The Estimation of the Fractional Coverage of Rainfall in Climate Models

by

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ABSTRACT

The fraction of the grid-cell area covered by rainfall, μ, is a very important parameter in the descriptions of land-surface hydrology in climate models. A simple procedure for estimating this fraction is developed consistent with extensive observations of storm areas and rainfall volumes. It is often observed that storm area and rainfall volume are linearly related. This relation is utilized in rainfall measurement to compute rainfall volume from the radar observation of storm area. It is suggested to use the same relation to compute the storm area from the volume of rainfall simulated by a climate model. The new formula for computing μ describes the dependence of the fractional coverage of rainfall on the season of the year, the geographical region, rainfall volume, and spatial and temporal resolutions of the model.

The new procedure is included into a 3-D climate model which is used in simulations of the regional climate of the Amazon basin. The results of these simulations indicate reasonable success in modeling land-surface hydrology in a rain-forest environment.

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The support and love of every one in my family is greatly appreciated; they remain the source of motivation in all of my work.

This Thesis is dedicated to
my wife
Shahinaz

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ATI	Area Time Integral
BATS	Biosphere-Atmosphere Transfer Scheme
ECMWF	European Center for Medium-range Weather Forecast
f^*	infiltration capacity of the top soil layer
f_R	conditional PDF of rainfall rate
GCM	General Circulation Model
g_R	statistical distribution of rainfall rate
MM4	Meso-scale Model version 4
PDF	Probability Density Function
R	rainfall rate
s	saturation of top soil layer
ΔX	spatial resolution of a climate model
ΔT	temporal resolution of a climate model
μ	fractional coverage of rainfall
ρ	mean of conditional distribution of rainfall rate

CHAPTER 1
Introduction

1.1 Scope

The basic physics of some hydrologic processes, such as runoff production and rainfall interception, are non-linear. Hence, the spatially averaged response of the land-surface to a spatially distributed rainfall field is very different from the response of the land-surface to the spatial average of that same rainfall field. These basic notions imply that the spatial variability of rainfall is a very important factor in the description of these hydrologic processes over large areas.

Recent parameterizations of hydrologic processes in climate models, e.g., Warrilow et al. (1986), Shuttleworth(1988 b), Entekhabi and Eagleson (1989), Famiglietti and Wood (1990) and Eltahir and Bras(1991), include explicit representations of rainfall spatial variability at the sub-grid scale. In all these schemes rainfall is modeled as a random variable which varies in space covering a prescribed fraction of the grid-cell area, μ. It is not resolved how to specify the value of this fraction for the different regions, in the different seasons, and whether μ should vary during the life cycle of a single storm.

The problem addressed in this study is the estimation of the fraction of the grid-cell area which receives rainfall amounts greater than zero. This fraction is computed by a climate model and used as input to the land-surface hydrology scheme. Many recent studies have demonstrated that simulations of land-surface hydrology are sensitive to the value of this fraction.

The solution developed for that problem is based on extensive observations of convective storms. It utilizes the observed linear relation between storm area and the rainfall volume produced by the storm. The new procedure for computing the fractional coverage of rainfall is simple; and the data needed for application consists of rainfall records at a point. It is encouraging that this kind of data is available for most of the regions around the world.

1.2 Outline

This thesis is organized in five chapters. Chapter 2 is a literature review. Chapter 3 deals with the problem of estimation of the fractional coverage of rainfall in climate models. Chapter 4 describes implementation of the new procedure using an off-line land-surface hydrology scheme, and a 3-D climate model. Chapter 5 includes summary and conclusions of the study.

CHAPTER 2

Literature Review

2.1 Introduction

This review covers two topics: the sensitivity of land-surface hydrology to the fractional coverage of rainfall, and the observations of rainfall fields in convective storms. The review of the first topic motivates the problem addressed in this study, and the review of the second topic motivates the solution to the problem, which is presented in Chapter 3.

2.2 Surface Hydrology and the Fractional Coverage of Rainfall

The early versions of land-surface hydrology schemes assume that runoff is related to rainfall and soil moisture by simple linear relations. For example the Goddard Institute for Space Studies (GISS) GCM assumes that runoff , r, is given by

$$r = R_f \ R \qquad (2.1)$$

and

$$R_f = 0.5 \ s \qquad (2.2)$$

where R_f is the runoff coefficient, R is rainfall, s is the level of saturation of the top soil layer which is defined as the ratio of the water in that layer to the soil field capacity. In the Geophysical Fluid Dynamics Laboratory (GFDL) GCM, Manabe (1969), runoff is represented using a "bucket" model, which is described by,

$$r = R - E \qquad s = 1 \qquad (2.3)$$

$$r = 0 \qquad\qquad s < 1 \qquad (2.4)$$

where E is evaporation from land surfaces; it is modeled by similar linear relations. Further, it is usually assumed that all these hydrologic variables are constant over the grid-cell area which typically has a scale of about 100-1000 Kilometers. Observations indicate that rainfall which is the main forcing of land-surface hydrology exhibits large spatial variability over these scales.

Recently, a new generation of land-surface schemes have been developed, e.g. the Biosphere-Atmosphere Transfer Scheme (BATS) which is described by Dickinson et al. (1986) and Simple Biosphere (SiB) which is described by Sellers et al. (1986). These schemes are characterized by their emphasis on the details of the vertical structure of the canopy. They involve sophisticated treatment of the energy fluxes at the surface but rather simple description of surface runoff and other hydrologic processes. For example BATS assumes that the runoff coefficient, R_f, is given by

$$R_f = s^4 \qquad (2.5)$$

This relation is not based on any physical grounds. The new generation of schemes are also characterized by their neglect of the sub-grid scale spatial variability.

Since hydrologic processes such as runoff and rainfall interception are basically non-linear; aggregation of these processes over large areas should be done carefully. Spatial variability in the hydrologic variables and forcings are important factors when considering these processes over large areas. Many recent schemes attempt to represent the effects due to sub-grid scale spatial variability by using a statistical approach, e.g., Warrilow et al. (1986), Shuttleworth(1988 b), Entekhabi and Eagleson (1989), Famiglietti and Wood (1990) and Eltahir and Bras(1991). It is often assumed that rainfall, which is the main forcing of surface hydrology is exponentially distributed in space covering a fraction of the grid-cell area, μ. For example the scheme of Shuttlewoth (1988 b) describes runoff production and rainfall interception; it assumes that rainfall is spatially variable but soil moisture and canopy storage are constant in space. Based on these assumptions Shuttlewoth (1988 b) derives the following relation,

$$R_f = \exp\left(-\frac{\mu F}{R}\right) \tag{2.6}$$

where F is the maximum infiltration rate of the top soil layer. Similar expressions, which are functions of μ, are developed for other hydrologic processes e.g. infiltration and throughfall. This treatment of spatial variability is likely to result in realistic descriptions of important processes such as runoff production and rainfall interception.

Some important questions which follow from the developments described above are : how to specify the value of μ ?, should μ be kept constant ? or, should it vary in space and time?. Entekhabi and Eagleson (1989) suggest that μ may be taken as 0.6 for convective rainfall, Warrilow et al. (1986) indicate that μ may be taken as 0.3 for convective rainfall. Both of these studies suggest that μ should be taken as a constant; in this study we will question the validity of this assumption.

Several recent studies focus on the sensitivity of large-scale surface hydrology to the value of μ. Pitman et al. (1990) study the sensitivity of runoff and evaporation to the value of μ. They use an off-line model of BATS with the surface hydrology modeled according to the scheme of Shuttleworth (1988 b). Figure 2.1 corresponds to Figure 1 of Pitman et al. (1990); it shows the results of their sensitivity experiments. It is evident that the specification of the value of μ has significant effects on the simulation of runoff and evaporation.

Johnson et al. (1991) use a 3-D climate model to study the sensitivity of land-surface hydrology to the value of μ; the scheme used in this study is that of Entekhabi and Eagleson(1989). Some of their results are shown in Table 2.1. They performed a control run to simulate the climate of the Earth for three years and with μ equal to 0.6; then μ is reduced from 0.6 to 0.15 and the simulation is repeated for another three years. As a result the runoff coefficient for South America increased from 0.13 to 0.44 and runoff coefficient for Africa increased from 0.10 to 0.45.

Similar significant sensitivity of land surface hydrology is demonstrated by Thomas and Henderson-Sellers (1991). They compared the results of two land-surface schemes and concluded that the modeling of land surface hydrology is sensitive to the specified value of the fraction μ.

Based on the results of all these sensitivity studies we conclude that the choice of μ affects significantly the representation of

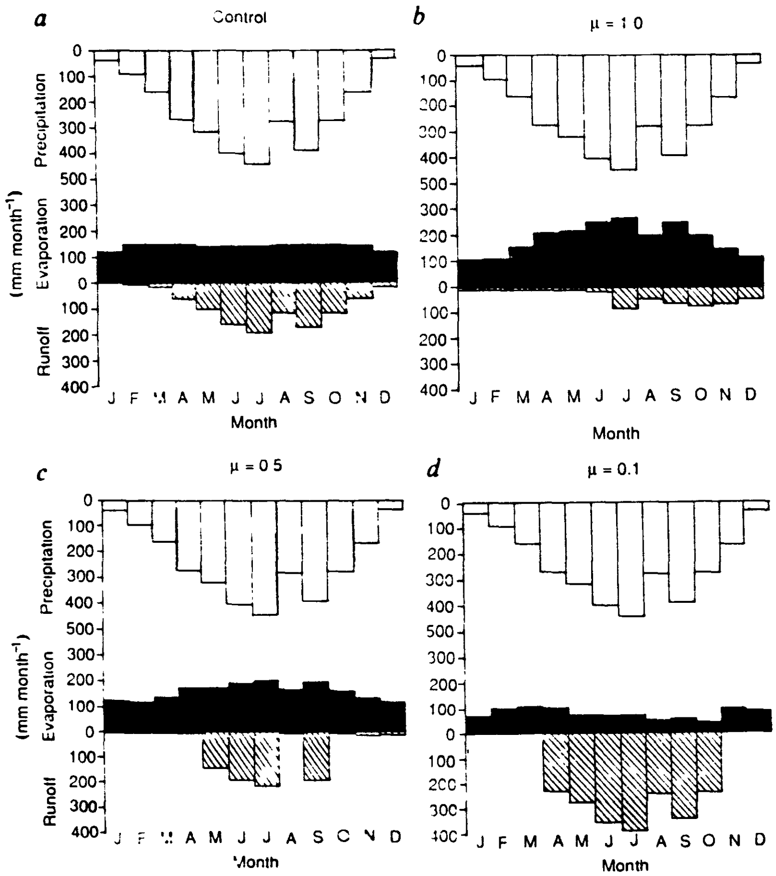

Figure 2.1 Sensitivity of land-surface hydrology to the fractional coverage of rainfall, from Pitman et al.(1990)

variable	South America	Africa
	μ = 0.6	
precipitation (mm/year)	1361	911
evaporation (mm/year)	1172	828
runoff (mm/year)	185	89
	μ = 0.15	
precipitation (mm/year)	1178	831
evaporation (mm/year)	682	478
runoff (mm/year)	516	373

Table 2.1 : Sensitivity of land-surface hydrology to the fractional coverage of rainfall,from Johnson et al. (1991)

of land-surface hydrology in climate models. There is a need for a new procedure for computing μ; this new procedure should be consistent with observations of rainfall fields. The next section presents some of these observations.

2.2 Observations of Rainfall Fields in Convective Storms

This section reviews some recent observational studies of rainfall areas and the corresponding volumes of rainfall produced by a storm. These observations will form the basis for the new procedure which will be developed in Chapter 3 for estimating the fractional coverage of rainfall, μ.

The earliest observations of the relation between the volume of rainfall produced by a convective storm and its size are reported by Byres (1948). Doneaud et al. (1981) were the first to suggest that the significant correlation between rainfall volume produced by a storm and the time integral of the area covered by rainfall can be utilized in rainfall measurement. They suggested to the use of radar observations of storm area to infer rainfall volume.

Doneaud et al. (1984) discuss the new method for measuring of rainfall over large areas; they verify the area-volume relation by estimating the correlation from one set of data and then validating the relation by using another independent set of data. They show that the area-volume method for measuring of rainfall over large areas is a fairly accurate technique. The time integral of the area covered by rainfall is referred to as Area-Time Integral (ATI). Figure 2.2 shows observations of ATI and rainfall volume; it shows significant correlation between the two quantities. It is suggested that this

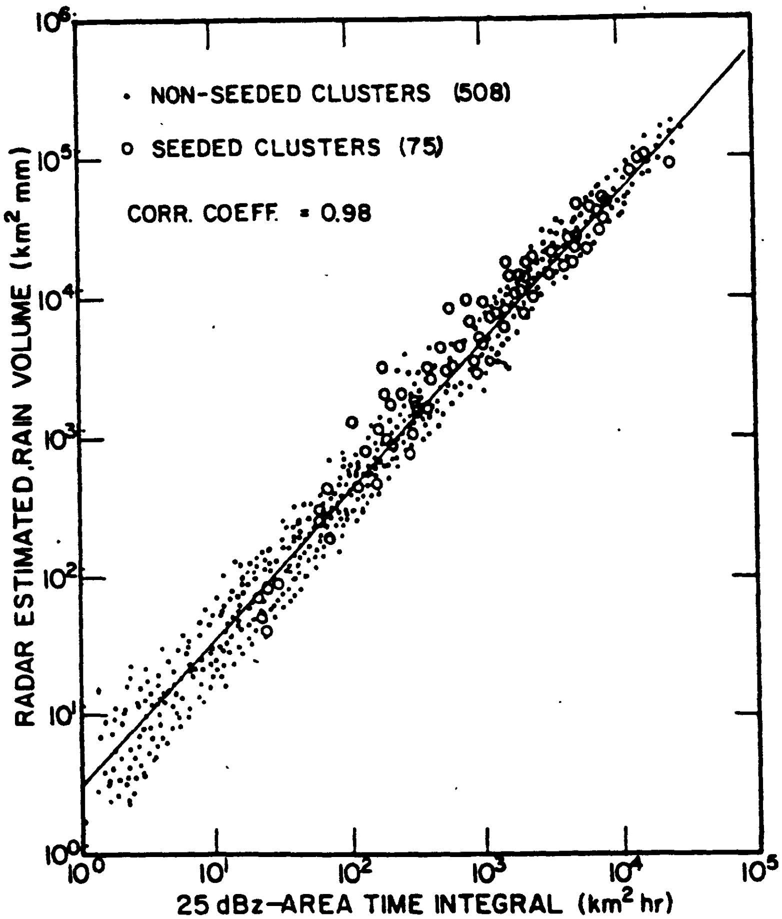

Figure 2.2 Observations of storm area and rainfall volume from Doneaud et al. (1984)

correlation can be used for measuring rainfall volume from radar observations of rain area.

Atlas et al. (1988) postulate that the existence of a well behaved Probability Density Function (PDF) of rainfall rate may explain the observed correlation between storm areas and the corresponding rainfall volumes. They show theoretically that the correlation between storm area and the corresponding rainfall volume depends on the threshold chosen for defining storm area. This dependence is observed in the data from previous studies e.g. Doneaud et al. (1984).

Lopez et al. (1989) study the correlation between storm area and the corresponding rainfall volume produced by the storm. They use data from a dense network of rain gages in Florida. Figure 2.3 shows some of their data. This study provides verification of the area-volume relation using measurements of surface rainfall which represent ground truth. Similar rain gage observations are presented by Short et al. (1989) using data from Darwin, Australia.

In a similar study, Kedem et al. (1990) analyzed rainfall data from GATE experiments; they show similar correlation as those reported by the previous studies; the data is shown in Figure 2.4. It is clear from their plots that the area-volume correlation depends on the threshold chosen for defining the raining area. The area-volume correlation increases with the value of the threshold. Hence, for the purposes of rainfall measurement the use of a large threshold value results in higher accuracy of the volume estimates. But even when the threshold chosen is zero the area-volume correlation is significant

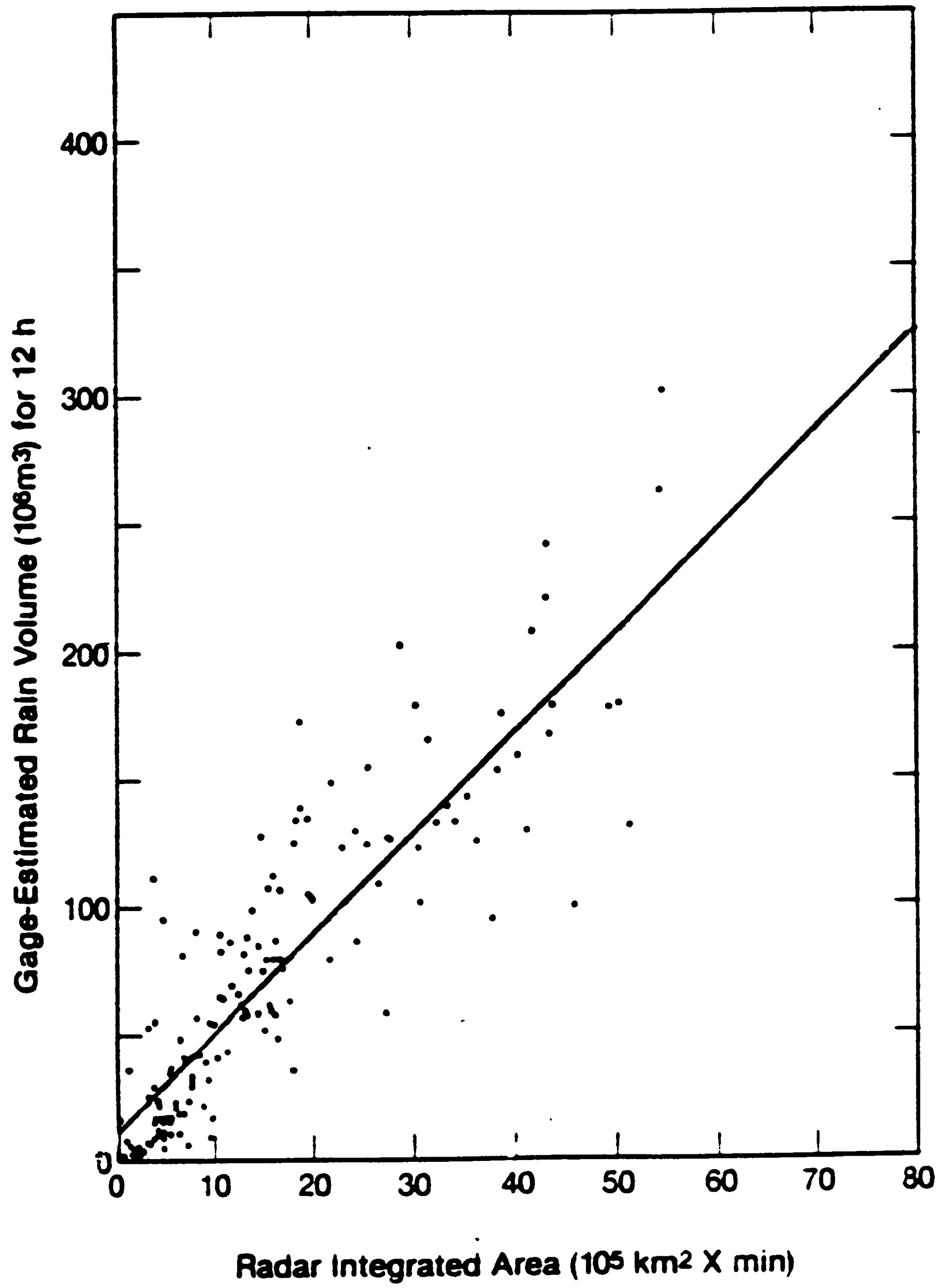

Figure 2.3 Observations of storm area and rainfall volume from Lopez et al. (1989)

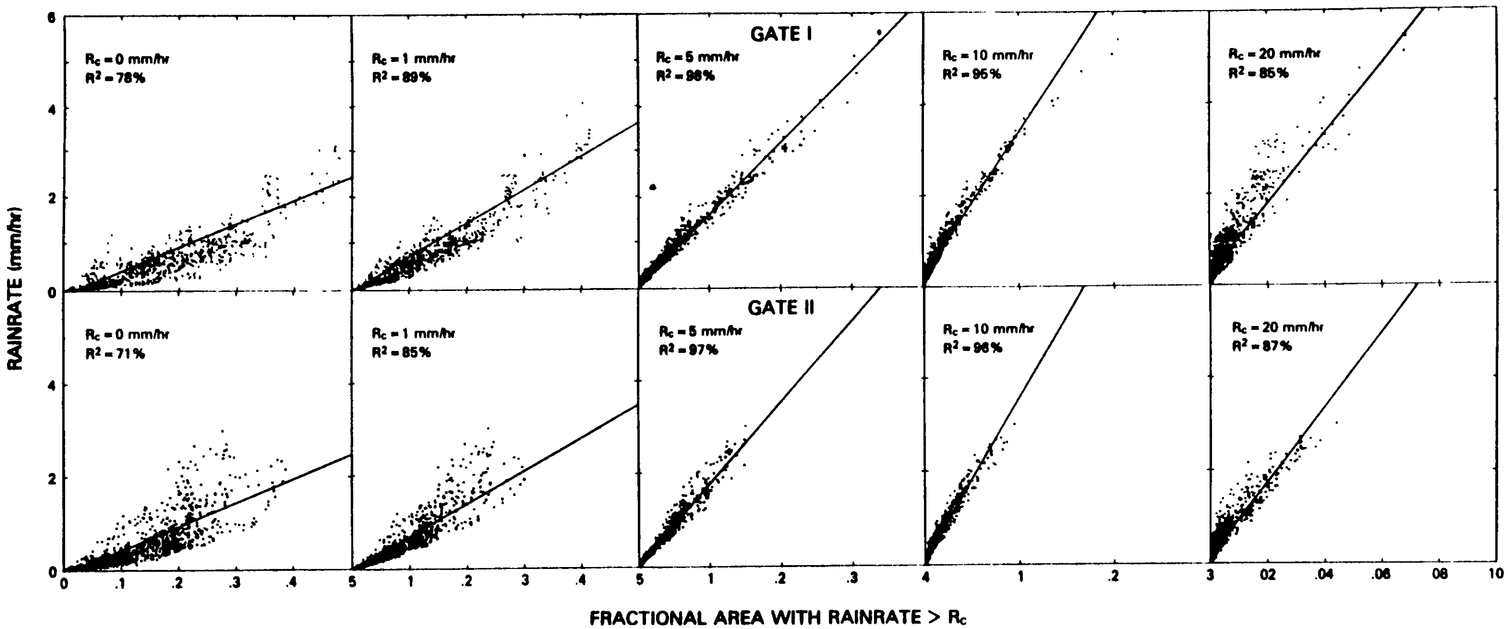

Figure 2.4 Observations of storm area and area -average rainfall rate from Kedem et al. (1990)

and explains a large percentage of the total variance in the data. In this study we are interested in the area-volume correlation when the threshold is zero; since by definition the fractional coverage of rainfall refers to the fraction of the area which receives rainfall greater than the threshold of zero.

In the same study, Kedem et al. (1990) present theoretical arguments, similar to those of Atlas et al. (1988), to explain these empirical observations. They suggest that the existence of a mixed PDF for the rainfall rate process may explain the correlation between rainfall area and the corresponding rain volume. Based on this assumption they develop the following relation,

$$E(R) = I(R > \tau)\ \beta(\tau) \tag{2.7}$$

where E denotes the expected value (or the spatial average); $I(R > \tau)$ is the fraction of the area which receives rainfall greater than the threshold τ; β is a constant which depends on the value of τ. In Chapter 3 we will develop in some detail a similar relation for the case of $\tau = 0$.

The observations of a significant correlation between storm area and the corresponding volume of rainfall produced by a storm will form the basis for the new procedure for computing μ, which will be introduced in the next chapter.

CHAPTER 3

Estimation of the Fractional Coverage of Rainfall in Climate Models

3.1 Introduction

A simple and consistent procedure for computing the fractional coverage of rainfall, μ, in convective storms is introduced in this chapter. It is based on extensive observations of convective storms. These observations are reviewed in Chapter 2.

The theoretical basis which explains the empirical observations is discussed in section 3.2. The new procedure for computing μ in convective storms is presented in section 3.3. This procedure requires the estimation of the conditional mean rainfall rate, ρ. The estimation of ρ is discussed in section 3.4. In Section 3.5, the new procedure is applied in estimation of μ in the Amazon region using a time series of area average rainfall rate, which is simulated by a climate model. The issue of estimation of the fractional coverage of rainfall in frontal storms is discussed in section 3.6. Some conclusions are presented in Section 3.7.

3.2 Theoretical Basis

The rainfall rate process can be described statistically by the following mixed distribution

$$g_R = (1 - \mu) \, . \, \delta(R - 0) + \mu \, . \, f_R \tag{3.1}$$

where δ is the Dirac delta function and f_R is the conditional Probability Density Function (PDF) of the rainfall rate, R, given that R is greater than zero. Since no assumptions are made about the conditional PDF, f_R, the description in Equation 3.1 is general and always valid.

For the rainfall rate process which has a *unique* conditional PDF, f_R, the total volume of rainfall is linearly related to the area of the storm which receives rainfall rate above a certain threshold. When the value of this threshold is zero the theory predicts the observed linear relation between the rainfall volume and storm area; this linear relation will be developed in the next section. A *unique* PDF means that f_R does not vary in time i.e. whenever it rains, the distribution of rainfall rate within the raining area is a realization from the same statistical distribution.

3.3 A New Procedure for Estimating μ in Convective Storms

The observed relation between storm area and rainfall volume is often used in estimating rainfall volume from the radar measurement of storm area. It is suggested that the same relation can be used to infer the storm area from the rainfall volume simulated by a climate model.

It is assumed that the rainfall rate process is described by a *unique* conditional PDF, f_R. The mean of this conditional distribution is denoted by ρ. It has a seasonal and geographical variability. The expected value of the rainfall rate process over the grid-cell area of a climate model is given by,

$$E(R) = \int_{R=0}^{\infty} R \,.\, g_R \,.\, dR$$

$$= (1-\mu)\,.0 + \mu \,. \int_{R=0^+}^{\infty} R \,.\, f_R \,.\, dR = \mu \,.\, \rho \qquad (3.2),$$

implying that

$$\mu = \frac{E(R)}{\rho} \tag{3.3}$$

According to Equation 3.3, the slope in the regression between E(R) and μ should be equal to ρ.

Figure 2.4 is a plot of E(R) versus μ which are measured during the GATE experiment. The slope of the regression line in Figure 2.4 is about 4.4 mm per hour; this value is very close to the observed conditional mean rainfall rate. Hence Equation 3.3 is consistent with the observations of Figure 2.4.

A climate model computes E(R) from the following relation,

$$E(R) = \frac{V}{((\Delta X)^2 . \Delta T)} \tag{3.4}$$

where V is the volume of rainfall simulated by the climate model within a grid cell area. ΔX and ΔT are the spatial and temporal resolutions of the model respectively. Substituting for E(R) from Equation 3.4 into Equation 3.3 results in,

$$\mu = \frac{V}{((\Delta X)^2 . \Delta T . \rho)}, \quad \mu \leq 1.0 \tag{3.5}$$

μ is, by definition, restricted to the range of values between zero and one.

Equation 3.5 is the formula for computing the fractional coverage of rainfall. It incorporates most of the important factors controlling μ. The regional climate is represented by the conditional mean of the rainfall rate process. The spatial and temporal resolutions of the climate model appear explicitly in Equation 3.5. The volume of rainfall simulated by the model during a time period ΔT varies between the different storms and within the life cycle of the same storm. According to the linear relation in Equation 3.5 μ should vary similarly. This variability is consistent with the observations of convective storms; the raining area initially increases during the early development of a convective storm and then slowly decreases while the storm dissipates. The seasonal variability of ρ is reflected in the estimate of μ through Equation 3.5. Figure 3.1 illustrates with a simple example the dependence of μ on the conditional mean of the rainfall rate process, the rainfall volume, the model spatial resolution and the model temporal resolution.

3.4 Estimation of the Mean Rainfall Rate ρ

The above procedure requires only one parameter which is the conditional mean rainfall rate, ρ. It is assumed that the conditional distribution of the rainfall rate process is ergodic. Hence ρ can be estimated by the climatological mean rainfall rate at a single location. This estimate is consistent with the assumption of a unique conditional PDF, f_R. According to this assumption every snapshot in every storm is a realization of the same statistical distribution. Invoking the ergodicity assumption, the mean of the conditional distribution can be estimated by the mean of the rainfall

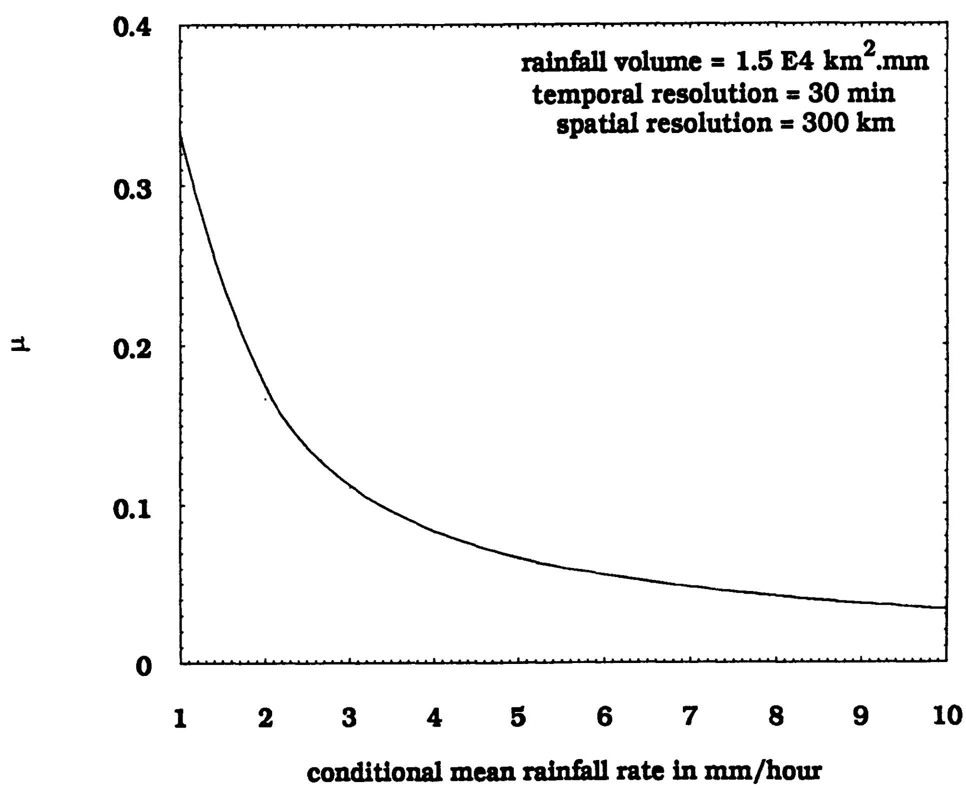

Figure 3.1 (a) Dependence of μ on the conditional mean rainfall rate

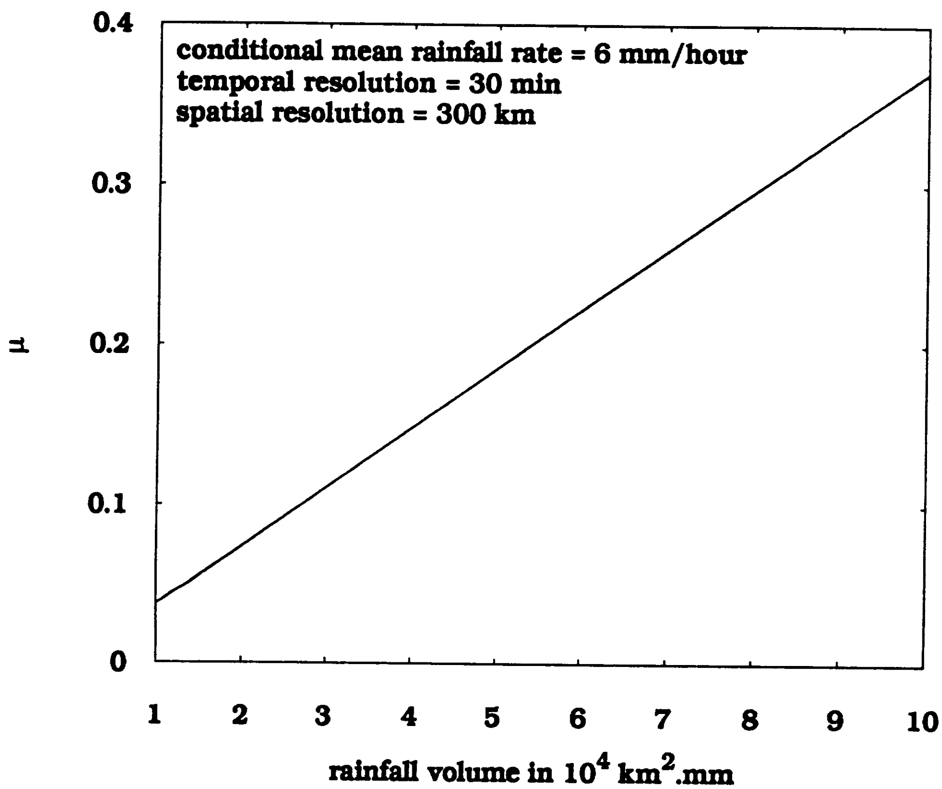

Figure 3.1 (b) Dependence of μ on the rainfall volume

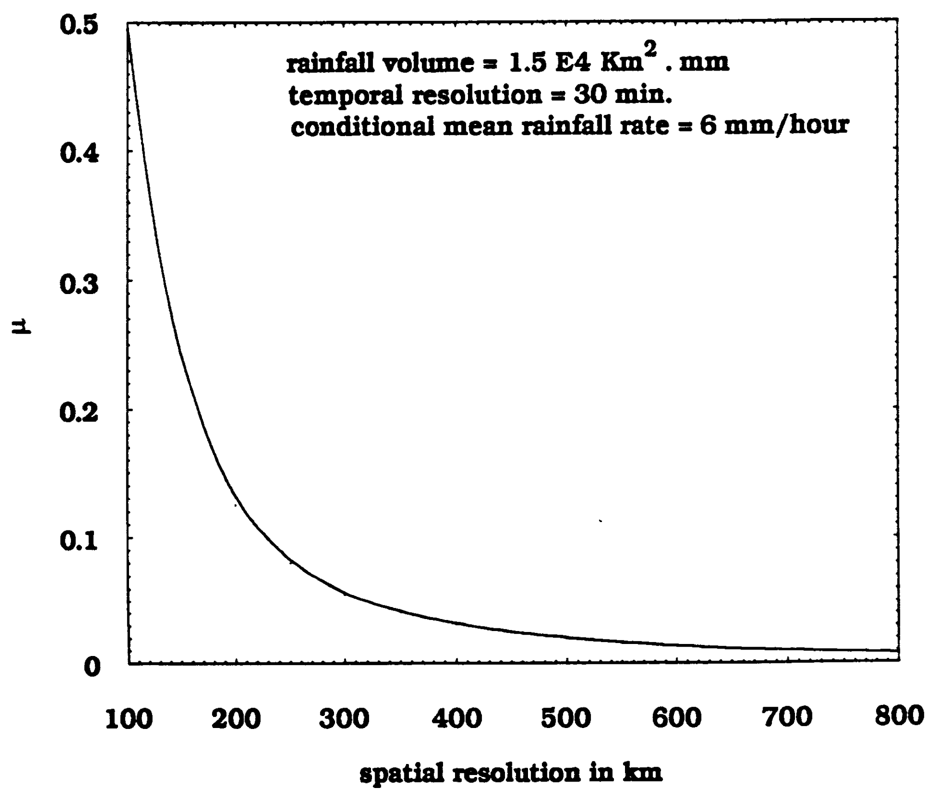

Figure 3.1 (c) Dependence of μ on the spatial resolution

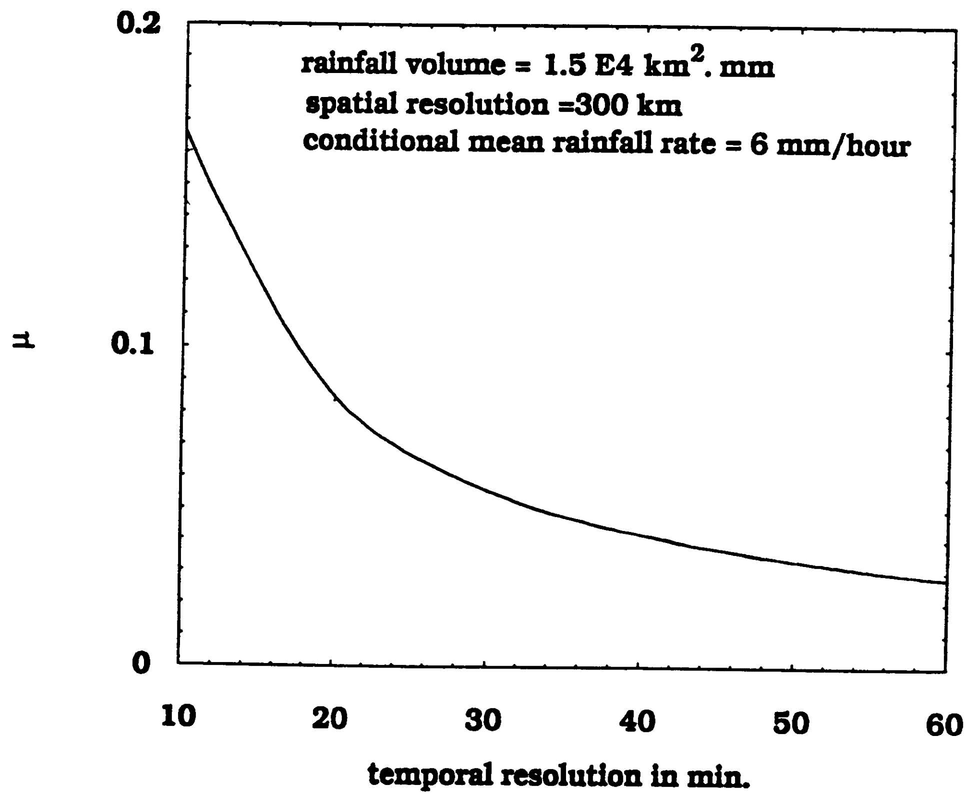

Figure 3.1 (d) Dependence of μ on the temporal resolution

rate process at a point when computed from a sufficiently large number of these realizations.

ρ is estimated from the rainfall records at a single location and for each month, M. It is estimated by

$$\rho\,(M) = \frac{\sum_{I=1}^{N} r\,(I, M)}{\sum_{I=1}^{N} t\,(I, M)} \tag{3.6}$$

where N is the total number of years with record of rainfall amounts. r is the monthly total rainfall amount. t is the monthly total duration of storms.

Table 3.1 shows estimates of ρ from different regions and for the different months of the year. The estimated values for ρ are larger in the tropics compared to midlatitudes and at each location those estimates are larger in summer compared to winter. These observations are consistent with the differences in the rainfall producing mechanisms. The information in Table 3.1 is sufficient for modeling μ at those locations.

3.5 Estimation of μ over the Amazon basin

The procedure introduced in the previous section is used to compute the fractional coverage of rainfall over the Amazon region. A time series of rainfall, averaged over a grid-cell area, is simulated by the climate model of the National Center for Atmospheric Research (NCAR) (CCM1). The location of the grid-cell corresponds to the Amazon region. The spatial resolution of the model is (approximately) 4.4 degrees in latitude by 7.5 degrees in

month	Wau (7N,28E)	Manaus (3S,60W)	Florence (44N,11E)	Boston (42N,71W)	Tucson (32N,111W)
Jan	7.3	5.5	1.1	1.2	1.1
Feb	6.1	6.1	1.3	1.3	1.0
Mar	6.0	6.5	1.4	1.3	1.0
Apr	9.2	6.5	1.3	1.3	1.2
May	9.7	6.1	1.7	1.4	1.0
Jun	11.1	3.9	2.1	1.7	1.6
Jul	10.5	4.3	3.4	2.0	2.3
Aug	10.9	3.8	3.3	2.2	2.4
Sep	10.0	3.8	2.8	1.8	2.5
Oct	9.6	5.6	2.1	1.6	1.7
Nov	6.0	5.3	2.0	1.6	1.2
Dec	6.6	6.6	1.3	1.3	1.2

Table 3.1 : Climatological mean rainfall rate (in mm/hour) from different regions of the world and for different months of the year.

longitude. The temporal resolution of the series is one half hour and the period covered is the first 300 days of a typical year. The same rainfall series is used in the study by O'Neill and Dickinson (1991).

Figure 3.2 shows the time series of μ which is obtained by applying Equation 3.3 to the rainfall series from the Amazon region. The conditional mean rainfall rate is estimated by the climatological mean rainfall rate at Manaus (see Table 3.1). The variability in μ is quite significant; μ is as variable as the areal average of rainfall. Parametrizations of surface hydrology in climate models often assume that μ is a constant. The significant variability in Figure 3.2 raises many questions about the accuracy of the current descriptions of land-surface hydrology in climate models. This issue will be discussed further in Chapter 4.

3.6 Estimation of μ for Frontal Storms

The observations described in the Chapter 2 are for convective storms. Hence, the procedure introduced in this Chapter is more accurate for modeling convective storms. Under those conditions, it should be very rare that μ approaches a value of 1. During the GATE experiment, which was conducted over the tropical ocean, the maximum observed value for μ is about 0.5 (see Figure 2.4). The observed area has a diameter of 400 Km. According to Equation 3.5 the value of μ is not allowed to exceed 1.

For non-convective storms associated with warm frontal systems the fractional coverage of rainfall approaches a value of 1

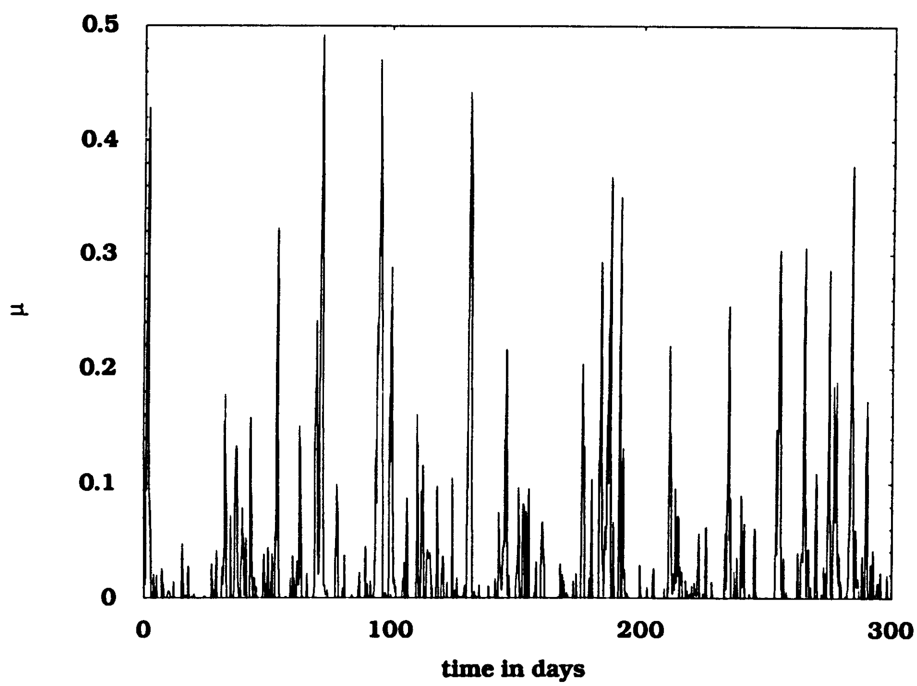

Figure 3.2 (a) Fractional coverage of rainfall in the Amazon region for the first 300 days of a typical year

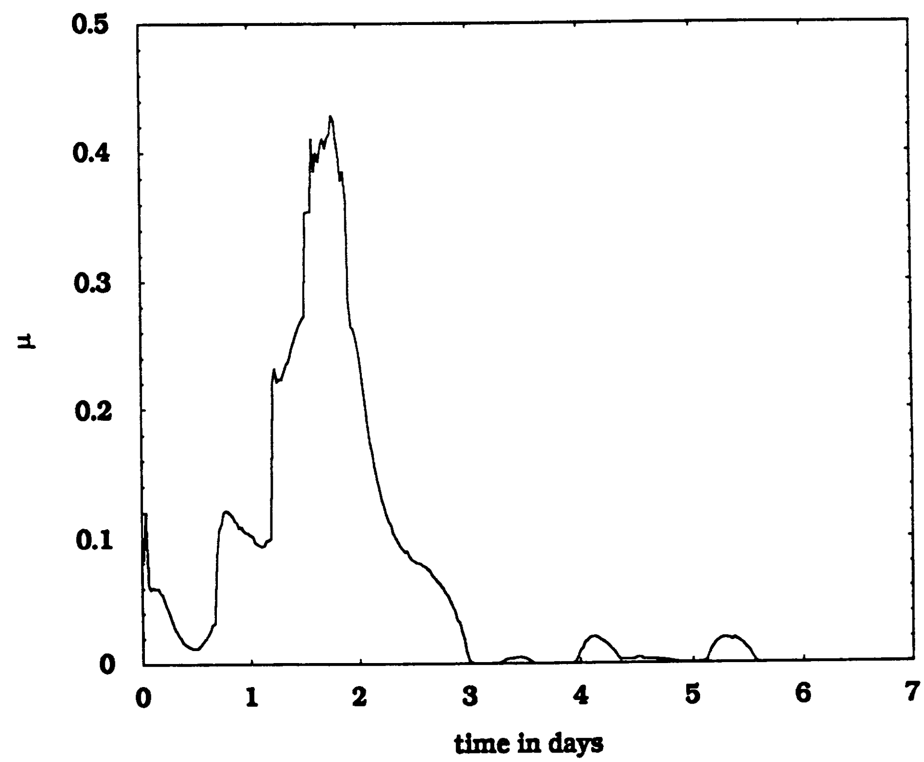

Figure 3.2 (b) Fractional coverage of rainfall in the Amazon region for the first week of the year

more often than for convective storms(depending on the relative size of the grid-cell compared to the typical area of the storm). The use of Equation 3.5 for estimation of the fractional coverage of rainfall in these storms may result in estimates of μ which exceed 1. That corresponds to the possibility of occurrence of a frontal storm covering the total grid cell area and with an average rainfall rate which exceeds the conditional mean rainfall rate. The probability of occurence for such an event is very small. Under those conditions μ should be reset to a value of 1.

The formula for computing μ is developed primarily for convective storms. Hence, it is less accurate when used for estimation of μ in non-convective storms. The new formula provides a better approximation than the assumption that μ is 1 for these kind of storms. This is particularly true when the model resolution is large compared to the typical scale of a frontal storm(few hundreds of kilometers). The assumption that μ is 1 for frontal storms is often made independent of the model spatial resolution and the rainfall volume.

3.7 Conclusions

A new procedure is introduced in this Chapter for estimation of the fractional coverage of rainfall in convective storms. The new procedure is easy to apply and needs data on rainfall at only one point.

In the next chapter the new procedure will be included into a land-surface scheme. It will be tested by comparing the results of simulations of land-surface hydrology with observations.

CHAPTER 4
Applications of the New Procedure for Estimation of the Fractional Coverage of Rainfall

4.1 Introduction

This chapter describes the application of the new procedure for estimation of the fractional coverage of rainfall. The new procedure is implemented as part of a modified version of the Biosphere-Atmosphere Transfer Scheme (BATS) , Dickinson et al. (1986). An off-line model of BATS is used in studying the sensitivity of land-surface hydrology to the value of μ. The model is used in simulating typical rainforest conditions. The new procedure is compared to the assumption of a constant μ, which is usually made in current climate models.

The modified version of BATS is included into a 3-D climate model which is used for simulating climate over the Amazon basin. These simulations represent a more rigorous test of the new procedure. It is found that the modified version of BATS, which utilizes the new procedure for estimation of μ, is capable of simulating rainfall interception and surface runoff with reasonable accuracy.

In the next section BATS is described briefly. Section 4.3 describes the off-line implementation of the new procedure for estimation of μ. Section 4.4 presents the application of the new procedure using a 3-D climate model. Some concluding discussion is presented in section 4.5.

4.2 The Biosphere-Atmosphere Transfer Scheme (BATS)

This section describes the original version of BATS and the modifications which are introduced to improve the surface hydrology. These modifications focus on surface runoff and rainfall interception.

4.2.1 The original BATS

BATS describes a land-surface which consists of a vegetation layer, a surface soil layer and a deep soil layer (root zone). A seasonally dependent fraction of the grid-cell area is covered by vegetation; the remaining fraction is assumed covered by bare soil. Soil temperature is predicted using a prognostic equation which describes the force-restore method of Deardorff (1972). The temperatures of the canopy and that of the air within the canopy are determined using diagnostic equations which describe conservation of energy and conservation of water mass at the land-surface. The energy balance equation includes radiative, latent and sensible heat fluxes.

The land-surface hydrology scheme consists of prognostic water balance equations which predict the water content of the surface layer and of the root zone. The components of this water balance are rainfall, throughfall, infiltration, evapotranspiration, surface runoff, groundwater runoff, infiltration below root zone and diffusive exchange of water between the two layers. The soil water movement formulation is parameterized to fit the results of detailed simulations using a high resolution soil model. It is assumed that the coefficient of surface runoff is equivalent to the fourth power of saturation in the surface soil layer. Soil saturation, s, is defined as the ratio of the water depth in the soil layer to the maximum capacity of the layer. The treatment of interception in BATS is very simple; whenever canopy storage exceeds canopy capacity the storage is restored back to the value of canopy capacity.

The fluxes of latent heat, sensible heat and momentum are calculated using the similarity theory approach. The drag coefficients are calculated based on surface roughness and atmospheric stability of the surface layer. For neutral and stable conditions turbulent vertical transport is modeled using an eddy diffusion formulation, for unstable conditions transport is modeled by a dry convective adjustment scheme. BATS is described in detail by Dickinson et al. (1986).

4.2.2 Modifications of BATS

BATS includes a detailed description of the vertical structure of the surface layer; in contrast the scheme assumes constant surface properties and uniform forcing in the horizontal. Sub-grid scale spatial variability in rainfall, canopy storage and soil moisture play a significant role in some important processes taking place in a rainforest environment. The partition of rainfall into throughfall and interception loss, and the subsequent partition of throughfall into infiltration and surface runoff are examples of these processes which are sensitive to the effects of sub-grid scale spatial variability. This sensitivity is basically due to the non-linearity involved in the interception and runoff processes. The BATS treatment of interception and runoff are modified in this study to account for the effects of the sub-grid scale spatial variability. These modifications are important because they balance the emphasis on *vertical* details by accounting for some of the important effects resulting from spatial variability in the *horizontal*. In the following the new descriptions of surface runoff and rainfall interception are presented.

Surface Runoff

A runoff scheme similar to that of Entekhabi and Eagleson (1989) is developed for modeling surface runoff. A slightly different form of the infiltration function is used; it is given by

$$f^* = \alpha\,(1 - s) \tag{4.1}$$

where f^* is infiltration capacity of the soil; α is infiltration capacity of the soil when completely dry; and s is saturation level of the surface layer(taken as the top 10 centimeters of the soil) . This infiltration function is a simplified form of the infiltration function used by Entekhabi and Eagleson (1989).

The spatial variability in rainfall and soil moisture are modeled using a statistical approach. It is assumed that rainfall is exponentially distributed in space according to

$$f_R = (1 - \mu)\,\delta\,(R - 0) + \frac{\mu^2}{E(R)}\, e^{-\frac{\mu R}{E(R)}} \tag{4.2}$$

Soil saturation of the top soil layer is assumed distributed in space according to

$$f_s = \frac{1}{E(s)}\, e^{-\frac{s}{E(s)}} \tag{4.3}$$

E(s) is the spatial average of soil saturation.

Runoff occurs at a point where rainfall intensity exceeds the infiltration capacity of the soil (Hortonian runoff) or where rainfall occurs on a saturated soil (Dunne runoff). The areally averaged runoff is then given by

$$r = \int_{s=0}^{1}\int_{R=f^*}^{\infty} (R - f^*) f_R \, dR \, f_s \, ds + \int_{s=1}^{\infty}\int_{R=0}^{\infty} R \, f_R \, dR \, f_s \, ds$$

$$= r_H + r_D \qquad (4.4)$$

where r_H is Hortonian runoff and r_D is Dunne runoff. Evaluating the above integrals results in

$$r = \frac{E(R)}{C\,E(s)} e^{-\frac{\mu\alpha}{E(R)}} (e^C - 1) + E(R)\, e^{-\frac{1}{E(s)}} \qquad (4.5)$$

where C is given by

$$C = \frac{\mu\alpha}{E(R)} - \frac{1}{E(s)}$$

The new formula for computing the fractional coverage of rainfall implies that the runoff coefficient , R_f , is given by

$$R_f = \frac{r}{E(R)} = \frac{e^{-\left(\frac{\alpha}{\rho}\right)} (e^C - 1)}{E(s)\ C} + e^{-\frac{1}{E(s)}} \qquad (4.6)$$

where C is given by

$$C = \frac{\alpha}{\rho} - \frac{1}{E(s)}$$

The runoff coefficient is computed based on this new formula instead of the empirical formula in the original BATS.

Rainfall Interception

A new interception scheme, Eltahir (1993), which accounts for the effects of spatial variability in rainfall and canopy storage is included into the BATS. The scheme is based on the Rutter model of interception, Rutter et al. (1971), and statistical description of the sub-grid scale spatial variability of canopy storage and rainfall. The details of this scheme are described in Appendix 4.1.

4.3 Off-line Application of the New Procedure

The off-line model of BATS is driven by the following forcings: solar radiation, above canopy temperature, above canopy humidity and a time series of surface rainfall. The forcings are designed to simulate a typical rainforest environment. The forcings are described in Table 4.1.

The rainfall series is generated using the stochastic model of Rodriguez-Iturbe and Eagleson (1987). The model simulates the rainfall rate process in space and time for each storm. The storm arrival process is described by a non-homogeneous Poisson process which favors occurrence of storms in the afternoons. This is consistent with the recent observations of Lloyd(1990) in the Amazon basin. The parameters of the model are selected to simulate convective storms which are characteristic of the rainforest environment. The rainfall simulated by the model is averaged in space over an area of ten thousands squared kilometers. The total duration of the simulation is two months.

The parameters of vegetation and soil are specified according to

maximum solar radiation at the surface	890 W/m^2
average above canopy temperature	300 K
daily range of above canopy temperature	6 K
relative humidity above the canopy	80%.
mean of the rainfall series	220 mm per month

Table 4.1 : Description of the model forcings

those of rainforest conditions from tables 2 and 3 of Dickinson, et al. (1986). The Rutter model parameters are specified according to those calibrated for an Amazonian rainforest and described in Shuttleworth (1988 a).

In the following the results of the simulations are presented. The new procedure for computing μ is compared to the alternative of taking μ as a constant. The results are presented for two hydrologic processes : surface runoff and rainfall interception. In these simulations the mean rain rate is taken as 5.5 mm/hour which is typical for Manaus in summer, see Table 3.1.

Figure 4.1 shows the average runoff coefficient computed for the two months of simulations. It is estimated that the runoff coefficient is about 0.53 . This ratio is close to the climatological runoff coefficient for the basin of about 0.44. The same figure shows the runoff coefficient resulting from assuming that the fractional coverage is constant. Figure 4.2 shows a similar comparison for the Hortonian runoff coefficient, which is explicitly related to the fractional coverage of rainfall . Figure 4.3 shows the sensitivity of Dunne runoff coefficient to the fractional coverage of rainfall. A smaller fractional coverage results in more total runoff and dryer soil conditions. The latter effect results in less Dunne runoff.

The results for rainfall interception are summarized in Figure 4.4.; it shows the dependence of interception loss on wind speed and the fractional coverage of rainfall. Wind speed is a surrogate for potential evaporation. Interception loss which is simulated using the new procedure for estimation of μ is compared to those obtained by

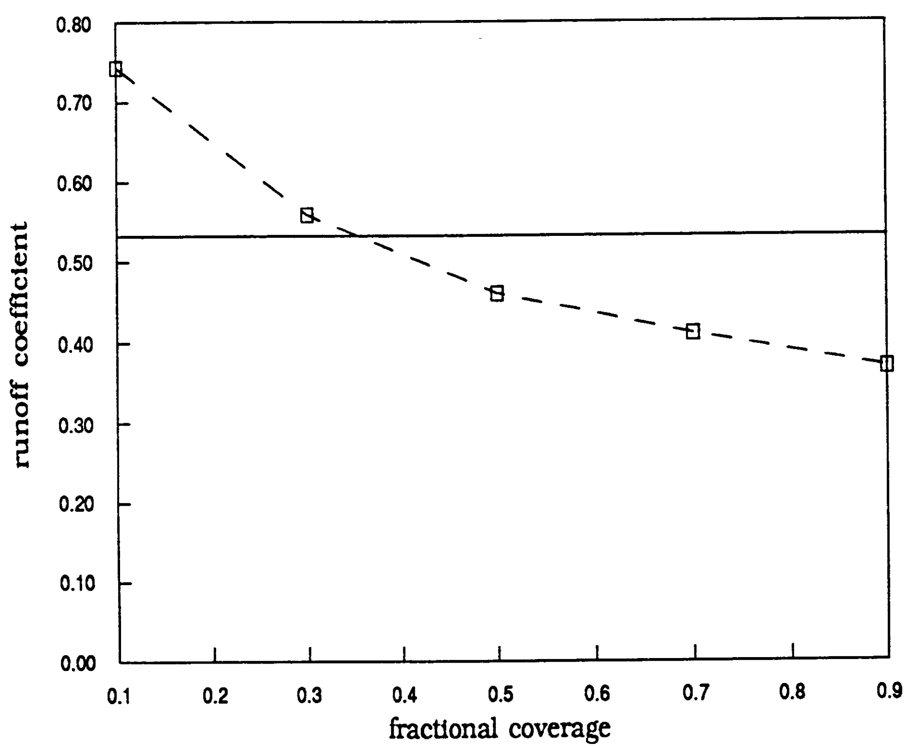

Figure 4.1 Sensitivity of total runoff coefficient to the fractional coverage of rainfall.(solid line: variable μ, dashed line:constant μ)

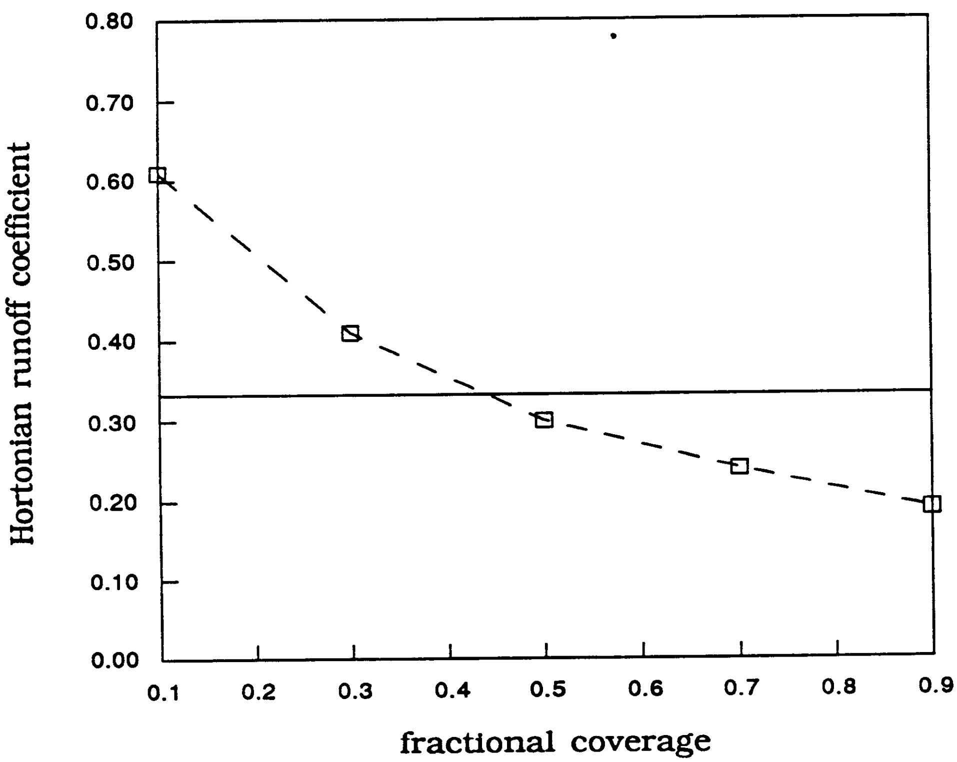

Figure 4.2 Sensitivity of Hortonian runoff coefficient to the fractional coverage of rainfall.(solid line: variable μ, dashed line:constant μ)

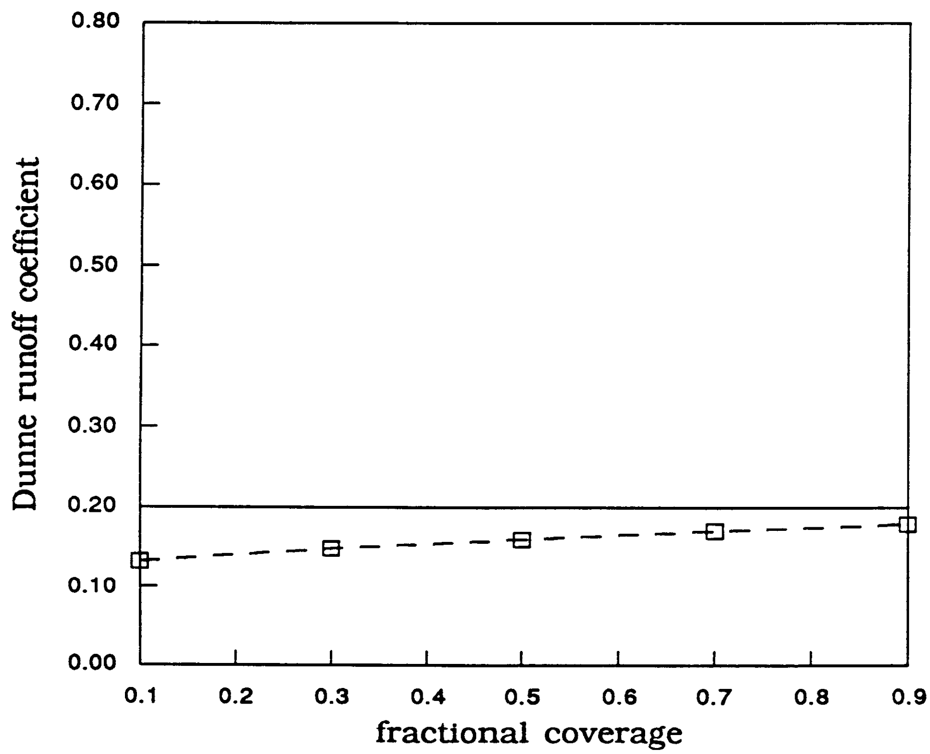

Figure 4.3 Sensitivity of Dunne runoff coefficient to the fractional coverage of rainfall.(solid line: variable μ, dashed line:constant μ)

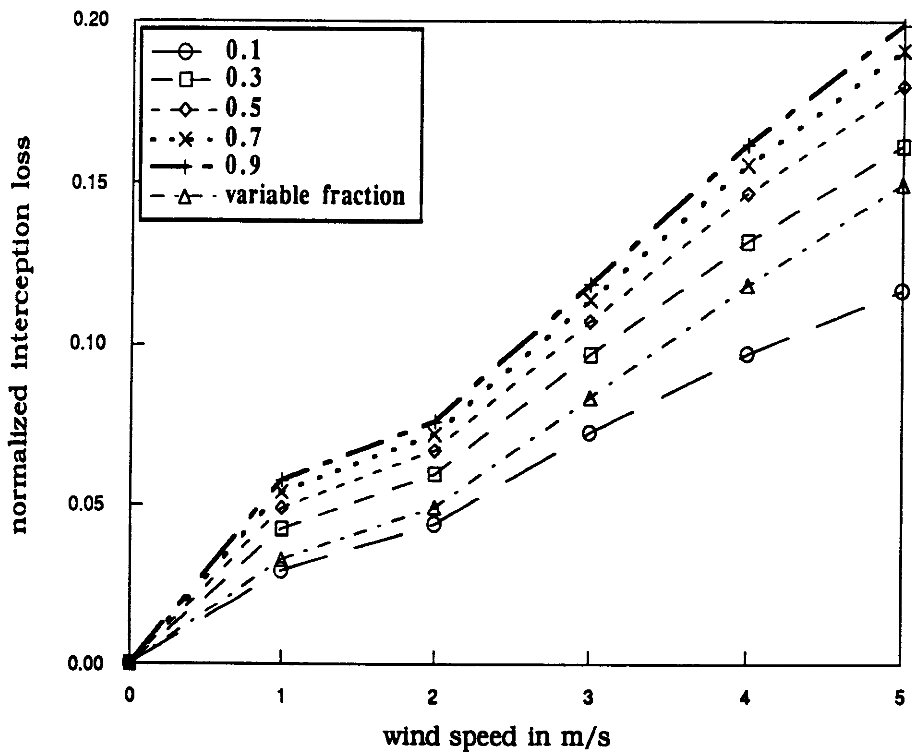

Figure 4.4 Sensitivity of interception ratio to the fractional coverage of rainfall.

assuming that the fractional coverage is a constant. In both cases the new interception scheme is used.

4.4 Implementation of the New Procedure in a 3-D Climate Model

This section describes the implementation of the modified version of BATS , which is described in section 4.2, in a 3-D climate model. The climate model is described in the section 4.4.1. The design of the experiments is described in section 4.4.2. The results are presented in section 4.4.3.

4.4.1 Description of the Climate Model

The original version of the model is known as the Pennsylvania State University/National Center for Atmospheric Research (PSU/NCAR) model. It is also referred to as the Meso-scale Model version 4 (MM4); it was originally developed for meso-scale meteorological studies. The climate model which is used in this study is an augmented version of MM4; it has been modified to suit climate studies.

The original MM4 is a compressible and hydrostatic model which solves the primitive equations in a terrain varying vertical coordinate. The model is driven by boundary conditions and solar radiation. It includes the bulk boundary layer parameterization of Deardorff (1972) and the cumulus parametrization of Anthes (1977). MM4 includes a simple long-wave radiative cooling scheme. The basic structure of the MM4 model is described by Anthes et al. (1987). The

MM4 model has been used successfully in a large number of meteorological studies.

For climate studies it is necessary to include into the model accurate descriptions of radiative transfer in the atmosphere and near the surface. The augmented version of MM4 model has the same structure as the original MM4 except that it includes a sophisticated surface physics and soil hydrology package, an explicit boundary layer formulation, and a more detailed treatment of radiative transfer.

The surface physics and soil hydrology package is the Biosphere-Atmosphere Transfer Scheme (BATS), Dickinson et al. (1986). The radiation parameterization is the same scheme as the one used in NCAR General Circulation Model (GCM); it performs separate calculations of atmospheric heating rates and surface fluxes for solar and infrared radiation for clear and cloudy skies. The solar clear sky scheme follows the parameterization of Lacis and Hansen (1974). The solar cloudy sky scheme accounts for reflection at the top of the clouds, multiple reflections between the clouds, and between the ground and clouds. Infrared radiative transfer scheme includes the contribution of atmospheric gases and clouds.

Two recent studies, Anthes et al. (1989) and Giorgi and Bates (1989), focus on the climatological skill of the MM4 model. They test the skill of the model in simulating observed climatology when driven by the corresponding observed boundary and initial conditions. The model performed reasonably good in these experiments.

The MM4 model has been used recently in many studies to simulate details of regional climates, Giorgi(1990). The model is driven with output from a General Circulation Model (GCM); the high resolution of the MM4 is utilized in resolving some physical effects which are not resolved by the G.C.M. , e.g., effects of topography. This one-way nesting procedure can be very useful in making predictions about changes in the regional climate. These predictions are not usually possible to make using G.C.M.s due to their coarse resolution. Implicit in this nesting procedure is the assumption that the improvements in description of the physical processes inside the model domain will not have any effect on the surrounding atmosphere.

4.4.2 Design of the Simulations

The sub-region of the Amazon rainforest considered in these simulations is centered at 6.5° S and 67.5° W, Figure 4.5. The scale of this region is 1600 kilometers each side. The simulations are performed for the months of January and July to represent typical summer and winter conditions respectively. For each of the two months the model is run to simulate the climate of the region.

The spatial resolution which is used in this study is 50 kilometers in the horizontal. 14 pressure levels are distributed between the surface and the tropopause in the vertical. The temporal resolution is 90 seconds.

The climate model is driven by solar radiation and boundary conditions from the European Center for Medium-range Weather Forecast (ECMWF) global data set. Temperature and pressure are specified at the boundaries according to the ECMWF data. Wind

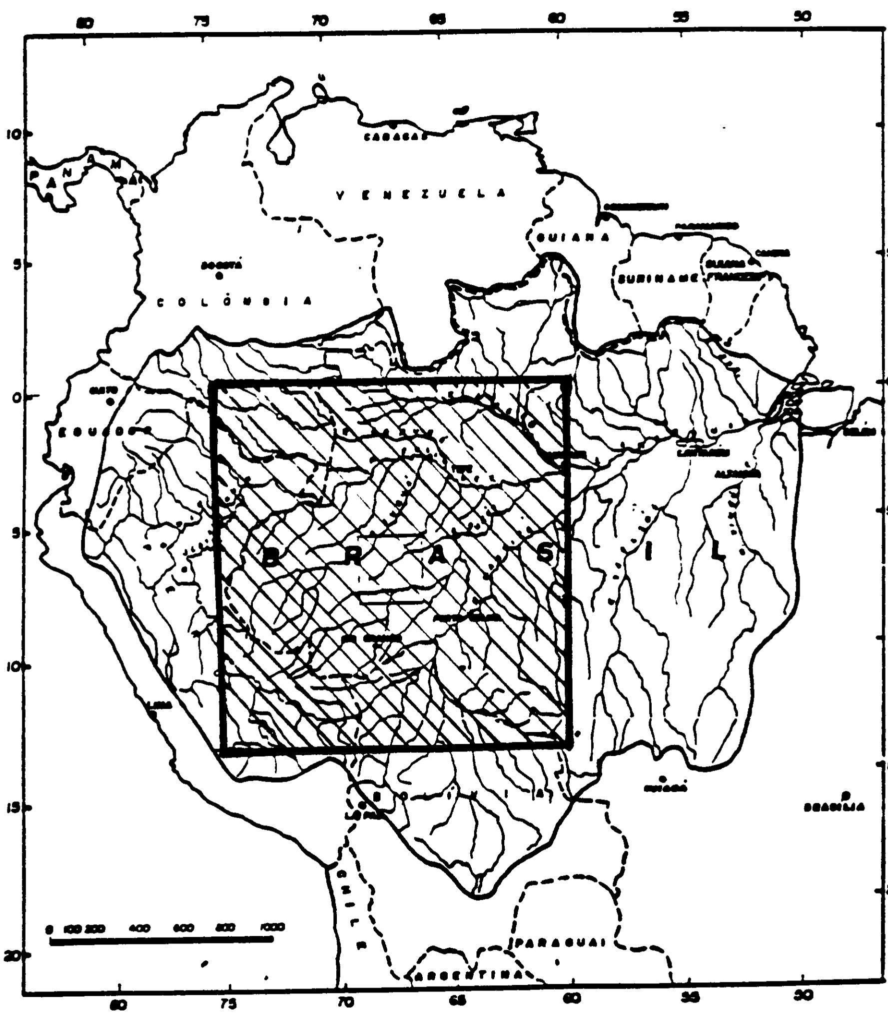

Figure 4.5 Location of the region considered in the climate simulation.

and specific humidity are specified at the inflow boundaries from the ECMWF analysis but the same two variables are predicted by the model solution at the outflow boundaries. This last condition is necessary for a smooth solution with insignificant boundary effects. The upper boundary condition is a no flow boundary.

For each month initial conditions are specified using the corresponding conditions from the ECMWF data. The initial soil moisture conditions are specified according to the standard values recommended by the original MM4 modeling system, Anthes et al. (1987). These values describe typical conditions for each season and land cover type.

The January and July climates are simulated by driving the model with the ECMWF data for January and July of the years 1985, 1986 and 1989. Since the years 1987 and 1988 include El Niño events, they are not included in these simulations. The average climate is estimated from the averages for January and July of the three years.

4.4.3 Results of the Simulations

The results of modeling surface runoff and rainfall interception in the Amazon basin using a 3-D climate model are shown Table 4.2. The results of the simulations indicate that the runoff coefficient and the interception loss ratio are simulated with reasonable accuracy compared with observations. This accuracy suggests that the modified version of BATS which utilizes the new procedure is reasonably successful in simulating land-surface hydrology in a rainforest environment.

Variable	Model	Observations	References
January			
evaporation	140	107	(1)
precipitation	184	270	(1)
runoff	86		
interception loss	22		
runoff ratio	47%	44%*	(2)
interception ratio	12%	10%	(1)
July			
evaporation	115	119	(1)
precipitation	64	110	(1)
runoff	27		
interception loss	9		
runoff ratio	42%	44%*	(2)
interception ratio	14%	20%	(1)

Table 4.2 : Results of the 3-D climate simulations.
* the observation of runoff coefficient is an annual value.
references are indexed as follows
(1) Shuttleworth (1988)a, and (2) Oltman (1967)

Table 4.2 shows that precipitation is underestimated for both months. This result is due to underestimation of atmospheric moisture convergence into the region by the ECMWF analysis which is used as boundary conditions for these simulations.

4.5 Concluding Discussion

The results of the simulations presented in section 4.4 indicate a reasonable success in modeling land-surface hydrology in a rainforest environment. The results of these simulations provide the necessary verification of the new procedure for estimation of the fractional coverage of rainfall in climate models. The accuracy of the simulations in Section 4.4 can not be attributed solely to the new procedure; this accuracy is achieved by using a physical-statistical approach in modeling surface runoff and rainfall interception and by adopting the basic structure of the original BATS. On the other hand the results of the off-line simulations in Section 4.3 suggest that rainfall interception and runoff are quite sensitive to the value of the fractional coverage of rainfall, μ. Hence, the combination of the new procedure for estimation of μ and the modified version of BATS seems to provide a consistent and accurate scheme for modeling surface hydrology in the rainforest environment.

CHAPTER 5

Conclusions

5.1 Introduction

This chapter presents a summary of the results, the general conclusions of the study, and some suggestions for future research.

5.2 Summary of the Results

(1) A new formula is developed for estimation of the fractional coverage of rainfall in convective storms. The new formula is suitable for use with any land-surface hydrology scheme as part of a climate model.

(2) The observations of convective storms which are the basis for the new formula are reviewed. It is shown that the new formula is consistent with observations from the tropics, subtropics, and midlatitudes.

(3) The new formula is applied in simulating the fractional coverage of rainfall over the Amazon basin. A time series of area average rainfall generated by a climate model is used in those simulations. It is found that the simulated fractional coverage of rainfall exhibits significant variability in time, which contradicts the assumption usually made in climate models that this fraction is a constant in time.

(4) The sensitivity of land-surface hydrology to the assumed fractional coverage of rainfall is studied using an off-line model of the Biosphere-Atmosphere Transfer Scheme (BATS). The scheme is modified to include the effects of sub-grid scale spatial variability

on land-surface hydrology. It is found that hydrologic processes such as rainfall interception and surface runoff are sensitive to the assumed fractional coverage of rainfall. This result is consistent with previous sensitivity studies which are reviewed in Chapter 2. The new formula for computing the fractional coverage is included into the off-line model and the results of the simulations using the new procedure are compared to those obtained by assuming constant fractional coverage of rainfall.

(5) The new procedure is implemented as part of the modified BATS in 3-D simulations of the regional climate of the Amazon basin. The results of these simulations indicate that a combination of the modified BATS and the new procedure for estimation of the fractional coverage of rainfall is capable of simulating the land-surface hydrology in a rainforest environment with reasonable accuracy. The surface runoff coefficient and the ratio of interception loss to total rainfall are predicted with reasonable accuracy.

5.3 General Conclusions

The new procedure for estimation of the fractional coverage of rainfall, μ, captures the seasonal and geographical variability in μ. It even describes the variability of μ from storm to storm and within the life cycle of a single storm. The new formula describes explicitly the dependence of the fractional coverage of rainfall on the spatial and temporal resolutions of the climate model. This advantage is significant since it allows for the possibility of using the new procedure in any climate run irrespective of the resolutions used. The procedure presented in this study is consistent with the observations of convective storms which indicate that μ is indeed a variable and not a constant parameter.

The procedure presented in this study is very easy to apply. Instead of specifying a constant value for μ to characterize rainfall in each region of the world, it is suggested to compute the climatological mean rainfall rate in every region and allow μ to vary in space and time. It is much easier to obtain information about the climatological mean rainfall rate from the records at a single rain gauge than to obtain information about the fractional coverage of rainfall.

From the results of applying the new procedure in computing μ in a rainforest environment it is concluded that inclusion of the variability of μ in land-surface hydrology parameterizations is crucial to the accuracy of those descriptions. It is unreasonable to neglect the effects of spatial variability on land-surface hydrology over large areas, but it is equally unreasonable to assume that all the convective storms, in every region of the world and in every season cover the same area.

The assumption of a *unique* conditional PDF for the rainfall rate process is an idealization of the process. In the real world that PDF may vary between the different storms and even within the life cycle of a single storm. Kedem et al. (1990) suggest that the relation between the total volume of rainfall and the area of the storm which receives rainfall rate above a certain threshold is robust when the threshold chosen is greater than zero. However the observations reviewed in Chapter 2 indicate that although the

PDF may vary in time, the range of this variability is small and hence the assumption of a *unique* PDF is a good working assumption.

5.4 Future Research

The observations presented in Chapter 2 describe convective storms. Future research may focus on observations of the area-volume relation in rainfall fields associated with frontal systems. These observations could be useful for rainfall measurement purposes as well as for verifying the new procedure developed in this study. Preliminary analysis of some limited data obtained using the MIT radar indicates that the new procedure can be used for estimation of the fractional coverage of rainfall produced by frontal systems. But further data collection and analysis are needed.

The data collection has to be planned to avoid snow storms and to obtain data at equal intervals throughout the life cycle of a storm and particularly the early stages. The data available from the MIT radar were collected for other purposes and do not satisfy this last condition.

Application of the new procedure requires data on rainfall at a point, which is available for many regions around the world. Future research may focus on processing data from different regions of the world to produce a map of the climatological rainfall intensity, such a map would provide sufficient information for applying the new procedure in simulations of the climate of planet Earth.

REFERENCES

Anthes , R.A., 1977. A cumulus parameterization scheme utilizing a one dimensional cloud model., *Mon. Wea. Rev.*, **106**, 1045-1078.

Anthes, R.A.,1990. Recent applications of the Penn. state/NCAR mesoscale model to synoptic, mesoscale and regional climate studies, *Bull. Amer. Meteor. Soc.*, **71**, 1610-1629.

Anthes, R.A., E. Y. Hsie and Y.H. Kuo, 1987. Description of the Penn. state/NCAR mesoscale model version 4 (MM4), *NCAR tech. note, NCAR/TN-282+STR*, 66 pages.

Anthes R.A. , Y.-H Kuo, E. -Y Hsie, S. Low-Nam and T. W. Bettge,1989. Estimation of skill and uncertainty in regional numerical models, *Quart. J. Roy. Meteor. Soc.*, **115**, 763-806.

Atlas, D., D. Rosenfeld and D.A. Short, 1990. The estimation of convective rainfall by area integrals, 1, the theoretical and empirical basis, *J. Geophys. Res.*, **95**, 2153-2160.

Byers, H.R., 1948. The use of radar in determining the amount of rain falling over a small area, *Trans. Amer. Geophys. Union*, **29**, 187-196.

Deardorff, J. W., 1972. Parameterization of the planetary boundary layer for use in general circulation models, *Mon. Wea. Rev.*, **100**, 93-106.

Dickinson R.E., A. Henderson-Sellers, P.J. Kennedy and M.F. Wilson, 1986. Biosphere-Atmosphere Transfer Scheme (BATS) for the NCAR community climate model, *NCAR tech. note, NCAR/TN-275+STR*, 69 pages.

Doneaud, A. A., P. L. Smith, A. S. Dennis, and S. Sengupta, 1981. A simple method for estimating convective rainfall over an area. *Water Resour. Res.*, **17**, 1676-1682.

Doneaud, A. A., S. I. Niscov, D. L. Priegnitz, and P.L. Smith, 1984. The area-time integral as an indicator for convective rain volumes, *J. Clim. Appl. Meteor.*, **23**, 555-561.

Eagleson, P.S., N.M. Fennessey, Q. Wang and I. Rodriguez-Iturbe, 1987 : Application of spatial Poisson models to air mass thunderstorm rainfall, *J. Geophys. Res.*, **92(D8)**, 9661-9678.

Eltahir, E.A.B., 1993. Interactions of Hydrology and Climate in the Amazon Basin, Doctorate Thesis, Massachusetts Institute of Technology, Cambridge, Mass., 188 pages.

Eltahir, E.A.B. and R.L. Bras, 1991. Spatial variability in the hydrology of the Biosphere-Atmosphere Transfer Scheme (BATS), American Geophysical Union, Abstracts of Fall Meeting, page 149.

Entekhabi, D., and P.S. Eagleson, 1989. Land surface hydrology parameterization for atmospheric general circulation models including sub-grid scale spatial variability, *J. Climate*, **2**, 816-831.

Famiglietti, J.S. and E.F. Wood, 1990. Evapotranspiration and runoff from large land areas, landsurface hydrology for atmospheric general circulation models, *Land surface-atmospheric interactions for climate modeling*, E.F. Wood (Ed.), Kluwer Academic Publishers, 179-205.

Giorgi F., 1990. Simulation of regional climate using a limited area model nested in a general circulation model, *J. Climate*, **3**, 941-963.

Giorgi, F. and G.T. Bates, 1989. The climatological skill of a regional model over complex terrain, *Mon. Wea. Rev.*, **117**, 2325-2347.

Johnson K.D., D. Entekhabi and P.S. Eagleson, 1991. The implementation and validation of improved landsurface hydrology in an atmospheric general circulation model, *Ralph M. Parsons laboratory technical report # 334*, Massachusetts Institute of Technology, Cambridge, 192 pages.

Kedem B., L.S. Chiu and Z. Karni, 1990. An analysis of the threshold method for measuring area average rainfall, *J. Appl. Meteor.*, **29**, 3-20.

Lacis, A.A and J.E. Hansen, 1974. A parameterization for the absorption of solar radiation in the Earth's atmosphere, *J. Atmos. Sci.*, **31**, 118-133.

Lloyd, C.R., 1990. The temporal distribution of Amazonian rainfall and its implications for forest interception, *Quart. J. Roy. Meteor. Soc.*, **116**, 1487-1494.

Lopez, R.E., D. Atlas, D. Rosenfeld, J.L. Thomas, D.O. Blanchard and R.L. Holle, 1989. Estimation of of areal rainfall using the radar echo area-time-integral. *J. Appl. Meteor.*, **28**, 1162-1175.

Manabe, S., 1969. Climate and the ocean circulation : I The atmospheric circulation and the hydrology of the Earth's surface, *Mon. Wea. Rev.*, **97**, 739-774.

Pitman A. J., A. Henderson-Sellers and Z-L Yang, 1990. Sensitivity of regional climates to localized precipitation in global models, *Nature*, **346**, 734-737.

Oltman, R.E., 1967. Reconnaissance investigation of the discharge and water quality of the Amazon, *Biota Amazonica*, **3**, 163-185.

O'Neill, C.A. and R.E. Dickinson, 1991. Temporal distribution of Amazon rainfall in CCM1/BATS climate simulation, American Geophysical Union, Abstracts of Fall Meeting, page 149.

Rodriguez-Iturbe, I. and P. S. Eagleson, 1987, Mathematical models for rainstorm events in space and time, *Water Resour. Res.*, **23**, 181-190.

Rutter A. J., K. A. Kershaw, P. C. Robins and A.J. Morton, 1971. A predictive model of rainfall interception in forests, I. derivation of the model from observations in a plantation of Corsican pine, *Agric. Meteorol.*, **9**, 367-384.

Rutter A. J., A.J. Morton and P. C. Robins , 1975. A predictive model of rainfall interception in forests, II. Generalization of the model and comparison with observations in some Coniferous and hardwood stands, *J. Appl. Ecol.*, **12**, 367-380.

Sellers, P. J., Y. Mintz, Y. C. Sud, and A. Delcher, 1986. A Simple Biosphere (SiB) model for use within general circulation models, *J. Atmos. Sci.*, **43**, 505-531.

Short, D. A., D. Rosenfeld, D. B. Wolff, and D. Atlas, 1989. A raingage, radar and satellite simulation study of the estimation of convective rainfall by area-time integrals, 4th conference on satellite meteorology and oceanography, American Meteorological Society, San Diego, CA, paper 10A.2,

Shuttleworth, W.J., 1988 a. Evaporation from Amazonian rainforest, *Phil. Trans. R. Soc. Lond.*, **B233**, 321-346.

Shuttleworth, W.J., 1988 b. Macrohydrology-the new challenge for process hydrology, *J. Hydrol.*, **100**, 31-56.

Thomas G. and A. Henderson-Sellers, 1991. An evaluation of proposed representations of subgrid hydrologic processes in climate models, *J. Climate*, **4**, 898-910.

Warrilow, D.A., A.B. Sangster and A. Slingo, 1986. Modelling of land surface processes and their influence on European climate, *DCTN 38*, *Dynamical Climatology*, Meteorological Office, Bracknell, Berkshire, 92 pages.

APPENDIX 4.1

New Interception Scheme

Introduction

This Appendix outlines the development of a new rainfall interception scheme which is based on the Rutter model of interception. This scheme is introduced by Eltahir (1993). In the following the Rutter model is described in some detail. The derivation of the new interception scheme is then presented.

Rutter Model of Interception

This model was introduced by Rutter et al. (1971) to provide a predictive tool of rainfall interception. The model specifies the functional dependence of canopy drainage and canopy evaporation on canopy storage. Canopy drainage is described by

$$D_r = K \cdot e^{\left(\frac{C}{b}\right)} \tag{A4.1}$$

where D_r is canopy drainage, C is canopy storage, K and b are constants characteristics of the canopy. It is important to note the exponential dependence of canopy drainage on canopy storage. This strong dependence results in rapid depletion of excessive local storage.

Evaporation from the canopy has two components: interception loss and transpiration. It is described by

$$e = \frac{C}{S} \cdot e_c + \left(1 - \frac{C}{S}\right) \cdot e_t, \quad 0 \leq C \leq S,$$

$$e = e_c, \; C \geq S \tag{A4.2}$$

where e_t is transpiration by the plant, e_c is evaporation from wet canopy and S is a constant characteristic of the canopy. S is the amount of water retained by the canopy after being completely wet and then drained for a "sufficiently" long period.

Canopy storage is added by rainfall and depleted by drainage and evaporation. The rate of change of canopy storage is given by,

$$\frac{\partial C}{\partial t} = (1-p).\ P - \frac{C}{S}.\ e_c - D_r \qquad \text{(A4.3)}$$

where p is fraction of rain falling directly to the ground and P is rainfall.

The exponential dependence of canopy drainage on canopy storage results in large drainage for large canopy storage. Hence when applying the model in describing interception processes using real data, e.g., Rutter et al. (1975), it is observed that canopy storage does not exceed a maximum of about 2 or 3 mm. The Rutter model is modified here to include a maximum limit for canopy storage, C_m, the maximum storage which the canopy can hold at any instant of time. This limit constrains primarily Equation A4.3 such that C does not exceed C_m. Equation A4.1 is also modified to

$$D_r = K.\ e^{\left(\frac{C}{b}\right)},\ C < C_m,$$

$$D_r = K.\ e^{\left(\frac{C_m}{b}\right)},\ C \geq C_m \qquad \text{(A4.4)}$$

A Description of Rainfall Interception over Large Areas

A new interception scheme is developed in this section. It combines the Rutter model and statistical description of the spatial variability in rainfall and canopy storage. It is assumed that rainfall is distributed in space according to

$$f_P = (1 - q_P).\ \delta(P - 0) + \frac{q_P^2}{E(P)}.\ e^{-\left(\frac{q_P . P}{E(P)}\right)} \qquad \text{(A4.5)}$$

where P is rainfall at any point in space, q_P is fraction of the area with $P > 0$,E( ) denotes the expected value of and δ denotes the Dirac delta function. The observations of Eagleson et al. (1987) support the assumption of exponential distribution for rainfall.

Canopy storage controls the local amounts of canopy drainage and evaporation. It is assumed that canopy storage is distributed in space according to an exponential distribution. In absence of any observations of the spatial distribution of canopy storage the choice of the exponential is a matter of convenience. The assumption is justifiable when rainfall variability is a major causal factor for variability in canopy storage. It is assumed that canopy storage is distributed in space according to

$$f_C = (1 - q_C).\ \delta(C - 0) + \frac{q_C^2}{E(C)}.\ e^{-\left(\frac{q_C . C}{E(C)}\right)} \qquad \text{(A4.6)}$$

where C is canopy storage at any point in space, E(C) is the spatially averaged canopy storage and q_C is the fraction of the area with $C > 0$.

The spatially averaged canopy drainage is obtained by taking the expected value of both sides in Equation A4.1. E(D_r) is given by

$$E(D_r) = \int_{C=0}^{\infty} D_r(C)\, f_C\, dC$$

$$= \left[1 - q_c + \frac{q_c^2.b}{(b.q_c - E(C))}\right].K + \left[q_c + \frac{q_c^2.b}{(b.q_c - E(C))}\right].K.\ e^{-\left[\frac{(b.q_c - E(C)).C_m}{b.E(C)}\right]} \qquad \text{(A4.7)}$$

The spatially averaged evaporation is obtained by taking the expected value of both sides in Equation A4.2. E (e) is given by

$$E(e) = \int_{C=0}^{\infty} e(C).\, f_C \, dC = e_t + (e_c - e_t).\, \frac{E(C)}{S}.\, \left(1 - e^{-\left(\frac{S.q_c}{E(C)}\right)}\right)$$

and

$$E(e') = e_c \, . \, \frac{E(C)}{S}.\, \left(1 - e^{-\left(\frac{S.q_c}{E(C)}\right)}\right) \qquad \text{(A4.8)}$$

where e' is interception loss.

Throughfall has three components: the fraction of rain falling directly to the ground through gaps in the canopy, drainage from the canopy and rainfall in excess of drainage at locations with maximum canopy storage. The spatial average throughfall is given by

$$E(T) = p\, .E(P) + E(D_r)$$

$$+ \int_{C=C_m}^{\infty} \int_{P=\frac{D_m}{(1-p)}}^{\infty} [(1-p).P - D_m - e'(C)].f_C.f_P \, . \, dP \, . \, dC$$

$$= p.\, E(P) + E(D_r)$$

$$+ [q_C \, . \, (1-p).\, E(P) - q_C.q_p.e_C] \, . \, e^{-\left[\frac{q_C.C_m}{E(C)} + \frac{q_p.D_m}{(1-p).E(P)}\right]} \qquad \text{(A4.9)}$$

where D_m is D_r (C_m).

BIOGRAPHICAL NOTE

Elfatih Eltahir was born in Omdurman, Sudan, in October of 1961. He received a Bachelor of Science with First Class Honors in Civil Engineering from the University of Khartoum in November 1985. He then started his graduate studies in Galway, Republic of Ireland, in August 1986. He received a Master of Science with First Class Honors in Hydrology from the National University of Ireland in August 1988. The title of the Master thesis was "Drought Frequency Analysis of Annual Rainfall Series in Central and Western Sudan".

In August 1988 he started his studies at the Massachusetts Institute of Technology working towards a Master of Science in Meteorology from the Department of Earth, Atmospheric and Planetary Sciences and a Doctor of Science in Hydroclimatology from the Department of Civil and Environmental Engineering. The title of the Doctorate thesis was " Interactions of Hydrology and Climate in the Amazon Basin ".

Elfatih Eltahir is the son of Ali Babiker Eltahir and Nafisa Hassan Musa. He has two brothers : Mohammed and Omer, and four sisters : Maha, Ilham, Lama, and Mai. He and his wife Shahinaz were married in December 1991.