN* rates. While these areas do not comprise a large percentage of the watershed area, they can produce a much larger proportion of total watershed *ET* and *PSN*, leading to greater areal flux rates compared with a model parameterization that does not incorporate the vegetation–landscape trends. At the same time, excessive distribution of soil water leads to the rapid expansion of soil moisture deficits in the driest portion of the basin and produces a significant drainage pathway by saturation runoff processes in the wettest portion, augmenting overall drainage rates. In addition to these impacts on areal average water and carbon flux rates, if there is interest in the patterns of surface flux and forest productivity over the landscape, the spatial distribution of land surface properties and soil water are essential to parameterize and model.

Second, if basin hydrological yield is considered an important model output, it is essential that the pathways of hillslope drainage by throughflow and partial area runoff generation be represented. It is well known that in steep forested catchments, surface overland flow by saturation mechanisms is restricted to a small area immediately around the drainage lines or in topographic hollows on the hillslopes. The hydrograph produced by the bucket model is not useful for validation against gauged flow.

The spatial patterns of model output in combination with the seasonal trajectories of catchment wide flux and state variables (e.g. outflow, soil water, ψ_{leaf}) represent a useful database for validation purposes. While many models can reproduce a basin outflow hydrograph without actually representing the significant processes (such as lumped parameterizations), setting a validation requirement including seasonal shifts in the spatial patterns of surface flux and state variables constrains the range of modeling strategies to those that adequately represent spatially distributed storage and flux, and the important interactions of adjacent land surface elements. We believe that our data and simulation system effectively provides the ability to both spatially parameterize the models and validate them at scales ranging from the entire watershed (by the hillslope to hillslope variation) to the patterns typically developed within the different hillslopes.

If these patterns could be captured with remotely sensed imagery at key points of the growing season, we would be able to test the model performance by comparison with the simulated flux patterns. The images of these processes and of the intermediate products of soil water and leaf water potential taken at specific time steps give the opportunity to structure a comparison with a remotely sensed image that shows an analogous process or quantity. This would need to be done with a series of such images through the year such that shifts in these patterns could be discerned on both simulated and remotely sensed images.

Finally, it should be noted that we have largely developed and tested our models for steep basins that endure a summertime drought, concentrating on the ability to capture the important interactions between surface ecological and hydrological processes in these environments. It is pointed out that the results shown here may be specific to water-limited ecosystems, and that the model year was one of significant drought. In more humid environments or in flatter topography, there will be a shift in some of the key controlling variables, and a required change in the manner in which some processes are represented. In flatter terrain, TOPMODEL may not be necessary or appropriate, or the stratification by hillslopes may not be necessary. The further development of RHESSys involves augmenting its modular structure, not just in the range of model forms and subroutines available for different circumstances, but the manner in which they are linked and run. It is expected that the range of conditions for which a given model form is valid may not be encountered uniformly through an entire basin (especially in very large basins). We are currently generalizing the ecological models used in RHESSys into a suite of different BIOME-BGC components (Running and Hunt, 1993) appropriate to a series of biomes. A similar diverse set of hydrological tools will be assembled.

ACKNOWLEDGMENTS

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APPENDIX

TOPMODEL assumes that there is an approximate balance between the rate of recharge in the contributing region of an element and the rate of throughflow across the element boundary such that

$$q_t = ar \tag{A1}$$

where $q_t(\mathrm{m^2\,s^{-1}})$ is the lateral soil water throughflow discharge per unit hillslope width, r $(\mathrm{m\,s^{-1}})$ is recharge rate to the saturated zone and a (m) is the surface area contributing drainage per unit hillslope width. With the assumption that surface gradients adequately approximate the hydraulic gradients in the saturated zone, q can also be modelled with a modified form of Darcy's Law

$$q_t = T \tan\beta \exp(-S/m) \tag{A2}$$

where $T(\mathrm{m^2\,s^{-1}})$ is the hydraulic transmissivity of the soil, (β) is the surface gradient, S(m) is the local soil water content measured as a deficit below saturation and m is a parameter controlling the decay rate of q with S. When S is zero (no soil deficit, or saturation), the exponential term drops out and eqn. (A2) is very close to Darcy's Law. Lower values of m model a larger exponential decay of q as S rises, as would be expected in soils in which the lateral hydraulic conductivity drops rapidly with depth. Hence, m is a function of soil structure and effectively controls soil water drainage and hydrograph recession. Equating eqns. (A1) and (A2) and rearranging terms gives an expression for the soil moisture deficit at point i

$$S_i = -m\,ln\left(\frac{ar}{T_i \tan\beta}\right) \tag{A3}$$

which can be integrated over the full catchment area, A, to give the mean catchment saturation deficit

$$S' = \frac{1}{A}\int_A -m\,ln\left(\frac{ar}{T_i \tan\beta}\right)\mathrm{d}A \tag{A4}$$

Substituting for lnr and rearranging gives

$$S_i = S' + m\lambda - m\,ln\left(\frac{aT_e}{T_i \tan\beta}\right) \tag{A5}$$

where

$$\lambda = \frac{1}{A}\int_A ln\left(\frac{a}{\tan\beta}\right)\mathrm{d}A \qquad lnT_e = \frac{1}{A}\int_A ln(T_i)\,\mathrm{d}A$$