Cambridge University Press
978-0-521-57125-8 - Scale Dependence and Scale Invariance in Hydrology
Edited by Garrison Sposito
Excerpt
More information

1

Scale Analyses for Land-Surface Hydrology

ERIC F. WOOD

1.1 Introduction

This chapter discusses some research problems associated with the scaling of land-surface hydrologic processes involved in the land-surface water-and-energy budget. As pointed out by Dubayah, Wood, and Lavallée (1996), the term "scaling" has come to have multiple definitions, depending not only on the general discipline (e.g., hydrology, meteorology, geography, physics) but also on the application within a discipline. Dubayah et al. (1996) refer to a process as exhibiting scaling "if no characteristic length scale exists; i.e. the statistical spatial properties of the field do not exhibit scale-dependent behavior." Blöschl and Sivapalan (1995) refer to scaling as the transfer of information between different spatial (or temporal) lengths. The "transfer of information" may consist in mathematical relationships, statistical relationships, or observations describing physical phenomena. This definition is consistent with that of Dubayah et al. (1996), albeit more qualitative. The term "upscaling" is often used by the remote-sensing community to describe going from observations on a small spatial (or temporal) scale to observations on larger scales, whereas "downscaling" consists in going from large-scale observations to smaller scales.

Interest in scaling as it relates to the land-surface water-and-energy budget arose from the more general problem of land-surface parameterization in climate models. In the 1970s, a series of Atmospheric General Circulation Model (AGCM) climate studies demonstrated the importance of land hydrology for the earth's climate: the sensitivity of albedo to climate (Charney et al., 1977), and the influence of soil-moisture anomalies (e.g., Walker and Rowntree, 1977). Those studies and related work resulted in the establishment of a Joint Scientific Committee Working Group on Land Surface Processes, with Peter S. Eagleson as chairman, under the World Meteorological Organization/International Council of Scientific Unions (WMO/ICSU) Joint Scientific Committee for the World Climate Research Programme (WCRP). That working group organized the Study Conference on Land Surface Processes in Atmospheric General Circulation Models in January 1981 help assess the state of the art in modeling land-surface hydrologic processes on the scale of AGCMs, to recommend research activities to improve knowledge in this area, and to suggest data requirements for initialization, validation, and parameter evaluation (Eagleson, 1982). The work and influence of that committee helped establish climatologic experimental programs like HAPEX (Hydrology Atmospheric Parameterization Experiment) and ISLSCP (International Satellite Land Surface Climatology Programme) to provide experimental data (however limited they may be) that have been used for more than a decade for scaling studies of land-surface processes. In the late 1980s the ICSU established the core project Biospheric Aspects of the Hydrological Cycle (BAHC). One research focus is spatial and temporal integration of biospheric–hydrospheric interactions. Even though such programs are under way, in general the land-surface research by the climate-modeling community has focused primarily on improved process models at small scales, using data collected through experiments like HAPEX-MOBILY and FIFE '87, with less emphasis on scaling point observations and the development of mathematical descriptions of land-surface processes at larger scales. Thus there remain opportunities for researchers to make significant contributions to the field.

My own interest in investigating land-surface hydrologic parameterizations and scaling arose through the influence of the seminal paper by Dooge (1982) that is part of the proceedings of the Study Conference on Land Surface Processes in Atmospheric General Circulation Models (Eagleson, 1982). This paper remains required reading for those who wish to understand the early foundational work in the scaling of land-surface hydrologic processes for AGCMs. Dooge clearly identifies the key role of soil moisture and its variability in the coupling of land-surface hydrology to atmospheric models and reviews the progress in estimating soil-moisture changes and evaporation, from field scales on the order of 10–100 hectares (ha) to catchment scales (100–1,000 km^2) to regional and AGCM grid scales (10,000–100,000 km^2). He comments that "in linking [soil moisture and evaporation] phenomena on that scale (10–100 ha.) to the usual scale (100–1000 sq. km.)... a number of approaches have been tried [but] this problem is largely an unsolved one" (Dooge, 1982). Though we have learned much over the past 15 years and certainly have gained an appreciation of the scaling problem, Dooge's comment is as valid today as in 1982. Over the past 15 years, considerable effort has been expended in trying to solve Dooge's problem. Much of that research has been reported in special conference and workshop proceedings that offer good overviews of current thinking and progress. Such publications include those by Gupta, Rodríguez-Iturbe, and Wood (1986), Bolle, Feddes, and Kalma (1993), Kalma and Sivapalan (1995), and Stewart et al. (1996).

The remainder of this chapter will be organized as follows: In the next section, the scaling problem as it relates to land-surface hydrologic processes will be more formally stated. Section 1.3 discusses the important role of variability in scaling. Section 1.4 presents selected research findings, first looking at simple analytical modeling results, followed by insights from analyses using complex land-surface models, and finally inferences derived from climatologic field experiments. Section 1.5 addresses the "approach of equivalent parameters," which is widely used by remote-sensing and micrometeorology scientists in scaling land-surface processes and fields. As described in Section 1.4, most of the work to date has been by empirical analysis. Sections 1.6 and 1.7 explore the potential of using statistical self-similarity theory for scaling land-surface processes. The final section provides a discussion of the progress to date and a look toward the future. Throughout these sections, an attempt has been made to indicate the areas of greatest potential for further research.

1.2 Statement of the Scaling Problem

The terrestrial water-and-energy-balance equations can be written as follows:

$$\left\langle \frac{\partial S}{\partial t} \right\rangle = \langle P \rangle - \langle E \rangle - \langle Q \rangle \tag{1.1}$$

and

$$\langle R_n \rangle = \langle \lambda E \rangle + \langle H \rangle + \langle G \rangle \tag{1.2}$$

where S is the moisture in the soil column, E is the evaporation from the land surface into the atmosphere, P is the precipitation from the atmosphere to the land-surface, Q is the net runoff from the control volume, R_n is the net radiation at the land-surface, λ is the latent heat of vaporization, H is the sensible heat, and G is the ground heat flux. The spatial average for the control volume is denoted by $\langle \cdot \rangle$. Equations (1.1) and (1.2) are valid over all scales, and only through the parameterization of individual terms does the water-and-energy balance become a "distributed" or "lumped" model.

By a "distributed" model I mean a model that accounts for spatial variability in inputs, processes, and parameters. This accounting can be either explicit – in which case the actual patterns of variability are represented, as, for example, in the European hydrologic system model SHE (Abbott, 1986a,b) and the three-dimensional finite-element catchment model of Paniconi and Wood (1993) – or statistical – in which case the patterns of variability are represented statistically [examples being models like TOPMODEL (Beven and Kirkby, 1979) and its variants (Moore, O'Laughlin, and Burch, 1984; Famiglietti and Wood, 1994; Liang et al., 1994), in which topography, soils, and vegetation play important roles in the distribution of water within a catchment].

By a "lumped" model I mean a model that represents the catchment (or grid) as being spatially homogeneous with regard to inputs and parameters. In engineering hydrology these models include the well-known unit hydrograph and its variants; in climate modeling these include the complex atmospheric–biospheric models such as the Biosphere Atmosphere Transfer Scheme (BATS) (Dickinson et al., 1986) and the simple-biosphere (SiB) model (Sellers et al., 1986).

Why do scaling problems exist? That is, why cannot data or model output be simply scaled up or down to meet the task at hand? Because the gridded output is inherently scale-limited by the grid spacing of the model or observation sensor, little information can be inferred below the smallest grid scale. As pointed out by Dubayah et al. (1996), this can hamper modeling over large areas, especially when the processes are spatially autocorrelated, because most hydrologic processes scale nonlinearly; that is, the moments (e.g., the mean and variance) obtained at one spatial scale may be significantly different from those obtained at a larger or smaller scale. Efforts to parameterize subgrid-scale variability are means for adjusting the statistical properties of fields to incorporate unmodeled fine-scale variability. In fact, very few models participating in the WCRP's Project of Intercomparison of Landsurface Parameters Scheme (PILPS) consider subgrid variability, either implicitly or explicitly. The model of Liang et al. (1994) considers variability in infiltration capacity (and soil moisture) through a distribution function, and the model of Koster and Suarez (1992) uses a mosaic approach for different vegetation types.

The terrestrial water balance, including infiltration, evaporation, and runoff, is known to be a highly nonlinear and spatially variable process, especially because of the key role that surface-soil moisture plays and its spatial and temporal variabilities due to variabilities in soil properties (e.g., Freeze, 1980). Thus the problem:

> What is the linkage between the parameterization of point-scale hydrologic processes and the parameterization of catchment-scale processes?

Study of this problem began in 1984. Figure 1.1, adapted from that early work, provides the framework. In general terms, the problem was stated as follows: Given a point representation of infiltration, drainage, and evapotranspiration, $g\{\theta(x), i(x,t)\}$, that is a function of spatially varying parameters $\theta(x)$ subjected to spatially and temporally varying inputs $i(x,t)$, which results in water-balance (and energy-balance) fluxes $o(x,t)$, we want to know how the integrated function of $g\{\theta(x), i(x,t)\}$, represented by $G(\Theta, I)$, changes as we consider the integrated function of $o(x,t)$ defined by

$$O = \iint_{t\in T, x\in A} o(x,t)\,dx\,dt \tag{1.3}$$

Here, $o(x,t)$ can be viewed as point fluxes, and O as the hillslope, catchment, or AGCM grid scale, depending on the size of the domain A. The degree to which

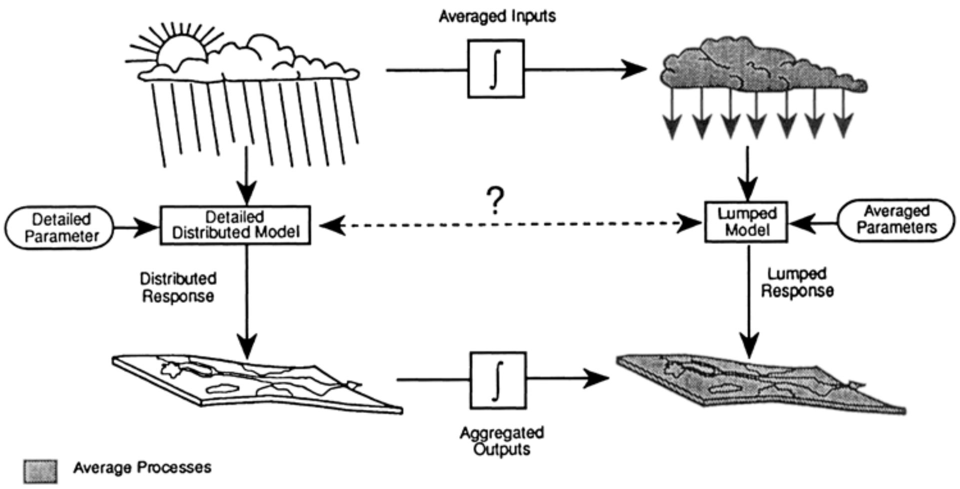

Figure 1.1. Schematic of aggregation and scaling in hydrologic modeling.

the aggregated outputs from distributed models will correspond to the output from lumped models will depend in part on the nonlinearity of the model process equations relative to the input fields, the spatial autocorrelation and scaling properties of those fields, and the amount of model spatial interaction.

The early research focused on the transition $g(\cdot) \to G(\cdot)$ under varying scales and stochastic spatial/temporal correlations. Perfect spatial correlation implies $g(\cdot) = G(\cdot)$. If $g(\cdot)$ is linear, then the stochastically varying variables can be represented by effective (or average) values $\langle O \rangle$ and $\langle I \rangle$, such that $g(\langle O \rangle, \langle I \rangle) = G(\langle O \rangle, \langle I \rangle)$. More discussion of effective parameters appears in Section 1.5. As nonlinearity in the point-scale processes increases, then averaging of the function $g(\cdot)$ occurs, and the functional form of $G(\cdot)$ will be significantly different from $g(\cdot)$.

1.3 Understanding the Role of Variability

Understanding the roles of spatial and temporal variabilities is central to understanding scaling. Variability in atmospheric forcings (rainfall and radiation), variability in land-surface characteristics (soil, vegetation, and topography), and variability in land-surface hydrologic processes all affect the transition $g(\cdot) \to G(\cdot)$. In an effort to understand the role of variability, Wood et al. (1988) carried out an empirical averaging experiment. Basically, their study consisted in averaging runoff over small subcatchments, aggregating the subcatchments into larger catchments, and repeating the averaging. By plotting the mean runoff against mean subcatchment area, they noted that the variance decreased until it was rather negligible at a catchment scale of about 1 km^2. That analysis has been repeated for the runoff ratio (Wood, 1995) and evaporation (Famiglietti and Wood, 1995) using data from Kings Creek, which was

Cambridge University Press
978-0-521-57125-8 - Scale Dependence and Scale Invariance in Hydrology
Edited by Garrison Sposito
Excerpt
More information

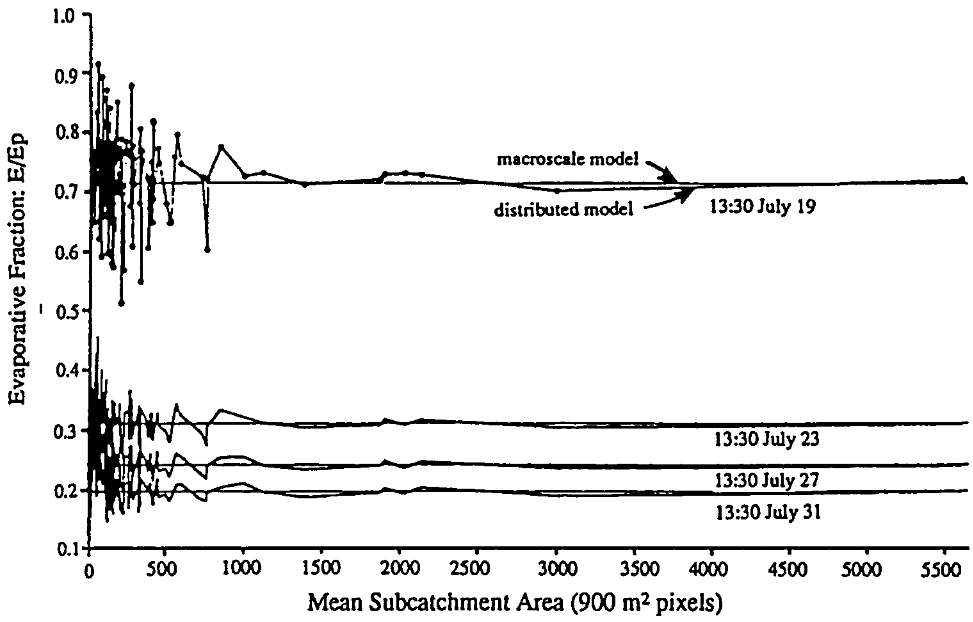

Figure 1.2. Comparison of inter-storm evapotranspiration values from a distributed hydrologic model and from a macroscale model for four times during the July 18–31, 1987, inter-storm period. The modeled catchment is the 11.7-km^2 Kings Creek catchment, which is part of the FIFE study area. The distributed model had a grid resolution of 30 m. The catchment was disaggregated into subcatchments, ranging from 66 to 5, depending on the scale of analysis. (From Wood, 1995, with permission.)

part of the FIFE '87 experiment. Figure 1.2 shows a typical result. The behavior of the catchment shows that at small scales there is extensive variability in both runoff and evaporation. This variability appears to be controlled by variabilities in soils and topography whose correlation length scales are on the order of 10^2–10^3 m – typical hillslope scales. At an increased spatial scale, the increased sampling of hillslopes leads to a decrease in the difference between subcatchment responses. At some scale, the variance between hydrologic responses for catchments of the same scale should reach a minimum. Wood et al. (1988) suggest that this threshold scale reflects a representative elementary area (REA), which they propose as a fundamental building block for hydrologic modeling and scaling. As defined by Wood et al. (1988), "the REA is the critical scale at which implicit continuum assumptions can be used without explicit knowledge of the actual pattern of topographic, soil, [vegetation,] or rainfall fields. It is sufficient to represent these fields by their statistical characterization." As pointed out by Beven (1995), the REA concept does *not* denote the scale at which average or equivalent parameters can be used in the continuum (macroscale) descriptions of the fluxes, because at the REA scale the distribution of characteristics may still be important – it is only the pattern of those characteristics that is unimportant.

The REA concept has raised some controversy since it was proposed. It is worth noting that several other hydrologic runoff studies have supported REA dimensions con

on the order of 1 km^2. In addition, a semivariogram from NS001-derived evaporative fraction for the EFEDA/Barrax site has a correlation length on the order of 750 m, which is quite consistent with the REA of 1 km^2 (see Fig. 5 of Bastiaanssen et al., 1996). Raupach (1993) describes three levels of heterogeneity from the perspective of the convective boundary layer (CBL). The smallest is the microscale heterogeneity (1–5 km), at which the turbulence and mixed-layer properties in the CBL cannot adjust to variations in the surface conditions. This may very well correspond to a sub-REA scale where pattern is important. Raupach defines macroscale heterogeneity as existing when the CBL turbulence adjusts fully from one "patch" to another, and he suggests that a homogeneous representation of the patch, using equivalent parameters (Raupach and Finnigan, 1995), can be made. I contend that at this scale, the macroscale patch representation may have to include the statistical representation of subgrid variability, as suggested by the REA concept.

Shuttleworth (1988) suggests the existence of "organized" and "disorganized" variabilities in surface cover (cover types A and B, respectively), depending upon whether or not the variabilities in surface conditions are "reflected in, and responsible for, associated changes in the mesoscale meteorology." Organized variability (cover type B) provides an organized response to the CBL and will have correlation lengths greater than about 10 km. As one can see, there is a loose correspondence concerning spatial variability and the scaling of land-surface processes. Nonetheless, as discussed throughout this chapter, I believe that the handling of variability is a critical problem and that the progress to date has been rather limited.

1.4 Results from Empirical Model Studies

The use of simplified models that can be solved analytically, or more complex models that must be solved numerically, can provide insights into scaling when used within a sensitivity framework. This allows the effects of spatial variability to be explored in a "what if" framework. The general approach is to develop the results for the model in hand and then make the leap of faith that the results will hold for the real world. All too often authors have taken that leap with great confidence, without the necessary caveats. A critical evaluation of papers using that approach is mandatory. In particular, the following questions should be asked: Are the dynamics too simple? Have critical processes been left out? Are the conditions tested reflective of observed conditions? In many cases, models that are just too simplistic are used under the guise that they "catch the essential features." In other cases the modeler ignores some physical process. One example of this is the paper by Collins and Avissar (1994), in which they conclude that stomatal resistance is the most important factor whose variability must be accounted for in upscaling. Their model has only limited soil-moisture representation, and stomatal resistance accounts for any surface control on evapotranspiration. Readers may misinterpret the essential conclusion that spatial variability in the *soil–vegetation* control is the critical process in understanding the scaling of surface hydrology and that the stomatal resistance is the surrogate for this in the Collins and Avissar (1994) paper.

Have such analyses provided essential insights? Can this type of approach provide essential insights into the scaling of land-surface processes? It is an appealing approach, and my belief is that progress can be made through carefully designed sensitivity studies using simplified models. Unfortunately, too many researchers have carried out such studies because they are relatively easy, often with models that appear mathematically elegant. It is critical that the findings from these models be further verified through observations and/or analyses with more complex land-surface models that have been validated through observations.

1.4.1 Results Using Simplified Models

Let us review some results from sensitivity studies in which I have been involved. Although my students and I learned a great deal from these studies, there was the belief that fundamental improvements to our land-surface models were needed before the conclusions reached with the simplified, analytical models could be generalized. One problem that we analyzed (Sivapalan and Wood, 1986; Wood, Sivapalan, and Beven, 1986) was the effect of spatial heterogeneity in soils and rainfall, with and without topographic effects, on the infiltration response of a catchment. Quasi-analytical expressions were derived for the statistics of the ponding time and infiltration rate. Examples of the derived mean areal infiltration rates are shown in Figures 1.3 and 1.4 for the different cases. These results show a strong bias (high) in mean infiltration rates if only mean values of soil properties and rainfall intensities are used in conjunction with point-scale infiltration rates. The results also show that the critical variable in understanding the spatial response is the time to ponding, its space–time distribution, and the resulting proportion of the catchment that is saturated. The upward bias has the effect of overestimating soil moisture and underestimating runoff.

The spatial correlation of the infiltration rate is an important variable in understanding the scale effects of runoff production. The findings of Sivapalan and Wood (1986) and Wood et al. (1986) indicate that the inclusion of soil variability has the effect of reducing the correlation and scale of variability in the infiltration rate. Thus, the results suggest opposing roles for rainfall and soil variability, with the importance of each depending on the relative magnitude of its variance. Whereas these results were derived for idealized catchments, the findings can provide important insights for actual catchments. For example, the constant-rainfall/variable-soil case may relate to medium-size, geologically variable catchments subjected to large-scale cyclonic events of relatively low rainfall intensities, whereas the variable-rainfall/constant-soil case could be related to small homogeneous catchments subjected to convective rain events.

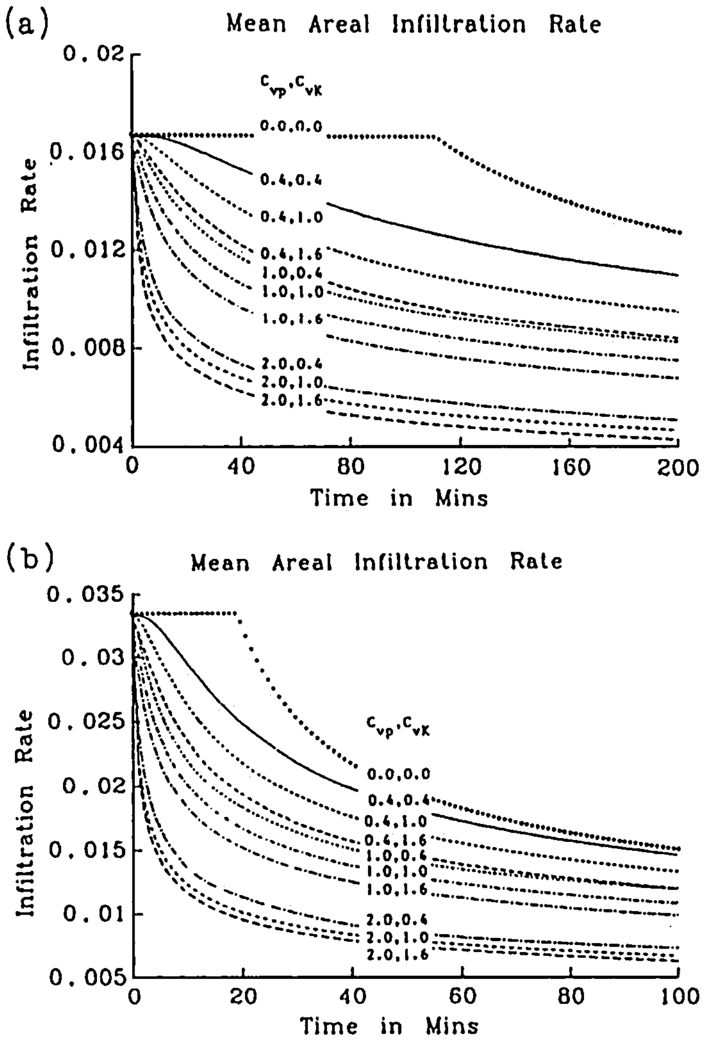

Figure 1.3. Mean areal infiltration rates due to spatially variable soils and rainfall. C_{vp} and C_{vK} refer to the coefficients of variation in precipitation and saturated-soil conductivity, respectively. The average saturated-soil conductivity was 0.008325 cm/min for both cases: (a) mean precipitation, 0.01665 cm/min; (b) mean precipitation, 0.0333 cm/min. (From Wood et al., 1986, with permission.)

This line of research was set aside (by myself) about 10 years ago to focus on improving the underlying process representation of the land surface, specifically the inclusion of vegetation and a surface energy balance, and to test these parameterizations against field data collected under FIFE and remote-sensing soil-moisture studies. As will be discussed further in a later section, Sellers, Heiser, and Hall (1992b), using analyses based on FIFE '87 data, suggested that land–atmosphere models are almost scale-invariant. Wood (1994) presented the counterargument, illustrating the mechanism for increased spatial variability in evapotranspiration under

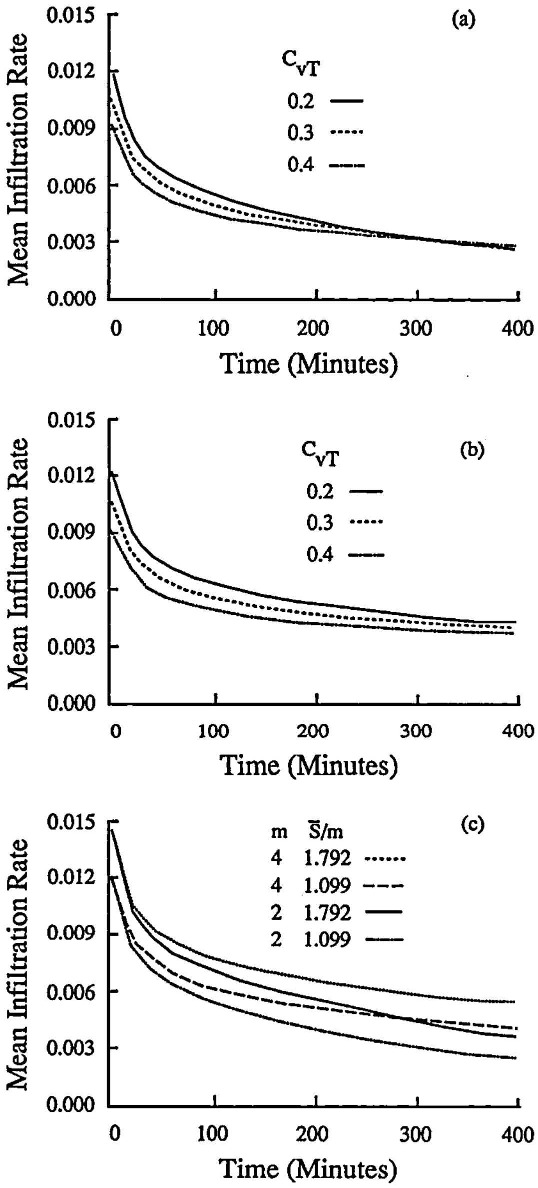

Figure 1.4. Mean areal infiltration rates due to spatially variable soils, rainfall, and topography. C_{vp}, C_{vK}, and C_{vT} refer to the coefficients of variation in precipitation, saturated-soil conductivity, and topographic index, respectively. The average precipitation was 0.01665 cm/min, and average saturated-soil conductivity 0.008325 cm/min, for all cases: (a) $\bar{S}/m = 1.0986$, $m = 2$; (b) $\bar{S}/m = 1.0986$, $m = 4$; (c) $C_{vT} = 0.5$. The variable $\bar{S}$ refers to the catchment-mean soil-moisture deficit, and m is a recession parameter within TOPMODEL (Beven and Kirkby, 1979). (From Wood et al., 1986, with permission.)