Observations of the transient characteristics of the hydrological balance

Matthew Czikowsky
University at Albany, State University of New York, matt@asrc.cestm.albany.edu

The University at Albany community has made this article openly available. Please share how this access benefits you.

Follow this and additional works at: https://scholarsarchive.library.albany.edu/legacy-etd

Part of the Environmental Sciences Commons, and the Hydrology Commons

Recommended Citation
https://scholarsarchive.library.albany.edu/legacy-etd/20

This Dissertation is brought to you for free and open access by the The Graduate School at Scholars Archive. It has been accepted for inclusion in Legacy Theses & Dissertations (2009 - 2024) by an authorized administrator of Scholars Archive.
Please see Terms of Use. For more information, please contact scholarsarchive@albany.edu.
OBSERVATIONS OF THE TRANSIENT CHARACTERISTICS
OF THE HYDROLOGICAL BALANCE

by

Matthew J. Czikowsky

A Dissertation
Submitted to the University at Albany, State University of New York
In Partial Fulfillment of
The Requirements for the Degree of
Doctor of Philosophy

College of Arts & Sciences
Department of Atmospheric and Environmental Sciences
2009
Acknowledgements

First, I would like to acknowledge my advisor, Dave Fitzjarrald, for his guidance, ideas, and insightful discussions through all stages of this work. Next, I thank my committee members, Gar Lala, Geoffrey Parker, and Karen Mohr, for their careful reading of and helpful comments on the thesis. Being a part of the Jungle Research Group at ASRC, I have had the opportunity to participate in many projects with and learn from the people who have been a part of the group during my stay in Albany: Ricardo Sakai, Luiz Medeiros, Alex Tsoyref, Jeff Freedman, Otávio Acevedo, Kathy Moore, Ralf Staebler, Gary Wojcik, and Dwayne Spiess. I thank John Sicker for his work keeping all the computer systems running here at ASRC. Much of the data used in this study was the result of a lot of field work that required the help of many people. Additional collaborators for the LBA studies included Osvaldo Moraes, Rodrigo da Silva, Júlio Tota, Troy Beldini, Lucy Hutyra, Steve Wofsy, Elaine Gottlieb, Scott Miller, Marc Kramer, Raimundo Cosme de Oliveira, Eleazar Brait, and Valdelirio Miranda. For the HVAMS study, additional collaborators included the NCAR-ISFF staff, MIPS staff, Jessica Neiles, along with the help of former work-study students for the Jungle Research Group, including Marina Virnik, Kim Sutkevich, Matt Doody, and Jason Herb. For assistance on working with CMORPH data, I acknowledge Júlia Cohen. And for assistance with data extractions and plotting of the ETA model run at ASRC, I thank Mark Beauharnois. The USGS is acknowledged for access to their high-rate streamflow data, as well as the NYCDEP for the Catskill weather station data. The NWS Albany office staff, including Vasil Kolec i, are acknowledged for their assistance with the detailed sounding and radar data. I thank my undergraduate advisor Ray Castillo for his guidance and support that led me to Albany. Finally I thank my family and friends for always standing behind me during every step of this work.
Abstract

The impact of transient input events on the hydrological balance components are studied over a wide range of spatial and temporal scales. First, interception evaporation and energy-balance changes during and following precipitation events are examined at the site spatial and event temporal scale. Second, the hydrological response to spatial and seasonal changes in precipitation and evapotranspiration due to land cover changes are examined on the small-to-large watershed scale, covering the event to seasonal time scales.

Estimates in the literature for interception evaporation using conventional methods vary widely even over the same forest type. We present an alternative method to estimate interception evaporation from eddy covariance measurements combined with a novel use of data-analysis techniques to form base state and event-based ensembles. Application of this method at a tropical rain forest site in the Amazon resulted in mean interception evaporation estimates of 11.6%, comparable to recent conventional studies in that region. Energy balance comparisons between dry and afternoon rain-days show an approximately 15% increase of evaporative fraction on the rain days, with the energy being supplied by a corresponding decrease in the canopy heat storage.

Large differences in the watershed response characteristics of streamflow peak and streamflow peak to watershed precipitation ratio in the Catskill-Hudson Valley Region of New York were observed relative to a watershed’s proximity to a precipitation shadow observed there. There were detectable changes in streamflow and soil moisture recession times, as well as in the diurnal streamflow amplitude over this watershed network during the autumn transition. This provides an independent estimate of autumn
or spring onset to complement phenological or satellite-derived measurements, and is important as this shows the timing of the hydrological impacts of abrupt changes in evapotranspiration forcing on the watersheds.

In a network of watersheds in the Amazon, precipitation and subsequent storage and subsurface drainage processes seem to have a greater influence on seasonal changes in soil moisture recession compared with seasonal variations in evapotranspiration due to vegetation state.
TABLE of CONTENTS

Acknowledgements ............................................................................................................. iii
Abstract .............................................................................................................................. iv
TABLE of CONTENTS .................................................................................................... vi
LIST of FIGURES ............................................................................................................. ix
LIST of TABLES ............................................................................................................. xvi

Chapter 1: Introduction ................................................................................................... 1
  1.1 Definitions and overview ......................................................................................... 1
    1.1.1 Surface water balance ......................................................................................... 1
    1.1.2 Evapotranspiration components ........................................................................... 2
    1.1.3 Transient terms of the hydrological balance ....................................................... 3
    1.1.4 Forest water budget ............................................................................................. 4
    1.1.5 Surface energy balance ...................................................................................... 5
    1.1.6 Spatial and temporal scales examined in this study .......................................... 7
    1.1.7 Closing the water balance in climate models .................................................... 9
  1.2 Outline and research questions ................................................................................. 10
  1.3 References .............................................................................................................. 13

Chapter 2: Detecting the evaporation of intercepted water over an old-growth rain
forest in the eastern Amazon using eddy flux measurements ........................................ 16
Abstract ......................................................................................................................... 16
  2.1 Introduction ............................................................................................................ 17
    2.1.1 The interception process .................................................................................... 17
    2.1.2 Definitions ......................................................................................................... 18
    2.1.3 Conventional interception-measurement methods .............................................. 19
    2.1.4 Interception estimation in global climate models ............................................... 27
    2.1.5 A new technique for measuring interception ...................................................... 28
    2.1.6 Outline .............................................................................................................. 30
  2.2 Location and Data ................................................................................................... 31
  2.3 Methods .................................................................................................................. 36
    2.3.1 Identification of rainfall events .......................................................................... 36
    2.3.2 Flux calculation methods .................................................................................... 40
    2.3.3 Flux datasets and ensemble formation ............................................................... 43
    2.3.4 Nighttime rainfall event methods ....................................................................... 45
    2.3.5 Individual daytime event LE baseline determination ....................................... 45
    2.3.6 Treatment of heavy rainfall-rate periods ........................................................... 48
  2.4 Results ..................................................................................................................... 51
    2.4.1 Nighttime precipitation events .......................................................................... 51
    2.4.2 Daytime precipitation events ............................................................................. 55
    2.4.3 Interception during the mornings following nighttime events ................................ 63
    2.4.4 Energy balance comparison for dry and rain days ............................................. 64
  2.5 Summary ................................................................................................................ 67
  2.6 Acknowledgements ................................................................................................. 71
Chapter 3: Spatial and seasonal changes in watershed response to rainfall events in two contrasting climates

Introduction (Foreword) ........................................................................................................ 77

Part I: HVAMS .................................................................................................................... 78

Abstract ............................................................................................................................... 78

3.1 Introduction .................................................................................................................. 79

3.1.1 Precipitation shadows ......................................................................................... 79

3.1.2 Catskill-Hudson Valley region precipitation shadow ............................................ 80

3.1.3 Watershed response .............................................................................................. 86

3.1.4 Catskill region watersheds ................................................................................... 87

3.1.5 Outline .................................................................................................................... 92

3.2 Location and Data/Instrumentation .......................................................................... 95

3.3 Methods ....................................................................................................................... 105

3.3.1 Watershed boundary determination ................................................................. 105

3.3.2 Watershed response stages ................................................................................. 107

3.3.3 Streamflow peaks ............................................................................................... 108

3.3.4 Recession ............................................................................................................. 109

3.3.5 Diurnal streamflow amplitude ............................................................................. 109

3.3.6 Model studies ........................................................................................................ 111

3.4 Analysis ....................................................................................................................... 114

3.4.1 September 23, 2003 precipitation case ............................................................ 114

3.4.1.1 Synoptic situation .......................................................................................... 114

3.4.1.2 Precipitation ................................................................................................. 117

3.4.1.3 Comparison of observations with remote sensing data .................................. 120

3.4.1.4 Watershed precipitation ................................................................................. 126

3.4.1.5 Streamflow peak analysis .............................................................................. 128

3.4.2 October 27, 2003 precipitation case ................................................................. 131

3.4.2.1 Synoptic situation .......................................................................................... 131

3.4.2.2 Precipitation ................................................................................................. 135

3.4.2.3 Comparison of observations with remote sensing data .................................. 139

3.4.2.4 Watershed precipitation ................................................................................. 144

3.4.2.5 Streamflow peak analysis .............................................................................. 146

3.4.3 Seasonal changes ................................................................................................. 148

3.4.3.1 Land cover and evapotranspiration changes ................................................. 148

3.4.3.2 Seasonal diurnal streamflow oscillation analysis .......................................... 153

3.4.3.3 Seasonal streamflow peak analysis ............................................................... 157

3.4.3.4 Seasonal streamflow recession analysis ....................................................... 157

3.4.4 Model sensitivity study ......................................................................................... 158

3.4.4.1 Model run setup ........................................................................................... 158

3.4.4.2 Model results: streamflow ............................................................................ 161

3.4.4.3 Model results: soil moisture ......................................................................... 164

3.4.4.4 Model results: evapotranspiration components ............................................ 166

3.5 Summary ...................................................................................................................... 168

3.6 Acknowledgements ..................................................................................................... 173
Chapter 3L: Hydrologic response to precipitation events in the eastern Amazon region

3.1L Introduction

3.1.1L Eastern Amazon region watersheds

3.1.2L Deforestation impacts on components of the Amazon climate/water balance

3.1.3L Outline

3.2L Location/Data

3.3L Methods

3.4L Results

3.4.1L Rainfall

3.4.2L Streamflow

3.4.3L Seasonal trends

3.5L Summary

3.6L References

Chapter 4: Conclusions

4.1 Summary and Conclusions

4.2 Future work

Appendix 1: Station location tables used in chapter 3
LIST of FIGURES

Chapter 1:

Figure 1.1 Components of evapotranspiration (ET) as shown in the context of the forest water balance. ................................................................. 2

Figure 1.2 Forms of transient responses to precipitation for: interception evaporation (a, lop left), soil moisture (b, top right), and streamflow (c, bottom). ......................3

Figure 1.3 Spatial and temporal scales of the processes analyzed and measurement methods considered in the thesis. .............................................. 7

Figure 1.4 Cloud base at the km67 old-growth rainforest site of the Large-Scale-Biosphere-Atmosphere Experiment in Amazonia (LBA), lifting condensation level (LCL) at the LBA km67 site, and LCL at the LBA km77 agricultural site during a dry season period in 2001 ......................... 8

Chapter 2:

Figure 2.1 Interception estimates reported in the literature using conventional methods for tropical rain forest sites.............................................. 24

Figure 2.2 Height sections of canopy surface area density along six 500m transects at an intact old-growth forest site, the km67 site in the Santarem region of the Large-Scale Biosphere-Atmosphere Experiment in Amazonia (LBA-ECO). ................................................... 26

Figure 2.3 Diagram illustrating the method used to estimate interception using eddy covariance. ................................................................. 28

Figure 2.4 Map of the weather stations and flux-measurement sites located in the Santarém region (STM) of LBA-ECO................................................ 32

Figure 2.5 Rain dials for the km67 site during the wet season (left) and dry season (right). ...................................................................................... 34

Figure 2.6 Median cloud cover fraction at km67 by hour of day for the wet season (February through May) and the dry season (September through December) for 2001 to 2003 ......................................................... 35

Figure 2.7 Top panel: Cloud base at km67, lifting condensation level (LCL) at km67, and LCL at km77 during a wet season period in 2001 (May 2-11, days 122-131). Bottom panel: As in top panel but for a dry season period in 2001 (October 2-12, days 275-285). .................................................... 36

Figure 2.8 Raw ceilometer backscatter (15-second samples) from 1300 to 1800 LT on December 10, 2001 at the LBA km67 site...................................... 38

Figure 2.9 Rainfall (mm), and average ceilometer backscatter up to half of the
cloud base height (units of log(10000*srad*km)$^{-1}$, dashed line) for days 338 to 345 in 2001.  

Figure 2.10 (Top) Km67 ensemble mean LE for dry days. (Bottom): Km67 ensemble mean –Q* for dry days.  

Figure 2.11 Diagram illustrating the method used to determine nighttime interception losses.  

Figure 2.12 Top: Precipitation event LE (W m$^{-2}$), dry-day ensemble LE, corrected dry-day baseline LE using method 1, method 2, and method 3. Bottom: Precipitation event –Q* (W m$^{-2}$), dry-day ensemble –Q*, and rain-day ensemble –Q*.  

Figure 2.13 Top panel: Ensemble mean latent heat flux (W m$^{-2}$) for all 54 nighttime precipitation events. Bottom panel: Mean wind speed (m s$^{-1}$) and u* (m s$^{-1}$) for the same 54 nighttime precipitation events.  

Figure 2.14 Top left: Histogram of all nighttime rainfall events by interception percentage. Top right: Histogram of all nighttime dry-season rainfall events by interception percentage. Bottom left: Scatterplot of event precipitation (mm) and interception percentage. Bottom right: Scatterplot of wind speed (m s$^{-1}$) and interception percentage.  

Figure 2.15 Top panel: Ensemble mean latent heat flux (W m$^{-2}$) for the 4 nighttime high-interception, high-wind events. Bottom panel: Mean wind speed (m s$^{-1}$) and u* (m s$^{-1}$) for the same 4 nighttime precipitation events.  

Figure 2.16 Top panel: Ensemble mean latent heat flux (W m$^{-2}$) for the 4 nighttime high-interception, low-wind events. Bottom panel: Mean wind speed (m s$^{-1}$) and u* (m s$^{-1}$) for the same 4 nighttime precipitation events.  

Figure 2.17 Mean departure from baseline LE (W m$^{-2}$) for daytime rainfall events with rainfall intensities <= 16 mm hr$^{-1}$ using method 3 (solid black line, standard error dashed).  

Figure 2.18 Mean departure from baseline LE (W m$^{-2}$) for daytime rainfall events with rainfall intensities > 16 mm hr$^{-1}$ using method 3 (solid black line, standard error dashed).  

Figure 2.19 Mean departure from baseline LE (W m$^{-2}$) for daytime Penman-Monteith-filled rainfall events with rainfall intensities > 16 mm hr$^{-1}$ using method 3 (solid black line, standard error dashed).  

Figure 2.20 Mean intercepted water binned by rainfall intensity for daytime events, with the standard error bars for each rain intensity bin shown.  

Figure 2.21 Mean intercepted water binned by rainfall intensity for Penman-Monteith-filled daytime events, with the standard error bars for each rain intensity bin shown.  

Figure 2.22 Total rainfall duration binned by rainfall intensity at the km67 site.  

Figure 2.23 Fraction of total rainfall at the km67 site binned by rainfall intensity.
Figure 2.24 Mean departure from baseline LE (W m⁻²) for early morning periods following nighttime rainfall events using method 3 (solid black line, standard error dashed). .................................................................64

Figure 2.25 Mean energy balance components for dry days (solid lines) and afternoon rain days (dashed lines). ..................................................................................................................65

Figure 2.26 Mean evaporative fraction (top pair of lines), sensible heat fraction (middle pair of lines), and storage fraction (bottom pair of lines) of –Q* for dry days (solid) and days with rain after 1400 GMT (dashed). ........................................66

Figure 2.27 Top: Mean Bowen ratio for dry days (solid black line) and days with rain after 1400 GMT (dashed line). Bottom: Same as top but Bowen ratio medians are plotted. .................................................................67

Figure 2.28 Interception estimates reported in the literature using conventional methods for tropical rain forest sites, with this study added.................................69

Chapter 3:

Figure 3.1 Annual average rainfall (in inches) for the Catskills, Hudson Valley and surrounding regions (Merriman 1907, Fig. 1)................................................ 81

Figure 3.2 850mb-level rain roses from the 12-km ETA model for selected HOBO, PAM, and DEP sites during the HVAMS Intensive Field Campaign (IFC) period. ........................................................................83

Figure 3.3 Surface-level rain roses for selected HOBO, PAM, and DEP surface weather stations during the HVAMS Intensive Field Campaign (IFC) period. ........................................................................ 84

Figure 3.4 Wind roses (black) and transmission factors (TF; gray lines) for the HOBO stations during the HVAMS IFC period ........................................................................................................ 85

Figure 3.5 Map of the major drainage basins in the Catskill-Hudson Valley region. ..................................................................................................................... 90

Figure 3.6 Time periods used in the analysis of streamflow recession and diurnal streamflow fluctuations over the network watersheds for the autumn transition season of 2003. The green bars labeled with L denote the streamflow recession periods used in the growing season in-leaf period. The brown bars labeled with NL denote the streamflow recession periods used in the dormant season no-leaf period. The yellow bars labeled with I, II, III, IV denote the periods used in the diurnal streamflow fluctuation analysis. Streamflow plotted is for Biscuit Brook, NY (units mm on watershed). ........................................................................ 94

Figure 3.7 Topography and data stations for the HVAMS study area of the Hudson Valley and Catskill uplands ........................................................................ 96

Figure 3.8 North-to-South topographic cross-section from South Albany Airport (station 8 in Fig. 3.7) to Phoenicia, NY ................................................................................................. 97
Figure 3.9 NYCDEP stations in the Catskill region, denoted as the sites with six-character identifiers on the western side of the map.............................................. 101

Figure 3.10 Map of the 12-km ETA model grid cell coverage over the HVAMS domain............................................................................................................. 103

Figure 3.11 Map of the 8-km CMORPH model grid cell coverage over the HVAMS domain.......................................................................................................................... 104

Figure 3.12 Map of the 40-km CMORPH model grid cell coverage over the HVAMS domain.......................................................................................................................... 105

Figure 3.13 Map of the USGS stream gauge sites (dots) and watershed boundaries in the Catskill-Hudson Valley region...................................................... 107

Figure 3.14 Diagram showing the stages of streamflow response to rainfall events.......................................................................................................................... 108

Figure 3.15 BROOK model flowchart (adapted from Federer 1995). .......................................................................................................................... 112

Figure 3.16 Northeastern US surface map for 1200Z on September 23, 2003........ 115

Figure 3.17 Northeastern US 850mb analysis for 1200Z on September 23, 2003. ...... 116

Figure 3.18 Albany (KENX) radar VAD wind profile for 1052-1146 Z on September 23, 2003. ........................................................................................... 117

Figure 3.19 Top panel: Rainfall for the PAM and HOBO weather stations for the September 23, 2003 event, covering the period from 0000Z on September 23, 2003 to 0000Z on September 24, 2003. Bottom panel: First row: Hourly ETA-estimated 850-mb winds for the September 23, 2003 event for the grid cell containing the Slide Mountain station. Second row: Hourly surface winds observed at Slide Mountain over the same time period. ................................................................................................................. 118

Figure 3.20 Observed Rainfall for the September 23, 2003 event, covering the period from 0000Z on September 23, 2003 to 0000Z on September 24, 2003..................................................................................................................... 120

Figure 3.21 Albany radar (KENX) estimated storm total precipitation (mm) for the period 0000Z on September 23, 2003 to 0000Z on September 24, 2003..................................................................................................................... 121

Figure 3.22 Scatterplot of observed rainfall ($P_{obs}$) vs. KENX radar-estimated rainfall ($P_{radar}$) for the September 23, 2003 precipitation event, with the regression line shown.......................................................................................... 122

Figure 3.23 CMORPH 8-km estimated rainfall (mm) for the September 23, 2003 precipitation event............................................................................................... 123

Figure 3.24 CMORPH 40-km estimated rainfall (mm) for the September 23, 2003 precipitation event...................................................................................... 124

Figure 3.25 Scatterplot of observed rainfall ($P_{obs}$) vs. CMORPH 8km-estimated rainfall ($P_{CM8}$) for the September 23, 2003 precipitation event, with the regression line shown.......................................................................................... 125
Figure 3.26 Scatterplot of observed rainfall ($P_{obs}$) vs. CMORPH 40km-estimated rainfall ($P_{CM40}$) for the September 23, 2003 precipitation event, with the regression line shown.......................................................................................... 126

Figure 3.27 Map of HVAMS watershed locations with respect to the precipitation shadow for the September 23, 2003 event.............................................................. 127

Figure 3.28 Observed streamflow peaks (in mm; normalized by watershed area) following the September 23, 2003 precipitation event.............................................. 129

Figure 3.29 Northeastern US surface map for 1200Z on October 27, 2003 …………. 132

Figure 3.30 Albany 1200Z sounding for October 27, 2003…………………………. 133

Figure 3.31 Northeastern US 850 mb analysis for 1200Z on October 27, 2003. ……… 134

Figure 3.32 Albany (KENX) radar VAD wind profile for 1227-1320 Z on October 27, 2003......................................................................................................................... 135

Figure 3.33 Top panel: Rainfall for the PAM and HOBO weather stations for the October 27-28, 2003 event, covering the period from 0000Z on October 27, 2003 to 1500Z on October 28, 2003. Bottom panel: First row: Hourly ETA-estimated 850-mb winds for the October 27-28, 2003 event for the grid cell containing the Slide Mountain station. Second row: Hourly surface winds observed at Slide Mountain over the same time period……….. 136

Figure 3.34 Rainfall for the October 27-28, 2003 event, covering the period from 0000Z on October 27, 2003 to 1500Z on October 28, 2003................................. 138

Figure 3.35 Albany radar (KENX) estimated storm total precipitation for October 27-28, 2003 precipitation event. ................................................................. 139

Figure 3.36 Scatterplot of observed rainfall ($P_{obs}$) vs. KENX radar-estimated rainfall ($P_{radar}$) for the October 27, 2003 precipitation event, with the regression line shown................................................................. 140

Figure 3.37 CMORPH 8-km estimated rainfall (mm) for the October 27, 2003 precipitation event................................................................................................. 141

Figure 3.38 CMORPH 40-km estimated rainfall (mm) for the October 27, 2003 precipitation event. ........................................................................................................ 142

Figure 3.39 Scatterplot of observed rainfall ($P_{obs}$) vs. CMORPH 8km-estimated rainfall ($P_{CM8}$) for the October 27, 2003 precipitation event, with the regression line shown. ......................................................................................... 143

Figure 3.40 Scatterplot of observed rainfall ($P_{obs}$) vs. CMORPH 40km-estimated rainfall ($P_{CM40}$) for the October 27, 2003 precipitation event, with the regression line shown. ......................................................................................... 144

Figure 3.41 Map of HVAMS watershed locations with respect to the precipitation shadow for the October 27-28, 2003 event. ................................................. 145

Figure 3.42 Observed streamflow peaks (in mm; normalized by watershed area) following the October 27-28, 2003 precipitation event. ........................................ 147
Figure 3.43 Example phenology over Harvard Forest, Massachusetts, calculated from MODIS interannual spectral mixture analysis (SMA; bottom line, diamonds) and Normalized Difference Vegetation Index (NDVI; top line, squares). ..........150

Figure 3.44 MODIS interannual phenology for regions in the Northeastern US. ....150

Figure 3.45 Flight track for the University of Wyoming King-Air aircraft on the afternoon of October 11, 2003 over the Catskill-Hudson Valley region. ..........151

Figure 3.46 Aerial photographs of land cover from the University of Wyoming King-Air aircraft over the area denoted in Figure 3.44 for October 3, 2003 (upper left), October 11, 2003 (upper right), October 25, 2003 (lower left), and October 30, 2003 (lower right). .................................................................152

Figure 3.47 PAM station network-averaged evaporation (W m⁻²) for September 10, 2003, October 8, 2003, and October 24, 2003. ...............................................153

Figure 3.48 (a): Streamflow stations with an observed diurnal streamflow signal (red X’s), and stations with no observed diurnal streamflow signal (black dots), day of year 234-236 (August 22-24). (b): same as in (a), but for day of year 252-254 (September 9-11). (c): same as in (a), but for day of year 281-283 (October 8-10). (d): same as in (a), but for day of year 296-298 (October 23-25). .................155

Figure 3.49 Histograms of stations observing a diurnal streamflow signal binned by drainage area for the time periods of August 22-24, 2003 (upper left), September 9-11, 2003 (upper right), and October 8-10, 2003 (lower left). Bottom right: Watershed area vs. diurnal signal time fraction (solid line) for a network of 151 stream-gauging stations in the eastern U.S. for the period 1989-2001, with the upper and lower 95% confidence intervals for the median shown. ..........156

Figure 3.50 Top: Biscuit Brook, NY observed streamflow (mm on watershed) for the September 23, 2003 event. Bottom: Brook model simulated streamflow (mm) for the September 23, 2003 event using the following run scenarios: Base precipitation with leaves (black solid line); Base precipitation without leaves (black dotted line); Doubled precipitation with leaves (blue solid line); Doubled precipitation without leaves (blue dotted line). ..............................................162

Figure 3.51 BROOK model total soil moisture (mm) for the following run scenarios: Base precipitation with leaves (black solid line); Base precipitation without leaves (black dotted line); Doubled precipitation with leaves (blue solid line); Doubled precipitation without leaves (blue dotted line). ..............................................164

Figure 3.52 BROOK model top layer soil moisture (mm) for the following run scenarios: Base precipitation with leaves (black solid line); Base precipitation without leaves (black dotted line); Doubled precipitation with leaves (blue solid line); Doubled precipitation without leaves (blue dotted line). ..............................................165

Figure 3.53 Brook model daily ET components (mm day⁻¹) for September 23-26, 2003, using the following model run scenarios: 1XP,L: Base precipitation with leaves; 1XP,NL: Base precipitation without leaves; 2XP,L: Doubled precipitation with leaves; 2XP,NL: Doubled precipitation without leaves. ................................167
Figure 3.1L Map of the watersheds, weather stations and flux-measurement sites located in the Santarém region (STM) of LBA-ECO. ..................................................182

Figure 3.2L Diagram showing the stages of streamflow response to rainfall events. ...183

Figure 3.3L Photographs of the landscape surrounding the km117 weather station (a, top) and the Mojui weather station (b, bottom). ..................................................185

Figure 3.4L Top panel: CMORPH-estimated precipitation for the grid cell containing the Mojui weather station in the Mojui watershed (black solid line) during the 2004 dry season, from August to December. The Mojui weather station measured rainfall is the red dashed line. Middle panel: same as top except the blue dashed line is the Belterra weather station precipitation. Bottom panel: CMORPH cumulative rainfall (black line) for the same 2004 period. Mojui rainfall is in red, Belterra rainfall in blue, and the Mojui watershed average measured cumulative rainfall (the average of the Mojui and Belterra weather station rainfall) is in pink. ..........................................................187

Figure 3.5L Streamflow (reported as stream depth in cm) for the Mojui River, Rio Branco, and Moju River for August to December 2004. ..........................189

Figure 3.6L Streamflow recession times (days) for the Mojui, Branco, and Moju Rivers, August to December 2004. ..................................................190

Figure 3.7L Left panel: Average monthly soil moisture recession times in days for the Mojui watershed weather stations (Mojui and Belterra) and the km117 weather station, for the years 2000-2005. Medians are indicated by X, and the bars indicate the upper and lower quartiles. The number of recessions analyzed for each month are listed above the plot. The pink line is the average soil moisture (m$^3$ m$^{-3}$) at the weather stations. The red line is the soil moisture at the km77 site, and the green and blue lines are the km83 soil moisture at 10 and 40cm depth respectively. Right panel: Monthly mean latent heat flux (LE in W m$^{-2}$, black) and precipitation (mm month$^{-1}$, red) at the km77 site. ..................................................192
LIST of TABLES

Chapter 2:

Table 2.1 Interception component measurement and estimation methods used by the references listed in Figure 2.1 ................................................................. 21
Table 2.2 Precipitation events with available data by season........................................ 40
Table 2.3 Mean interception estimates for daytime rainfall events classed by rainfall rate............................................................................................................ 61
Table 2.4 Mean interception estimates for Penman-Monteith-filled daytime rainfall events classed by rainfall rate. .............................................................. 61
Table 2.5 Mean evaporative, sensible heat, and storage fractions of $-Q^*$ for dry and rain days for the pre-rain period 1200 – 1400 GMT and rain periods 1400 – 1800 GMT and 1800 – 2200 GMT. ......................................................... 66

Chapter 3:

Table 3.1 Schoharie Creek drainage basin rain-shadow coverage statistics for the September 23, 2003 precipitation event. .........................................................128
Table 3.2 Schoharie Creek drainage basin streamflow peak statistics for the September 23, 2003 precipitation event. .................................................................130
Table 3.3 Schoharie Creek drainage basin rain-shadow coverage statistics for the October 27-28, 2003 precipitation event. .........................................................146
Table 3.4 Schoharie Creek drainage basin streamflow peak statistics for the October 27, 2003 precipitation event. .................................................................148
Table 3.5 Average streamflow recession times from discharge data for the entire watershed network, as well as those watersheds in the peak-precipitation area, transition-precipitation area, and rain shadow region (in days, mean and standard error given) for the 2003 leaf period and the 2003 no-leaf period. .......... 158
Table 3.6 Streamflow peak statistics for the BROOK model runs. .............................163
Table 3.7 Soil moisture recession times (days) for the BROOK model runs. ...............164
Table 3.8 Top-layer soil moisture recession times (in days, mean values with standard errors in parentheses) during leaf and no-leaf periods for all observed HOBO weather stations (OBS: All); observed HOBO weather stations in sandy soil (OBS: Sand); observed HOBO weather stations in clay soil (OBS: Clay); BROOK model runs with base precipitation (BROOK 1XP); and BROOK model runs with doubled precipitation (BROOK 2XP). ........................................ 166
Appendix 1:

Table A1.1 NCAR-ISFF PAM station numbers, names, latitude, longitude, and elevation. .................................................................208

Table A1.2 HOBO station numbers, names, coordinates, and elevations for the HVAMS IFC period deployment. .................................................208

Table A1.3 NYCDEP station names, coordinates, and elevations. ......................208

Table A1.4 Location, elevation and watershed area of the USGS stream gauge sites used in this study......................................................209
Chapter 1: Introduction

1.1 Definitions and overview

In this thesis, hydrological and micrometeorological approaches are used to examine transient characteristics of the water balance components. The interception evaporation process is examined using the micrometeorological approach, employing a novel use of data-analysis methods using the eddy covariance technique. The hydrological approach is used in analyzing watershed responses to changes in precipitation inputs and land cover over space and time. Evapotranspiration (ET) is the process that links both of these approaches.

1.1.1 Surface water balance

The surface water balance used in hydrology is written as follows:

\[ ET = P - R - \Delta S \]  (1.1)

where \( ET \) is evapotranspiration, \( P \) precipitation, \( R \) runoff, and \( \Delta S \) the change in storage term.

Advantages of the hydrological budget approach include the precipitation \( P \) and runoff \( R \) being directly measured, and widely available. Disadvantages include that the evapotranspiration \( ET \) is not directly measured in this approach, but is instead found as a residual by assuming that annually \( \Delta S \) is equal to zero at a site, as is done in water-balance catchment studies (e.g. Bosch and Hewlett 1982; Hornbeck et al. 1993). Evapotranspiration in other studies has also been estimated by other means, such as by the Penman-Monteith equation (e.g. Stannard 1993). The effective spatial scale for this approach is from the small watershed (1-10 km²) to large watershed size (~500 km²).
1.1.2 Evapotranspiration components

The total evapotranspiration (ET) can be broken down into the following components:

\[ ET = Tr + E_i + E_s \]  

(1.2)

where \( Tr \) is transpiration, \( E_i \) interception evaporation, and \( E_s \) evaporation from the bare soil and forest floor litter. On an annual basis in a forested environment, transpiration is the dominant component of ET, followed by interception evaporation and then bare-soil evaporation (Fig. 1.1). However, during precipitation events, interception evaporation becomes the dominant component of ET during rainfall and during the hours immediately following precipitation.

**Components of Evapotranspiration (ET)**

![Diagram of components of evapotranspiration](image)

Figure 1.1: Components of evapotranspiration (ET) in the annual water balance of a mature mixed-species deciduous forest on the Maryland Coastal Plain. Values given are in percent of annual incident precipitation of about 100 cm (G. Parker, personal communication).
1.1.3 **Transient terms of the hydrological balance**

In this thesis, we focus on analysis of the transient terms in the hydrological balance. These include the runoff, or streamflow \((R)\); soil moisture, part of the storage term \(\Delta S\) in the surface water balance (equation 1.1); and the interception evaporation component \((E_i)\) of evapotranspiration (equation 1.2). All three terms show a similar response form to the transient forcing of a precipitation event (Fig. 1.2). In all cases, there is a sharp increase, or pulse in the measured variable in direct response to rainfall, whether it is an increase in latent heat flux due to interception evaporation (Fig. 1.2a), or an increase to a peak in soil moisture or streamflow, followed by a decline, which is called recession with regards to soil moisture and streamflow (Fig. 1.2b and Fig. 1.2c respectively).

**Forms of transient responses**

a) Interception evaporation \((E_i)\)

b) Soil moisture \((\Delta S)\)

c) Streamflow \((R)\)

Figure 1.2: Forms of transient responses to precipitation for: interception evaporation (a, top left), soil moisture (b, top right), and streamflow (c, bottom).
Changes in evapotranspiration resulting from altered land cover or leaf state modulate the responses of soil moisture and streamflow to rainfall with regards to two characteristics. First, soil moisture and streamflow recession times may change due to a change in land cover or leaf state. Second, the presence of near-stream transpiring vegetation in the growing season results in the appearance of diurnal streamflow fluctuations (Fig. 1.2c), providing an indicator of hydrologic function in a watershed.

1.1.4 Forest water budget

The forest water budget on the timescale of a precipitation event is expressed as follows:

\[ P = T_h + E_i + S_t \]  

(1.3)

where \( T_h \) is throughfall, and \( S_t \) is stemflow.

Here the incident precipitation \( P \) falls into the forest canopy and may reach the forest floor either directly as throughfall \( T_h \) or runs down the stems and trunks of the trees reaching the ground as stemflow \( S_t \). The precipitation that does not reach the ground at first is intercepted by the leaf surfaces and may fall off the leaf surface and eventually reach the ground, contributing to the throughfall term. However the majority of the water stored on the leaves will evaporate off the leaf surfaces as interception evaporation \( E_i \) until the forest canopy’s water storage capacity is reached.
Some of the net precipitation received at the forest floor (the sum of throughfall and stemflow) will be stored in the leaf litter and will be available for evaporation from the leaf litter or bare-soil evaporation $E_s$. The remaining water is available for infiltration $I$ into the soil:

$$I = Th + St - E_s$$  \hspace{1cm} (1.4)

The importance of interception on the timescale of a precipitation event on the water balance has been recognized and studied in depth at least since the early detailed observations and review by Horton (1919), who stated “It is evident that interception losses, which may amount to one-third or more of the precipitation, should not be disregarded in estimating run-off or yield of drainage basins”. It was also acknowledged that large variations may exist among interception measurements taken at sites located in similar forest environments, even within the same site. Horton (1919) further wrote “The subject is one which is somewhat difficult to experiment in a satisfactory manner, and it is not surprising that the conclusions hitherto drawn by different authorities are sometimes at variance, and many of the data are seemingly discordant”. These challenges still remain today (e.g. Manfroi et al., 2006), and provide a challenge for inputs to large-scale models.

### 1.1.5 Surface energy balance

The surface energy balance used in micrometeorology is expressed as follows:

$$A = -(Q* - G) = H + LE + S_{bc} + Adv$$  \hspace{1cm} (1.5)

where $A$ is the available energy, $Q^*$ the net radiation, $G$ the ground heat flux, $H$ the sensible heat flux, $LE$ the latent heat flux, $S_{bc}$ the biomass and air canopy storage
term, and \( Adv \) the advection term. The sign convention is upward fluxes are positive so that daytime \( Q^* \) is negative. The term linking the surface energy and water balances is evapotranspiration, expressed as \( LE \) in the surface energy balance (equation 1.5) and \( ET \) in the surface water balance (equation 1.1).

One advantage of the micrometeorological approach is that the latent heat flux \( LE \) is directly measured by the eddy-covariance method. Another advantage is that one can define in time or space how the ensemble average, or average of many events occurring under similar conditions, is formed. We introduce a novel use of the eddy-covariance approach in this thesis by determining the ensemble average in the form of \textit{event-based ensembles} based on days with and without rain. Composing the ensemble average in this way allows us to infer what would have happened had it not rained, which is an important consideration in determining interception evaporation using eddy covariance.

However, there are limitations to using the eddy-covariance method. These limitations include the failure of the method during calm nights with low turbulence (e.g. Goulden et al. 1996; Blanken et al. 1997). In addition, this method may also fail during some heavy rainfall periods due to water droplet collection on the sonic anemometer transducers used for the measurements (e.g. Mizutani et al. 1997). Also, the flux footprint may change with stability, wind speed and direction (e.g. Horst and Weil 1994; Finn et al. 1996; Schmid and Lloyd 1999). The effective spatial scale for this approach can be as large as the small watershed size (\(< 10 \text{ km}^2\)).
1.1.6 Spatial and temporal scales examined in this study

The link between both the hydrological-based and micrometeorological-based approaches is on the small watershed scale (Fig. 1.3). It is therefore useful to employ both approaches at this scale when assessing transient features of the hydrological balance.

Figure 1.3: Spatial and temporal scales of the processes analyzed and measurement methods considered in this thesis. Methods are indicated in italics; the main scales considered here are in bold.

The data analysis methods presented in the thesis take advantage of the following two ideas. First, nature effectively conducts a set of natural repeatable experiments that, when a sufficient amount of observational samples are gathered, are suitable to be used as case studies for ensembling. This natural, repeatable experiment may take the form of a day without precipitation in the dry season at a given location in the tropics, where there is little day-to-day variation in properties such as net radiation, timing of first cloud
appearance, amount of cloud cover, and the height of the lifting condensation level (LCL; see Fig 1.4). Furthermore, surface climate properties, such as temperature $T$, specific humidity $q$, and incoming shortwave radiation $K_{dn}$, induce feedbacks in the cloud environment that help to maintain a nearly constant LCL during the dry season in the tropics (Fig 1.4) or during the growing season in the mid-latitudes (Freedman et al. 2001).

Second, nature runs a series of real-world sensitivity studies that can be understood with careful analysis of observational data, sometimes by exploiting data in ways that may not have been intended when they were first collected. These observational sensitivity studies may take the form of the response to a natural transient event, one that may lead to well-defined changes in the energy partition for the duration of the event. Examples include a transient precipitation event over a forested site, which changes the surface water and energy balance partition at that site during and after the event. Another example is the abrupt change in leaf state that occurs during autumn leaf drop in mid-latitude regions with deciduous forests, and the sharp reduction in ET forcing.

Figure 1.4: Cloud base at the km67 old-growth rainforest site of the Large-Scale-Biosphere-Atmosphere Experiment in Amazonia (LBA) (black), lifting condensation level (LCL) at the LBA km67 site (blue), and LCL at the LBA km77 agricultural site (pink) during a dry season period in 2001 (October 2-12, days 275-285).
the site or watershed experiences as a result. When examined closely, signals exist in datasets such as streamflow records that are indicators of how a watershed responds to such forcing events (e.g. Federer 1973; Lundquist and Cayan 2002; Czikowsky and Fitzjarrald 2004). These transient events occur over a range of spatial and temporal scales. In this thesis, transient forcing caused by precipitation is examined over the event-to-synoptic time scale over the site-to-large watershed spatial scale (Fig. 1.3). The effect of change in leaf state due to phenology is studied on the seasonal time scale over the small-to-large watershed spatial scale. Land cover change effects due to reforestation or deforestation may also cover the successional time scale (Czikowsky, 2003).

1.1.7 Closing the water balance in climate models

Further motivation for this study on the smaller spatial and temporal scales comes on these fronts. First, global climate models do not close the water balance on regional scales. The importance of seasonal and successional vegetation effects while trying to connect atmospheric and hydrologic models from regional to global scales has been acknowledged (e.g. Betts and Ball 1999). Projects such as the Global Energy and Water Cycle Experiment (GEWEX) have attempted to close the water and energy budgets on these large spatial scales through observation and modeling (e.g. Stewart et al. 1998; Raschke et al. 2001). For example, Roads et al. (2002), using the NCEP-DOE Reanalysis II (NCEPRII) model, reported annual water budget errors as large as the associated runoff over some continental areas. To keep this model close to observations and from drifting into its own climate, soil moisture was added by adding the previous 5-day difference between the reanalysis precipitation and the observed precipitation to the top soil layer
(10cm), based on the predicted runoff. If the top soil layer saturated or dried, the lower soil layer was adjusted, lagged by 5 days. However in nature, this lag is not constant. This correction resulted in a surface water residual term that was as large as the associated runoff in midlatitude regions such as the Mackenzie Basin. The residual was largest in the late-winter, early-spring season and rapidly decreased in magnitude during the spring season, when there is a rapid ET increase with leaf emergence. It appears the model is drying the soil out too fast and not explicitly taking into account dynamic processes such as streamflow recession that are functions of bulk watershed parameters controlling soil moisture storage and transport and change with ET forcing and vegetation state. The failure of continental-scale climate models to close the water budget provides motivation for working on their improvement and understanding the underlying smaller-scale processes.

1.2 Outline and research questions

Observations of the transient characteristics of the surface water and energy balances to transient events are examined starting at small temporal and spatial scales in chapter 2, increasing in scale in chapter 3. In chapter 2, a new methodology for estimating interception evaporation at a site using micrometeorological measurements is introduced and described. This new approach addresses some of the shortcomings of existing interception-estimation techniques. An application of this method at an old-growth tropical rain forest site in Brazil operated during the Large-Scale-Biosphere-Atmosphere Experiment in Amazonia (LBA) is presented. The interception evaporation process considered in the chapter is studied at the site-to-small watershed spatial scale
over event and synoptic time scales (Fig. 1.3). Chapter 3 contains an observational analysis focusing on transient effects on the surface water balance over larger spatial and temporal scales. Studies were conducted in two locations: in a network of watersheds located in New York State in the domain of the Hudson Valley Ambient Meteorology Study (HVAMS), with the second location being a small network of watersheds located in the LBA operational domain. In both studies, watershed response is analyzed spatially in a network of small-to-large watersheds on a precipitation-event basis. Temporal changes in watershed response to precipitation events are also analyzed seasonally. In the HVAMS portion of the study the seasonal change takes the form of land-cover change and reduced evapotranspiration due to fall-season leaf drop. The BROOK watershed model (Federer 1995) is run to conduct a sensitivity study on watershed response to changes in input precipitation amount induced by orographic effects and changes in vegetation state due to phenology, and to assist in interpreting the observations. In the LBA portion of the study the seasonal change is the transition from the wet to dry season, resulting in a sharp decrease in precipitation input into the ecosystem. Chapter 4 provides an overview of the main conclusions of the thesis, with a discussion on outstanding issues and questions left for future work.

Specific questions to be addressed in the thesis include:

- Through the expansion of eddy flux-measurement networks such as Fluxnet (Baldocchi et al. 2001), the number and coverage of long-term eddy flux measurement sites has grown to over 460 worldwide sites, distributed over a wide range of land cover types (http://www.fluxnet.ornl.gov). What
additional information can be obtained from these sites beyond the standard reports of half-hour average fluxes?

- Under what conditions would micrometeorological-based interception evaporation measurements be expected to work, and under what conditions would such a technique be of limited use?

- Watersheds in the Catskill-Hudson Valley Region of New York State (HVAMS) receive much different inputs of precipitation due to orographic effects, and similar effects are seen in nearby upland regions of the Northeast United States (e.g. Brady and Waldstreicher 2001). As the frequency of extreme precipitation events may be expected to increase in this region as a result of climate change (e.g. Burns et al. 2007), such orographic effects on precipitation input may be enhanced. What is the watershed response to such rainfall events?

- Underlying the individual watershed response in the HVAMS region to differing amounts of precipitation is the seasonal land cover change due to transpiring vegetation in leaf during the growing season transitioning to leaf drop and the beginning of the dormant season, thereby changing the amount of ET forcing the watersheds receive. Can any detectable changes in watershed response be seen during such transition periods?

- Watersheds in the LBA portion of the study are located in differing landuse/landform settings, from undisturbed upland forest to a lowland setting under agricultural use. There is also an underlying seasonal change from the wet season to dry season in the region that changes the precipitation inputs the
watersheds receive. What changes in watershed response can be detected resulting from these factors?

1.3 References


Chapter 2: Detecting the evaporation of intercepted water over an old-growth rain forest in the eastern Amazon using eddy flux measurements

Abstract

We introduce and demonstrate a new method to estimate rainfall interception in an old-growth rain forest in the eastern Amazon using eddy covariance evaporation measurements. The approach is to estimate the ‘excess’ evaporation that occurs following individual events, using baseline evaporation time series obtained from long time series of flux data and creating ensemble averages from these precipitation events and base-state dry days. One advantage of this method over the traditional techniques of estimating interception using rain gauges alone is that the interception evaporation is directly measured and not determined as the residual of incident precipitation and throughfall. This method would also be useful in cases where rain gauge measures of precipitation are suspect, such as in fog or wind-driven conditions. Furthermore, the large differences in interception that can occur on a site due to varying forest canopy density, structure and the appearance of canopy gaps is smoothed out using the eddy covariance method as the size of the flux footprint area incorporates these variations, and provides an average interception value over the flux footprint area. Identification of light rainfall events not detected by an on-site tipping bucket rain gauge was aided by the use of a ceilometer. For daytime events, interception percentages decrease with rainfall intensity, with mean interception for light (0-2 mm hr\(^{-1}\)), moderate (2-16 mm hr\(^{-1}\)), and heavy (> 16 mm hr\(^{-1}\)) rainfall-rate events being 18.0, 9.9, and 7.8 percent of incoming precipitation respectively. The mean interception for all events in the study (daytime and
nighttime) was 11.6%. Energy balance comparisons between dry and afternoon rain-days show an approximately 15% increase of evaporative fraction on the rain days, with the energy being supplied by a corresponding decrease in the canopy heat storage. This method may be applicable to other tower sites worldwide in varying types and climates as net radiation is used to scale the evaporation in this method.

2.1 Introduction

2.1.1 The interception process

Interception of rainfall by the forest canopy and the subsequent evaporation into the atmosphere constitute an important part of the hydrological balance over forests. On an annual basis in a forest environment, transpiration is the dominant component of evapotranspiration (ET), followed by interception evaporation and then bare-soil and litter evaporation. However, during and following transient precipitation events, re-evaporation of intercepted water exceeds transpiration as the dominant component of ET, resulting in a shift in the hydrological balance. During the process of interception evaporation, the leaves are wet, increasing the stomatal conductance depending on the fraction of the canopy that is wet. Under such conditions, when surface (physiological) controls are reduced, enhanced rates of evaporation of intercepted water can be expected from forests compared to shorter vegetation, in all climatic zones (Newson and Calder 1989). Evaporation from a wet forest canopy can proceed at a greater rate than a dry one, up to five times that of the transpiration of surface-dry vegetation (Hewlett 1982). During interception-loss periods, two-thirds of total ET can be evaporation of intercepted water from the leaf surfaces (Stewart 1977).
An appreciable fraction of water vapor in the Amazon is recycled through ET, with 25 to 50 percent of Amazon precipitation having been previously evaporated from the forest (Salati and Vose 1984; Eltahir and Bras 1994; Hutyra et al. 2005). Lawrence et al. (2007) estimated the annual evapotranspiration partition over the Amazon to be 58% transpiration, 33% interception, and 9% soil evaporation. Thus the interception evaporation process is a critical part of the water budget for the Amazon.

2.1.2 Definitions

The surface energy balance is expressed as follows:

\[ A = -(Q^* - G) = H + LE + S_{bc} + Adv \]  

(2.1)

where \( A \) is the available energy, \( Q^* \) the net radiation, \( G \) the ground heat flux, \( H \) the sensible heat flux, \( LE \) the latent heat flux, \( S_{bc} \) the biomass and air canopy storage term, and \( Adv \) the advection term. The sign convention is upward fluxes are positive.

The surface water balance is written as follows:

\[ ET = P - R - AS \]  

(2.2)

where \( ET \) is evapotranspiration, \( P \) precipitation, \( R \) runoff, and \( AS \) the change in storage term. The term linking the surface energy and water balances is evapotranspiration, expressed as \( LE \) in the surface energy balance and \( ET \) in the surface water balance.

The components of evapotranspiration (ET) are as follows:

\[ ET = Tr + E_i + E_s \]  

(2.3)

where \( Tr \) is transpiration, \( E_i \) interception evaporation, and \( E_s \) the evaporation from the bare soil and forest floor litter.
During a rainfall event over a forest, precipitation \( (P) \) either:

(a) falls through the canopy and reaches the ground as throughfall \( (Th) \);

(b) may be caught by leaf surfaces and then fall to the ground, contributing to the throughfall;

(c) may be caught by tree branches and stems and be routed down the tree trunks to the ground as stemflow \( (St) \);

(d) may be caught by the forest canopy to be temporarily stored and then evaporated back into the atmosphere as interception evaporation \( (E_i) \).

Therefore, the forest water budget with respect to a rainfall event may be expressed as follows:

\[
P = Th + E_i + St
\]

\[\text{(2.4)}\]

2.1.3 Conventional interception-measurement methods

The most commonly-used method to estimate interception is to set up a series of rain gauges, one or more at or above the top of the forest canopy and/or in a nearby clearing to catch the total incident precipitation and several at the forest floor to measure the precipitation reaching the ground, known as throughfall. The forest-floor rain gauges may remain in a fixed location or be periodically relocated (roving) to attempt to improve spatial representation of throughfall. Troughs have also been deployed at the forest floor to catch throughfall. Stemflow is collected through the use of collars placed around the tree stems and then routed into collector bins. The interception is not directly measured,
but found as the residual of the total incident precipitation and the sum of throughfall and stemflow (if measured) and is usually expressed as a percentage of total precipitation.

The models most commonly used to calculate interception at a site are the Rutter et al. (1971, 1975) numerical model and Gash analytical model (Gash 1979; Gash et al. 1995). These models use the Penman-Monteith equation to calculate wet-canopy evaporation (Monteith 1965). All require a value for the canopy storage capacity, a quantity that is commonly estimated through linear regression analysis of an interception (or throughfall) vs. precipitation scatterplot. However, these canopy-storage estimates have been found to underestimate directly observed canopy water storage by a factor of two (Klaassen et al. 1998). A listing of the conventional methods used in published interception studies in tropical rain forest regions is shown in Table 2.1.
Table 2.1: Interception component measurement and estimation methods used by the references listed in Figure 2.1. The italicized reference numbers are as in Figure 2.1.

<table>
<thead>
<tr>
<th>Reference</th>
<th>Gross precipitation measurement method</th>
<th>Throughfall measurement method</th>
<th>Stemflow measurements available (Yes/No)</th>
<th>Model used</th>
</tr>
</thead>
<tbody>
<tr>
<td>Franken et al. (1982a) (1B)</td>
<td>Chart-recording rain gauge</td>
<td>Fixed funnel rain gauges, recorded weekly</td>
<td>Yes Stemflow collectors used following Lima (1976)</td>
<td>None</td>
</tr>
<tr>
<td>Franken et al. (1982b) (2B)</td>
<td>Funnel rain gauge</td>
<td>Fixed funnel rain gauges, recorded weekly</td>
<td>No</td>
<td>None</td>
</tr>
<tr>
<td>Schubart et al. (1984) (3B)</td>
<td>Funnel and chart-recording rain gauges</td>
<td>Funnel rain gauges (no other details specified)</td>
<td>No</td>
<td>None</td>
</tr>
<tr>
<td>Leopoldo et al. (1987) (4B)</td>
<td>Rain gauges (type unspecified)</td>
<td>Rain gauges (type unspecified)</td>
<td>Yes Stemflow collectors used following Lima (1976)</td>
<td>None</td>
</tr>
<tr>
<td>Lloyd and Marques (1988) (5B)</td>
<td>Tipping-bucket rain gauge</td>
<td>Roving tipping-bucket and bottle rain gauges</td>
<td>Yes Gash (1979) analytical model; Rutter et al. (1971,1975) model</td>
<td>None</td>
</tr>
<tr>
<td>Imbach et al. (1989) (6C)</td>
<td>Chart-recording rain gauge</td>
<td>Fixed troughs, recorded after each rainfall event</td>
<td>No</td>
<td>None</td>
</tr>
<tr>
<td>Ubarana (1996) (7B, 8B)</td>
<td>Tipping-bucket rain gauge</td>
<td>Roving funnel rain gauges, recorded weekly</td>
<td>Yes</td>
<td>Rutter et al. (1971,1975) model</td>
</tr>
<tr>
<td>Study</td>
<td>Rain Measurement Details</td>
<td>Fixed Rain Gauges</td>
<td>Stemflow Collectors</td>
<td>Rainfall Estimation Model(s)</td>
</tr>
<tr>
<td>---------------------------</td>
<td>------------------------------------------------------------------------------------------</td>
<td>-------------------</td>
<td>---------------------</td>
<td>-----------------------------</td>
</tr>
<tr>
<td>Cavelier et al. (1997)</td>
<td>Funnel rain gauge, recorded daily; chart-recording rain gauge</td>
<td>Yes</td>
<td>None</td>
<td></td>
</tr>
<tr>
<td>Arcova et al. (2003)</td>
<td>Rain gauge (type unspecified)</td>
<td>Yes</td>
<td>None</td>
<td>Stemflow collectors used following Likens and Eaton (1970)</td>
</tr>
<tr>
<td>Ferreira et al. (2005)</td>
<td>Funnel rain gauges</td>
<td>No</td>
<td>None</td>
<td></td>
</tr>
<tr>
<td>Manfroi et al. (2006)</td>
<td>Tipping-bucket rain gauge</td>
<td>Yes</td>
<td>None</td>
<td></td>
</tr>
<tr>
<td>Holwerda et al. (2006)</td>
<td>Tipping-bucket rain gauge</td>
<td>Yes</td>
<td>None</td>
<td></td>
</tr>
<tr>
<td>Germer et al. (2006)</td>
<td>Troughs (manually read and with tipping-bucket)</td>
<td>Yes</td>
<td>Revised Gash et al. (1995) model</td>
<td></td>
</tr>
<tr>
<td>Cuartas et al. (2007)</td>
<td>Tipping-bucket rain gauge (recorded every 5 or 30-minutes)</td>
<td>Yes</td>
<td>Revised Gash et al. (1995) model; Rutter et al. (1971,1975) model</td>
<td></td>
</tr>
</tbody>
</table>
Other methods of indirectly estimating interception include the use of a loadcell-based weighing system for measuring precipitation and throughfall (Lundberg et al. 1997), use of strain gauges to measure the amount of water intercepted by individual branches (Huang et al. 2005), and the use of microwave-transmission techniques to measure canopy water storage (Bouten et al. 1991). However, these have not been widely used in prior studies conducted in tropical rain forests.

Past studies of interception in tropical rain forest sites using conventional methods have yielded a wide range of interception estimates, from 8 to nearly 40 percent of total annual precipitation (Fig. 2.1). Furthermore, large annual interception differences can be found within plots in the same forest. Manfroi et al. (2006) reported interception estimates ranging from 3 to 25% in 23 subplots over a 4-ha area, depending on the canopy structure and density of the subplots.
Figure 2.1: Interception estimates reported in the literature using conventional methods for tropical rain forest sites. Studies done in Brazil are labeled in black with ‘B’, Central America in pink with ‘C’, Malaysia in green with ‘M’, Australia in blue with ‘A’, and Puerto Rico in red with ‘P’. References are as follows: 1B, 2B: Franken et al. (1982a, b); 3B: Schubart et al. (1984); 4B: Leopoldo et al. (1987); 5B: Lloyd and Marques (1988); 6C: Imbach et al. (1989); 7B, 8B: Ubarana (1996); 9C: Cavelier et al. (1997); 10B: Arcova et al. (2003); 11B: Ferreira et al. (2005); 12M: Manfroi et al. (2006); 13A: Wallace and McJannet (2006); 14P: Holwerda et al. (2006); 15B: Germer et al. (2006); 16B: Cuartas et al. (2007).
Given typical heterogeneous, complex canopy and subcanopy structure and the random appearance of canopy gaps seen in tropical rain forests (Fig. 2.2), conducting throughfall measurements to deploy the appropriate number and distribution of gauges to accurately sample the inherent spatial variation in throughfall to obtain a representative area-average is a major challenge. For example, Kimmins (1973) reported that up to 100 or more rain gauges would be required to reduce the error in mean throughfall to below 5 percent at the 95% confidence interval. Roving rain-gauge setups help in reducing error in estimating throughfall. Deploying approximately 30 to 50 rain gauges over a concentrated area (such as 100 m x 100 m) and relocated weekly for a time period of a year or longer, estimated mean throughfall error has been reported to be below 5 percent (Lloyd and Marques 1988; Ubarana 1996). However, the necessary relocations are intensive and may be impractical, especially in remote locations for long periods of time.

Troughs can be used to reduce the number of collectors needed to sample spatial variation in throughfall. Crockford and Richardson (1990) in a comparison of troughs to rain gauges concluded that the number of troughs deployed could be reduced by about one-fifth as opposed to the number of rain gauges to obtain the same error in mean throughfall. However, troughs also require maintenance, and suffer from splash-out and greater adhesive losses as opposed to rain gauges (Lundberg 1997). Adhesive losses refer to the water collected on the trough side walls and evaporated without reaching the bottom of the trough.
Figure 2.2. Height sections of canopy surface area density along six 500m transects at an intact old-growth forest site, the km67 site in the Santarem region of the Large-Scale Biosphere-Atmosphere Experiment in Amazonia (LBA-ECO). (G. Parker, personal communication).
2.1.4 *Interception estimation in global climate models*

There has been a high interception bias reported in estimates from global climate models (e.g. Fernandes et al. 2008; Wang et al. 2007; Lawrence et al. 2007). Identified factors that have produced large errors in interception estimates include a parameter that represents the fractional area of a leaf that collects water (Lawrence et al. 2007), and subgrid-scale precipitation coverage (Wang et al. 2007). Accounting for these factors has led to some improvement. For example, after accounting for subgrid-scale precipitation coverage and precipitation type when calculating interception in the Community Land Model, Version 3 (CLM3), Wang et al. (2007) reported a decrease in estimated interception over the Amazon from 45 percent to 24 percent of annual precipitation. Although much improved, this estimate is still greater than most of the published observational interception estimates in this region (Fig. 2.1).

Incorrect estimation of interception in land-surface models has consequences for other hydrological-balance components and quantities in these models. For example, Fernandes et al. (2008), in running surface water budgets with the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis (ERA40) over the Amazon River basin, reported that a high model interception bias contributes to higher rainy-season evapotranspiration than observed and reduces the wet-season storage to the point that the soil moisture reservoir is not fully recharged by the end of the wet season. Consequently, there is an insufficient soil moisture supply in the model to start the dry season, and a large amount of soil moisture nudging in the reanalysis to compensate. Other reanalysis models in simulating large-basin scale water budgets suffer from the necessity to have a large soil moisture nudging term to stay close to observations and
keep from drifting into an unrealistically dry climate (e.g. Roads et al. 2002, NCEP-DOE Reanalysis II).

2.1.5 *A new technique for measuring interception*

We introduce and describe a new, alternate method for observing interception using eddy-covariance data that could be applied to other tower flux sites worldwide in varying forest types. The approach is to estimate the ‘excess’ evaporation that occurs during and following individual events, using baseline evaporation time series obtained from long time series of flux data (Fig. 2.3). To achieve this, we combine the eddy-covariance technique with a novel use of data-analysis methods.

![Diagram illustrating the method used to estimate interception using eddy covariance. A base state ensemble LE is composed using dry days. The interception loss for a precipitation event is the difference between the base state and event LE.](image)

Rainfall over a forest perturbs the energy and hydrological budgets during and following a rain event due to interception evaporation. One advantage of using this micrometeorological-based approach is that one can define in time or space how the
ensemble average is formed. We introduce a novel use of the eddy-covariance approach in this chapter by determining the ensemble average in the form of two *event-based ensembles*, collected from data over one forest, to quantify this perturbation in the form of the excess evaporation that occurs as a result of the precipitation falling into and being temporarily stored in the forest canopy. The base state ensemble is composed of days without rain, under the same radiative conditions as rain-days. The precipitation event ensemble is composed of days with rain. One assumption made is that transpiration does not stop during and following the rainfall event, so the difference between the base state and precipitation event ensembles represents interception evaporation. Composing the ensemble averages in this way allows us to infer what would have happened had it not rained, which is an important consideration in determining interception evaporation using eddy covariance.

Another advantage of this method over the traditional techniques of estimating interception is that the interception evaporation is directly measured and not determined as the net precipitation, the residual of incident precipitation and throughfall and stemflow. Furthermore, the large differences in interception that can occur on a site due to varying forest canopy density, structure and the appearance of canopy gaps is smoothed out using the eddy covariance method as the size of the flux footprint area incorporates these variations, and provides an average interception value over the flux footprint area. This also provides a more suitable input for models requiring such data.

Fernandes et al. (2008) state that interception estimates are not available from conventional observations on the basin scale for land-surface model comparisons. The new technique outlined in this chapter could improve this situation.
Savenije (2004) argues that there is a broader definition for interception than just the difference between total precipitation and the sum of throughfall and stemflow. Interception also includes the part of the rainfall captured by the ground surface that is evaporated before it can take part in any subsequent runoff, drainage or transpiration processes. Therefore, traditional interception estimates based on net precipitation would be biased low since the wet-surface evaporation contribution to the total interception was neglected. In estimating interception using the eddy covariance method, the total evaporation is measured. Thus, both the interception evaporation contributions from the wet forest canopy and the wet ground surface are included in the measurement.

2.1.6 Outline

In this chapter, a new methodology for estimating interception evaporation at a site using micrometeorological measurements is introduced and described. This new approach addresses some of the shortcomings of existing interception-estimation techniques. An application of this method at an old-growth tropical rain forest site in Brazil operated during the Large-Scale-Biosphere-Atmosphere Experiment in Amazonia (LBA) is presented. The interception evaporation process considered in the chapter is studied at the site-to-small watershed spatial scale over event and synoptic time scales. The interception evaporation process is examined using the micrometeorological approach, employing a novel use of data-analysis methods using the eddy covariance technique.

First, we introduce the study area and instrumentation used in the study. Next, we discuss the data analysis methods employed, starting with rainfall-event identification
methods followed by flux calculation and ensemble formation techniques. We then
discuss the methods used to estimate interception during nighttime and daytime rainfall
events. This is followed by an energy-balance comparison for dry and wet days. Finally,
we conclude with a comparison of our results to the conventional results reviewed in the
introduction.

Specific questions to be addressed in this chapter include:

- Through the expansion of eddy flux-measurement networks such as Fluxnet
  (Baldocchi et al. 2001), the number and coverage of long-term eddy flux
  measurement sites has grown to over 460 worldwide sites, distributed over a
  wide range of land cover types (http://www.fluxnet.ornl.gov). What
  additional information can be obtained from these sites beyond the standard
  reports of half-hour average fluxes?
- Under what conditions would micrometeorological-based interception
  evaporation measurements be expected to work, and under what conditions
  would such a technique be of limited use?

2.2 Location and Data

The data used in this study were collected in an old-growth forest site that was
operated as part of the Large-Scale-Biosphere-Atmosphere Experiment in Amazonia
(LBA-ECO, km67 site). This site is located in the Tapajos National Forest south of
Santarém, Brazil in the eastern Amazon region (Fig. 2.4). The site coordinates are
2.88528°S, 54.92047°W at an elevation of 117 m. The height of the forest canopy at the site is approximately 43 m.

An eddy-covariance system that included a Campbell CSAT 3-D sonic anemometer (Campbell Scientific, Inc.) and a Licor 6262 CO$_2$/H$_2$O analyzer was operating at a frequency of 8 Hz at a height of 57.8 m, near the top of the flux tower at the km67 site. Net radiation was measured at 64.1 m height using a Kipp and Zonen
CNR1 net radiometer, which measured the upward and downward longwave and shortwave radiation components separately. A tipping bucket rain gauge was installed at a height of 42.6 m on the tower, and reported precipitation at 1-minute intervals with 0.1 mm resolution. A Vaisala CT-25K laser ceilometer was operating at the site from April 2001 to July 2003. Along with cloud base measurements, the ceilometer provided 15-second measurements of a backscatter profile from the surface to 7500 m at 30-m resolution. Cloud cover fraction was obtained by the fraction of time the ceilometer reported cloud base. The presence of forced cumulus clouds was identified by noting the proximity of the surface lifting condensation level (LCL) to cloud base. Temperature and humidity profile measurements were also taken at eight heights spanning the tower. Further site details can be found in Hutyra et al. (2005) and Saleska et al. (2003).

A large number of precipitation events need to be analyzed under similar conditions to form a sufficient ensembles for this approach to work. One advantage of using this approach to estimate interception at this site is that there is a marked diurnal pattern in precipitation and cloudiness, especially in the dry season. Precipitation frequently occurs during the same times of the day, helping to build a large ensemble of similar cases. At the km67 site there is an afternoon convective peak in rainfall in both the dry and wet seasons, and a nighttime synoptic peak in the wet season (Fig. 2.5).
Boundary layer cumulus clouds regularly form during the dry season in late morning and dissipate after nightfall (Fig. 2.6, 2.7), aiding to form a large ensemble of dry-day latent heat flux. Furthermore, there is little day-to-day variation in cloud fraction and cloud base during the dry season (Fig. 2.6, 2.7).
Figure 2.6: Median cloud cover fraction at km67 measured by the ceilometer by hour of day for the wet season (February through May, plotted in blue) and the dry season (September through December, plotted in black) for 2001 to 2003. The quartiles are indicated by the bars. Note the presence of convective cloudiness during the day in the dry season and the absence of clouds at night in the dry season. Cloud cover fraction peaks during the morning in the wet season.
Figure 2.7: Top panel: Cloud base at km67 (black), lifting condensation level (LCL) at km67 (blue), and LCL at km77 (pink) during a wet season period in 2001 (May 2-11, days 122-131). Bottom panel: As in top panel but for a dry season period in 2001 (October 2-12, days 275-285).

2.3 Methods

2.3.1 Identification of rainfall events

Precipitation events were identified two ways: using the rain gauge and the ceilometer backscatter profile. First, a storm separation time needed to be chosen so that a clear start and end time could be defined for each rainfall event. This is important because a rainfall event is often composed of many irregularly-spaced rainfall tips. A rainfall tip is defined as the amount of water required to activate a counter on the rain gauge (0.1 mm in this study). The length of the storm separation time should be long enough that the precipitation event has finished and the canopy has ample time to dry (provided it is daytime). The storm separation time should not be too long as to combine rainfall of two separate events if the canopy had dried in the interim. We chose a 4-hour
storm separation time in our analysis, a value successfully used by Wallace and McJannet (2006) in an Australian rain forest and Van Dijk et al. (2005) in a West Javan rain forest. This separation time worked well at this site given the regularity of the daily timing of the precipitation at this site.

From the 1-minute precipitation data from the rain gauge, a precipitation event was identified in the following manner. The precipitation file was scanned until the first rain tip was found, which constituted the rain event start time. The rain event end time was defined as the time that a rain tip was reached where there were no further tips for the following 4 hours. This process was repeated for all rain events.

Ceilometers have been used to observe boundary-layer aerosols (Zephoris et al. 2005). We found that the ceilometer is also able to detect rain droplets (diameters 0.5 – 6.0 mm) quite well through the use of the ceilometer backscatter profile (Fig. 2.8). Due to the large amounts of data contained in each backscatter profile, we averaged the raw 15-second ceilometer data to 5 minutes to perform the rain-identification analysis. This had little impact on the event fluxes calculated later since the minimum flux-calculation length used was 15 minutes.
Figure 2.8: Raw ceilometer backscatter (15-second samples) from 1300 to 1800 LT on December 10, 2001 at the LBA km67 site. Backscatter units are log(10000*srad*km)^-1. Red dots indicate cloud bases (m). The pink line is the incoming shortwave radiation (Sdown, units of W m^-2). The light blue line is the photosynthetically active radiation (PARdown, units of μmol m^-2 s^-1). Precipitation fell during two periods. The first event occurred in the early afternoon from 1325 to 1400 LT. A second, lighter rain shower occurred for a brief period from 1640 to 1655 LT. The on-site rain gauge recorded 0.76 mm of precipitation for the first rain event, but none for the second rainfall.

The same storm separation time and scanning method were used with the ceilometer data as with the rain gauge data, but two additional things needed to be determined. These were the threshold value for precipitation, and the heights at which to average the backscatter profile. Based on the review of many days of rainfall, possible values for the rain identification threshold were between 1.2 and 1.5 (units of backscatter are log(10000*srad*km)^-1). Histograms of backscatter intensity showed the largest decrease in the number of observed backscatter intensities between 1.2 and 1.3 (not shown). This is an indicator of the rain threshold, since most of the time it is not raining. The rain threshold value of 1.3 log(10000*srad*km)^-1 was chosen because of the result of
the histogram inspection and because this value agreed best with visual inspection of the ceilometer records for rainfall. A plot of ceilometer backscatter data with the range of rain-identification thresholds and rain gauge data is shown (Fig. 2.9).

Figure 2.9: Rainfall (mm, solid line at bottom), and average ceilometer backscatter up to half of the cloud base height (units of $\log(10000*srad*km)^{-1}$, dashed line) for days 338 to 345 in 2001. The horizontal solid lines indicate the range of rain-identification threshold values used for the ceilometer backscatter data.

Before the ceilometer data could be used for identifying rainfall events, the range of backscatter profile heights to average needed to be determined to ensure backscatter returns from clouds were not included in the rainfall-identification process. We averaged the ceilometer backscatter profile three ways. First, the lowest 90 m (three range gates) of the backscatter profile were averaged. Second, the backscatter profile was averaged from the lowest level to 50 percent of the cloud base height. Third, the backscatter profile was averaged from the lowest level to 75 percent of the cloud base height. Using the lowest 90 m only for the average backscatter led to false rainfall returns due to fog or smoke. Using the backscatter profile averaged to 75 percent of the cloud base height also led to false rainfall returns due to clouds. Averaging the backscatter profile up to 50 percent of the cloud base height yielded the best results when comparing to events identified by the ceilometer using this criterion to events recorded by the rain gauge.
One advantage of using both the ceilometer backscatter data and rain gauge to identify precipitation events over the rain gauge alone is that the ceilometer detects all rainfall events, including light ones when the rain gauge may not catch any rainfall or not enough to force a tip. Second, the ceilometer gives the instantaneous start time for rainfall, whereas with the tipping bucket rain gauge, light precipitation may have been falling for several minutes before a tip is recorded. The two methods differ in sensitivity to rain and to its timing.

A total of over 200 events were identified using the tipping bucket rain gauge over the April 2001 to July 2003 time period (Table 2.2). The on-site ceilometer detected nearly 40 light precipitation cases in the dry season that were not detected by the tipping bucket rain gauge.

Table 2.2: Precipitation events with available data by season. The wet season is defined as the months January through June, the dry season July through December. Both daytime and nighttime events are included.

<table>
<thead>
<tr>
<th></th>
<th>Wet</th>
<th>Dry</th>
<th>All</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tipping Bucket (2001-03)</td>
<td>143</td>
<td>63</td>
<td>206</td>
</tr>
<tr>
<td>Ceilometer (2001-02)</td>
<td>80</td>
<td>102</td>
<td>182</td>
</tr>
</tbody>
</table>

2.3.2 Flux calculation methods

The latent heat flux $LE$ and the sensible heat flux $H$ were determined by the eddy covariance method using the following:

$$LE = \rho L_e w'q'$$  \hspace{1cm} \text{(2.5)}

$$H = \rho C_p w'T'$$  \hspace{1cm} \text{(2.6)}
where $L_v$ is the latent heat of vaporization, $C_p$ the specific heat of air, and $\rho$ the air density; $\overline{w'q'}$ and $\overline{w'T'}$ are the latent and sensible kinematic heat fluxes, and the overbars indicate Reynolds averaging. In the Reynolds averaging procedure, the variables to be averaged are separated into mean and turbulent components (denoted by the primed variables in Eq. 2.5 and 2.6) following a set of rules (e.g. Kaimal and Finnigan 1994).

In other studies using eddy covariance in wet canopy conditions, the sensible heat flux was directly measured along with the net radiation, but the latent heat flux was not directly measured (van der Tol et al. 2003; Herbst et al. 2008). It was determined as a residual in the energy balance. Here we directly measure the latent heat flux through eddy covariance.

Following standard practice, four mean-removal methods were initially employed to make the Reynolds average for the eddy flux calculation: block-average, linear trend removal, centered running mean removal, and smoothed mean removal. Procedures for performing these mean-removal methods are found in e.g. Kaimal and Finnigan (1994). These methods have been used in standard practice in the flux measurement community, including over flux tower-measurement networks such as AmeriFlux (e.g. Massman and Lee 2002). The block average, linear trend, and centered running mean removal calculations follow that of Sakai et al. (2001); see also Kaimal and Finnigan (1994). The smoothed mean removal employed here uses a locally-weighted regression smoothing function, run in the Splus software package as the function supsmu (Mathsoft, Inc.; function details given in Fitzjarrald et al. 2001) to detrend the time series.

Raw data points that were recorded during calibration cycles, data points out of range for the sonic anemometer, and points with missing data were flagged. Any flux-
calculation period with greater than 2 percent of its raw data points identified as flagged points was discarded from the analysis.

The sensible and latent heat fluxes ultimately used in the analysis are the average of the smoothed mean removal, linear trend removal, and running mean removal methods. The block-averaged method was very sensitive to the flagged points in the data, whereas the other three methods were much more insensitive to flagged points.

Fluxes were initially calculated at 15-minute and 30-minute intervals. The 15-minute fluxes were used for further analysis for two reasons. First, the 30-minute fluxes are insufficient to fully resolve event detail, given the transient nature of the evaporation pulses that occur during a precipitation event. Second, the use of a 30-minute averaging window for the fluxes resulted in larger quantities of good data being discarded surrounding calibration periods opposed to using a 15-minute averaging window.

For each 15-minute period the friction velocity \( u \) was calculated along with the mean and standard deviations of the net radiation \(-Q^*\), wind speed, temperature, and humidity. The biomass and canopy air storage term \( S_{bc} \) in the energy balance was calculated following the empirical relation of Moore and Fisch (1986) developed in a similar rain forest setting in Manaus, Brazil. The same relation was also used by da Rocha et al. (2004) at a rain forest site (km83 site of LBA-ECO) near the location of this study. The \( S_{bc} \) term was calculated as:

\[
S_{bc} = 16.7 \Delta T_r + 28.0 \Delta q_r + 12.6 \Delta T_r^* \tag{2.7}
\]

where \( \Delta T_r \): hourly air temperature change (C)

\( \Delta q_r \): hourly specific humidity change (g/kg)

\( \Delta T_r^* \): 1-hour lagged hourly air temperature change (C)
2.3.3 Flux datasets and ensemble formation

Taking advantage of the regularity of dry-day weather conditions in this region, we define the composition of the ensemble average in terms of an event-based ensemble, based on whether or not precipitation fell on a given day. Days without rainfall and with sufficient data were composed to form an ensemble average representing the “base state” or baseline latent heat flux; the evapotranspiration that would have occurred had precipitation not fallen (Fig. 2.3). Likewise, ensemble averages were formed using days with rainfall at any time of day and afternoon rainfall (to separate afternoon convective precipitation from nighttime squall-line-associated rainfall) to represent the baseline latent heat flux for those days.

The effect of wet-canopy transpiration on the base state latent heat flux is minimal, since the evapotranspiration from wet-canopy surfaces is several times than the transpiration alone (Larsson 1981 and references therein).

To facilitate the composition of these ensembles, two flux datasets were created, each using different sets of starting and ending times for the flux-calculation periods. In the first dataset, fluxes were calculated for consecutive 15-minute periods for the entire dataset, with the first calculation period of each day beginning at midnight, regardless of the occurrence of precipitation events. This flux dataset was used in the composition of the dry-day, rain-day, afternoon rain-day baseline ensembles. The dry-day baseline ensembles for $LE$ and $-Q^*$ are shown (Fig. 2.10).
In the second dataset, fluxes were calculated relative to the timing of each precipitation event. The starting flux-calculation time \( t=0 \) for a given precipitation event depended on the manner which the event was detected. For rain-gauge recorded events, \( t=0 \) was the time of the first recorded tip by the tipping bucket rain gauge. For ceilometer-detected events, \( t=0 \) was the time of the first ceilometer backscatter return detected that was beyond the threshold backscatter value. For events detected by both the rain gauge and the ceilometer, the rain gauge event times were used. Fluxes were calculated at consecutive 15-minute intervals for each event starting four hours before the start of the event until four hours after the end of the event, which was defined as the time of the last recorded tip by the rain gauge or the last detected above-threshold ceilometer backscatter return. The precipitation event flux ensembles described below were calculated using this dataset.
2.3.4 Nighttime rainfall event methods

For nighttime cases, the situation is simpler because the base state LE is nearly zero at night (Fig. 2.11). Therefore, the nighttime portion of event LE can be integrated directly and converted to an equivalent water depth. The remaining amount of water stored in the canopy that does not get evaporated the night of the event will evaporate the following morning, and this portion has to be addressed separately.

![Diagram illustrating the method used to determine nighttime interception losses.](image)

Figure 2.11: Diagram illustrating the method used to determine nighttime interception losses.

The individual event departures from the base state LE (the interception losses) were used to form an ensemble average of interception evaporation occurring during nighttime rainfall events with respect to the starting time of each rain event.

2.3.5 Individual daytime event LE baseline determination

For daytime events, the process of determining the interception evaporation is more complex because the base state LE is not zero during the daytime (Fig. 2.3). To determine the baseline dry-day LE for an individual rainfall event, the net radiation must be taken into account. The net radiation for a given rainfall event is less than what
would be observed on a dry day at the same time of day (Fig. 2.12). The dry-day baseline LE should represent the latent heat flux that would occur on a dry day under the same radiative conditions as a day with rain. Three methods are outlined below to determine the dry-day baseline LE. All methods start with the ensemble LE for all dry days ([LE]_{dry}). The brackets indicate the ensemble average was taken.

**Method 1**

Divide the dry-day ensemble LE ([LE]_{dry}) by the dry-day ensemble –Q* ([\mathbf{-Q^*}]_{dry}) to get the dry-day ensemble evaporative fraction ([EF]_{dry}) for the corresponding time of day covering the precipitation event. Then for each data point during the event, multiply the dry-day ensemble evaporative fraction by the event –Q* (-Q*_{ev}) to arrive at the baseline LE:

\[
[LE]_{baseline} = [EF]_{dry} \times -Q^*_{ev}
\]  

(2.8)

**Method 2**

Divide the dry-day ensemble LE ([LE]_{dry}) by the dry-day ensemble –Q* ([\mathbf{-Q^*}]_{dry}) to get the dry-day ensemble evaporative fraction ([EF]_{dry}) for the corresponding time of day covering the precipitation event. Then for each data point during the event, multiply the dry-day ensemble evaporative fraction by the rain-day ensemble –Q* ([\mathbf{-Q^*}]_{rain}) for the same time of day to arrive at the baseline LE:

\[
[LE]_{baseline} = [EF]_{dry} \times [-Q^*]_{rain}
\]  

(2.9)
Method 3: Divide the mean of the event –Q* (–Q*ev) by the mean of the dry-day baseline –Q* ([–Q*]dry) for the time of day of the precipitation event to get the radiative fraction (–Q*frac) for the corresponding time of day covering the precipitation event. Multiply this event radiative fraction by the raw dry-day baseline LE ([LE]dry) for the same time of day to get the baseline LE:

\[
-\text{Q*}_{\text{frac}} = \frac{\sum (\text{–Q*}_{\text{ev}})}{n_{\text{ev}}} / \frac{\sum ([\text{–Q*}]_{\text{dry}})}{n_{\text{dry}}}
\]  \hspace{1cm} (2.10)

\[
[\text{LE}]_{\text{baseline}} = -\text{Q*}_{\text{frac}} \times [\text{LE}]_{\text{dry}}
\]  \hspace{1cm} (2.11)

Figure 2.12: Top: Precipitation event LE (W m⁻², solid line), dry-day ensemble LE (long dashed line), corrected dry-day baseline LE using method 1 (alternating dashed and dotted line), method 2 (short dashed line), and method 3 (dotted line). Bottom: Precipitation event –Q* (W m⁻², solid line), dry-day ensemble –Q* (long dashed line), and rain-day ensemble –Q* (short dashed line).
Applying all three methods to the individual daytime precipitation events, it was found that Method 3 most effectively separated the interception LE from the base-state LE. The approach in Method 3 to use the available energy for each individual rainfall event to scale the dry-day ensemble LE resulted in a better representation of the base-state LE during a given rainfall event, closest to the base-state LE outlined in Fig. 2.3. Method 3 was used for further analysis.

2.3.6 Treatment of heavy rainfall-rate periods

During periods of heavy rainfall, the sensors used in the eddy covariance system may fail and therefore directly-measured latent heat fluxes cannot be obtained (e.g. Mizutani et al. 1997). Therefore, during these periods, we use the Penman-Monteith equation to estimate event latent heat flux (Monteith 1965).

The Penman-Monteith equation is shown below:

\[
Q_E = \frac{\varepsilon A + \rho L_v \delta}{\varepsilon + 1 + \frac{r_s}{r_a}}
\]  

(2.12)

where

\(Q_E\) : Latent heat flux

\(A\) : Available energy

\(\varepsilon\) : \(L_v S_v / C_P\), where \(S_v\) is the slope of the saturation vapor-pressure curve; and

\(C_P\) is the specific heat of air at constant pressure.
\( \delta \): Saturation deficit

\( r'_s, r'_a \): Stomatal, aerodynamic resistances

\( L_V \): Latent heat of vaporization

\( \rho \): Air density

The available energy was determined directly from the radiation measurements near the tower top, and temperature, humidity measurements near the tower top for the saturation deficit. The aerodynamic resistance was calculated using the following relation (e.g. Wallace and McJannet 2006; Mizutani and Ikeda 1994):

\[
 r'_a = \frac{1}{k^2 u(z)} \left( \ln \frac{z - d}{z_0} \right)^2
\]  

(2.13)

where

\( z \): anemometer height

\( z_0 \): roughness length

\( d \): displacement height

\( u(z) \): wind speed at height \( z \)

\( k \): von Karman constant

In this study \( z \) was 58 m, the height of the top-level wind speed measurement. Following Monteith and Unsworth (1990), the displacement height was equal to 0.75 \( h \), and the roughness length equal to 0.1 \( h \), with \( h \) being the canopy height (43 m in this study).
Typical values for stomatal resistance used over forests are 150 s m$^{-1}$ for dry conditions and 0 s m$^{-1}$ for wet-canopy conditions (e.g. Raupach and Finnigan 1988). However, in our dataset, using a stomatal resistance of 0 during and following rain events in the Penman-Monteith calculation resulted in latent heat fluxes approximately three times the values for observed events where eddy covariance data were available. At the km67 study site, ensemble stomatal resistances found as a residual term in the Penman-Monteith equation shows that on rain-days, the stomatal resistance was not 0, but approximated 40 s m$^{-1}$ during rainfall periods. Using this stomatal resistance of 40 s m$^{-1}$ yielded much better agreement with observed LE during light and moderate rainfall-rate events, and was used in the heavy rain-rate cases. On dry days, stomatal resistance exceeds 100 s m$^{-1}$ in the afternoon at the study site.

Further justification for imposing some resistance for the wet canopy is that the stomatal resistance observations at the km67 study site indicate that the entire forest canopy is not wetted during these rain events. Since the formulation of the Penman-Monteith equation used here is for a uniform canopy, a stomatal resistance of zero would imply that the entire canopy surface is completely wet. Furthermore, Lloyd et al. (1988) explains that in using the Rutter (1971, 1975) model, evaporation from a saturated canopy is calculated from the Penman-Monteith equation with the stomatal resistance set to zero. When the depth of water stored on the canopy $C$ is less than the canopy storage capacity $S$, indicating a partially-wet canopy, the evaporation is reduced in proportion to the calculated value of $C/S$. Imposing a stomatal resistance comparable to ensemble observations in a uniform-canopy scheme has a similar impact on the resulting
evaporative flux as scaling back zero-stomatal-resistance Penman Monteith evaporation by the fraction of the canopy that is wet.

An issue arises when using the Penman-Monteith equation in rainfall conditions with regards to the available energy term, and net radiation. Water droplet collection on the dome of a net radiometer during rainfall can result in a decrease in the measured net radiation. In a comparison study of several types of net radiometers, Brotzge and Duchon (2000) show examples of a net radiation decrease of approximately 25 W m\(^{-2}\) during rainfall by the Kipp & Zonen CNR-1 net radiometer (used in this study) due to water droplet collection, which was the least amount of decrease for all the radiometers in the study. Over the range of meteorological conditions experienced at the km67 site, this 25 W m\(^{-2}\) decrease in available energy results in about a 2 to 5 % decrease in latent heat flux calculated by the Penman-Monteith equation.

2.4 Results

2.4.1 Nighttime precipitation events

For the nighttime precipitation events, ensemble means of the latent heat flux based on the rain start-time show a pulse of interception evaporation starting as the precipitation begins to wet the forest canopy, even before the first recorded tip at \(t=0\) of the rain events (Fig. 2.13).
Figure 2.13: Top panel: Ensemble mean latent heat flux (W m$^{-2}$) for all 54 nighttime precipitation events. Dotted lines indicate the standard error. The time axis refers to the number of hours before/after the first rain tip. Bottom panel: Mean wind speed (m s$^{-1}$, solid line) and $u^*$ (m s$^{-1}$, dotted line) for the same 54 nighttime precipitation events. The mean interception ($\pm$ standard error) for these events is 4.7% ± 0.9%. The mean precipitation ($\pm$ standard error) for these events is 3.32mm ± 0.59mm. The mean amount of water intercepted per event ($\pm$ standard error) is 0.09mm ± 0.03mm.

The nighttime interception evaporation pulse continues for about two hours following the event start, decreasing in magnitude with time. In the two hours following the precipitation event start, the mean interception estimate was just under 5% of the total precipitation. Then, the evaporation pulse stopped, possibly due to the air near the ground stabilizing as a result of the nocturnal interception evaporation.
A few of the nighttime events were associated with interception estimates of over 15% (Fig. 2.14). These high-interception nighttime events are broken into two categories. First, there were high-interception, high-wind events occurring around midnight, between 2300 LT and 0100 LT. Ensemble mean wind speeds for these events reached nearly 5 m s$^{-1}$, (Fig. 2.15) and are consistent with the timing of nocturnal squall lines at the site. The second type of high-interception nighttime event is a high-interception, low-wind event that occurs near the time of evening transition (around 1800 LT). Wind speeds during these events remain below 3 m s$^{-1}$ (Fig. 2.16).
Figure 2.15: Top panel: Ensemble mean latent heat flux (W m$^{-2}$) for the 4 nighttime high-interception, high-wind events. The time axis refers to the number of hours before/after the first rain tip. Bottom panel: Mean wind speed (m s$^{-1}$, solid line) and $u^*$ (m s$^{-1}$, dotted line) for the same 4 nighttime precipitation events. The mean interception (± standard error) for these events is 20.6% ± 5.7%. The mean precipitation (± standard error) for these events is 1.40mm ± 0.81mm. The mean amount of water intercepted per event is 0.28mm.
2.4.2 Daytime precipitation events

For the daytime rainfall events, ensembles of departure from baseline LE (representing interception evaporation) were constructed for different classes of rainfall rates with respect to the rain-event starting times. For light-to-moderate rainfall rates (<= 16 mm hr\(^{-1}\)), there was a steady pulse of interception evaporation for four hours following the event start (Fig. 2.17), with departure from baseline LE values maximizing at around 30 W m\(^{-2}\) around the rain-start time for the events.
Figure 2.17: Mean departure from baseline LE (W m\(^{-2}\)) for daytime rainfall events with rainfall intensities \(<= 16\) mm hr\(^{-1}\) using method 3 (solid black line, standard error dashed). A total of 104 events are included in the ensemble, and missing event data points were filled. The time \(t=0\) indicates the time of the first recorded tip by the rain gauge for tipping bucket rain gauge-recorded events or the first precipitation echoes detected by the ceilometer for ceilometer-detected events.

For the heavy rainfall-rate events \((>16\) mm hr\(^{-1}\)), the LE departure ensemble from direct event LE observations (Fig. 2.18) shows that the eddy-covariance system fails during the first hour after rainfall. For these events, the Penman-Monteith filled event LE was used to form the ensemble LE (Fig. 2.19). The departure from baseline LE was around 100 W m\(^{-2}\) for the first three hours after rainfall before decreasing to near zero.
Figure 2.18: Mean departure from baseline LE (W m⁻²) for daytime rainfall events with rainfall intensities > 16 mm hr⁻¹ using method 3 (solid black line, standard error dashed). A total of 25 events are included in the ensemble, and missing event data points were not filled. The time \( t=0 \) indicates the time of the first recorded tip by the rain gauge for tipping bucket rain gauge-recorded events or the first precipitation echoes detected by the ceilometer for ceilometer-detected events.

Figure 2.19: Mean departure from baseline LE (W m⁻²) for daytime Penman-Monteith-filled rainfall events with rainfall intensities > 16 mm hr⁻¹ using method 3 (solid black line, standard error dashed). A total of 25 events are included in the ensemble, and missing event data points were not filled. The time \( t=0 \) indicates the time of the first recorded tip by the rain gauge for tipping bucket rain gauge-recorded events or the first precipitation echoes detected by the ceilometer for ceilometer-detected events.
The mean intercepted water binned by rainfall intensity for daytime events (Fig. 2.20) shows an increase in the amount of water intercepted per event with increasing rainfall rate, towards a canopy capacity. The measurements from the events in the rainfall-rate bins less than 16 mm hr$^{-1}$ are directly from observations, with the greater than 16 mm hr$^{-1}$ rainfall bin using Penman-Monteith estimated event latent heat flux. A greater number of events at the high rainfall rates would be needed to split the high-rainfall rate bin into smaller bins to determine whether the amount of intercepted water would continue to increase towards a canopy capacity for rainfall rates of, for instance, 25 mm hr$^{-1}$ or greater.
Figure 2.20: Mean intercepted water binned by rainfall intensity for daytime events, with the standard error bars for each rain intensity bin shown. The rain intensity bins are as follows: \( \leq 2 \text{ mm hr}^{-1} \), 2-7 mm hr\(^{-1} \), 7-16 mm hr\(^{-1} \), and > 16 mm hr\(^{-1} \). The events in the \( \leq 2 \text{ mm hr}^{-1} \), 2-7 mm hr\(^{-1} \), and 7-16 mm hr\(^{-1} \) intensity bins used event LE, whereas the > 16 mm hr\(^{-1} \) bin used Penman-Monteith-filled LE. The numbers along the bottom of the plot indicate the number of events included in each rain intensity bin.
The mean intercepted water binned by rainfall intensity for daytime Penman-Monteith LE filled events (Fig. 2.21) shows a similar pattern to that found directly from observations, but with a slightly higher magnitude for the light to moderate rainfall-rate bins.

Figure 2.21: Mean intercepted water binned by rain intensity for Penman-Monteith-filled daytime events, with the standard error bars for each rain intensity bin shown. The rain intensity bins are as follows: <= 2 mm hr\(^{-1}\), 2-7 mm hr\(^{-1}\), 7-16 mm hr\(^{-1}\), and > 16 mm hr\(^{-1}\). The numbers along the bottom of the plot indicate the number of events included in each rain intensity bin.

The mean interception estimate for light rainfall-rate events (<= 2 mm hr\(^{-1}\)) using observed event LE was 18% (Table 2.3), with estimates for moderate rainfall rates (2-16 mm hr\(^{-1}\)) decreasing to about 10%. The percentage of daytime light-to-moderate rainfall
events in our sample with good data (80.7%) is close to the percentage of all light-to-moderate rainfall events detected in our dataset (77.4%).

Table 2.3: Mean interception estimates for daytime rainfall events classed by rainfall rate. The events in the <= 2 mm hr\(^{-1}\) and 2-16 mm hr\(^{-1}\) rate categories used observed LE, whereas the event LE in the > 16 mm hr\(^{-1}\) was filled with Penman-Monteith LE.

<table>
<thead>
<tr>
<th>Rainfall rate</th>
<th>Mean interception (standard error)</th>
<th>Number of events</th>
</tr>
</thead>
<tbody>
<tr>
<td>&lt;= 2 mm hr(^{-1})</td>
<td>18.0% (12.2%)</td>
<td>46</td>
</tr>
<tr>
<td>2-16 mm hr(^{-1})</td>
<td>9.9% (2.6%)</td>
<td>58</td>
</tr>
<tr>
<td>&gt; 16 mm hr(^{-1})</td>
<td>7.8% (1.6%)</td>
<td>25</td>
</tr>
</tbody>
</table>

The mean interception estimates for light rainfall-rate events (<= 2 mm hr\(^{-1}\)) and moderate rainfall rates (2-16 mm hr\(^{-1}\)) using Penman-Monteith filled event LE were 21.5% and 14.7% respectively (Table 2.4), slightly higher from the corresponding observed values. Mean interception for the heavy rainfall events was 7.8%.

Table 2.4: Mean interception estimates for Penman-Monteith-filled daytime rainfall events classed by rainfall rate.

<table>
<thead>
<tr>
<th>Rainfall rate</th>
<th>Mean interception (standard error)</th>
<th>Number of events</th>
</tr>
</thead>
<tbody>
<tr>
<td>&lt;= 2 mm hr(^{-1})</td>
<td>21.5% (12.2%)</td>
<td>46</td>
</tr>
<tr>
<td>2-16 mm hr(^{-1})</td>
<td>14.7% (3.5%)</td>
<td>58</td>
</tr>
<tr>
<td>&gt; 16 mm hr(^{-1})</td>
<td>7.8% (1.6%)</td>
<td>25</td>
</tr>
</tbody>
</table>
At the km67 study site, nearly half of the time rainfall rates are at the lightest category, with the heaviest rain rates only occurring about 6% of the time (Fig. 2.22). However, due to the convective nature of the heavy precipitation, these heavy rainfall rates do contribute one-third of the total rainfall amount at this site (Fig. 2.23).

Figure 2.22: Total rainfall duration binned by rainfall intensity at the km67 site.
2.4.3 *Interception during the mornings following nighttime events*

In the early morning following a nighttime rainfall event, the remaining water stored in the canopy from the previous night’s rain event will be evaporated. The ensemble departure from baseline latent heat flux for the mornings after nighttime rainfall events show a pulse of evaporation starting at about one hour after sunrise, peaking between two and three hours after sunrise at about 60 W m\(^{-2}\) (Fig. 2.24). This evaporation pulse amounts to an additional a mean amount of 0.05 mm of water evaporated during the early morning, with a standard error of 0.02 mm. Therefore, the combined mean (± standard error) interception for all nighttime rainfall events is 7.2% (± 1.0%).

Figure 2.23: Fraction of total rainfall at the km67 site binned by rainfall intensity.
Figure 2.24: Mean departure from baseline LE (W m\textsuperscript{-2}) for early morning periods following nighttime rainfall events using method 3 (solid black line, standard error dashed). A total of 54 events are included in the ensemble, and missing event data points were not filled. The time $t=0$ indicates the approximate time of sunrise following the rain event (0600 LT).

2.4.4 Energy balance comparison for dry and rain days

To compare the energy balance components for dry and rain days, ensembles of each of the components $-Q^*$, LE, and $H$ were assembled for dry days (189 days) and days with rain that started after 1400 GMT (100 days total). At the km67 site, the number of rainfall cases observed starts to increase in the late morning around 1400 GMT, with the greatest number of cases occurring at the afternoon convective peak. Choosing this time effectively separates the rainfalls associated with the nighttime/early morning peak from the afternoon convective peak. In the early morning, before 1200 GMT, $-Q^*$ for dry and rain days are close to each other. However, $-Q^*$ decreases on the rain days starting around 1200 GMT, about two hours before the rain period on the rain days, and representing the onset of cloudiness (Fig. 2.25). For the remainder of the day, all of the energy balance components are greater in magnitude for the dry days than the rain days.
Figure 2.25: Mean energy balance components for dry days (solid lines) and afternoon rain days (dashed lines). –Q* is in black (top pair of lines), LE in blue (second pair of lines from top), H in red (third pair of lines from top), and S_{bc} in green (bottom pair of lines). The dry-day ensemble includes 189 days, with 100 days in the rain-day ensemble. The vertical dashed line indicates the time (1400 GMT) after which rain fell during the rain days.

Dividing the energy balance components for the dry and rain days by their respective –Q* values results in energy balance component fractions that can directly be compared for dry and rain days. When scaled by radiative energy, the effect of the rain on the energy balance components becomes more apparent (Fig. 2.26). During the early morning pre-rainfall period, the evaporative fraction is greater on the dry days than the rain days. Once the rainfall period is reached, the evaporative fraction becomes greater on the rain days than the dry days and remains so for the balance of the day. During the late afternoon period (1800 – 2200 GMT), the rain-day evaporative fraction is over 5 percent greater than the dry-day evaporative fraction (Table 2.5). On the rain days, the evaporative fraction increases over 16 percent from the pre-rain period to the late afternoon period, while the sensible heat fraction falls by 7 percent. Most of the energy required for the evaporative fraction increase on the rain days appears to be supplied by the storage term. From the pre-rainfall morning period to the late afternoon period, the storage fraction decreases by over 15 percent, falling to negative values indicating a
release of energy. The storage fraction decrease nearly offsets the evaporative fraction increase on the rain-days.

![Figure 2.26: Mean evaporative fraction (blue, top pair of lines), sensible heat fraction (red, middle pair of lines), and storage fraction (green, bottom pair of lines) of \(-Q^*\) for dry days (solid) and days with rain after 1400 GMT (dashed). The vertical dashed line indicates the 1400 GMT rain cutoff time. There are 189 days included in the dry-day ensemble and 100 days in the rain-day ensemble.](image)

Table 2.5: Mean evaporative, sensible heat, and storage fractions of \(-Q^*\) for dry and rain days for the pre-rain period 1200 – 1400 GMT and rain periods 1400 – 1800 GMT and 1800 – 2200 GMT.

<table>
<thead>
<tr>
<th></th>
<th>1200 – 1400 GMT</th>
<th>1400 – 1800 GMT</th>
<th>1800 – 2200 GMT</th>
</tr>
</thead>
<tbody>
<tr>
<td>([\text{LE}]/[-Q^*]) (%)</td>
<td>47.5</td>
<td>49.3</td>
<td>51.8</td>
</tr>
<tr>
<td>([\text{H}]/[-Q^*]) (%)</td>
<td>19.9</td>
<td>20.3</td>
<td>18.7</td>
</tr>
<tr>
<td>([\text{Sbc}]/[-Q^*]) (%)</td>
<td>10.2</td>
<td>4.3</td>
<td>4.3</td>
</tr>
</tbody>
</table>

Bowen ratio (H/LE) values during the pre-rainfall morning period are nearly the same for the dry and rain days (Fig. 2.27). Following the onset of rainfall, the rain-day Bowen ratio becomes lower than the dry-day Bowen ratio and remains so for the rest of the day. For the period following rainfall (1400 GMT – 2100 GMT), mean Bowen ratios for the dry and rain days are 0.34 and 0.28 respectively.
Figure 2.27: Top: Mean Bowen ratio for dry days (solid black line) and days with rain after 1400 GMT (dashed blue line). The vertical dashed line indicates the 1400 GMT rain cutoff time. Bottom: Same as top but Bowen ratio medians are plotted.

2.5 Summary

Reviewing the questions put forth at the beginning of the chapter:

- Through the expansion of eddy flux-measurement networks such as Fluxnet (Baldocchi et al. 2001), the number and coverage of long-term eddy flux measurement sites has grown to over 460 worldwide sites, distributed over a wide range of land cover types (http://www.fluxnet.ornl.gov). What additional information can be obtained from these sites beyond the standard reports of half-hour average fluxes?
We have introduced a methodology by which one can directly observe the amount of interception evaporation by using eddy-covariance data that are available at a number of worldwide flux tower sites, combined with novel data-analysis techniques to form base-state and event-based ensembles of latent heat flux.

Mean interception for moderate daytime rainfall-rate events was about 10%, with light events at 18% and heavy events at 7.8%. The mean interception for all daytime and nighttime events combined was 11.6%. Some conventional interception estimates in tropical rain-forest environments exceed our results by two times (Fig. 2.28).

Energy balance comparisons between dry and afternoon rain-days show an approximately 15% increase of evaporative fraction on the rain days, with the energy being supplied by a nearly equivalent decrease in the canopy heat storage.
Figure 2.28: Interception estimates reported in the literature using conventional methods for tropical rain forest sites. Studies done in Brazil are labeled in black with ‘B’, Central America in pink with ‘C’, Malaysia in green with ‘M’, Australia in blue with ‘A’, and Puerto Rico in red with ‘P’. References are as follows: 1B, 2B: Franken et al. (1982a, b); 3B: Schubart et al. (1984); 4B: Leopoldo et al. (1987); 5B: Lloyd and Marques (1988); 6C: Imbach et al. (1989); 7B, 8B: Ubarana (1996); 9C: Cavelier et al. (1997); 10B: Arcova et al. (2003); 11B: Ferreira et al. (2005); 12M: Manfroi et al. (2006); 13A: Wallace and McJannet (2006); 14P: Holwerda et al. (2006); 15B: Germer et al. (2006); 16B: Cuartas et al. (2007). The mean interception estimate for all rainfall events in this study is denoted by 17B.

An inherent advantage of using this method to estimate interception evaporation is that the footprint area of the eddy-covariance measurement incorporates the spatial variability in throughfall and interception that is a challenge to adequately sample using conventional methods, providing an average representative interception value over the entire flux footprint, and therefore be more suitable for grid-cell model input.
Another advantage of using the method presented here is that once baseline data are established, it can be used for long-term interception monitoring without intensive effort beyond the maintenance of the eddy-covariance system at a site, and not the degree of effort required to maintain ongoing interception measurements using conventional methods, such as maintaining long-term adequate sampling of throughfall measurements with roving rain gauges, for example.

- Under what conditions would micrometeorological-based interception evaporation measurements be expected to work, and under what conditions would such a technique be of limited use?

Tests of the method over an eastern Amazon old-growth rain forest show the method to be effective using direct observations under light-to-moderate rainfall rates (\(\leq 16 \text{ mm hr}^{-1}\)). For events with heavy rainfall rates (\(> 16 \text{ mm hr}^{-1}\)), Penman-Monteith estimated evaporation can be used to substitute for LE during periods when eddy-covariance does not work. For determination of the dry-day daytime base-state ensembles, the method is applicable in nearly all daytime turbulent conditions. At night, base state latent heat flux is zero, so formation of base-state nighttime ensembles is not necessary and the well-documented eddy covariance failures during calm, low-turbulent conditions are not applicable.
2.6 Acknowledgements

This work was supported as part of the LBA-ECO project, supported by the NASA Terrestrial Ecology Branch under grants NCC5-283 and NNG-06GE09A (Phase 3 of LBA-ECO) to the authors’ institutions. We acknowledge the Harvard University group who ran instrumentation at the km67 site, and Lucy Hutyra and Elaine Gottlieb with providing information on the dataset and calibrations.

2.7 References


Manfroi, O., and Coauthors, 2006: Comparison of conventionally observed interception evaporation in a 100-m² subplot with that estimated in a 4-ha area of the same Bornean lowland tropical forest. *J. Hydrol., 329*, 329-349.


Chapter 3: Spatial and seasonal changes in watershed response to rainfall events in two contrasting climates

Introduction (Foreword)

In this chapter, we examine spatially and seasonally varying streamflow response to rainfall by performing an observational and model sensitivity studies in two regions. The first location is a network of watersheds located in New York State in the Hudson Valley Ambient Meteorology Study (HVAMS) study area. In this region, precipitation patterns are strongly influenced by topography, and evapotranspiration forcing changes as leaf state changes seasonally. The second location includes a small group of watersheds located in Brazil, part the study area of the Large-Scale-Biosphere-Atmosphere Experiment in Amazonia (LBA). In this region of the equatorial tropics, precipitation is seasonal, with distinct wet and dry seasons. This study focuses on some of the hydrological consequences of dry season convective rainfall. Evapotranspiration and soil properties in this region change in response to land cover changes, primarily the result of deforestation. We examine components of the surface water balance to transient forcing events at the small-to-large watershed scale in both regions. We begin with the HVAMS study, which forms the bulk of the chapter, followed by a smaller section that discusses the LBA watershed study.
Part I: HVAMS

Abstract

Response to rainfall is examined in a network of watersheds in the Catskill-Hudson Valley region, the source area for New York City’s water supply. We conduct a sensitivity study that combines observations and the BROOK hydrological model (Federer 1995). The focus is on the impacts of spatial and seasonal changes in two variables in the watersheds: a) the effect of topography, as it relates to spatial changes in precipitation due to the presence of a rain shadow in the lee of the Catskills; b) phenology, how seasonal changes in leaf state affect regional evapotranspiration. The two detailed case studies presented take advantage of a suite of atmospheric, hydrological, and remote-sensing data collected during the Hudson Valley Ambient Meteorology Study (HVAMS) Intensive Field Campaign (IFC) in September and October 2003 to characterize spatial and seasonal differences in watershed precipitation and the subsequent response. Seasonal variation in watershed response to rainfall is analyzed observationally using streamflow and soil moisture data in three stages: the period when streamflow peaks following input precipitation, the streamflow recession following the peak, and diurnal streamflow fluctuations that may occur during the dry periods between rainfall events once the streamflow has receded to its baseflow. Analysis of the spatial and temporal changes in the diurnal streamflow fluctuations observed over the network during the autumn transition season illustrates how the major change in regional evapotranspiration at the end of the growing season alters watershed response to rainfall.
3.1 Introduction

3.1.1 Precipitation shadows

In the northeast US, Brady and Waldstreicher (2001) documented mountain-induced precipitation shadows in the Appalachian Mountain-Wyoming Valley region of Pennsylvania, where precipitation amounts were reduced within and up to 15 km downstream of the Wyoming Valley. The phenomenon is more frequently documented in regions with more rugged terrain, such as the Cascades of the northwestern US (e.g., Smith et al., 2005) and the Alps in Europe (e.g., Pfister et al., 2004). In many studies, for spatial scales greater than 40 km and for mountain ranges with altitude exceeding 1500 m, the predominant mechanism responsible for the orographic precipitation is stable upslope ascent. Precipitation maxima on the windward side of the topographic barrier result from forced mechanical lifting and resulting cooling and condensation (Roe, 2005). Descent of the air on the leeward side results in adiabatic warming and drying of the air, and reduced precipitation rates. However, orographic precipitation may occur as a result of one or a combination of several other dynamic mechanisms at a given location, even during a given rainfall event. Roe (2005) lists these mechanisms as: the partial blocking of an impinging airmass, down-valley flow induced by evaporative cooling, lee-side convergence, convection triggered by solar heating, convection owing to mechanical lifting above the level of free convection, and the “seeder-feeder” mechanism, used to describe the process by which precipitation falling from a higher, large-scale “seeder” cloud accretes additional moisture when falling through a lower “feeder” cloud generated by upslope ascent. Thus, precipitation rates are enhanced in the terrain below. Each of
these topographically-induced forms of precipitation may result in very different spatial patterns of precipitation in a given area than the classic lee-side precipitation shadow. For instance, higher amounts of precipitation have consistently been observed on the leeward side of Mount Carmel, Israel due to horizontal convergence of the lee-side flow (Goldreich et al., 1997). According to Roe (2005), orographic precipitation is “intrinsically a transient phenomenon”, with precipitation rates varying substantially as synoptic conditions change during a single rainfall event.

3.1.2 Catskill-Hudson Valley region precipitation shadow

The Catskill-Hudson Valley region of New York State lies in an area of complex terrain that influences the magnitude and spatial variability of the local precipitation, wind, and water-drainage patterns. The presence of a lee-of-Catskills precipitation shadow has long been known, and occurs frequently enough to appear in mean annual precipitation totals (Merriman 1907; Fig. 3.1).
Figure 3.1: Annual average rainfall (inches) for the Catskills, Hudson Valley and surrounding regions (Merriman 1907, Fig. 1). Precipitation totals are based on data collected at a network of stations in the map region from varying periods of time from the early 1800s to 1905. The Rondout Creek, Esopus Creek, Catskill Creek, and upper Schoharie Creek drainage basins are outlined. The method by which these watershed boundaries were drawn was not described. The lower portion of the Schoharie Creek drainage basin (not outlined) extends further north from the upper Schoharie basin into the main portion of the precipitation shadow.
Mean annual precipitation totals in the peak precipitation region of the Catskills west of Poughkeepsie were more than 350 mm greater than in the precipitation shadow located in the lower Schoharie Creek drainage basin west of Albany, a difference of greater than 30 percent from the regional-average mean annual precipitation (Fig. 3.1). For individual precipitation events, the difference may be greater, with approximately twice the precipitation falling in the peak precipitation region than in the rain shadow. Thaler’s (1996) Catskill precipitation climatology covering the years 1961-1990 places the peak-precipitation area and rain shadows in the same general area as did Merriman, but the average annual precipitation in the peak-precipitation region is about 250 mm greater.

Rain roses were constructed for a number of stations in the Catskill-Hudson Valley region during the period of the Hudson Valley Ambient Meteorology Study (HVAMS) Intensive Field Campaign (IFC) period of September and October 2003 (Figs. 3.2, 3.3).
The rain roses constructed using ETA model-estimated 850-mb flow, a few hundred meters above the highest Catskill ridge tops, indicate that the majority of precipitation at all Catskill-Hudson Valley stations fell with the 850-mb wind from the south-to-southwesterly directions (Fig. 3.2). Details of the ETA model and the weather stations are discussed in Section 3.2 below. The general agreement with the south-to-north precipitation gradient observed in Figure 3.1, and with the precipitation shadow on the northern downwind side of the Catskills, indicates that our case studies are representative of the general situation.
Rain roses constructed from surface wind observations (Fig. 3.3) show a variety of local effects. Channeling along the Hudson Valley is seen, with the dominant wind directions during rain slightly changing to follow the orientation of the valley at a given station (e.g., stations 3, 6, and 9 on Fig. 3.3). A similar channeling up the Schoharie Creek basin is seen (stations CSM038 and CSM039 on Fig. 3.3). For the other upland stations, local site characteristics such as sheltering by nearby obstructions to the station appear to play a large role in the wind directions observed during rainfall. The degree to which a station is sheltered was quantified through the use of the transmission factor (TF; Fujita and Wakimoto 1982, Acevedo and Fitzjarrald 2003). For a given site, TF is
determined by azimuth as the average wind from a given direction divided by the maximum average wind observed in the station network from that direction. Therefore, TF = 1 denotes an open direction, and TF = 0 represents a completely obstructed direction. For example, the surface rain roses from stations H2 and H3 (Fig. 3.3) show dominant wind directions during rainfall that correspond to the well-exposed directions indicated by each station’s transmission factors (Fig. 3.4).

Figure 3.4: Wind roses (black) and transmission factors (TF; gray lines) for the HOBO stations during the HVAMS IFC period. The outer dotted ring represents TF=1 (an open direction).

It is important to know the precise distribution of rainfall and orientation of any precipitation shadows in these topographically-complex drainage basins. First, better
knowledge of the present regional precipitation distribution and the subsequent watershed response would assist in improving and implementing a sound management strategy for maintaining the quantity and quality of the area’s water resources. Second, this knowledge would benefit flood forecasting in the region, especially since many of the area watersheds are small and in upland topographic landform settings, and have a proportionately large contributing variable source-area (Hewlett and Hibbert, 1967) that results in a large fraction of the overland flow resulting from rainfall going directly into runoff. As a result, these streams in these watersheds are quick-responding and more susceptible to flash flooding.

3.1.3 Watershed response

Individual watersheds may be expected to respond differently to changing climatic conditions; however there are many factors complicating the response. Pfister et al. (2004) reported that an increase in westerly circulation types during the last 50 years has resulted in increased winter precipitation totals in a group of watersheds in the Mosel River basin of Luxembourg. The spatial variability of the trends affecting wintertime precipitation amount, duration, and intensity in the watersheds have resulted in spatially-varying positive trends in maximum daily water levels that are strongly related to the topography of that study area. Drogue et al. (2004) in a model study simulating watershed response in the Alzette River basin of Luxembourg, Belgium and France, concluded that the impact of mesoscale climate change is highly variable in the watersheds, with the spatial variability of the streamflow responses largely conditioned by hydro-physiographic characteristics of the watersheds, such as drainage area, relief,
drainage density, and soil properties. They indicated some sub-basins were particularly sensitive to altered climate conditions in terms of changes in high or low flows. The climate zone the watershed is situated in may also dramatically alter the magnitude of watershed response to climate change (Mohseni and Stefan 2001).

Although precipitation input and the resulting streamflow peak are the elements most discussed in the literature, it represents only the initial stage of watershed response to rainfall. Equally important are the following stages of watershed response, streamflow recession following precipitation (Federer 1973) and the diurnal streamflow fluctuations observed during extended periods between rainfall (Czikowsky and Fitzjarrald 2004). Bulk streamflow properties such as streamflow recession and diurnal streamflow fluctuations are observable indicators of watershed response that may used to test more complex watershed models. Both properties are influenced by changes in forcing characteristics such as evapotranspiration, itself affected by land-use/land-cover characteristics such as vegetation amount, cover, and state, all of which vary seasonally. In regions that may experience more extreme weather conditions such as prolonged dry periods or droughts, knowledge of how these properties operate on the watershed scale becomes increasingly important.

3.1.4 Catskill region watersheds

The over-1300 km² area of watersheds draining the Catskill-Hudson Valley region provides a major source for New York City’s water supply, with the Catskill upland region alone supplying 40 percent of the city’s daily water demand. Combined with the adjacent Delaware River watersheds, approximately 90 percent of the city’s water is
supplied (O’Melia et al. 2000). Therefore, proper management of the water and land resources in this region is a top priority, especially in the face of a changing climate (e.g. Burns et al. 2007, Rosenzweig et al. 2007) that would likely include changes in regional precipitation, evapotranspiration, water quality, and water yield.

The Schoharie Creek drainage basin (Fig. 3.5) provides a focus area for the analysis. Its headwaters are located in the area of highest precipitation in the region, but the areas downstream lie in a rain shadow. In the downstream region of the drainage basin lies the Gilboa Dam, which is no longer structurally sound, raising concerns that a major precipitation/flooding event in the area could lead to the dam’s failure (Owings, 2008 and references therein), and is another reason to consider small watershed response. Burns et al. (2007) indicated from a network of 12 precipitation stations in the Catskills that regional annual mean precipitation had increased at a rate of 136 mm per 50 years during the 1952-2005 period, an increase of greater than 10 percent in regional annual precipitation. The greatest increases occurred in May through September. During the same period, mean regional air temperature increased at a rate of 0.6°C per 50 years, with the largest increase observed at the high-elevation Slide Mountain station (1.8°C per 50 years).

Projections of future climate scenarios in the region point to a greater likelihood of extreme events such as flooding and drought. Blake et al. (2000; cited in Burns 2007) projected that regional mean air temperature will increase 1.5-3.0°C by 2050 and 3.3-5.8°C by 2100, with the potential for increased drought frequency. On the other hand, Frumhoff et al. (2007; cited in Owings 2008) showed the area is likely to experience a greater occurrence of heavy precipitation events that may enhance the effect of the
orographic precipitation received and increase the likelihood of a dam failure. This region has suffered what had been estimated to be 100-year floods four times in the past 50 years. Furthermore, average daily discharge in Schoharie Creek has increased in the most recent 10-year period (1998-2007) when compared with the previous 50 years (1940-1997), with an increase in the average annual discharge of approximately 20 percent (Kern 2008; cited in Owings 2008). Increases in discharges in the recent period were most prominent in the months October through January, with possible reasons for the change being a greater amount of total precipitation in these months or a greater proportion of the precipitation as rain instead of snow in the winter months. The largest decrease in daily average streamflow in the later period occurred in May. Possible reasons for the shift are an earlier spring snowmelt period in recent years or a diminished winter snowpack due to more winter precipitation falling as rain. Burns et al. (2007) reported a shift in the timing of peak snowmelt runoff from the first week of April to the last week of March from the period 1952-2005 in a study consisting of 8 streamflow stations in the Catskills.
A more detailed characterization of the Catskills-Hudson Valley precipitation shadow is hindered by the lack of adequate ground-based precipitation observation sites operating at the present time in the shadow region, especially in the downstream portions of the Schoharie Creek drainage basin. The only Automated Surface Observing System (ASOS) stations are at Albany and Poughkeepsie. The New York City Department of Environmental Protection (NYCDEP) operates a network of weather stations, but these are mainly concentrated in the Catskill upland region, not in the precipitation-shadow region. The National Weather Service Cooperative Observer (NWS COOP) network has a limited amount of stations in this region. However, cooperative observers report only
daily precipitation totals, insufficient temporal resolution to describe short-duration storms.

A consequence of the lack of in situ observations in this region is an over-reliance on remotely sensed data to determine the orientation and magnitude of the regional precipitation shadow. The primary tool used is the NWS Doppler Radar (NEXRAD WSR-88D), whose local installation is at the Albany (KENX) site. Its broad spatial sampling capabilities make it an appealing choice as a data source for hydrological models. Previous comparisons of precipitation estimated from the NEXRAD precipitation algorithm and ground-based observations provided by rain gauges often show an underestimation from the NEXRAD-estimated precipitation that may change based on the distance from the radar site (Jayakrishnan et al. 2004 and references therein). Part of the calibration process for the radar data includes data correction from ground-measured rainfall (Fulton et al. 1998), but this can be complicated due to the very different sampling areas in space and time from the radar and rain gauges (Jayakrishnan et al. 2004). Radar beam blockage by the terrain surrounding the radar site complicates radar estimates of precipitation and their correction (Young et al. 1999), and this is a factor that affects data from the KENX radar site. As a result, “caution should be employed with using NEXRAD-estimated rainfall] data for hydrologic modeling” (Over et al. 2007).

Gridded precipitation data are also available through the NOAA Climate Prediction Center Morphing (CMORPH) data product (Joyce et al. 2004). The CMORPH data are a satellite-based passive microwave precipitation estimate, available at 8- and 40-km resolution. The 40-km resolution product has been shown to describe the
aspects of the medium- and large-scale spatial variability in rainfall in the tropics (Fitzjarrald et al., 2008). To date, the CMORPH product has not been tested in the Catskill-Hudson Valley region and will be investigated here to assess its suitability as a possible additional data source for the precipitation-shadow region. The HVAMS deployment allowed for additional ground-based rainfall observations in this observationally-sparse rain shadow region, useful for the comparison of ground-based and remotely-sensed precipitation data in case studies.

3.1.5 Outline

We conduct an observational and model sensitivity study concentrating on spatial and seasonal effects on two forcing parameters in the Catskill-Hudson Valley watersheds, precipitation and evapotranspiration. First, we examine spatial differences in precipitation, as the presence of the Catskill-Hudson Valley precipitation shadow results in the region’s watersheds receiving very different amounts of precipitation during storms with southwesterly winds aloft. Second, we examine seasonal changes in evapotranspiration, as the land-cover change that occurs at the end of the growing season due to leaf drop results in a marked decrease in evapotranspiration in the region’s watersheds.

The impact of these large differences in forcing inputs on the regional watershed response has not been studied comprehensively in this region on a case-study basis. The aim here is to assemble a comprehensive set of atmospheric, hydrological, and remote-sensing observations in conducting these natural sensitivity studies.
The outline for this part of the chapter is as follows. An observational analysis and mechanistic study focusing on transient effects on components of the surface water balance over the small-to-large watershed scale is presented for a network of watersheds located in New York State in the domain of the Hudson Valley Ambient Meteorology Study (HVAMS). We begin with a description of the study-area location, instrumentation, and data sources. We then outline the analysis methods used in the paper to assess watershed response. Next, we present two case studies of heavy precipitation events and associated watershed response during the HVAMS Intensive Field Campaign (IFC) during September and October 2003, with emphasis on the Schoharie Creek basin, which encompasses the precipitation shadow region. The days of the event case studies were for the September 23, 2003 and October 27, 2003 precipitation events. The watershed response indicators of streamflow recession and diurnal streamflow fluctuations are analyzed during the autumn transition season of 2003. Streamflow recessions were calculated following the September 2, September 23, September 27, and October 27, 2003 precipitation events. The September 2 and 23 events were binned together for the growing season, in-leaf period of the autumn transition (marked as periods with an L on Fig. 3.6). The September 27 and October 27 periods were binned together for the dormant season, no-leaf period (marked as periods with an NL on Fig. 3.6). Diurnal streamflow fluctuations were examined during four dry periods with sufficient time between precipitation events: a) Period I: August 22-24; b) Period II: September 9-11; c) Period III: October 8-10; d) Period IV: October 23-25. These periods are marked as I through IV on Fig. 3.6.
Figure 3.6: Time periods used in the analysis of streamflow recession and diurnal streamflow fluctuations over the network watersheds for the autumn transition season of 2003. The green bars labeled with L denote the streamflow recession periods used in the growing season in-leaf period. The brown bars labeled with NL denote the streamflow recession periods used in the dormant season no-leaf period. The yellow bars labeled with I, II, III, IV denote the periods used in the diurnal streamflow fluctuation analysis. Streamflow plotted is for Biscuit Brook, NY (units mm on watershed).

In the study, watershed response is analyzed spatially in a network of small-to-large watersheds on a precipitation event basis. Temporal changes in forcing on watershed response to precipitation events are also analyzed seasonally. In this study, the seasonal change in the forcing takes the form of land-cover change and reduced evapotranspiration due to fall-season leaf drop. The BROOK watershed model (Federer 1995) is run to conduct a sensitivity study on watershed response to changes in input precipitation amount induced by orographic effects and changes in vegetation state due to phenology, and to assist in interpreting the observations.

Specific questions to be addressed in the chapter include:

- Watersheds in the Catskill-Hudson Valley Region of New York State (HVAMS) receive much different inputs of precipitation due to orographic effects, and similar effects are seen in nearby upland regions of the Northeast United States (e.g. Brady and Waldstreicher 2001). As the frequency of extreme precipitation events may be expected to increase in this region as a
result of climate change (e.g. Burns et al. 2007), such orographic effects on precipitation input may be enhanced. What is the watershed response to such rainfall events, and are there any detectable changes in the response due to changes in precipitation forcing?

- Underlying the individual watershed response in the HVAMS region to differing amounts of precipitation is the seasonal land cover change due to transpiring vegetation in leaf during the growing season transitioning to leaf drop and the beginning of the dormant season, thereby changing the amount of ET forcing the watersheds receive. Can any detectable changes in watershed response be seen during such transition periods?

3.2 Location and Data/Instrumentation

The study area encompasses the mid-Hudson River Valley between Albany and Poughkeepsie, New York, extending westward into the Catskill mountain region (Fig. 3.7). The approximate latitude and longitude bounds for the study region are 41.6°N to 42.8°N, and 73.5°W to 75.0°W respectively. The width of the Hudson River Valley in this region ranges between 20 and 30 km. The valley walls are approximately 200 to 300 m in elevation, with the peak elevation in the Catskills exceeding 1000m. The Catskills are technically not a mountain range but an eroded plateau. The edge of this plateau forms an escarpment oriented in a north-south direction with walls near 1000 m along the west side of the valley between Kingston and Catskill (approximately between stations 6 and M in Fig. 3.7). Further north, the escarpment turns to the northwest between Freehold (H2 in Fig. 3.7) and East Jewett (H3 in Fig. 3.7).
Figure 3.7: Topography and data stations for the HVAMS study area of the Hudson Valley and Catskill uplands. Streamflow stations are the black dots. Precipitation sites include the NCAR-PAM sites (1-9), Hobo weather stations (H1-H5), MIPS station (M), ASOS stations (A1 and A2), and the Cooperative observer network (c). Units for the elevation on the legend are in meters. The white dotted line is the topographic transect line for the cross-section shown in Fig. 3.8.

Following a topographic transect line from South Albany Airport (station 8 in Fig. 3.7) to the northeast, through Freehold (station H2) and East Jewett (station H3) to Phoenicia, NY, to the southwest, the elevation increases approximately 700 m crossing the escarpment in a distance of under 10 km (Fig. 3.8). The escarpment provides the
focus area for the orientation of the orographically-induced rain shadow observed in the region.

Figure 3.8: Northeast-to-Southwest topographic cross-section from South Albany Airport (station 8 in Fig. 3.7) to Phoenicia, NY. The topographic transect line is shown in Fig. 3.7.

A network of weather and flux-measurement stations was deployed in the study area as part of the Hudson Valley Ambient Meteorological Study (HVAMS). The HVAMS surface network included nine NCAR Integrated Surface Flux Facility (ISFF) Portable Automated Mesonet (PAM) weather stations (1 through 9 on Figure 3.7, station names listed in Table A1.1), deployed along the Hudson Valley during the months of September and October 2003, which was the time period of the HVAMS Intensive Field Campaign (IFC). These stations provided flux measurements of heat, moisture, and momentum, as well as wind, temperature, humidity and solar radiation. Precipitation was recorded at one-minute intervals at these sites. At six of these stations (2, 3, 4, 6, 8, and 9) microbarometers were operated. At four sites (PAM stations 2, 3, 6, and 9) CO₂ sensors were installed, and at three sites (PAM stations 2, 3, and 9) ozone sensors were installed. These stations were deployed along the Hudson Valley ranging from 25 to 156 m elevation. At the PAM stations, precipitation was measured using an MRI
(Meteorological Research, Inc., Altadena, CA) model 302/303/304 tipping bucket rain
gauge, with no wind screen. Wind speed and direction were measured at two levels: at
10-m height using an RM Young model 9101 propeller vane anemometer, and at 7-m
height using a Campbell CSAT-3 sonic anemometer. Periodic calibrations were
performed during their deployment by NCAR technicians. Specifics regarding the other
sensors used at the PAM stations can be found at the NCAR-ISFF website
(http://www.eol.ucar.edu/rtf/projects/HVAMS03).

Five standard weather stations (HOBO, Onset Computer Corp., Bourne, MA; H1
through H5 on Fig. 3.7) were deployed in the highlands surrounding the Hudson Valley.
The HOBO weather stations were located at sites H1 through H5 during the HVAMS IFC
period. Station names and locations during the IFC period are listed in Table A1.2. After
the IFC period, station H1 was moved to the site of PAM station 4 (Green Acres Airport),
station H3 to Schodack State Park (S on Fig. 3.7), and station H4 to PAM station 6 (Van
Orden Farm).

The HOBO measurement suite included precipitation, temperature, humidity,
wind, pressure, solar radiation, and soil moisture/temperature, sampled at one-second
intervals and recorded at one-minute intervals during the IFC period. Following the IFC
period, the recording interval at the HOBO stations was increased to five minutes.
Precipitation was recorded using HOBO tipping bucket rain gauges (from Texas
Electronics, Inc.) at 0.2 mm resolution. The calibration accuracy of these gauges is ±1%
at rainfall rates of up to 20 mm hr⁻¹. Soil moisture at the HOBO sites was measured at 5
cm depth using an EC-20 Dielectric Aquameter from Decagon Devices, Inc. The soil
moisture sensor has a resolution of ± 0.04% with an accuracy of ±3%. The soil moisture
readings were calibrated by taking periodic soil core samples at each site and performing
gravimetric analysis on the samples. Wind speed and direction were measured using a
cup anemometer/wind vane at 3.4-m height (Onset Computer Corp. wind speed/direction
sensor S-WCA-M003). Intercomparisons among all these sensors conducted two weeks
prior to the IFC showed good agreement with regards to precipitation and wind
speed/direction.

The University of Alabama-Huntsville Mobile Integrated Profiling System
(MIPS) station installed at Kingston-Ulster Airport during the IFC period (M on Figure
3.7) included a surface weather station, wind profiler, ceilometer, and radiometer.
Precipitation was recorded at one-minute intervals with a tipping-bucket rain gauge at the
MIPS station at 0.2mm resolution (Texas Electronics model TR-525M). The calibration
accuracy of these gauges is ±1% at rainfall rates of up to 50 mm hr⁻¹. Further information
regarding the MIPS instrumentation can be found at the MIPS website
(http://vortex.nsstc.uah.edu/mips/system/system.html).

During the latter part of the IFC and afterwards a flux tower “Anchor station” was
installed at the Greig Farm in Dutchess County, NY (G on Figure 3.7), and included
wind/temperature profile measurements, pressure, radiation, fluxes, soil
moisture/temperature, and precipitation. A tipping bucket rain gauge (Texas Electronics
model TR-525M) was used at the site, recording at 0.2mm resolution. Low-level wind
profiles were obtained from two sodars (Remtech PA0 and PA1) at Schodack Island State
Park (S on Fig. 3.7) during the IFC period. High-resolution sounding data were obtained
from the Albany National Weather Service (NWS) during the IFC period for the 00Z and
12Z daily launches. Additional launches were conducted during the IFC period. A
NOAA Environmental Technology Laboratory (ETL) wind profiler was installed at Schenectady Airport (L on Fig. 3.7). Finally the University of Wyoming King-Air aircraft was deployed during the IFC period, with over twenty flights in the HVAMS domain and collecting data including temperature, humidity, flight-leg fluxes, and aerial imagery.

Additional data sources to supplement the HVAMS data were used for the study. The Albany and Poughkeepsie Automated Surface Observing Stations (ASOS; A1 and A2 on Fig. 3.7) provided standard weather station data at one-minute intervals, with precipitation data recorded at 15-minute intervals. Daily precipitation data were obtained from the NWS Cooperative weather observers network, with reporting stations spread throughout the Hudson Valley-Catskill region. (COOP; stations marked as “c” on Fig. 3.7), and were useful in supplementing storm-total precipitation. The New York City Department of Environmental Protection (NYCDEP) operated a network of surface weather stations over the Catskill region, from which data were obtained for the IFC period. NYCDEP station names and locations are listed in Table A1.3, with locations plotted on the map on Figure 3.9. Hourly measurements of precipitation, temperature, humidity, wind, pressure, and solar radiation were included in the dataset.
Figure 3.9: NYCDEP stations in the Catskill region, denoted as the sites with six-character identifiers on the western side of the map. The Albany radar site is labeled KENX, and the Millbrook weather station MBK. Other stations are as in Fig. 3.7.

Hourly data were also available from a weather station located at the Institute of Ecosystem Studies in Millbrook, Dutchess County, NY (station MBK on Fig. 3.9). The site is located at 41.7858°N, 73.7414°W at an elevation of 128.0 m. Meteorological variables measured there included temperature, wind, humidity, precipitation, and radiation. Precipitation was measured with using a Belfort Instrument Co. Universal Recording rain gauge, Series 5-780, with a sensitivity of ±1.3 mm. Specifics regarding other Millbrook instrumentation can be found at the Institute for Ecosystem studies website (http://www.ecostudies.org/emp_meteorological.html).

Radar data in the form of level II reflectivity and radial velocity data were obtained during the IFC period from the Albany NWS KENX WSR-88D radar site.
(station KENX on Fig. 3.9). The radar site is located at 42.5864°N, 74.0639°W at an elevation of 557.0 m. Using the WSR-88D Algorithm Testing And Display System (WATADS; NSSL 2000) support software, regional event precipitation maps were produced as well as wind profiles over given locations in the radar domain.

There were 51 streamflow stations (listed in Table A1.4; locations shown as dots on Fig. 3.7) operated by the United States Geological Survey (USGS) that recorded streamflow data at 15-minute intervals. Three of the stations are located in the Hudson Valley, with the remainder in the Catskill Mountain region. The valley station elevations ranged between 10 m and 56 m with the valley station drainage areas ranging between 463 and 1779 km$^2$. The mountain stations range in elevation from 189 m and 628 m. Although upland station drainage areas range from 2 to 1100 km$^2$, over 30 watersheds have drainage areas under 200 km$^2$. We obtained data from these stations from day of year 232 (August 20) to day of year 319 (November 15) in 2003. Four additional streamflow stations located in the study region were excluded because of large amounts of flow regulation.

Topographic data were obtained from the USGS National Elevation Dataset (NED), which were merged quadrangle Digital Elevation Model (DEM) files at 30-m resolution for the Catskill-Hudson Valley Region.

Model output was available from the University at Albany Atmospheric Sciences Research Center (ASRC) Air Quality Forecast Modeling System (AQFMS; Cai 2006). This model uses the University of Athens 12-km ETA-SKIRON model (Nickovic et al. 2001) to drive the meteorology. This 12-km ETA model gives hourly forecast output for
meteorological variables at 14 levels ranging from the surface to 4000m. The 12-km ETA grid-cell coverage over the HVAMS domain is shown (Figure 3.10).

Gridded precipitation data was available through the NOAA Climate Prediction Center Morphing (CMORPH) data product (Joyce et al. 2004). The CMORPH data are a
satellite-based passive microwave precipitation estimate, available at 8- and 40-km resolution. The CMORPH 8-km and 40-km grid maps over the HVAMS domain are shown (Fig. 3.11 and Fig. 3.12 respectively). Given the complex topography of the region and the effect this has on spatial precipitation totals, output from both grid resolutions are analyzed.

Figure 3.11: Map of the 8-km CMORPH model grid cell coverage over the HVAMS domain. Station labels are as in Figure 3.7.
Figure 3.12: Map of the 40-km CMORPH model grid cell coverage over the HVAMS domain. Station labels are as in Figure 3.7.

3.3 Methods

3.3.1 Watershed boundary determination

For each USGS stream-gauging station, the drainage area for the watershed upstream of the gauging point is reported. Although there are available data containing the major regional drainage-basin boundaries (Figure 3.5), these drainage basin boundaries are not detailed enough to outline the contributing drainage area at each individual stream gauge site. Furthermore, there are no watershed boundaries available for these individual watersheds in electronic form. Therefore, using the USGS NED topographic data, watershed boundaries were drawn for the each of the available stream-gauging stations in the Catskill-Hudson Valley region using the Open Source Geographic
Resources Analysis Support System Geographic Information System software (GRASS-GIS; Neteler and Mitasova 2004). This process was done in two steps. First, the GRASS function *r.watershed* was run using the elevation data as input. This resulted in a drainage map that contained the drainage direction for each grid cell on the elevation map. To expedite this calculation, the grid resolution for the input elevation map was resampled from 30 m to 90 m. Testing both resolutions on a small portion of the elevation-map domain showed very little effect on the final watershed boundaries and drainage areas. The second step was to run the GRASS function *r.water.outlet* using the drainage-direction map as input and specifying as the drainage outlet points the coordinates of each stream-gauge station to generate watershed boundary maps for each individual stream-gauging station. These watershed boundary maps were then merged to result in the regional watershed boundary map shown in Figure 3.13. Drainage areas calculated by GRASS were comparable to those drainage areas reported in the USGS stream gauge descriptions.
Figure 3.13: Map of the USGS stream gauge sites (dots) and watershed boundaries in the Catskill-Hudson Valley region. Elevation (m) is shaded. Station labels are as in Figure 3.7.

3.3.2 Watershed response stages

The hydrologic response of a watershed to precipitation events is broken down into three stages with respect to the streamflow hydrograph (Figure 3.14): a) the peak in streamflow following the input precipitation; b) the streamflow recession, the decrease in streamflow following the precipitation peak; c) the evapotranspiration-driven diurnal streamflow fluctuations that may occur after the streamflow has receded to its baseflow.
While diurnal oscillations correspond to relatively small contributions to streamflow, they do indicate the presence of evapotranspiration. The same processes apply to the near-surface soil moisture. In this study we examine the input watershed precipitation and the recession stage of the watershed response for streamflow and soil moisture.

![Diagram showing the stages of streamflow response to rainfall events. The same process applies to soil moisture. In this study we focus on the input precipitation and recession components of the response.](image)

**Figure 3.14**: Diagram showing the stages of streamflow response to rainfall events. The same process applies to soil moisture. In this study we focus on the input precipitation and recession components of the response.

### 3.3.3 Streamflow peaks

The streamflow peak is defined as the maximum value in the stream hydrograph following rainfall from the 15-minute stream discharge data (Fig. 3.14). Statistics for the magnitude of the peak streamflow, time to peak streamflow following precipitation onset, and ratio of the streamflow peak to event precipitation were kept for the network stations, with the aim of detecting any changes in these characteristics coincident with the reduced evapotranspiration that occurs around the time of leaf senescence.
3.3.4 Recession

The streamflow recession time \( \tau \) is defined as the streamflow e-folding time to return to base flow the following a rain event (Fig. 3.14). Federer (1973) used 12 years of data from the 42-ha Hubbard Brook, New Hampshire watershed and found streamflow recession proceeds more quickly with the onset of transpiration in the spring and slows with the leaf drop in autumn. Streamflow recessions were found to proceed more quickly in the spring in a large network of U.S. east-coast stations (Czikowsky and Fitzjarrald 2004). Streamflow recession times were calculated from streamflow discharge data using \( \tau \). Four rainfall events were chosen for analysis; day of year 245 (September 2); day of year 266 (September 23); day of year 270 (September 27); and day of year 302 (October 29). These were events with precipitation throughout the network and with no precipitation in the days immediately following the event. For these events, network-wide averages of streamflow recessions were compiled.

From the HOBO weather stations, near-surface soil moisture data were compiled and analyzed for the 2003-2005 time period to look for seasonal changes in soil moisture recessions. For the soil moisture recessions in this study, to maximize the number of recessions available in this limited dataset without fitting, we report recessions for soil moisture data using a shorter threshold point of \( \frac{1}{0.4} \tau \), \( \frac{1}{0.4e} \) of the soil moisture peak value from a precipitation event (Fig. 3.14).

3.3.5 Diurnal streamflow amplitude

A well-defined diurnal streamflow oscillation is observed in some small watersheds during dry periods in the growing season. During the day, transpiring
vegetation draws upon the groundwater supply that feeds the baseflow of a gaining stream, thereby reducing the stream inflow and total streamflow. Transpiration is at a minimum at night, resulting in increased stream inflow and total streamflow. The signal is slightly asymmetric; a gradual nighttime to morning rise is followed by a more abrupt afternoon to evening decline in streamflow (Lundquist and Cayan 2002). However, approximating the diurnal streamflow signal using a simple, symmetric sine curve has been found to be adequate for determining the presence and amplitude of the diurnal streamflow signal (Czikowsky and Fitzjarrald 2004).

A station was deemed to exhibit the diurnal streamflow signal using the following objective procedure, which identified the same cases as did a subjective review of the data. First, three-day windows of streamflow data were least-squares fitted using a simple sine curve, with the trend, amplitude, and phase statistics kept. Next, precipitation or recession periods were removed by retaining only the data for which the sum of the residuals normalized by streamflow was less than one, an objective way to retain the data that best fit the sine curve. Streamflow periods sampled with diurnal variations greater than or equal to two percent of the total streamflow were considered to have the diurnal streamflow signal.

A review of the precipitation and streamflow data resulted in the selection of four periods to examine the presence of the diurnal streamflow signal. Each period was in at least a four-day span for which no appreciable rainfall was observed in the network. Four time periods were examined: Period I (day of year 234-236, August 22-24); Period II (day of year 252-254, September 9-11); Period III (day of year 281-283, October 8-10); and Period IV (day of year 296-298, October 23-25).
3.3.6 Model studies

The BROOK watershed model (Federer 1995; Federer et al., 2003) simulates the surface water budget for a small, catchment-scale size watershed with a uniform land cover. The model was run with a 15-minute time step. In this model, potential evapotranspiration is calculated using the Shuttleworth and Wallace (1985) revision of the Penman-Monteith equation, separating evapotranspiration into bare-soil evaporation and plant transpiration components. The evapotranspiration algorithm is run at a daily time-step. Several model parameters that may be changed depending on land-cover type and phenological state. These include LAI, albedo, canopy height, surface roughness, and leaf stomatal conductance. Net throughfall may infiltrate into the surface soil layer and move vertically according to Darcy’s Law and the Richards equation for saturated or unsaturated flow. In the model, Darcy’s Law is used to describe vertical water flow (infiltration) through the soil at a point. Darcy’s Law for vertical unsaturated flow is:

\[
q_z = -K_h(\theta) \left[ 1 + \frac{d\psi(\theta)}{dz} \right]
\]

(3.1)

where \( q_z \) is the volumetric water flow rate in the vertical direction

\( K_h \) is the hydraulic conductivity of the soil

\( \theta \) is the soil moisture content

\( \psi \) is the pressure head

The Richards equation is derived by combining Darcy’s Law with the conservation of mass is:

\[
\frac{\partial \theta}{\partial t} = -\frac{\partial K_h(\theta)}{\partial z} + \frac{\partial}{\partial z} \left[ K_h(\theta) \frac{\partial \psi(\theta)}{\partial z} \right]
\]

(3.2)
where \( \theta \) is the soil moisture content

\[ K_h \] is the hydraulic conductivity of the soil

\( \psi \) is the pressure head

In the model, net throughfall may also move directly into deeper soil layers or streamflow as stormflow in a variable source area or through macropore flow. Several soil layers are included, and the model contains linear groundwater storage. The flowchart for the BROOK model is shown in Figure 3.15.

Figure 3.15: BROOK model flowchart (adapted from Federer 1995).
The BROOK model hydrograph is made up of four components, source-area flow, bypass flow, downslope flow, and groundwater flow. Source-area flow is runoff generated from areas that become saturated during storms, and includes overland flow (direct flow over the land into the stream). This is a fast-response flow component, also termed quickflow. Bypass flow is runoff generated during storms that infiltrates into the soil and enters a stream through large subsurface channels, or macropores. Downslope flow is water moving vertically through the soil column, eventually reaching the stream channel. Groundwater flow is the water that replenishes the stream’s baseflow.

The BROOK model was run to conduct a sensitivity study on watershed response to changes in input precipitation amount induced by orographic effects and changes in vegetation state due to phenology. Four model runs were performed, with the following adjustments made to the model forcing input parameters. The base case contained input precipitation from station H2 (Freehold; located in the precipitation shadow) with leaves. Run 2 had base precipitation, with no leaves, which was done by setting LAI to 0 in the model. Run 3 used twice the base precipitation as input with leaves. This approximately gives the amount of precipitation received at station H3 (East Jewett; located in the peak-precipitation region). Run 4 used twice the base precipitation without leaves.
3.4 Analysis

3.4.1 September 23, 2003 precipitation case

3.4.1.1 Synoptic situation

At 0000Z on September 23, 2003, a 993-mb low-pressure center was located over the northern shore of Lake Huron moving north-northeastward, with an occluded front extending south to a triple point situated near the northwestern Pennsylvania shoreline of Lake Erie. A south-southeasterly surface flow was felt at Albany, along with other observing stations in the Northeast US ahead of the system. By 1200Z the same day, the surface low had strengthened to 988 mb crossing the James Bay region of Ontario, with the attending occluded front extending southward to a triple-point just to the southwest of Albany, New York (Fig 3.16). A cold front extended south of the triple point, west of the mid-Hudson Valley, which was then entering the warm sector.
At 0000Z on September 23, 2003, the Albany sounding showed winds veering with height, starting with south-southeasterly winds near the surface veering to south-southwesterly with a speed of approximately 15 m s$^{-1}$ at the 850mb level (not shown). The 1200Z sounding on September 23 was not available at Albany. The northeastern US 850mb analysis on that day indicated a south-southwesterly flow over the region at that level (Fig. 3.17).
Figure 3.17: Northeastern US 850 mb analysis for 1200Z on September 23, 2003.

The winds aloft at Albany were estimated using the velocity azimuth profile (VAD), a product from the National Weather Service Doppler Radar (NWS WSR-88D) KENX Albany radar site. During the 1100Z hour the VAD profiles showed south-southeasterly winds at the lowest levels veering with height to the south-southwest at a speed of 20 to 30 m s\(^{-1}\) at the 800-1200 m levels, the approximate height of the Catskill Escarpment peaks (Fig 3.18). The south-southwesterly winds increased to 30 to 40 m s\(^{-1}\) at the 1200-1500 m levels, an increase from the 0000Z sounding. Winds at this level according to earlier VAD profiles had increased after 0500Z, near the onset time of the precipitation in the network (Fig 3.18), with subsequent VAD profiles showing a similar pattern to the 1100Z hour profiles.
By 0000Z on September 24, 2003, the frontal system had cleared the New England coast, leaving a west-northwesterly surface flow at Albany in its wake.

3.4.1.2 Precipitation

Precipitation began falling at approximately 0200 GMT on September 23, 2003 over the DEP stations in the western part of the network, with precipitation overspreading the rest of the network by 0500 GMT (Fig. 3.19). Initially rainfall rates were light, only a few mm per hour at all stations. Most of the precipitation fell in a heavier band over a 2-to-4 hour period at each station, spanning the times from 0930 to 1530 GMT. Light
amounts of rain fell after this period, ending over the entire network before 0000 GMT on September 24 (Fig. 3.19).

During the period of heaviest precipitation, hourly rainfall rates were greater than 10 mm hr\(^{-1}\) in the peak-precipitation areas of the Catskills, peaking at over 20 mm hr\(^{-1}\) at
East Jewett and Slide Mountain (station DNM148 on Fig. 3.9). Meanwhile, rainfall rates in the shadowed region were much less, with maximum rates of less than 5 mm hr\(^{-1}\) at Freehold. Throughout the event, 850-mb winds were out of the south-southwest between 20 and 25 m s\(^{-1}\), shifting to westerly near the end of the event (Fig. 3.19, bottom panel). Surface winds at Slide Mountain were southeasterly early in the event, shifting to the south-southeast and then to the south during the heaviest period of precipitation. Finally, winds shifted to westerly following the frontal passage near the end of the event. At Freehold, winds were near calm early in the event, then easterly during the period of heaviest rainfall, shifting to west-northwest near the time the precipitation ended.

Storm rainfall totals ranged from approximately 20 to 70 mm. The southwestern Catskills received the greatest amount of rainfall, over 60 mm, with the entire Catskill region received at least 50 mm of rainfall (Fig. 3.20). Slide Mountain recorded 61.7 mm of precipitation and East Jewett 50.4 mm for the event. On the northeastern leeward side of the Catskill Escarpment at its base, rainfall amounts decreased considerably to a minimum of 23.8 mm at Freehold, located just 15 km northeast of East Jewett. The precipitation shadow extended northward, with a broad area receiving less than 35 mm of rainfall west of Albany and in the lower Schoharie Creek drainage basin. The southwest-to-northeast orientation of the rain shadow followed the 850 mb wind crossing the escarpment for the event.
3.4.1.3 Comparison of observations with remote sensing data

The KENX radar-estimated storm total precipitation for the September 23 event (Fig. 3.21) shows general agreement with observations with regards to the placement of the precipitation shadow. However, the scatterplot of the KENX radar estimated rainfall against observations show that the radar underestimated the observed precipitation totals
Radar estimates of rainfall are based on a radar reflectivity ($Z$) to rainfall rate ($R$) relationship ($Z-R$) that differs based on precipitation type (e.g. Rosenfeld et al. 1993). Operationally, the $Z-R$ relationship at NEXRAD radar sites, including KENX, is changed seasonally from convective in the summer to stratiform in the fall and winter. During storms with stratiform precipitation, the radar underestimates the storm total precipitation if it was set with a convective $Z-R$ relationship (Ulbrich et al. 1999). Based on the underestimation in this case it appears that the KENX radar may still have been set with the convective $Z-R$ relationship during this stratiform precipitation event.

Figure 3.21: Albany radar (KENX) estimated storm total precipitation (mm) for the period 0000Z on September 23, 2003 to 0000Z on September 24, 2003. HVAMS deployment stations are labeled.
Figure 3.22: Scatterplot of observed rainfall ($P_{\text{obs}}$) vs. KENX radar-estimated rainfall ($P_{\text{radar}}$) for the September 23, 2003 precipitation event, with the regression line shown. The regression equation is $P_{\text{radar}} = 0.57 P_{\text{obs}} + 9.17$, $r^2 = 0.70$.

The CMORPH 8-km estimated total storm rainfall shows a precipitation shadow; however it is placed in the incorrect location over the Catskills in the observed peak-precipitation instead of on the leeward side of the Catskills, and moreover has the wrong spatial extent (Fig. 3.23). For example, CMORPH places the East Jewett station in the rain shadow. Secondly, the magnitude of the precipitation shadow depicted by CMORPH is much less than the observed magnitude of the precipitation shadow, with CMORPH showing a difference of less than 10 mm between the shadowed and peak-precipitation areas. The observed difference from the shadow to peak-precipitation area was about 30 mm.
Figure 3.23: CMORPH 8-km estimated rainfall (mm) for the September 23, 2003 precipitation event. Station labels are as in Fig. 3.7.

The 40-km CMORPH estimated storm total precipitation map shows similar, but smoother features than does the 8-km resolution map (Fig. 3.24).
Scatterplots of the CMORPH 8- and 40-km storm total rainfall (Fig. 3.25 and Fig. 3.26 respectively) show no positive correlation with the observed rainfall for the September 23 event. This is likely because of the improper placement of the rain shadow to the south in the Catskills than in its observed location to the north on the leeward side of the Catskills. Although qualitatively CMORPH crudely resembles the general spatial pattern of measured precipitation amounts, CMORPH does not quantitatively agree with the measured precipitation at each observation point, and the observed position of the rain shadow is missed by CMORPH.
Figure 3.25: Scatterplot of observed rainfall ($P_{obs}$) vs. CMORPH 8km-estimated rainfall ($P_{CM8}$) for the September 23, 2003 precipitation event, with the regression line shown. The regression equation is $P_{CM8} = -0.02P_{obs} + 36.32$, $r^2 = 0.00$. 
Figure 3.26: Scatterplot of observed rainfall ($P_{obs}$) vs. CMORPH 40km-estimated rainfall ($P_{CM40}$) for the September 23, 2003 precipitation event, with the regression line shown. The regression equation is $P_{CM40} = -0.06 \, P_{obs} + 38.04$, $r^2 = 0.03$.

3.4.1.4 Watershed precipitation

Precipitation in the individual watersheds was determined by averaging the storm total precipitation for all stations located in a given watershed. Watersheds were then classified into three categories based on the amount of precipitation received. Watersheds with greater than 60% of the maximum precipitation in the network were classified in the peak-precipitation region. Watersheds receiving between 40 and 60% of the maximum network precipitation were determined to be in a transition area between the peak precipitation and the rain shadow; and watersheds with less than 40% of maximum network precipitation were placed into the rain shadow (Fig. 3.27). Total
watershed precipitation for the September 23 event shows the upland Catskill watersheds in the peak-precipitation region, with the lowland watersheds to the north in the precipitation shadow.

Figure 3.27: Map of HVAMS watershed locations with respect to the precipitation shadow for the September 23, 2003 event. Watersheds within the precipitation shadow (defined as those watersheds receiving less than 40% of the maximum observed precipitation in the network) are denoted by the upper-left to lower-right diagonal lines. Watersheds outside of the precipitation shadow (defined as those watersheds receiving greater than 60% of the maximum observed precipitation in the network) are shown by the lower-left to upper-right diagonal lines. Watersheds located in the precipitation shadow transition area (defined as those watersheds receiving between 40 and 60% of the maximum network-observed precipitation) are indicated by the vertical lines.
For the Schoharie Creek drainage basin, (Table 3.1) the upstream watersheds in the basin (the watersheds with the smallest watershed areas in Table 3.1) received much greater amounts of precipitation than those downstream. The most upstream portion of the basin lies entirely in the transition precipitation region, with the percentage of the basin in the rain shadow increasing steadily going downstream until the majority of the basin is in the rain shadow at the most downstream gauge at Burtonsville.

Table 3.1: Schoharie Creek drainage basin rain-shadow coverage statistics for the September 23, 2003 precipitation event.

<table>
<thead>
<tr>
<th>Schoharie Ck watershed information</th>
<th>Percentage of watershed in</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>USGS ID</strong></td>
<td><strong>Gauge Location</strong></td>
</tr>
<tr>
<td>01349705</td>
<td>Lexington</td>
</tr>
<tr>
<td>01350000</td>
<td>Prattsville</td>
</tr>
<tr>
<td>01350101</td>
<td>Gilboa</td>
</tr>
<tr>
<td>01350180</td>
<td>North Blenheim</td>
</tr>
<tr>
<td>01350355</td>
<td>Breakabeen</td>
</tr>
<tr>
<td>01351500</td>
<td>Burtonsville</td>
</tr>
</tbody>
</table>

3.4.1.5 Streamflow peak analysis

Event streamflow peaks over the entire Catskill region exceeded 20 mm, normalized to the watershed area (Fig. 3.28). There were areas in the southwestern, central, and eastern Catskills where peaks exceeded 40 mm. The highest streamflow peak was recorded at Sugarloaf Brook, located in the eastern Catskills in Tannersville, at 158.5 mm. For gauged locations in the precipitation shadow region, streamflow peaks were at 10 mm or less.
For the Schoharie Creek drainage basin, streamflow peaks were much higher upstream in areas located mostly in the transition precipitation region, with the Lexington and Prattsville gauges recording peak streamflows of 54.8 and 28.3 mm, respectively (Table 3.2). Further downstream in the basin, locations whose watersheds were predominately in the rain-shadow region observed much lower streamflow peaks. The
Breakabeen and Burtonsville gauges recorded peak streamflows of 10.1 and 7.1 mm respectively during the event. Streamflow peak to watershed precipitation ratios were highest at Lexington (1.09), decreasing to a low of 0.21 at Burtonsville. A greater overland flow contribution to the streamflow likely accounts for the higher ratios observed upstream. Times to streamflow peak from the start time of precipitation ranged from 9 hours at the most upstream gauge at Lexington, increasing downstream to just over one day at Burtonsville.

Table 3.2: Schoharie Creek drainage basin streamflow peak statistics for the September 23, 2003 precipitation event. The peak to precip ratio is the ratio of the streamflow peak to the average watershed precipitation. The time to streamflow peak is defined as the time required to reach peak streamflow following the onset of precipitation.

<table>
<thead>
<tr>
<th>Schoharie Ck watershed information</th>
<th>Watershed</th>
</tr>
</thead>
<tbody>
<tr>
<td>USGS ID</td>
<td>Gauge Location</td>
</tr>
<tr>
<td>--------</td>
<td>----------------</td>
</tr>
<tr>
<td>01349705</td>
<td>Lexington</td>
</tr>
<tr>
<td>01350000</td>
<td>Prattsville</td>
</tr>
<tr>
<td>01350101</td>
<td>Gilboa</td>
</tr>
<tr>
<td>01350180</td>
<td>North Blenheim</td>
</tr>
<tr>
<td>01350355</td>
<td>Breakabeen</td>
</tr>
<tr>
<td>01351500</td>
<td>Burtonsville</td>
</tr>
</tbody>
</table>
3.4.2 October 27, 2003 precipitation case

3.4.2.1 Synoptic situation

At 1800Z on October 26, 2003, a 1013-mb low pressure area was situated midway along the north shore of Lake Ontario moving eastward. A cold front extended southwest through westernmost New York and northwest Pennsylvania with a warm front snaking through central New York and northeastern Pennsylvania. Surface winds at Albany were out of the south-southeast ahead of the system. By 0000Z on October 27, 2003, the frontal system had moved east with a 1012-mb low pressure center analyzed over Lake Champlain with a trailing cold front across central New York and Pennsylvania. The south-southeasterly surface flow continued at Albany as precipitation began in the Hudson Valley-Catskill region at this time. At 1200Z on October 27, 2003, about halfway through the precipitation event, the frontal boundary was positioned from northeast to southwest through northern New England, and extending southwest across the Albany region and into northeast Pennsylvania (Fig. 3.29). Surface winds at Albany were light southerly.
At 1800Z on October 27, 2003, the stationary front remained nearly in the same position, oriented northeast to southwest across the Albany region, with a 1004-mb low pressure center analyzed near Monticello, New York. Surface winds at Albany were out of the southeast, and precipitation continued over the region. The frontal system then began to move through the region, with precipitation ending in the region by 0000Z on October 28, 2003. At 1200Z on October 28, 2003, the frontal system was located well offshore and high-pressure building back into the region.

Near the onset of the precipitation event, the 0000Z Albany sounding on October 27, 2003 showed a veering wind profile with height starting from the south below 300 m, shifting to a southwesterly direction at 1500 m (not shown). Wind speeds were steady at
about 15 m s\(^{-1}\) from 300 to 1500 m height. The 1200Z Albany sounding showed a similar wind profile, with winds starting from the south at about 300 m, veering to the south-southwest at 1500 m (above 1200 m) (Fig. 3.30). Wind speeds were 8 m s\(^{-1}\) at 300 m increasing to 15 m s\(^{-1}\) in the 1200 to 1500 m layer. The 850 mb flow at 1200Z on October 27 was out of the south-southwest over most of the northeastern US (Fig. 3.31).

Figure 3.30: Albany 1200Z sounding for October 27, 2003.
The VAD profile over Albany for the hour surrounding 1300Z showed southerly winds near the surface near 10 m s$^{-1}$, shifting to the south-southwest near 15 m s$^{-1}$ at the 1500 m level (Fig. 3.32). This wind profile remained consistent for the majority of the event (not shown).
3.4.2.2 Precipitation

Rain began falling throughout the network at approximately 0000 GMT on October 27, 2003 (Fig. 3.33). This was a long-duration event, with steady precipitation falling for the entire day, but with less intensity than the September 23 event. Rainfall rates were highest during two periods in the event, between 0430 and 0700 GMT and between 1630 and 1900 GMT. During the second band of heavier precipitation, the highest hourly rainfall totals were observed. Areas in the peak-precipitation region in the Catskills had peak rainfall rates of 10 to 15 mm hr⁻¹. Meanwhile, stations in the precipitation shadow recorded much lower rainfall rates, generally < 5 mm hr⁻¹. Freehold had a peak rainfall rate for one hour near 8 mm hr⁻¹. Following the second
band of heavier precipitation, rainfall tapered off and ended in the early hours of October 28.

Figure 3.33: Top panel: Rainfall for the PAM and HOBO weather stations for the October 27-28, 2003 event, covering the period from 0000Z on October 27, 2003 to 1500Z on October 28, 2003. Bottom panel: First row: Hourly ETA-estimated 850-mb winds for the October 27-28, 2003 event for the grid cell containing the Slide Mountain station. A full wind barb line indicates 10 m s\(^{-1}\), with a half-barb indicating 5 m s\(^{-1}\). Second row: Hourly surface winds observed at Slide Mountain over the same time period.
ETA-estimated winds at 850 mb were from the southwest at the beginning of the event, becoming more southerly and increasing to over 30 m s\(^{-1}\) during the second period of heavy precipitation (Fig 3.33, bottom panel). Winds shifted to westerly after the precipitation stopped. Surface winds at Slide Mountain were also from the southwest early in the event, shifting to the south during the heaviest precipitation prior to the frontal passage. Surface winds at Freehold were near calm during rainfall.

Storm total rainfall ranged from approximately 30 mm to over 100 mm. The highest amounts were recorded in the southwestern Catskills, over 100 mm, with the entire Catskill region receiving over 80 mm of rainfall (Fig. 3.34). Slide Mountain received 86.6 mm, with East Jewett recording 91.4 mm of rainfall. A well-defined rain shadow exists to the north-northeast of the Catskill Escarpment, following the 850 mb event wind. Freehold received 53.0 mm, with the minimum precipitation totals for the event occurring further north into the rain shadow near Thacher Park (H4), which received 31.4 mm of rainfall. The longer duration of this event likely contributed to the greater range in precipitation amounts from the peak-precipitation region to the rain-shadow area than the September 23 event.
Figure 3.34: Rainfall for the October 27-28, 2003 event, covering the period from 0000Z on October 27, 2003 to 1500Z on October 28, 2003. Station labels are as in Fig. 3.7.
3.4.2.3 Comparison of observations with remote sensing data

The KENX radar-estimated storm total precipitation for the October 27 event (Fig. 3.35) depicts the general position of the peak-precipitation region and rain shadow. Two defined peak regions of precipitation are seen in the radar estimate as in the observations, with the rain shadow downwind of the escarpment. However, the KENX radar overestimated the magnitude of the precipitation shadow. The radar indicates a region in the precipitation shadow west of the Thacher Park site (H4) with precipitation totals of around 15 mm instead of around the observed 30 mm. Furthermore, the KENX radar overestimated the precipitation in the peak precipitation region, with peak radar-
estimated precipitation at over 130 mm, whereas the highest observed precipitation total in the network was 108 mm. Inconsistency between estimated against observed precipitation totals show the problems and limitations present in the radar calibration. Also note the areas of beam blockage present in the estimate heading southwestward through the Catskills. The plot of observed vs KENX radar rainfall shows (Fig. 3.36) appreciable scatter, with the r-squared value of the regression being 0.52.

![Figure 3.36: Scatterplot of observed rainfall ($P_{\text{obs}}$) vs. KENX radar-estimated rainfall ($P_{\text{radar}}$) for the October 27, 2003 precipitation event, with the regression line shown. The regression equation is $P_{\text{radar}} = 0.88 P_{\text{obs}} - 1.52$, $r^2 = 0.52$.](image)

The KENX rainfall in this case is closer to observations in this case than in the September 23 case, indicating the radar Z-R relationship may have been set back to stratiform precipitation for the fall/winter season.
The CMORPH 8- and 40-km estimated storm total rainfall (Fig. 3.37 and Fig. 3.38 respectively) performed better in estimating the magnitude of the precipitation shadow than it did in the September 23 case, with the CMORPH estimates in the same range as the observations. However, the placement of the shadow in both the 8- and 40-km data is incorrect. The peak-precipitation region depicted by CMORPH extends too far north and northeast into the observed rain-shadow area.

Figure 3.37: CMORPH 8-km estimated rainfall (mm) for the October 27, 2003 precipitation event.
For the October 27 event, scatterplots of the 8- and 40-km CMORPH estimated rainfall (Fig. 3.39 and Fig. 3.40 respectively) against observed rainfall show a positive correlation, with $r^2$ values of 0.48 and 0.51 respectively. CMORPH tended to overestimate the lower observed rainfall totals and underestimate the higher observed rainfall amounts.

These comparisons of CMORPH-estimated to observed precipitation illustrate some of the current shortcomings in the morphing techniques used to derive this product, and possible limitation on its effectiveness in certain precipitation types, such as the stratiform precipitation that occurred in these cases.
Figure 3.39: Scatterplot of observed rainfall ($P_{\text{obs}}$) vs. CMORPH 8km-estimated rainfall ($P_{CM8}$) for the October 27, 2003 precipitation event, with the regression line shown. The regression equation is $P_{CM8} = 0.35 P_{\text{obs}} + 47.74$, $r^2 = 0.48$. 

```
slope 0.35  int 47.74  r^2 0.48
```
3.4.2.4 Watershed precipitation

The watershed precipitation distribution map (Fig. 3.41) is similar to the map for the September 23 event, with the exception that the watersheds experiencing near-peak precipitation amounts extended further north. In the October 27 case, the most upstream gauged watershed in the Schoharie Creek drainage basin at Lexington was entirely in the peak-precipitation region (Table 3.3). The amount of the drainage basin in the precipitation shadow quickly changes as one heads downstream. At Gilboa, 25% of the watershed is in the rain shadow, and at the most downstream gauge at Burtonsville, nearly 75% of the watershed is in the rain shadow.

Figure 3.40: Scatterplot of observed rainfall ($P_{obs}$) vs. CMORPH 40km-estimated rainfall ($P_{CM40}$) for the October 27, 2003 precipitation event, with the regression line shown. The regression equation is $P_{CM40} = 0.36 P_{obs} + 47.71$, $r^2 = 0.51$. 

Observed vs. CMORPH 40–km rainfall: 10/27/2003

<table>
<thead>
<tr>
<th>Observed rainfall (mm)</th>
<th>CMORPH-estimated rainfall (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td>40</td>
<td>40</td>
</tr>
<tr>
<td>60</td>
<td>60</td>
</tr>
<tr>
<td>80</td>
<td>80</td>
</tr>
<tr>
<td>100</td>
<td>100</td>
</tr>
<tr>
<td>120</td>
<td>120</td>
</tr>
</tbody>
</table>

slope 0.36  int 47.71  $r^2 0.51$
Figure 3.41: Map of HVAMS watershed locations with respect to the precipitation shadow for the October 27-28, 2003 event. Watersheds within the precipitation shadow (defined as those watersheds receiving less than 40% of the maximum observed precipitation in the network) are denoted by the upper-left to lower-right diagonal lines. Watersheds outside of the precipitation shadow (defined as those watersheds receiving greater than 60% of the maximum observed precipitation in the network) are shown by the lower-left to upper-right diagonal lines. Watersheds located in the precipitation shadow transition area (defined as those watersheds receiving between 40 and 60% of the maximum network-observed precipitation) are indicated by the vertical lines.
Table 3.3: Schoharie Creek drainage basin rain-shadow coverage statistics for the October 27-28, 2003 precipitation event.

<table>
<thead>
<tr>
<th>Schoharie Ck watershed information</th>
<th>Percentage of watershed in</th>
</tr>
</thead>
<tbody>
<tr>
<td>USGS ID</td>
<td>Location</td>
</tr>
<tr>
<td>-----------</td>
<td>----------</td>
</tr>
<tr>
<td>01349705</td>
<td>Lexington</td>
</tr>
<tr>
<td>01350000</td>
<td>Prattsville</td>
</tr>
<tr>
<td>01350101</td>
<td>Gilboa</td>
</tr>
<tr>
<td>01350180</td>
<td>North Blenheim</td>
</tr>
<tr>
<td>01350355</td>
<td>Breakabeen</td>
</tr>
<tr>
<td>01351500</td>
<td>Burtonsville</td>
</tr>
</tbody>
</table>

3.4.2.5 Streamflow peak analysis

Event streamflow peaks over the entire Catskill region exceeded 20 mm, normalized to the watershed area (Fig. 3.42). Streamflow peaks of 80 mm or greater were recorded near the peak-precipitation area in the eastern Catskills. The highest streamflow peak was recorded at Sugarloaf Brook, located in the eastern Catskills in Tannersville, at 174.5 mm. For gauged locations in the precipitation shadow region, streamflow peaks were less than 20 mm.
Figure 3.42: Observed streamflow peaks (in mm; normalized by watershed area) following the October 27-28, 2003 precipitation event. Dots indicate the stream-gauge stations used. Station labels are as in Fig. 3.7.

For the Schoharie Creek drainage basin, the stream gauges located in watersheds located either entirely or mostly in the peak-precipitation region received much higher peak streamflows than areas further downstream, with Lexington and Prattsville recording streamflow peaks of 98.5 and 54.6 mm respectively (Table 3.4). For the stream gauging stations located in the watersheds with the greatest fraction of rain-
shadow area (Breakabeen and Burtonsville), streamflow peaks were around 10 mm.

Times to streamflow peak from the precipitation starting time were longer in the October 27 event compared to the September 23 event, as the October event was of longer duration. Streamflow peak to watershed precipitation ratios were similar to the September 23 event, with ratios of around 1 at the furthest upstream gauge, decreasing to around 0.2 downstream. Times to streamflow peak ranged from between 16 and 30 hours between Lexington and Breakabeen, with the longest time to peak observed, 39 hours, at the most downstream gauge at Burtonsville.

Table 3.4: Schoharie Creek drainage basin streamflow peak statistics for the October 27, 2003 precipitation event. The peak to precip ratio is the ratio of the streamflow peak to the average watershed precipitation. The time to streamflow peak is defined as the time required to reach peak streamflow following the onset of precipitation.

<table>
<thead>
<tr>
<th>Schoharie Ck watershed information</th>
<th>Watershed</th>
</tr>
</thead>
<tbody>
<tr>
<td>USGS ID</td>
<td>Location</td>
</tr>
<tr>
<td>01349705</td>
<td>Lexington</td>
</tr>
<tr>
<td>01350000</td>
<td>Prattsville</td>
</tr>
<tr>
<td>01350101</td>
<td>Gilboa</td>
</tr>
<tr>
<td>01350180</td>
<td>North Blenheim</td>
</tr>
<tr>
<td>01350355</td>
<td>Breakabeen</td>
</tr>
<tr>
<td>01351500</td>
<td>Burtonsville</td>
</tr>
</tbody>
</table>

3.4.3 Seasonal changes

3.4.3.1 Land cover and evapotranspiration changes

During the autumn season transition from the growing season to the dormant season, there is an abrupt, marked decrease in evapotranspiration in the forested upland
watersheds in the northeastern US. This region is largely covered by deciduous forests (U. S. Forest Inventory and Analysis; http://fia.fs.fed.us), so that the autumn season decrease in evapotranspiration represents a widespread land cover change. As evapotranspiration decreases, a smaller amount of groundwater is withdrawn, allowing more groundwater to feed into streams and result in higher streamflows. Following leaf drop in autumn, streamflow has been observed to increase even in the absence of precipitation (Doyle 1991).

In the northeastern US, the end of the growing season brings a marked, abrupt change in land cover due to leaf drop and a corresponding sharp decrease in evapotranspiration. This sharp land cover change can be observed through remote-sensing techniques. Independent estimates of phenology can be obtained through the Moderate Resolution Imaging Spectrometer (MODIS), the Normalized Difference Vegetation Index (NDVI), and Landsat, among other techniques. Depending on the method used, estimates of spring and fall onset may vary from a week to more than 10 days (Fisher and Mustard 2007). At Harvard Forest, Massachusetts, approximately 200 km east-northeast of the Catskills, the NDVI estimate of average autumn onset (defined as half the maximum senescence) is on October 12, as compared to the MODIS estimate of October 22 (Fig. 3.43). The MODIS estimate also gives a sharper result. Note the raw values that are averaged for the lines in Fig. 3.43 are extremely variable. This possible range of days in the estimate also complicates interpretation of interannual variability in the autumn onset date, which has varied more than two weeks during the 2000-2006 period according to the MODIS data, ranging from October 3 to October 19 (Fig. 3.44).
Figure 3.43: Example phenology over Harvard Forest, Massachusetts, calculated from MODIS interannual spectral mixture analysis (SMA; bottom line, diamonds) and Normalized Difference Vegetation Index (NDVI; top line, squares). Spring onset and autumn onset (half maximum green-up and senescence, respectively) are marked with vertical lines and labeled. NDVI shows consistently elevated values due to fractional coniferous cover, and smoothed green-up and senescence relative to the sharper SMA result. Adapted from Fisher and Mustard (2007, Fig. 10).

Figure 3.44: MODIS interannual phenology for regions in the Northeastern US. The MODIS sample region covered of southern and central New England as well as adjacent parts of New York State, Pennsylvania, and New Jersey. Spring onset and autumn onset (half maximum green-up and senescence, respectively) are marked with vertical lines and labeled. Detail from Fisher and Mustard (2007, Fig. 2c).
Over the Hudson Valley-Catskill region, this autumn land cover transition was observed through a series of aerial photographs taken during flights by the University of Wyoming King-Air aircraft in October 2003. The location for the land-cover photo sequence is shown in the aircraft flight-track (Fig. 3.45), and is representative of the land cover change that occurred throughout the region.

Figure 3.45: Flight track for the University of Wyoming King-Air aircraft on the afternoon of October 11, 2003 over the Catskill-Hudson Valley region. The area pictured in the land-cover photo sequence in Figure 3.46 is indicated by the arrow. The coordinates for the land cover photo area were 41.8912°N, 74.0728°W.
The aerial photographs of the land cover show that on October 3, the foliage was still green (Fig. 3.46). The autumn land cover transition can be seen in the October 11 photo, with the leaf-color change apparent. Leaf drop was nearly complete by October 25, with fully bare deciduous vegetation and completion of the autumn transition in the October 30 photo.

Figure 3.46: Aerial photographs of land cover from the University of Wyoming King-Air aircraft over the area denoted in Figure 3.45 for October 3, 2003 (upper left), October 11, 2003 (upper right), October 25, 2003 (lower left), and October 30, 2003 (lower right).

PAM station network-averaged evaporation shows the sharp drop in evaporation that occurs during the early-to-mid October leaf-senescence period (Fig. 3.47). Midday values of latent heat flux decrease from around 200 W m\(^{-2}\) in early September to 100–
150 W m\(^{-2}\) in early October, decreasing to below 100 W m\(^{-2}\) in late October, about half its growing-season value.

![Figure 3.47: PAM station network-averaged evaporation (W m\(^{-2}\)) for September 10, 2003 (black dots), October 8, 2003 (red dots), and October 24, 2003 (blue dots).](image)

3.4.3.2 Seasonal variation of the diurnal streamflow oscillation

We have shown the abrupt changes in land cover and evapotranspiration that occur during autumn in the study region. We seek indicators in the streamflow and soil moisture records of this transition to determine how and when the hydrological response of these watersheds is altered under changed forcing. In this section, we examine spatial and temporal changes in the prevalence of the evapotranspiration-modulated diurnal streamflow oscillation observed in these watersheds.
The presence of a diurnal streamflow oscillation signal in the network watersheds during the growing to dormant season transition is seen (Fig. 3.48). Period I in the sequence (August 22-24) falls well within the growing season. Nearly half of the stations in the network observed a diurnal streamflow oscillation during this period (Fig. 3.48a). One of the large valley stations (area > 1000 km$^2$) observed the signal during this period. All of the remaining stations observing the diurnal streamflow signal have drainage areas less than 200 km$^2$, and the mean and standard deviation of the drainage area of these stations are 48.2 km$^2$ and 39.7 km$^2$ respectively.

Period II (September 9-11) still falls within the growing season. Little change is noted as 48% of the stations in the network observed a diurnal streamflow signal for this period (Fig. 3.48b). No valley stations reported a diurnal streamflow oscillation during this period, and all of the stations observing the diurnal streamflow signal have drainage areas less than 200 km$^2$. The mean and standard deviation of the drainage area of these stations are 51.7 km$^2$ and 45.3 km$^2$ respectively.

During Period III of the sequence (October 8-10), a much lower percentage (11%) of network stations observe a diurnal streamflow signal (Fig. 3.48c). This period occurs during the transition from the growing season to the dormant season in this region, and is concurrent with ET decreasing to near dormant season values at Harvard Forest, a long-term flux measurement site considered to be representative of the region (Fitzjarrald et al. 2001). The presence of the diurnal streamflow fluctuation is confined to even smaller watersheds, with all stations observing the signal having drainage areas less than 80 km$^2$. 

154
The mean and standard deviation of the drainage area of these stations are $39.9 \text{ km}^2$ and $23.8 \text{ km}^2$ respectively.

During Period IV (October 23-25), none of the network stations observe a diurnal streamflow oscillation, an indication of the onset of the dormant season in the region (Fig. 3.48d).

Figure 3.48: (a): Streamflow stations with an observed diurnal streamflow signal (red X’s), and stations with no observed diurnal streamflow signal (black dots), Period I (day of year 234-236; August 22-24). (b): same as in (a), but for Period II (day of year 252-254; September 9-11). (c): same as in (a), but for Period III (day of year 281-283; October 8-10). (d): same as in (a), but for Period IV (day of year 296-298; October 23-25).
Histograms of stations observing a diurnal streamflow oscillation by drainage area indicate that the smallest watersheds (less than 100 km$^2$) exhibit diurnal streamflow fluctuations most frequently and longer into the autumn season than larger watersheds (Fig. 3.49). This agrees with the observation that the smallest watersheds containing the greatest riparian area fraction exhibit diurnal streamflow fluctuations most frequently during the growing season (Fig. 3.49, lower right).

Figure 3.49: Histograms of stations observing a diurnal streamflow oscillation, binned by drainage area for Period I (August 22-24, 2003; upper left), Period II (September 9-11, 2003; upper right), and Period III (October 8-10, 2003; lower left). Bottom right: Watershed area vs. the time fraction diurnal streamflow oscillations are observed (solid line) for a network of 151 stream-gauging stations in the eastern U.S. for the period 1989-2001, with the upper and lower 95% confidence intervals for the median shown. Values listed above the x-axis indicate the number of station-years included for each watershed size range. Dashed curves represent the power-law relationship between watershed-area ($A_w$) and riparian-area fraction ($F_{rip}$), $F_{rip} \sim A_w^{\beta-1}$. Labels above the curves denote $\beta$ values from 0.5 to 0.8. From Czikowsky and Fitzjarrald (2004; Fig. 12a).
3.4.3.3 **Seasonal streamflow peak analysis**

Little change in streamflow peak, time to peak, and streamflow peak–to–precipitation ratio could be attributed to seasonal change. The differences in the magnitude of streamflow peak were mainly due to the differing amounts of precipitation received during the two events, and time to streamflow peak differences were related to precipitation event duration. Streamflow peak-to-watershed precipitation ratios were in the same range for both events. These results were to be expected as the streamflow peak is portion of the streamflow hydrograph with the fastest response, and so is mostly governed by overland flow from the precipitation itself. Slower-response terms such as streamflow recession may be expected to change seasonally.

3.4.3.4 **Seasonal streamflow recession analysis**

Widespread rainfall events with sufficient streamflow data were taken and separated into two categories: a) the growing season leaf period, which contained rainfall events from September 2 and 23; b) the transition to dormant season leaf drop and no-leaf period, which contained rainfall events from September 27 and October 27. Network-wide averages of streamflow recession were taken for the two periods and are shown (Table 3.5). There was an observed increase in the streamflow recession time of approximately one-half day from the leaf period to the no-leaf period. This also indicates an abrupt shift in the hydrological conditions in these watersheds during the last week of September. Even though the land-cover transition and leaf drop has not been completed by the end of September, transpiration by the vegetation has decreased enough to show a detectable effect on streamflow response.
Table 3.5: Average streamflow recession times from discharge data for the entire watershed network, as well as those watersheds in the peak-precipitation area, transition-precipitation area, and rain shadow region (in days, mean and standard error given) for the 2003 leaf period and the 2003 no-leaf period.

<table>
<thead>
<tr>
<th>Period</th>
<th>Entire network</th>
<th>Peak-precip area</th>
<th>Transition-precip area</th>
<th>Rain shadow area</th>
</tr>
</thead>
<tbody>
<tr>
<td>Leaf period</td>
<td>0.91 (0.07)</td>
<td>0.83 (0.11)</td>
<td>0.98 (0.09)</td>
<td>0.86 (0.20)</td>
</tr>
<tr>
<td>No-leaf period</td>
<td>1.40 (0.10)</td>
<td>1.33 (0.18)</td>
<td>1.53 (0.25)</td>
<td>1.15 (0.13)</td>
</tr>
</tbody>
</table>

Average streamflow recession times were consistently longer during the no-leaf period than during the leaf period in the peak-precipitation region, transition-precipitation region, and the rain-shadow region (Table 3.5). The difference was smallest in the rain shadow region (0.28 days) and larger in the peak- and transition-precipitation regions (0.50 and 0.55 days respectively).

3.4.4 Model sensitivity study

3.4.4.1 Model run setup

We conduct a sensitivity study using the BROOK hydrological model (Federer, 1995) to assess how changes in precipitation and land cover alter the watershed response characteristics of streamflow and soil moisture peaks and recessions. Four model runs were conducted. Input meteorological variables included 15-minute precipitation, maximum and minimum temperature, solar radiation for the day, average daily vapor pressure, and average wind speed. The observations from the Freehold station were used for the input meteorological data for the four runs. The precipitation and land cover variables were changed in the four runs as follows:
Run 1: (base case): Freehold observed precipitation (in rain-shadow) at 15-minute input intervals for the September 23, 2003 case with leaves (done by setting the leaf-area index, LAI, to 6.

Run 2: Base precipitation, no leaves (done by setting LAI to 0).

Run 3: Twice the base precipitation (gives approximately the higher precipitation values seen in the network, such as East Jewett) with leaves.

Run 4: Twice the base precipitation (gives approximately the higher precipitation values seen in the network, such as East Jewett) without leaves.

Initial guidance for parameter selection was taken from looking at a simpler, first-order illustrative model of ET effects on water table and stream and stream inflow fluctuations (Czikowsky and Fitzjarrald 2004). The expression for the watershed time constant $\tau_w$ in that model is:

$$\tau_w = \frac{d_w A_{\text{rip}}}{K A_d}$$

where $d_w$ is the watershed diameter for the riparian zone and river

$K$ is the hydraulic conductivity of the soil

$A_{\text{rip}}$ is the riparian area in the watershed

$A_d$ is the drainage cross-sectional area

and the riparian area in the watershed $A_{\text{rip}}$ is equal to the product of the riparian width $d_{\text{rip}}$ and the mainstream length $l$. In that model, the dynamics of ET effects on
runoff characteristics suggest that the recession time constant and diurnal amplitude depend on the bulk watershed characteristics of riparian area and hydraulic conductivity.

For a sample Catskill watershed, Biscuit Brook (area 10 km²), using the geometric relation that the mainstream length $l$ scales to the watershed area $A_w$ as $l \sim A_w^h$; $h \approx 0.5$ (Eagleson 1970), and assuming a watershed diameter $d_w$ of 50m, a riparian width $d_{rip}$ of 30m, and a hydraulic conductivity $K$ for wet, sandy soil (approximately $1.76 \times 10^{-4}$ m s$^{-1}$), we arrive at a watershed time constant $\tau_w$ of 2.0 days, which is similar to observations.

In the BROOK model, the two parameters that can be easily changed that are analogous to the mainstream length and hydraulic conductivity are LENGTH and DRAIN, respectively. We found that the model was insensitive to changes in the LENGTH parameter. Part of the reason of the insensitivity may be that the BROOK model vegetation and rooting parameters are uniform over the whole watershed, and there are no separate riparian-zone vegetation and rooting parameters. Therefore, changes in fractional riparian area and associated water use in the watershed with mainstream length are inconsequential to the model. As input, we ultimately used as the LENGTH parameter the mainstream length for the Biscuit Brook watershed for the model runs. The DRAIN parameter for a given soil controls the outflow to groundwater from the deepest soil layer. The default value of 1 produces vertical drainage under a gravity gradient, with a value of 0 preventing drainage from the bottom of the soil column. Ranges of values for DRAIN were tested and the default value was ultimately used in the model runs, as decreasing the value of DRAIN lowered baseflow portion of the hydrograph to the point that the resulting streamflow hydrographs were unrealistic, with only contributions from the fast-response overland flow and quickflow.
The BROOK model default values for other internal soil, rooting, and drainage parameters were calibrated for Hubbard Brook, New Hampshire, a small upland watershed with similar physiographic and vegetation characteristics to the Catskill watersheds under consideration, and were used in this study.

3.4.4.2 Model results: streamflow

The BROOK model hydrograph is composed as the sum of four components, source-area flow, bypass flow, downslope flow, and groundwater flow. In these model runs, the contribution of the bypass and downslope flows was negligible. The source-area flow component dominates the early stages of the model hydrograph (Fig. 3.50). For comparison, the Biscuit Brook, NY observed streamflow is shown as representative of streamflow in the transition precipitation region during this event.
Figure 3.50: Top: Biscuit Brook, NY observed streamflow (mm on watershed) for the September 23, 2003 event. Bottom: BROOK model simulated streamflow (mm) for the September 23, 2003 event using the following run scenarios: Base precipitation with leaves (black solid line); Base precipitation without leaves (black dotted line); Doubled precipitation with leaves (blue solid line); Doubled precipitation without leaves (blue dotted line). The time for both panels is in GMT.

Model streamflow peaks for the doubled-precipitation cases (Table 3.6) were comparable to what was observed for some of the watersheds receiving the most precipitation in the network. However, model streamflow for the base precipitation cases (approximately 45 mm) was greater than what was observed in the precipitation-shadow regions (less than 20 mm). The model appears to discharge the source-area flow too quickly into the stream, as the timing of the peak is from 4–21 hours earlier than observed (see Table 3.6; Table 3.2 for observations). Second, the source-area and groundwater flow can be seen as two distinct components in the model hydrograph, not a smooth transition between the two as is observed.
Table 3.6: Streamflow peak statistics for the BROOK model runs.

<table>
<thead>
<tr>
<th>Case</th>
<th>Streamflow peak (mm on watershed)</th>
<th>Streamflow peak to precipitation ratio</th>
<th>Time to streamflow peak (hours)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Base precipitation, leaf</td>
<td>43.4</td>
<td>1.82</td>
<td>5.5</td>
</tr>
<tr>
<td>Base precipitation, no-leaf</td>
<td>47.0</td>
<td>1.97</td>
<td>5.5</td>
</tr>
<tr>
<td>2x precipitation, leaf</td>
<td>103.1</td>
<td>2.17</td>
<td>5.5</td>
</tr>
<tr>
<td>2x precipitation, no-leaf</td>
<td>110.9</td>
<td>2.33</td>
<td>5.5</td>
</tr>
</tbody>
</table>

Modeled streamflow peaks from the in-leaf cases were about 7 to 8 percent less than the corresponding no-leaf cases, with similar differences in the streamflow peak to precipitation ratio (Table 3.6). This seasonal increase in the streamflow peak to precipitation ratio was not observed (Tables 3.2, 3.4). Modeled streamflow peaks for the doubled-precipitation cases resulted were about 2.3 times greater than for the base precipitation cases, with a nearly 20 percent increase in streamflow peak to precipitation ratio (Table 3.6). Observationally, streamflow peak to precipitation ratios were higher in the upstream areas of the Schoharie Creek drainage basin receiving greater precipitation than in the downstream areas of the basin in the rain shadow (Tables 3.2, 3.4). Times to streamflow peak were insensitive to precipitation input and leaf state, suggesting the model has a fixed shape for the hydrograph.

Model streamflow recessions for all runs were much faster than observations (3.25 to 3.5 hours), a result of the source-area flow component acting too quickly in the model. Since in reality the recession process for streamflow and soil moisture operates on a longer time scale than just the source-area flow portion of the hydrograph, we examine the soil moisture recessions in the model to look at any seasonal changes that may be apparent.
3.4.4.3 Model results: soil moisture

The BROOK model was run with seven soil layers. The top soil layer covers the uppermost 8 cm of the soil profile, with the bottom layer extending to 90 cm depth. Default rooting profiles were used for all runs.

For all soil layers, the doubled precipitation cases resulted in a slightly higher peak soil moisture content than the corresponding base precipitation cases (Fig. 3.51). Total soil moisture recessions proceeded more quickly for the in-leaf cases than the corresponding no-leaf cases. The difference in recession time was greater for the in-leaf case (0.67 days) than the no-leaf case (0.24 days) (Table 3.7). The seasonal difference in observed streamflow recession over the network between the growing and dormant season, at about 0.5 days, falls within this range (Table 3.5).

![Figure 3.51: BROOK model total soil moisture (mm) for the following run scenarios: Base precipitation with leaves (black solid line); Base precipitation without leaves (black dotted line); Doubled precipitation with leaves (blue solid line); Doubled precipitation without leaves (blue dotted line).](image)

<table>
<thead>
<tr>
<th>Case</th>
<th>Recession time (days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Base precipitation, leaf</td>
<td>2.94</td>
</tr>
<tr>
<td>Base precipitation, no-leaf</td>
<td>3.61</td>
</tr>
<tr>
<td>2x precipitation, leaf</td>
<td>3.5</td>
</tr>
<tr>
<td>2x precipitation, no-leaf</td>
<td>3.74</td>
</tr>
</tbody>
</table>
For a more direct comparison with recessions from near-surface soil moisture observations, we examine the top soil layer output from the model. The doubled-precipitation model runs show a slightly higher soil moisture peak than the corresponding base precipitation cases (Fig. 3.52). Soil moisture recessions proceeded more quickly for the in-leaf runs than the corresponding no-leaf model runs.

Differences in model recession times were greater for the top soil layer than for the entire soil column. For the top soil layer, the no-leaf recession times were longer than the in-leaf recession times for the base and doubled-precipitation cases by 1.16 and 0.61 days respectively (Table 3.8).

Observations of near-surface soil moisture (at 5 cm depth) were taken at the HOBO weather stations. Soil moisture recession times were calculated at these stations for the years of 2003 to 2006, resulting in a total of 166 recessions. These were grouped into two periods, a late-growing season leaf period (July to September) and an early dormant season period (October and November). The weather station sites were grouped
by soil type, with the Schodack, Green Acres, and Chatham stations with sandy soils, and the Van Orden and Freehold stations with clay soils.

Table 3.8: Top-layer soil moisture recession times (in days, mean values with standard errors in parentheses) during leaf and no-leaf periods for all observed HOBO weather stations (OBS: All); observed HOBO weather stations in sandy soil (OBS: Sand); observed HOBO weather stations in clay soil (OBS: Clay); BROOK model runs with base precipitation (BROOK 1XP); and BROOK model runs with doubled precipitation (BROOK 2XP).

<table>
<thead>
<tr>
<th>Period</th>
<th>OBS: All</th>
<th>OBS: Sand</th>
<th>OBS: Clay</th>
<th>BROOK 1XP</th>
<th>BROOK 2XP</th>
</tr>
</thead>
<tbody>
<tr>
<td>Leaf</td>
<td>2.00 (0.10)</td>
<td>1.76 (0.11)</td>
<td>2.34 (0.18)</td>
<td>2.61</td>
<td>3.44</td>
</tr>
<tr>
<td>No-leaf</td>
<td>2.88 (0.22)</td>
<td>2.56 (0.31)</td>
<td>3.44 (0.29)</td>
<td>3.76</td>
<td>4.05</td>
</tr>
</tbody>
</table>

Observations with all HOBO weather stations combined show the top-layer soil moisture recessions for the leaf period proceeding 0.88 days faster than for the no-leaf period, within the range of the top-layer soil moisture recession differences predicted by the model (Table 3.8). Recessions at the sandy-soil sites were quicker than at the clay-soil sites, with the recession time difference between the leaf and no-leaf period 0.3 days less than at the clay sites.

3.4.4.4 Model results: evapotranspiration components

Finally, we look at the components of daily evapotranspiration for the model runs (Fig. 3.53). During the first day in which the precipitation fell on September 23, interception evaporation is the dominant component of evapotranspiration, accounting for about 88 and 60 percent of total evapotranspiration for the leaf and no-leaf cases respectively. Total evapotranspiration dropped slightly during the next day for the leaf runs, with transpiration replacing interception as the dominant evapotranspiration component. For the no-leaf cases, day 2 (September 24) evapotranspiration was entirely
from bare-soil evaporation. For the following two days (September 25 and 26), the partition of the evapotranspiration components were similar to day 2, but total evapotranspiration decreased from day 2 values in the leaf and no-leaf runs by approximately 30 and 45 percent respectively.

Figure 3.53: BROOK model daily ET components (mm day\(^{-1}\)) for September 23-26, 2003, using the following model run scenarios: 1XP,L: Base precipitation with leaves; 1XP,NL: Base precipitation without leaves; 2XP,L: Doubled precipitation with leaves; 2XP,NL: Doubled precipitation without leaves. Transpiration is indicated by the green vertical lines in the barplot, interception by the lower-left to upper-right diagonal lines, and soil evaporation by the upper-left to lower-right diagonal lines.

Doubling the precipitation in the model runs resulted in only a 0.1 mm day\(^{-1}\) increase in total evapotranspiration over the four days of the run in the leaf case. The modeled fractionation of evaporation was insensitive to the amount of input precipitation. Interception increased by 0.05 mm day\(^{-1}\), indicating that the canopy was already at
capacity from the base precipitation, approximately 2.5 mm. On a percentage basis, interception was 10.7% of total precipitation for the base-precipitation leaf case and 5.4% for the doubled-precipitation leaf case. For the no-leaf runs, total evapotranspiration increased by 0.4 mm day$^{-1}$ over the four days of the run, or 12 percent of the total evapotranspiration as a result of doubling the precipitation. The component with the largest increase was bare-soil evaporation, at 0.35 mm day$^{-1}$.

Model-estimated interception was over 4 times greater for the leaf runs than the no-leaf runs. This resulted in very low model-estimated interception values of 2.4% and 1.3% of total precipitation for the base-precipitation and doubled-precipitation no-leaf runs respectively. These values are lower than what has been observed for hardwood deciduous forests during the dormant season, and range further from growing-season values than observations. For example, Helvey and Patric (1965), in a review of interception studies in eastern US deciduous forests to that time, reported average dormant- and growing-season interception at approximately 6 and 9 percent of total precipitation respectively.

3.5. Summary

We return to the questions posed at the beginning of this part of the chapter:

- Watersheds in the Catskill-Hudson Valley Region of New York State (HVAMS) receive much different inputs of precipitation due to orographic effects, and similar effects are seen in nearby upland regions of the Northeast United States (e.g. Brady and Waldstreicher 2001). As the frequency of extreme precipitation events may be expected to increase in this region as a
result of climate change (e.g. Burns et al. 2007), such orographic effects on precipitation input may be enhanced. What is the watershed response to such rainfall events, and are there any detectable changes in the response due to changes in precipitation forcing?

The observational case studies presented indicate that the peak precipitation region in the Catskills receives as much as two to three times the precipitation as does the rain-shadowed area on the leeward side of the Catskills. Comparisons with ground-based observations show that remote-sensing methods of estimating precipitation fail with either the positioning or magnitude of the precipitation shadow. Radar-estimated precipitation did capture the positioning of the precipitation shadow, but incorrectly estimated its magnitude. CMORPH-estimated rainfall failed in both aspects of characterizing the precipitation shadow. This underscores the need to have adequate, permanent ground-based observations, particularly in the precipitation shadow region.

Characteristics of streamflow response in the network watersheds are related to proximity to the rain shadow. Streamflow peak values were highest in the upland Catskill watersheds located in the peak-precipitation region for both events. Streamflow peaks decreased rapidly over a short distance as the fraction of a watershed in the rain shadow increased, as was observed in the Schoharie Creek drainage basin. Streamflow peak to watershed precipitation ratios were greatest in the upland Catskill watersheds, with ratios at or above 1.0. The streamflow peak to watershed precipitation ratio decreased moving north into the precipitation shadow region, with ratios dropping to around 0.2 in the downstream portions of the Schoharie Creek drainage basin. This
observed pattern in streamflow peak to watershed precipitation ratio was consistent for both case study events, even though the October 27 event featured greater precipitation (about 40 mm greater in the peak-precipitation region) than the September 23 event.

Times to streamflow peak from the start of precipitation increased as one moved to the larger downstream watersheds. Overall, times to streamflow peak were longer in the October 27 event; however this was a consequence that the precipitation lasted over a longer duration in that event.

Observed streamflow recession times were slightly longer in the watersheds located in the transition-precipitation region than both the peak-precipitation and rain-shadow regions.

The BROOK hydrological model was used to conduct a sensitivity study of the influence of added input precipitation and changes in foliage cover on streamflow and soil moisture responses. The sensitivity of the BROOK model to increased precipitation input showed an increase in streamflow peaks in the doubled-precipitation cases of about 2.3 times from the base-precipitation cases. Streamflow-peak to precipitation ratios were around 2.0, about twice the ratios observed in the upland Catskill watersheds. This is a result of the large, very rapid source area flow contribution to the modeled streamflow hydrograph. Ratios also increased by about 0.35 for the doubled-precipitation cases from their corresponding base-precipitation cases. The time to streamflow peak from the start time of the precipitation was much faster than observations, and did not change with respect to changes in precipitation forcing.
Although BROOK model streamflow recession time was insensitive to the amount of input precipitation, model soil moisture recession times increased in length in the doubled precipitation cases for all layers.

- Underlying the individual watershed response in the HVAMS region to differing amounts of precipitation is the seasonal land cover change due to transpiring vegetation in leaf during the growing season transitioning to leaf drop and the beginning of the dormant season, thereby changing the amount of ET forcing the watersheds receive. Can any detectable changes in watershed response be seen during such transition periods?

We observed changes in the watershed response indicators of streamflow recession and diurnal streamflow fluctuations in the network during the autumn transition period. Observed streamflow recessions over the entire network were approximately one-half day faster during the in-leaf growing season period than in the no-leaf dormant season period. The observed transition time in the recessions occurred rather abruptly during the last week of September. Observed near-surface soil moisture recession times for all sites increased in length from the growing season to the dormant season by a comparable amount to that predicted in the BROOK model for the top soil layer.

The seasonal change in the presence of the diurnal streamflow oscillation in the network was evident, with the signal appearing in about half of the network watersheds until the growing to dormant season transition in early October, when only the smallest watersheds exhibited the diurnal streamflow signal. The abrupt disappearance of the
diurnal streamflow signal shown here in fall is similar to the speed at which the diurnal streamflow signal appears in the spring at the onset of the growing season (Czikowsky and Fitzjarrald 2004). This feature can be exploited to assess the phenological condition of vegetation in given watersheds, shown here to be an important control on streamflow recessions.

BROOK model streamflow recession times, streamflow peaks, time to streamflow peak, and streamflow peak to watershed precipitation ratio were insensitive to leaf state, since overland flow was the dominant component of the BROOK model hydrograph. Soil moisture recession times in the BROOK model were sensitive to leaf state, and closer to observations.

Interception evaporation on the day of rainfall for the leaf cases is exaggerated in the BROOK model, and transpiration is suppressed during and after rainfall, in contrast to observations in the Amazon (Chapter 2). In the days following rainfall, transpiration is the dominant component of ET in the leaf cases, with bare-soil evaporation the dominant ET component in the no-leaf cases.

Analysis of the streamflow data using these methods presented is an indicator of seasonal change in hydrological response in these watersheds. This provides an independent estimate of autumn or spring onset from phenological or satellite-derived measurements, and is important as this shows the timing of the hydrological impacts of abrupt changes in evapotranspiration forcing on the watersheds.
3.6 Acknowledgements

This work was supported by National Science Foundation grant ATM-0313718. We thank the USGS for providing the streamflow data. We also acknowledge the work of the MIPS staff from the University of Alabama-Huntsville, as well as the NCAR-ISFF personnel in operating the PAM stations during the HVAMS intensive period.

3.7 References


Kern, A., 2008: Discharge of Schoharie Creek. USGS internal report.


Part II: LBA

Chapter 3L: Hydrologic response to precipitation events in the eastern Amazon region

3.1L Introduction

3.1.1L Eastern Amazon region watersheds

The watersheds in the eastern Amazon region are characterized by contrasting topography and land use/land cover histories, each of which influences the hydrologic response of these watersheds to precipitation events. We examine the streamflow response to rain events in three mesoscale watersheds in the eastern Amazon region near Santarém (2.3333°S, 54.7000°W) as part of the Large-Scale-Biosphere-Atmosphere Experiment in Amazonia (LBA-ECO). One of these watersheds (Mojui) is in a lowland/plateau landform setting where land use change has been extensive. Much of the original forest has been converted to secondary forest, pasture, and croplands. The Moju and Branco watersheds are both in an upland more incised (steeper) topographic setting, with the Rio Branco watershed undergoing moderate forest conversion to secondary forest and croplands and Rio Moju still dominated by intact primary forest. The watersheds are of a similar size, with the Mojui, Branco, and Moju watershed areas being 123, 94, and 102 km² respectively (Kramer et al. 2005). Relatively little has been published up to now comparing land-use land-cover changes in medium-sized watersheds in the Amazon directly from observations, of a size like the ones studied here. Observational studies in the Amazon have generally been restricted to small watersheds (e.g. Moraes et al. 2006; Tomasella et al. 2008; Chaves et al. 2008).
3.1.2L Deforestation impacts on components of the Amazon climate/water balance

The Mojui watershed is a good example of what is occurring in an increasing number of areas in the eastern Amazon region, with land use change and disturbance from forest to agriculture, and provides a focus area for this study. This change has impacts on the local climate, influencing local cloud patterns, surface fluxes (e.g. Sakai et al. 2007) and evapotranspiration and drought susceptibility in the region (e.g. Hutyra et al. 2005). There has long been concern as to how increased, widespread land use/land cover change over Amazonia would affect the hydrological response of the larger tributaries and ultimately the main stem of the Amazon (Sternberg 1987). This concern has grown as the number and spatial extent of small-to-medium sized watersheds that are disturbed continue to increase over Amazonia. In a review of water-balance experiments performed over the Amazon, deforestation has been observed to decrease evapotranspiration over all spatial scales (D’Almeida et al. 2006, and references therein). Localized deforestation in the Amazon has been observed to increase runoff in the disturbed watersheds, as these disturbances are not at a large enough scale to influence precipitation patterns. However, complete deforestation results in decreased precipitation and runoff regionally. Model simulations of the water balance over the Amazon show that although localized forest disturbances result in increased runoff, mesoscale forest disturbances ($10^2$ to $10^5$ km$^2$) may or may not speed up the water cycle, depending on the patchiness of the deforestation and the spatial heterogeneity of the landscape resulting from the deforestation, with patchy, widespread deforestation more likely to have increased precipitation and runoff as opposed to areas of complete deforestation (D’Almeida et al. 2006). The question then arises to what extent such land
cover/land use changes would have to other observational measures of watershed response, such as streamflow and soil moisture recessions (the decrease of soil moisture following a precipitation event), and how this information can be represented in models.

3.1.3L Outline

The following research question is addressed in this section:

- Watersheds in the LBA portion of the study are located in differing landuse/landform settings, from undisturbed upland forest to a lowland setting under agricultural use. There is also an underlying seasonal change in the precipitation inputs the watersheds receive. What is the sensitivity of watershed response to these factors?

To address these issues, we conduct an observational sensitivity study in the LBA watersheds on two fronts. First is the land use/land cover change in the watersheds, ranging from intact primary forest to deforested land now used for agriculture, and associated water-balance changes that may affect the hydrological response of the watersheds. Second is to examine how the different watersheds respond to the sharp precipitation decrease during the transition from wet to dry season.

We begin by examining input precipitation to the Mojui watershed by comparing ground-based precipitation measurements in the watershed with precipitation totals estimated by remote sensing over the 2004 dry season. Using remote sensing methods to estimate precipitation over small-to-medium size Amazon watersheds is important because most of the climate stations in the region are located in towns near the larger, main river banks with road access, and not in the more remote inland regions. This
introduces a bias in the location of the stations, especially since river stations have different diurnal precipitation patterns than stations further inland (Fitzjarrald et al. 2008). Since many of these small-to-medium size watersheds are located in these inland regions away from the larger rivers, use of remote sensing products provide a promising tool to estimate precipitation in more remote, ungauged catchments.

To assess watershed response, we use streamflow data to examine the seasonal trend in the streamflow recessions over the three watersheds over the course of the 2004 dry season. Several years of soil moisture data at weather stations in the study area allowed for the calculation and analysis of seasonal trends in the soil moisture recessions over the disturbed Mojui watershed. This mesoscale soil-moisture array is not common in the Amazon region, and is a longer-duration dataset than other studies have reported on to date.

### 3.2 Location/Data

The study area is located in the eastern Amazon region of Brazil, in the Santarém region of the Large-Scale-Biosphere-Atmosphere Experiment in Amazonia (LBA-ECO; see Fig. 3.1L for a map of the study area and station locations). As part of this project a series of surface weather stations were deployed and included measurements of temperature, humidity, precipitation, wind, pressure, solar radiation, and near-surface soil moisture and temperature. The first weather stations were deployed in 1998 with measurements continuing through 2005. Precipitation at the weather stations was measured with a Texas Instruments model TE525 tipping bucket rain gauge at 0.5 m height. Further details about the weather station precipitation measurements can be
found in Fitzjarrald et al. (2008). Soil moisture was measured at the weather stations using a Campbell Scientific model CS615 water content reflectometer at 0.2 m depth. Soil moisture and precipitation data from three of these stations are used in this study. The Belterra and Mojui stations are used since their locations bound the Mojui watershed, and the km117 station is used as it is sited in a similar landform setting as the Mojui watershed. Furthermore, soil moisture data from two flux tower sites are utilized. These are the km77 site, an agricultural site (Sakai et al. 2004), and km83, a selectively-logged forest site (da Rocha et al. 2004). At km77, soil moisture was measured with a Campbell Scientific CS615 water content reflectometer at 0.29 m depth, and precipitation was measured with a Texas Instruments model TE525 tipping bucket rain gauge at 0.5 m height. Streamflow was measured at three sites, one in each of the three study watersheds (Kramer et al. 2005). Continuous streamflow was determined by recording water level every 15 minutes using pressure transducers for a 90-day period and recording streamflow and water level using current velocity meters. Stream discharge-water level curves were then constructed (Kramer et al. 2005). Precipitation was estimated for the 40-km grid cell covering the Mojui watershed by remote sensing using the NOAA Climate Prediction Center Morphing (CMORPH) data product, a satellite-based passive microwave estimate of precipitation (Joyce et al. 2004). The watershed boundary data were provided by Kramer et al. (2005), derived from 90-m Shuttle Radar Topography Mission (SRTM) digital elevation (DEM) data (Farr et al. 2007). Land cover and land use maps were derived from the Landsat-Thematic Mapper (TM) product and used for the watershed landform classification.
Figure 3.1L: Map of the watersheds, weather stations and flux-measurement sites located in the Santarém region (STM) of LBA-ECO. Watershed boundaries are delineated by the red lines. Watershed names are in red. The black lines are the major rivers. The weather stations and flux-measurement sites are labeled in black. The red dots indicate stream gauge locations. The dashed grid boxes indicate the boundaries of the CMOPRH 40-km data grid cells (data are for the center of each grid box). Elevation (m) is shaded.

3.3L Methods

The hydrologic response of a watershed to precipitation events is broken down into three stages with respect to the streamflow hydrograph (Fig. 3.2L). The first stage is the peak in streamflow following the input precipitation. The second stage is the streamflow recession, the decrease in streamflow following the precipitation peak. The
third stage is the evapotranspiration-driven diurnal streamflow fluctuations that may occur after the streamflow has receded to its baseflow. The same processes apply to the near-surface soil moisture. In this study we examine the input watershed precipitation and the recession stage of the watershed response for streamflow and soil moisture.

![Diagram showing the stages of streamflow response to rainfall events. The same process applies to soil moisture. In this study we focus on the input precipitation and recession components of the response.](image)

To determine watershed precipitation, we combined data from two sources: measured precipitation from the network of weather stations covering the study area, and precipitation estimated by remote sensing using the CMORPH data. Watershed response to rain events was assessed in two parts. First, stream response was determined by analyzing the timing and magnitude of streamflow peaks and streamflow recession times at stream gauges in each of the three watersheds in relation to rainfall events. Second, the shallow soil moisture response to rainfall was assessed by analyzing the characteristic soil moisture values and soil moisture recession times following rainfall as recorded at the weather stations which were located nearby, or in topographic settings with comparable soils and land cover as do the watersheds. Input precipitation for the Mojui watershed was estimated as the average of the precipitation at the Belterra and Mojui
weather stations, which are at the western and eastern boundaries of the watershed respectively. The CMORPH-estimated precipitation for the Mojui watershed was taken to be that from the 40-km grid square containing the Mojui weather station in the Mojui watershed shown in Fig. 3.1L. Input precipitation was not estimated from observations for the Branco and Moju watersheds, as no stations measuring precipitation were located within the boundaries of those watersheds.

The streamflow recession time $\tau$ is commonly defined as the e-folding time for the streamflow following a rain event (Fig. 3.2L). In this study, to maximize the number of recessions available in this limited dataset without fitting the recession curves, we report recessions for streamflow using a shorter threshold point of $\frac{1}{0.4}\tau, \frac{1}{0.4e}$ of the streamflow peak value from a precipitation event (Fig. 3.2L). Streamflow recession values were calculated following precipitation events over the three watersheds from August to December 2004, which covers the course of the dry season in this region.

The shallow soil moisture response for the Mojui watershed was assessed by analyzing the characteristic soil moisture values and soil moisture recession times for the weather stations located in the watershed (Belterra and Mojui) as well as any stations located outside the watershed in a topographic setting with comparable soils and land cover. The km117 weather station, located in a plateau geomorphic setting similar to the Mojui watershed, met this criterion. The land-cover shift from forest to agricultural land at these sites is typical of recent changes in the Mojui region (Fig. 3.3L)
The soil moisture recession times were calculated in the same manner as the streamflow recession times described above. Since several years of soil moisture data were available, seasonal trends in the soil moisture recessions were examined.
3.4L Results

3.4.1L Rainfall

The observed Mojui watershed rainfall at the Belterra and Mojui weather stations, the watershed-averaged rainfall, and the CMORPH-estimated Mojui watershed rainfall are shown for the August to December 2004 time period (Fig. 3.4L). Note that the CMORPH data does not capture all of the lighter rainfall events that were not experienced over the entire watershed (see between days 220 and 240). However, CMORPH does capture the larger rain events (see near days 260 and 280) but overestimated actual precipitation in the heaviest event. This overestimation led to the cumulative dry-season precipitation estimated by CMORPH to be over 20 percent higher than the observed Mojui watershed-averaged precipitation (taken as the average of the Belterra and Mojui station rainfall) for the same period. The overestimation in the CMORPH-estimated cumulative period precipitation was lower, about 13 percent, compared to only the Mojui station rainfall, the station within the CMORPH grid cell. The highly localized, convective nature of rainfall in this region during the dry season results in irregular amounts of rainfall even within a 40-km grid cell, complicating the process of estimating precipitation over a watershed.
Figure 3.4L: Top panel: CMORPH-estimated precipitation for the grid cell containing the Mojui weather station in the Mojui watershed (black solid line) during the 2004 dry season, from August to December. The Mojui weather station measured rainfall is the red dashed line. Middle panel: same as top except the blue dashed line is the Belterra weather station precipitation. Bottom panel: CMORPH cumulative rainfall (black line) for the same 2004 period. Mojui rainfall is in red, Belterra rainfall in blue, and the Mojui watershed average measured cumulative rainfall (the average of the Mojui and Belterra weather station rainfall) is in pink.
Note that the CMORPH precipitation estimates are much closer to observations for these convective-precipitation events than the stratiform-precipitation cases examined in the HVAMS portion of this chapter. Although the 8-km resolution CMORPH data were not analyzed here, comparisons with observations should not be appreciably different from the 40-km data as results from HVAMS comparisons earlier in this chapter show the 40-km data to be a smoothed product of the 8-km data.

3.4.2L Streamflow

The observed streamflow for the Mojui, Branco, and Mojui stream gauge sites are shown for the period of record (Fig. 3.5L). The upland Branco and Moju rivers show a much greater peak response to rainfall than does the lowland/plateau Mojui River.
Figure 3.5L: Streamflow (reported as stream depth in cm) for the Mojui River (black), Rio Branco (blue), and Moju River (red) for August to December 2004. Time is in GMT.

For the precipitation events shown in Fig. 3.4L, streamflow recessions were calculated for the three stream gauge sites (Fig. 3.6L). Recession times are shorter in the upland watersheds of Rio Branco and Moju than in the lowland/plateau Mojui watershed. Recession times also appear to get shorter in length during the course of the dry season, shortening by approximately three days in both the Rio Branco and Moju River during the two month span from September to November. There is a sharp decrease in the latent heat flux from August to October, the early dry season period (Fig. 3.7L). The shortening of the streamflow recession time constant during the dry season with decreased latent heat flux is contrary to the expected result in the midlatitudes, where streamflow recession times increase in length with decreased latent heat flux.
3.4.3L Seasonal trends

Monthly precipitation in the region is highest during the January-to-June wet season, illustrated by the km77 site where precipitation exceeds 100 mm for each of the wet season months, with a maximum of over 300 mm in February (Fig. 3.7L). There is a sharp decrease in monthly precipitation through June, when precipitation remains below 100 mm per month until the end of the dry season in December. Monthly latent heat flux at km77 increases during the wet season, maximizing during the late wet season and early dry season (April to August). During the dry season, there is a sharp drop in monthly latent heat flux in September and October to nearly half the value observed in the preceding months, and remaining below 60 W m\(^{-2}\) through the end of the dry season.
in December. Note that the conversion of the latent heat flux to evapotranspiration (ET, in water-balance units of mm month\(^{-1}\), directly comparable to precipitation) is approximately 1.04. This conversion was used to allow the comparison of ET to precipitation discussed below.

The average monthly soil moisture (in units of m\(^3\) m\(^{-3}\)) for the Mojui watershed weather stations (Belterra and Mojui) as well as the km117 weather station, km77 and km83 are shown in Fig. 3.7L. Similar patterns in the seasonal trends in the soil moisture appear at these stations. At all sites the soil moisture increases during the early wet season months, peaking between April and June. This lags a few months behind the actual month of greatest precipitation, primarily because the difference between precipitation and ET is sufficiently large enough to sustain increasing soil moisture even though precipitation is decreasing. At all sites the soil moisture has its most rapid decline over the year between the months of June and August. These are the transition months from the wet season to the dry season and the months at km77 where the monthly-averaged ET exceeds the monthly-averaged precipitation. Soil moisture values are at their lowest levels of the year at all sites in the late dry season from October onward, which is the time of year where the lowest ET values are observed at km77. The deeper soil moisture (40cm) shows a similar seasonal cycle as the shallow soil moisture (10cm) at km83, but with a lower amplitude and a time lag.

The soil moisture recession times for the Belterra, Mojui, and km117 weather stations were gathered together and binned by month (Fig. 3.7L).
Figure 3.7L: Left panel: Average monthly soil moisture recession times in days for the Mojui watershed weather stations (Mojui and Belterra) and the km117 weather station, for the years 2000-2005. Although the km117 weather station is outside the watershed, it is in a similar plateau geomorphic setting. Medians are indicated by X, and the bars indicate the upper and lower quartiles. The number of recessions analyzed for each month are listed above the plot. The pink line is the average soil moisture (m³ m⁻³) at the weather stations at 20cm depth. The red line is the soil moisture at the km77 site at 29cm depth, and the green and blue lines are the km83 soil moisture at 10 and 40cm depth respectively. Right panel: Monthly mean latent heat flux (LE in W m⁻², black) and precipitation (mm month⁻¹, red) at the km77 site. Conversion of LE to evapotranspiration in the same units as the precipitation (mm month⁻¹) is 1.04.

Note the soil moisture recession times rapidly increase during the transition from the wet season to the dry season, peaking in August, in the early part of the dry season. This is contrary to the expectation that the recession lengths should become shorter since ET is at nearly the highest levels observed over the year, exceeding precipitation during the June-to-August wet-to-dry season transition period. For instance, recession times
have been observed to shorten in midlatitude deciduous forests as a result of increased ET following leaf onset and the start of the growing season (Federer et al. 1973; Czikowsky and Fitzjarrald 2004).

It appears that different processes are may be responsible for the peak near-surface soil-moisture recession times observed in the early dry season. Tomasella et al. (2008) observed in the Asu catchment, a small central Amazonian watershed known locally as an *igarapé*, that groundwater storage peaks annually during the early portion of the dry season, declining thereafter. Furthermore, Tomasella et al. (2008) observed that during the early dry season, precipitation events wetted the upper soil (0 to 2 m), but not the lower soil layers on a plateau portion of the catchment away from the higher-slope areas. There was little further storage change in the lower soil layers until the start of the following wet season. They concluded that there was no significant root uptake in the deeper soil layers, since there was sufficient water above. Also, their data indicate there was no drainage out of the soil layers during this period. In another study in a similar *igarapé* catchment in the far eastern Amazon, Moraes et al. (2006; Fig. 7f) observed that light rainfall events at the end of the wet season and transitioning into the dry season were only sufficient to wet the uppermost ~ 30 cm of the soil, with no drainage to and change in soil moisture content in the deeper layers down to 1 m.

In our study, we observe a similar phenomenon, with light rainfall events at the beginning of the dry season, approximately days 220 to 250 (Fig. 3.4 L). This is precisely the time when the near-soil soil moisture recession times are the longest, likely because only the uppermost soil layers are being wetted, and with little drainage below.
This maintained level of soil moisture storage and lack of drainage out of the soil profile during the early dry season may be responsible for the longer soil moisture recession times observed at that time, since soil water may possibly be held in the upper soil profile for a longer period of time. Although the dry season begins one month earlier in the Asu catchment (June) than in the Santarém region watersheds in this study and the Moraes et al. (2006) study (July); all three sites are in otherwise similar climate regimes.

With land cover conversion from forest to pasture, the soil’s hydraulic conductivity and infiltration rates can be greatly reduced, thus slowing vertical drainage through the soil profile and shifting streamflow-generation flowpaths towards overland flow. Chaves et al. (2008) observed over two adjacent small southwestern Amazon watersheds (one in pasture, the other undisturbed forest) that the pasture watershed had much greater overland flow contributions to streamflow, consistent with the greatly reduced hydraulic conductivity and reduced infiltration rates observed in the pasture soil.

It is only after the early dry-season period that the soil moisture recession times then become shorter, by approximately three days, through the end of the dry season in December. The watershed appears to be showing a delayed response to the water balance deficit where ET exceeds precipitation. Tomasella et al. (2008) observed that there is a strong memory effect in the deeper soil and groundwater system that works on a long enough time scale to maintain seasonal anomalies in storage and recharge, and affect hydrological response into the following year. The recessions we report are an observational indicator for this memory effect.

The trend of decreasing soil moisture recession lengths during the mid-to-late dry season and its magnitude is consistent with what is observed in the streamflow recession
trends. Furthermore, this mid-to-late dry season recession trend of shortening recession times holds for both the disturbed and intact forest land cover types, as the soil moisture recessions were recorded in the disturbed Mojui watershed whereas the streamflow recession data came from the intact forest Branco and Moju watersheds.

3.5L Summary

We now return to the question put forth at the beginning of this section:

- Watersheds in the LBA portion of the study are located in differing landuse/landform settings, from undisturbed upland forest to a lowland setting under agricultural use. There is also an underlying seasonal change from the wet season to dry season in the region that changes the precipitation inputs the watersheds receive. What changes in watershed response can be detected resulting from these factors?

The response of three eastern Amazon watersheds to precipitation events was examined on an event and seasonal basis. Comparison of observed watershed precipitation to watershed precipitation estimated by CMORPH showed that although CMORPH did not capture the lightest precipitation events, there was an overestimation of rainfall by CMORPH for the heaviest events. This resulted in an overestimation of watershed rainfall by CMORPH for the period of record. Nevertheless, given the nature of the convective precipitation in the region, CMORPH does appear to be a promising tool to provide general estimates of precipitation over remote, inland watersheds where ground-based measurements of precipitation are unavailable.
Streamflow peaks resulting from the input precipitation events were more pronounced for the upland Branco and Moju watersheds as opposed to the lowland/plateau Mojui watershed. We observed a decrease in the length of the mid-to-late dry season soil moisture and streamflow recession values that were similar in magnitude and timing. This was observed even though the soil moisture and streamflow measurements were taken in watersheds with different land use/land cover types. From the seasonal trends in the streamflow and soil moisture recession data presented here, there are indications that the recession watershed response characteristic appears not to change with respect to changes in land use/land cover type.

Two unexpected results emerged from this study when comparing seasonal changes in streamflow and soil moisture recessions to what occurs in the mid-latitudes. First, streamflow recession times decreased in length even though latent heat flux had also decreased, contrary to what is observed in midlatitudes (increased streamflow recession times with decreased latent heat flux). Second, there appears to be different processes controlling soil moisture recessions when comparing the LBA data with results from the HVAMS watersheds presented earlier in the chapter. In the LBA watersheds studied, precipitation and subsequent storage and subsurface drainage processes seem to have a greater influence on seasonal changes in soil moisture recession times than in the HVAMS watersheds, where such changes are dominated by seasonal variations in evapotranspiration due to vegetation state.

We note that although our streamflow dataset available in this study was limited, the soil moisture dataset was of a sufficiently long duration to examine seasonal changes in hydrologic response in a disturbed watershed. Analysis of the recession time constants
for streamflow and soil moisture is a first step toward more efficient use of LBA-ECO data to provide information directly useful to those modeling the hydrology of the Amazon. In particular, these time constants are the observational equivalent of the soil moisture memory that is appearing solely as a model concept elsewhere (e.g., Mahanama, et al., 2003; Wu and Dickinson, 2004). If studies similar to ours are done throughout the Amazon Basin, the results will help to constrain models (e.g., Zeng et al., 2008).

3.6L References


Chapter 4: Conclusion

4.1 Summary and questions

In this thesis, we have examined the impact of transient impulse events on the features of the hydrological balance components, covering a wide range of spatial and temporal scales. The transient terms studied include the interception component of evapotranspiration (Chapter 2), and the runoff and soil moisture storage terms in the surface water balance (Chapter 3). Two common themes are present in this work. First, the transient responses to precipitation events of the processes studied in this work are similar in form, with a sharp increase in the term (interception evaporation, runoff, or soil moisture) resulting from a rainfall event, followed by a more gradual decline (Fig. 1.2). The timing of the response is modulated by changes in input precipitation (influenced by topography and season) and by changes in evapotranspiration forcing (influenced by land cover and leaf state). Second, one can take advantage of perturbations occurring in nature by combining data in space and time collected during both perturbed and unperturbed states to conduct an observational sensitivity study. For example, in this work eddy covariance data taken over a rain forest were separated into dry and rainy periods, and then combined into event-based ensembles to assess the shifts in the hydrological and energy balances that occur during and after rainfall (Chapter 2). Also, spatial and seasonal changes in hydrological response to precipitation events were examined splitting the datasets into periods based on leaf and land-cover states (Chapter 3).
Reviewing the questions posed the introduction, and addressed in chapters 2 and 3, starting with chapter 2:

- Through the expansion of eddy flux-measurement networks such as Fluxnet (Baldocchi et al. 2001), the number and coverage of long-term eddy flux measurement sites has grown to over 460 worldwide sites, distributed over a wide range of land cover types (http://www.fluxnet.orl.gov). What additional information can be obtained from these sites beyond the standard reports of half-hour average fluxes?

We have introduced and demonstrated the utility of a new method to estimate interception evaporation using eddy covariance measurements combined with novel data analysis techniques at an old-growth tropical rain forest site. Mean interception for moderate daytime rainfall-rate events was about 10%, with light events at 18% and heavy events at 7.8%. The mean interception for all daytime and nighttime events combined was 11.6%. Some conventional estimates of interception in tropical rain-forest environments exceed our results by two times. Our observations suggest transpiration continues during and after rainfall (stomatal resistance \( r_s > 0 \)), contrary to model predictions over wet canopies. Energy balance comparisons between dry and afternoon rain-days show an approximately 15% increase of evaporative fraction on the rain days, with the energy being supplied by a nearly equivalent decrease in the canopy heat storage.
One advantage of using this method to estimate interception evaporation is that the footprint area of the eddy-covariance measurement incorporates the spatial variability in throughfall and interception that is a challenge to adequately sample using conventional methods, providing an average representative interception value over the entire flux footprint, and therefore be more suitable for grid-cell model input.

Another advantage of using the method presented here is that once baseline data are established, it can be used for long-term interception monitoring without intensive effort beyond the maintenance of the eddy-covariance system at a site.

- Under what conditions would micrometeorological-based interception evaporation measurements be expected to work, and under what conditions would such a technique be of limited use?

For determination of the dry-day daytime base state ensemble values, the method is applicable in nearly all daytime turbulent conditions. Recall that at night, base state latent heat flux is zero, so formation of base-state nighttime ensembles are not necessary and the well-documented eddy covariance failures during calm, low-turbulent conditions are not applicable here. For rainfall events, direct measurements using eddy covariance can be made for events with light-to-moderate rainfall-rate conditions. In this study, events with average rainfall rates of 16 mm hr\(^{-1}\) or less could be used.
From the HVAMS section of chapter 3:

- Watersheds in the Catskill-Hudson Valley Region of New York State (HVAMS) receive much different inputs of precipitation due to orographic effects, and similar effects are seen in nearby upland regions of the Northeast United States (e.g. Brady and Waldstreicher 2001). As the frequency of extreme precipitation events may be expected to increase in this region as a result of climate change (e.g. Burns et al. 2007), such orographic effects on precipitation input may be enhanced. What is the watershed response to such rainfall events, and are there any detectable changes in the response due to changes in precipitation forcing?

The large variation in precipitation received in the Catskill-Hudson Valley region resulting from the presence of the lee-of-the-Catskills precipitation shadow results in rapid changes in streamflow response depending on the location of a watershed with respect to the peak-precipitation region and precipitation shadow. The highest streamflow peaks and streamflow peak to watershed precipitation ratios were found in the Catskill peak-precipitation region, rapidly decreasing as one moves into the rain-shadowed watersheds.

Comparisons of ground-based observations of event regional precipitation to that estimated from remote sensing methods show problems with the remote sensing methods in correctly determining the position and magnitude of the precipitation shadow. This underscores the need to have adequate, permanent ground-based observations, particularly in the precipitation shadow region.
Observed streamflow recession times were slightly longer in the watersheds located in the transition-precipitation region than both the peak-precipitation and rain-shadow regions. Although no detectable change was output from the BROOK model streamflow recessions as a result of doubling the precipitation, model soil moisture recession times increased in length in the doubled precipitation cases for all layers.

Model evapotranspiration increased about 12 percent from doubling the precipitation in the no-leaf case, with bare-soil evaporation accounting for most of that increase. Little change in evapotranspiration was output from the model after doubling the precipitation in the leaf case.

- Underlying the individual watershed response in the HVAMS region to differing amounts of precipitation is the seasonal land cover change due to transpiring vegetation in leaf during the growing season transitioning to leaf drop and the beginning of the dormant season, thereby changing the amount of ET forcing the watersheds receive. Can any detectable changes in watershed response be seen during such transition periods?

Detectable changes in the watershed response indicators of streamflow recession and diurnal streamflow oscillations over the network during the autumn transition season. Observed streamflow recessions over the entire network were approximately one-half day faster during the in-leaf growing season period than in the no-leaf dormant season period. The observed transition time in the recessions occurred rather abruptly during the last week of September. Observed near-surface
soil moisture recession times for all sites increased in length from the growing season to the dormant season by a comparable amount to that predicted in the BROOK model for the top soil layer.

The seasonal change in the presence of the diurnal streamflow oscillation in the network was evident, with the signal appearing in about half of the network watersheds until the growing to dormant season transition in early October, when only the smallest watersheds exhibited the diurnal streamflow signal. This feature can be exploited to assess the phenological condition of vegetation in given watersheds, shown here to be an important control on streamflow recessions.

From the LBA section of chapter 3:

- Watersheds in the LBA portion of the study are located in differing landuse/landform settings, from undisturbed upland forest to a lowland setting under agricultural use. There is also an underlying seasonal change from the wet season to dry season in the region that changes the precipitation inputs the watersheds receive. What changes in watershed response can be detected resulting from these factors?

The response of three eastern Amazon watersheds to precipitation events was examined on an event and seasonal basis. Comparison of observed watershed precipitation to watershed precipitation estimated by remote sensing (CMORPH) showed that although there was an overestimation of precipitation by CMORPH over the period of record, CMORPH did perform much better in the Amazon, a region
with a significant amount of convective precipitation as opposed to New York State, which has more stratiform precipitation. In the Amazon region, CMORPH does appear to be a promising tool to provide general estimates of precipitation over remote, inland watersheds where ground-based measurements of precipitation are unavailable.

Streamflow peaks resulting from the input precipitation events were more pronounced for the upland Branco and Moju watersheds as opposed to the lowland/plateau Mojui watershed. We observed a decrease in the length of the mid-to-late dry season soil moisture and streamflow recession values that were similar in magnitude and timing. This was observed even though the soil moisture and streamflow measurements were taken in watersheds with different land use/land cover types. From the seasonal trends in the streamflow and soil moisture recession data presented here, there are indications that the recession watershed response characteristic appears not to change with respect to changes in land use/land cover type.

Different processes control soil moisture recessions in the LBA and HVAMS watersheds. In the LBA watersheds studied, precipitation and subsequent storage and subsurface drainage processes seem to have a greater influence on seasonal changes in soil moisture recession times than in the HVAMS watersheds, where such changes are dominated by seasonal variations in evapotranspiration due to vegetation state.
4.2 Future work

Application of the method described in chapter 2 of this thesis to estimate interception evaporation from eddy-covariance measurements opens the opportunity to test and implement the method at other long-term flux-measurement sites in a wide range of land-cover types and climate regimes, such as those in Fluxnet, as the instrumentation and measurements required are already in place at these sites. Furthermore, many of these sites have been in operation long enough so that the observational base-state ensembles employed in the method can be formed. Longer time periods than what was available in this study may be required to form the ensembles in regions with more variable day-to-day weather patterns than this study site in the tropics. The raw high-rate data collected at a site should be kept/archived as reprocessing at 15-minute intervals is necessary to obtain greater event detail. The use of an on-site ceilometer increased the number of events to analyze, and some light events would be missed without its use; however the method would still work and the base-state ensembles could be adequately formed without its use. Combining this new approach with conventional interception-estimation methods would be useful for a comparison study at the same site.

The seasonal and spatial analysis of diurnal streamflow oscillations, and streamflow and soil moisture recessions presented in chapter 3 provide an indicator of seasonal change in hydrological response over the watersheds analyzed. This provides an independent estimate of autumn or spring onset from phenological or satellite-derived measurements, and is important as this shows the timing of the hydrological impacts of abrupt changes in evapotranspiration forcing on the watersheds. This analysis could be
extended to other regions worldwide to assess the timing of seasonal changes in hydrological response.

Furthermore, these time constants presented are the observational equivalent of the soil moisture memory that appears solely as a model concept elsewhere. The results from further studies of this type may help to constrain models.
Appendix 1: Station location tables used in chapter 3

The following section of the appendix includes tables that indicate the locations of the station used in the HVAMS portion of the study in chapter 3.

Table A1.1: NCAR-ISFF PAM station numbers, names, latitude, longitude, and elevation.

<table>
<thead>
<tr>
<th>Stn#</th>
<th>Name</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Alexander Farm</td>
<td>42.5806</td>
<td>-73.6162</td>
<td>156.0</td>
</tr>
<tr>
<td>2</td>
<td>Black Horse Farm</td>
<td>42.2959</td>
<td>-73.8508</td>
<td>47.0</td>
</tr>
<tr>
<td>3</td>
<td>Southlands Farm</td>
<td>41.8863</td>
<td>-73.9127</td>
<td>45.0</td>
</tr>
<tr>
<td>4</td>
<td>Green Acres</td>
<td>42.1481</td>
<td>-73.7522</td>
<td>94.0</td>
</tr>
<tr>
<td>5</td>
<td>Fix Bros. Farm</td>
<td>42.1800</td>
<td>-73.8300</td>
<td>108.0</td>
</tr>
<tr>
<td>6</td>
<td>Van Orden Farm</td>
<td>42.1790</td>
<td>-73.8960</td>
<td>25.0</td>
</tr>
<tr>
<td>7</td>
<td>Zena Cornfield</td>
<td>42.0397</td>
<td>-74.0850</td>
<td>133.2</td>
</tr>
<tr>
<td>8</td>
<td>South Albany Airport</td>
<td>42.5607</td>
<td>-73.8339</td>
<td>53.0</td>
</tr>
<tr>
<td>9</td>
<td>Pertgen</td>
<td>42.4670</td>
<td>-73.7500</td>
<td>76.0</td>
</tr>
</tbody>
</table>

Table A1.2: HOBO station numbers, names, coordinates, and elevations for the HVAMS IFC period deployment.

<table>
<thead>
<tr>
<th>Stn#</th>
<th>Name</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>H1</td>
<td>Lake Taghkanic State Park</td>
<td>42.0900</td>
<td>-73.7200</td>
<td>202.0</td>
</tr>
<tr>
<td>H2</td>
<td>Freehold Airport</td>
<td>42.3643</td>
<td>-74.0660</td>
<td>133.0</td>
</tr>
<tr>
<td>H3</td>
<td>East Jewett</td>
<td>42.2438</td>
<td>-74.1812</td>
<td>559.0</td>
</tr>
<tr>
<td>H4</td>
<td>Thacher State Park</td>
<td>42.6591</td>
<td>-74.0435</td>
<td>414.0</td>
</tr>
<tr>
<td>H5</td>
<td>Chatham</td>
<td>42.4475</td>
<td>-73.5844</td>
<td>128.0</td>
</tr>
</tbody>
</table>

Table A1.3: NYCDEP station names, coordinates, and elevations.

<table>
<thead>
<tr>
<th>Stn#</th>
<th>Name</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CAM001</td>
<td>Boiceville</td>
<td>41.9497</td>
<td>-74.2000</td>
<td>182.9</td>
</tr>
<tr>
<td>CAM002</td>
<td>Chichester</td>
<td>42.1181</td>
<td>-74.2758</td>
<td>612.6</td>
</tr>
<tr>
<td>CSM038</td>
<td>Gilboa Dam</td>
<td>42.3976</td>
<td>-74.4502</td>
<td>292.6</td>
</tr>
<tr>
<td>CSM039</td>
<td>Prattsville Airport</td>
<td>42.2807</td>
<td>-74.3911</td>
<td>545.5</td>
</tr>
<tr>
<td>DCM074</td>
<td>Stilesville</td>
<td>42.0672</td>
<td>-75.3782</td>
<td>356.6</td>
</tr>
<tr>
<td>DCM076</td>
<td>Walton</td>
<td>42.2494</td>
<td>-75.0936</td>
<td>634.0</td>
</tr>
<tr>
<td>DCM077</td>
<td>Stamford</td>
<td>42.4099</td>
<td>-74.6420</td>
<td>682.8</td>
</tr>
<tr>
<td>DCM080</td>
<td>East Delhi</td>
<td>42.2986</td>
<td>-74.9072</td>
<td>490.7</td>
</tr>
<tr>
<td>DNM146</td>
<td>Neversink</td>
<td>41.8274</td>
<td>-74.6395</td>
<td>445.0</td>
</tr>
<tr>
<td>DNM147</td>
<td>Claryville</td>
<td>41.8974</td>
<td>-74.6117</td>
<td>670.6</td>
</tr>
<tr>
<td>DNM148</td>
<td>Slide Mountain</td>
<td>42.0132</td>
<td>-74.3965</td>
<td>1106.4</td>
</tr>
<tr>
<td>DPM111</td>
<td>Margaretville</td>
<td>42.0783</td>
<td>-74.6169</td>
<td>673.6</td>
</tr>
</tbody>
</table>
Table A1.4: Location, elevation and watershed area of the USGS stream gauge sites used in this study.

<table>
<thead>
<tr>
<th>USGS ID</th>
<th>Station Name</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation (m)</th>
<th>Area (km²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>01334500</td>
<td>Hoosic R near Eagle Bridge</td>
<td>42.9387</td>
<td>-73.3771</td>
<td>108.3</td>
<td>1315.80</td>
</tr>
<tr>
<td>01349150</td>
<td>Canajoharie Ck near Canajoharie</td>
<td>42.8762</td>
<td>-74.6029</td>
<td>195.1</td>
<td>154.03</td>
</tr>
<tr>
<td>01349541</td>
<td>Sugarloaf Bk S of Tannersville</td>
<td>42.1451</td>
<td>-74.1229</td>
<td>591.3</td>
<td>2.89</td>
</tr>
<tr>
<td>01349700</td>
<td>East Kill near Jewett Center</td>
<td>42.2493</td>
<td>-74.3026</td>
<td>442.6</td>
<td>91.85</td>
</tr>
<tr>
<td>01349705</td>
<td>Schoharie Ck near Lexington</td>
<td>42.2370</td>
<td>-74.3401</td>
<td>408.4</td>
<td>249.74</td>
</tr>
<tr>
<td>01349711</td>
<td>West Kill below Hunter Bk near Spruceton</td>
<td>42.1851</td>
<td>-74.2768</td>
<td>630.9</td>
<td>12.82</td>
</tr>
<tr>
<td>01349810</td>
<td>West Kill near West Kill</td>
<td>42.2304</td>
<td>-74.3929</td>
<td>438.9</td>
<td>69.66</td>
</tr>
<tr>
<td>01349840</td>
<td>Batavia Kill near Maplecrest</td>
<td>42.2895</td>
<td>-74.1160</td>
<td>658.4</td>
<td>5.24</td>
</tr>
<tr>
<td>01349950</td>
<td>Batavia Kill at Red Falls near Prattsville</td>
<td>42.3084</td>
<td>-74.3899</td>
<td>384.0</td>
<td>176.99</td>
</tr>
<tr>
<td>01350000</td>
<td>Schoharie Ck at Prattsville</td>
<td>42.3195</td>
<td>-74.4365</td>
<td>344.9</td>
<td>611.46</td>
</tr>
<tr>
<td>01350032</td>
<td>Toad Hollow Bk near Grand Gorge</td>
<td>42.3329</td>
<td>-74.4935</td>
<td>536.4</td>
<td>1.96</td>
</tr>
<tr>
<td>01350035</td>
<td>Bear Kill near Prattsville</td>
<td>42.3381</td>
<td>-74.4515</td>
<td>347.5</td>
<td>66.31</td>
</tr>
<tr>
<td>01350080</td>
<td>Manor Kill at West Conesville near Gilboa</td>
<td>42.3770</td>
<td>-74.4129</td>
<td>382.8</td>
<td>83.59</td>
</tr>
<tr>
<td>01350101</td>
<td>Schoharie Ck at Gilboa</td>
<td>42.3973</td>
<td>-74.4504</td>
<td>286.4</td>
<td>815.28</td>
</tr>
<tr>
<td>01350120</td>
<td>Platter Kill at Gilboa</td>
<td>42.4062</td>
<td>-74.4471</td>
<td>329.2</td>
<td>28.12</td>
</tr>
<tr>
<td>01350140</td>
<td>Mine Kill near North Blenheim</td>
<td>42.4290</td>
<td>-74.4729</td>
<td>323.1</td>
<td>41.80</td>
</tr>
<tr>
<td>01350180</td>
<td>Schoharie Ck at North Blenheim</td>
<td>42.4659</td>
<td>-74.4621</td>
<td>243.8</td>
<td>923.64</td>
</tr>
<tr>
<td>01350355</td>
<td>Schoharie Ck at Breakabeen</td>
<td>42.5370</td>
<td>-74.4104</td>
<td>209.3</td>
<td>1145.52</td>
</tr>
<tr>
<td>01351500</td>
<td>Schoharie Ck at Burtonsville</td>
<td>42.8001</td>
<td>-74.2629</td>
<td>154.8</td>
<td>2285.88</td>
</tr>
<tr>
<td>01362192</td>
<td>Panther Mtn Trib to Espous Ck near Oliverea</td>
<td>42.0337</td>
<td>-74.4204</td>
<td>603.5</td>
<td>3.97</td>
</tr>
<tr>
<td>013621955</td>
<td>Birch Ck at Big Indian</td>
<td>42.1090</td>
<td>-74.4518</td>
<td>378.0</td>
<td>32.25</td>
</tr>
<tr>
<td>01362200</td>
<td>Esopus Ck at Allaben</td>
<td>42.1170</td>
<td>-74.3801</td>
<td>304.2</td>
<td>164.35</td>
</tr>
<tr>
<td>01362342</td>
<td>Hollow Tree Bk at Lanesville</td>
<td>42.1423</td>
<td>-74.2649</td>
<td>451.1</td>
<td>5.03</td>
</tr>
<tr>
<td>01362380</td>
<td>Stony Clove Ck near</td>
<td>42.0981</td>
<td>-74.3170</td>
<td>274.3</td>
<td>81.27</td>
</tr>
<tr>
<td>Station Number</td>
<td>Location Description</td>
<td>Longitude</td>
<td>Latitude</td>
<td>Altitude</td>
<td>Temperature</td>
</tr>
<tr>
<td>---------------</td>
<td>----------------------------------------------</td>
<td>------------</td>
<td>-----------</td>
<td>----------</td>
<td>-------------</td>
</tr>
<tr>
<td>01362465</td>
<td>Beaver Kill Trib above Lake Hill</td>
<td>-74.1831</td>
<td>42.0831</td>
<td>396.2</td>
<td>2.53</td>
</tr>
<tr>
<td>01362497</td>
<td>Little Beaver Kill at Beechford nr Mt Tremper</td>
<td>-74.2663</td>
<td>42.0195</td>
<td>201.2</td>
<td>42.57</td>
</tr>
<tr>
<td>01362500</td>
<td>Esopus Ck at Coldbrook</td>
<td>-74.2707</td>
<td>42.0142</td>
<td>189.4</td>
<td>495.36</td>
</tr>
<tr>
<td>01363382</td>
<td>Bush Kill below Maltby Hollow Bk at West Shokan</td>
<td>-74.2929</td>
<td>41.9656</td>
<td>213.4</td>
<td>41.80</td>
</tr>
<tr>
<td>01364959</td>
<td>Rondout Ck above Red Bk at Peekamoose</td>
<td>-73.3746</td>
<td>41.9370</td>
<td>530.4</td>
<td>13.83</td>
</tr>
<tr>
<td>01365000</td>
<td>Rondout Ck near Lowes Corners</td>
<td>-74.4871</td>
<td>41.8665</td>
<td>266.6</td>
<td>98.81</td>
</tr>
<tr>
<td>01365500</td>
<td>Chestnut Ck at Grahamsville</td>
<td>-74.5404</td>
<td>41.8451</td>
<td>267.9</td>
<td>53.92</td>
</tr>
<tr>
<td>01367500</td>
<td>Rondout Ck at Rosendale</td>
<td>-74.0860</td>
<td>41.8431</td>
<td>10.0</td>
<td>988.14</td>
</tr>
<tr>
<td>01371500</td>
<td>Wallkill R at Gardiner</td>
<td>-74.1651</td>
<td>41.6862</td>
<td>56.6</td>
<td>1793.10</td>
</tr>
<tr>
<td>01372500</td>
<td>Wappinger Ck near Wappingers Falls</td>
<td>-73.8726</td>
<td>41.6531</td>
<td>34.9</td>
<td>466.98</td>
</tr>
<tr>
<td>01413088</td>
<td>East Brànch Delaware R at Roxbury</td>
<td>-74.5597</td>
<td>42.2917</td>
<td>451.1</td>
<td>34.83</td>
</tr>
<tr>
<td>01413398</td>
<td>Bush Kill near Arkville</td>
<td>-74.6013</td>
<td>42.1509</td>
<td>420.6</td>
<td>120.49</td>
</tr>
<tr>
<td>01413408</td>
<td>Dry Bk at Arkville</td>
<td>-74.6232</td>
<td>42.1468</td>
<td>408.4</td>
<td>210.43</td>
</tr>
<tr>
<td>01413500</td>
<td>East Brànch Delaware R at Margaretville</td>
<td>-74.6535</td>
<td>42.1448</td>
<td>397.0</td>
<td>420.54</td>
</tr>
<tr>
<td>01414000</td>
<td>Platte Kill at Dunraven</td>
<td>-74.6954</td>
<td>42.1331</td>
<td>394.6</td>
<td>90.04</td>
</tr>
<tr>
<td>01414500</td>
<td>Mill Bk near Dunraven</td>
<td>-74.7304</td>
<td>42.1062</td>
<td>395.8</td>
<td>65.02</td>
</tr>
<tr>
<td>01415000</td>
<td>Tremper Kill near Andes</td>
<td>-74.8185</td>
<td>42.1201</td>
<td>391.9</td>
<td>85.66</td>
</tr>
<tr>
<td>01420500</td>
<td>Beaver Kill at Cooks Falls</td>
<td>-74.9796</td>
<td>41.9464</td>
<td>351.0</td>
<td>621.78</td>
</tr>
<tr>
<td>01421610</td>
<td>West Brànch Delaware R at Hobart</td>
<td>-74.6690</td>
<td>42.3715</td>
<td>499.9</td>
<td>39.99</td>
</tr>
<tr>
<td>01421618</td>
<td>Town Bk SE of Hobart</td>
<td>-74.6621</td>
<td>42.3612</td>
<td>509.0</td>
<td>36.89</td>
</tr>
<tr>
<td>01421900</td>
<td>West Brànch Delaware R upstream from Delhi</td>
<td>-74.9071</td>
<td>42.2804</td>
<td>414.5</td>
<td>345.72</td>
</tr>
<tr>
<td>01422389</td>
<td>Coulter Bk near Bovina Center</td>
<td>-74.7360</td>
<td>42.2387</td>
<td>609.6</td>
<td>1.96</td>
</tr>
<tr>
<td>01422500</td>
<td>Little Delaware R near Delhi</td>
<td>-74.9016</td>
<td>42.3523</td>
<td>422.3</td>
<td>128.48</td>
</tr>
<tr>
<td>01432900</td>
<td>Mongaup R at Mongaup Valley</td>
<td>-74.7810</td>
<td>41.6684</td>
<td>359.7</td>
<td>197.63</td>
</tr>
<tr>
<td>01434017</td>
<td>East Brànch Neversink R near Claryville</td>
<td>-74.5402</td>
<td>41.9254</td>
<td>530.4</td>
<td>59.08</td>
</tr>
<tr>
<td>01434025</td>
<td>Biscuit Bk above Pigeon Bk at Frost Valley</td>
<td>-74.5010</td>
<td>41.9954</td>
<td>627.9</td>
<td>9.60</td>
</tr>
<tr>
<td>01435000</td>
<td>Neversink R near Claryville</td>
<td>-74.5899</td>
<td>41.8901</td>
<td>464.0</td>
<td>171.83</td>
</tr>
</tbody>
</table>