

DISSERTATION

A POLARIMETRIC RADAR ANALYSIS OF CONVECTION OBSERVED DURING
NAME AND TIMREX

Submitted by

Angela Kay Rowe

Department of Atmospheric Science

In partial fulfillment of the requirements

For the Degree of Doctor of Philosophy

Colorado State University

Fort Collins, Colorado

Summer 2011

Doctoral Committee:

Advisor: Steven Rutledge

Richard Johnson
Sue van den Heever
Timothy Lang
Richard Eykholt

ABSTRACT

A POLARIMETRIC RADAR ANALYSIS OF CONVECTION OBSERVED DURING NAME AND TIMREX

The mountainous regions of northwestern Mexico and southwestern Taiwan experience periods of intense rainfall associated with the North American and Asian monsoons, respectively, as warm, moist air is ushered onshore due to a reversal of mean low-level winds. Potentially unstable air is lifted along the steep topography, leading to convective initiation over the high peaks and adjacent foothills in both regions. In addition, an enhancement of convection in preexisting systems is observed due to interaction with the terrain, leading to localized heavy rain along the western slopes. The predictability of warm-season rainfall in these regions is limited by the lack of understanding of the nature of these precipitating features, including the diurnal variability and elevation-dependent trends in microphysical processes. Using polarimetric data from NCAR's S-band, polarimetric radar (S-Pol), deployed during the North American Monsoon Experiment (NAME) and Terrain-influenced Monsoon Rainfall Experiment (TiMREX), individual convective elements were identified and tracked, allowing for an analysis of hydrometeor characteristics within evolving cells. Furthermore, a feature classification algorithm was applied to these datasets to compare

characteristics associated with isolated convection to cells contained within organized systems.

Examples of isolated cells from a range of topography during NAME revealed the presence of Z_{DR} columns, attributed to the lofting of drops above the melting level, where subsequent freezing and growth by riming led to the production of graupel along the western slopes of the Sierra Madre Occidental (SMO) and adjacent coastal plain. Melting of large ice hydrometeors was also noted over higher terrain, leading to short-lived yet intense rainfall despite truncated warm-cloud depths compared to cells over the lower elevations. Cells embedded within mesoscale convective systems (MCSs) during NAME also displayed the combined roles of warm-rain and ice-based microphysical processes as convection organized along the terrain. In addition to enhancing precipitation along the western slopes of the SMO, melting ice contributed to the production of mesoscale outflow boundaries, which provided an additional focus mechanism for convective initiation over the lower elevations and resulted in propagation of these systems toward the coast.

Intense rainfall was also observed along the Central Mountain Range (CMR) in Taiwan; however, in contrast to the systems during NAME, this enhancement occurred as MCSs moved onshore within the southwesterly flow and intercepted the CMR's steep slopes. Elevated maxima in polarimetric variables, similar to observations in convection during NAME, indicated a contribution from melting ice to rainfall at these higher elevations. Vertical profiles of ice mass, however, revealed greater amounts throughout the entire vertical depth of convection during NAME. In addition, isolated cells during TiMREX were relatively shallow compared to organized convection in both regions.

Nonetheless, instantaneous rain rates were comparable during both experiments, suggesting efficient warm-rain processes within convection observed in the TiMREX radar domain and emphasizing a range of microphysical processes in these two regions. In addition, the greatest contribution to hourly accumulated rain mass in these regions was associated with deep organized systems along the western slopes, posing threats along the steep topography due to flash flooding and subsequent landslides, emphasizing the need for accurate prediction and understanding of the processes that lead to intense rainfall in these vulnerable regions.

ACKNOWLEDGEMENTS

I would like to express sincere gratitude to my advisor, Dr. Steven Rutledge, for the support and guidance he offered for my research, as well as for the encouragement to participate in additional activities to enhance my graduate experience and future career. I appreciate the comments and suggestions offered by my committee members, Dr. Richard Johnson, Dr. Sue van den Heever, Dr. Timothy Lang, and Dr. Richard Eykholt, who graciously accepted a last-minute request to serve on my committee. I also want to thank my former committee member, Dr. Jorge Ramirez, for his feedback in the early stages of this research.

My research experience involving TiMREX was greatly enhanced due to the National Science Foundation's East Asia and Pacific Summer Institute (EAPSI), which provided support for me to spend 10 weeks in Taiwan to analyze this data set under the supervision of Dr. Ben Jou of National Taiwan University. I am grateful to Professor Jou for serving as my host and providing valuable guidance throughout this research experience. I would also like to thank the members of his research group for their willingness to help me become accustomed to their lab environment and culture.

I would also like to acknowledge past and current members of the CSU Radar Meteorology group for their advice, support, and friendship. Special thanks is extended to Paul Hein for patiently responding to my programming questions and providing

continuous computer support. Finally, I want to express my appreciation of the support and encouragement offered by my family and friends throughout all these years.

The NAME research was funded by the National Science Foundation Grant ATM-0733396.

TABLE OF CONTENTS

PART I: INVESTIGATION OF MICROPHYSICAL PROCESSES OCCURRING IN ISOLATED CONVECTION DURING NAME	1
1. Introduction	2
2. Data and methodology	6
2.1 S-Pol data	6
2.2 Cell identification and tracking.....	7
2.3 Cell properties.....	11
2.4 Topographic data	12
3. Case studies	15
3.1 Coastal plain.....	15
3.2 Western slopes	18
3.3 SMO	21
4. General statistics	37
4.1 Rain rate and echo-top heights.....	37
4.2 Polarimetric observations.....	39
5. Discussion and conclusions	54
References	58
PART II: INVESTIGATION OF MICROPHYSICAL PROCESSES OCCURRING IN ORGANIZED CONVECTION DURING NAME	63
1. Introduction	64
2. Data and methodology	68
2.1 S-Pol data	68
2.2 Cell identification and tracking.....	69
2.3 Cell properties.....	70
2.4 Topographic data	71
3. Case studies	74
3.1 12 July.....	74
3.2 5 August	78

3.3 Additional features of interest.....	83
3.4 Case comparison	85
4. Comparison with isolated convection	106
4.1 Rain rate and echo-top heights.....	106
4.2 Polarimetric observations.....	107
5. Discussion and conclusions	116
References	120
 PART III: A STUDY OF MICROPHYSICAL PROCESSES	
OCCURRING IN CONVECTION DURING TIMREX.....	126
1. Introduction	127
2. Data and methodology	131
2.1 S-Pol data.....	131
2.2 Cell identification and tracking.....	133
2.3 Cell properties.....	134
2.4 Topographic data	135
2.5 Case selection.....	135
3. Case studies	138
3.1 30 May 0100–1200 UTC	139
3.2 31 May 0100–1100 UTC	141
3.3 1 June 0300–1100 UTC	145
3.4 13 June 1700 UTC – 14 June 1800 UTC.....	147
4. General statistics	177
4.1 Type comparison.....	177
4.2 Elevation comparison.....	178
5. Comparison with NAME	196
6. Discussion and conclusions	214
References	218
 PART IV: OVERALL SUMMARY AND CONCLUSIONS.....	
References	230

LIST OF TABLES

PART I

3.1	Median values of specific variables for all isolated cells in a particular elevation group.....	25
------------	---	----

LIST OF FIGURES

PART I

2.1	Topography of northwestern Mexico with locations of Mazatlan, S-Pol, and a wind profiler during NAME.....	13
2.2	Gridded composite reflectivity for 2209 and 2217 UTC on 10 July 2004 including cell numbers and ellipses	14
3.1	Frequency of occurrence of initial locations of isolated cells.....	26
3.2	An RHI through an isolated cell on 10 July 2004 at 2159 UTC	27
3.3	Vertical cross sections through a cell on 10 July 2004 at (a) 2146, (b) 2152, (c) 2202, and (d) 2209 UTC.....	28
3.4	Vertical cross sections through a cell on 10 July 2004 at 2146 UTC	29
3.5	Vertical cross sections through a cell on 19 August 2004 at (a),(b) 0000, (c),(d) 0015, and (e),(f) 0100 UTC	30
3.6	Contoured frequency by altitude diagrams (CFADs) of Z and Z_{DR} for a cell on 19 August 2004 at (a),(b) 0000, (c),(d) 0015, (e),(f) 0100, and (g),(h) 0130 UTC	31
3.7	Frequency contours of HID-identified graupel fractions for three isolated cells over the SMO on 14 August 2004.....	32
3.8	Vertical cross section through a cell on 14 August 2004 at 2215 UTC.....	33
3.9	Reflectivity CFADS for the mature stages of two isolated cells on 14 August 2004 at (a) 2215 UTC and (b) 2353 UTC	34
3.10	Vertical cross section through a cell on 20 July 2004 at 2015 UTC.....	35
3.11	Reflectivity CFADS for the mature stage of an isolated cell on 20 July 2004 at (a) 2015 and (b) 2030 UTC.....	36
4.1	Cumulative distribution functions of maximum rain rate for all isolated cells partitioned by elevation groups.....	44
4.2	Cumulative distribution functions of maximum rain rate for all isolated cells partitioned by echo-top height	45
4.3	Histogram of echo-top heights for isolated cells as a function of terrain	46
4.4	Cumulative distribution functions of the maximum height of 40-dBZ reflectivity in isolated cells as a function of terrain height	47
4.5	Cumulative distribution functions of (a) liquid water mass and (b) ice mass as a function of terrain height	48
4.6	Cumulative distribution functions of water mass fraction as a function of terrain height	49
4.7	Vertical profiles of mean ice mass as a function of terrain height.....	50
4.8	Probability distribution functions of (a) Z_{DR} and (b) K_{DP} in isolated cells as a function of terrain height	51
4.9	Contoured frequency diagrams of Z_{DR} and K_{DP} for isolated cells.....	52

4.10 Cumulative distribution functions of positive Z_{DR} identified above the melting level in isolated cells as a function of terrain	53
---	----

PART II

2.1 Gridded composite reflectivity for 2054 UTC on 5 August 2004 including cell numbers and ellipses	72
2.2 Frequency of occurrence of initial locations of organized cells	73
3.1 Vertical cross sections through cells at 0845 UTC on 12 July 2004	87
3.2 An RHI through a cell on 12 July 2004 at 0912 UTC	88
3.3 An RHI through a cell on 12 July 2004 at 0925 UTC	89
3.4 PPI images of Z and V_R on 12 July 2004 at 0929 UTC.....	90
3.5 Vertical cross sections through cells on 12 July 2004 at 1008 UTC	91
3.6 Vertical cross sections through cells on 12 July 2004 at 1018 UTC	92
3.7 Distributions for tracks over the low terrain and over water on 12 July 2004.....	93
3.8 Vertical cross sections through cells on 5 August 2004 at (a),(b) 1951 UTC and (c),(d) 2002 UTC.....	94
3.9 Vertical cross sections through cells on 5 August 2004 at (a),(b) 2021 UTC (c),(d) 2042 UTC, and (e),(f) 2047 UTC	95
3.10 PPI images of Z and V_R on 5 August 2004 at 2055 UTC	96
3.11 An RHI through a cell on 5 August 2004 at 2108 UTC	97
3.12 An RHI through a cell on 5 August 2004 at 2123 UTC	98
3.13 Vertical profiles of mean ice mass for organized cells on 5 August 2004 as a function of terrain height	99
3.14 A time series of IWP ratio and maximum rain rate for a track on 5 August 2004 as a function of terrain height	100
3.15 PPI images of Z and V_R on 21 July 2004 at 0152 UTC	101
3.16 An RHI through a cell on 21 July 2004 at 0208 UTC	102
3.17 Vertical profiles of mean ice mass for organized cells during 16–17 August as a function of terrain height	103
3.18 Distributions of maximum Z_{DR} for organized cells during case days as a function of terrain height	104
3.19 Cumulative distribution functions of (a) liquid water mass and (b) ice mass for organized cells during case days as a function of terrain height	105
4.1 Cumulative distribution functions of maximum rain rate for all cells divided by terrain height and cell type	111
4.2 Distributions of echo-top height for all cells divided by terrain height and cell type	112
4.3 Distributions of maximum Z_{DR} for all cells divided by terrain height and cell type	113
4.4 Contoured frequencies of K_{DP} as a function of Z_{DR} for (a) isolated cells over low terrain, (b) organized cells over low terrain, (c) organized cells over water, and (d) organized cells over land	114
4.5 Profiles of (a) mean ice mass and (b) mean liquid water mass for all cells divided by terrain height and cell type.....	115

PART III

2.1	Topography of southwestern Taiwan with radar locations during TiMREX	137
3.1	Vertical cross sections of K_{DP} through a cell on 30 May 2008 at (a) 0422, (b) 0430, (c) 0437, and (d) 0445 UTC	150
3.2	Vertical cross sections of HID through a cell on 30 May 2008 at (a) 0422, (b) 0430, (c) 0437, and (d) 0445 UTC	151
3.3	Vertical profiles of mean Z_{DR} corresponding to the cell evolution of Fig. 3.2	152
3.4	An RHI through cells on 30 May 2008 at 0450 UTC	153
3.5	Vertical cross sections through cells on 30 May 2008 at (a) 0715 and (b) 0730 UTC	154
3.6	Cumulative distribution functions of maximum rain rate for all cells on 30 May 2008 as a function of elevation	155
3.7	Vertical cross sections through a cell on 31 May 2008 at 0422 UTC	156
3.8	Vertical cross sections through cells on 31 May 2008 at (a) 0815 UTC, (b) 0822, (c) 0830, and (d) 083 UTC	157
3.9	Time series of IWP ratio, maximum rain rate, mean drop diameter, and maximum terrain height for a cell on 31 May 2008	158
3.10	An RHI through a cell on 31 May 2008 at 0758 UTC	159
3.11	An RHI through a cell on 31 May 2008 at 0806 UTC	160
3.12	An RHI through a cell on 31 May 2008 at 0813 UTC	161
3.13	An RHI through a cell on 31 May 2008 at 0821 UTC	162
3.14	Cumulative distribution functions of maximum rain rate for all cells on 31 May 2008 as a function of elevation	163
3.15	Vertical cross sections of K_{DP} for a cell on 1 June 2008 at (a) 0530, (b) 0537, (c) 0545, (d) 0552 UTC	164
3.16	Vertical cross sections of HID for a cell on 1 June 2008 at (a) 0530, (b) 0537, (c) 0545, (d) 0552 UTC	165
3.17	An RHI through a cell on 1 June 2008 at 0551 UTC	166
3.18	Distributions of Z_{DR} and K_{DP} for the cells shown in Figs. 3.16 and 3.17	167
3.19	Vertical cross sections through a cell on 14 June 2008 at 0507 UTC	168
3.20	An RHI through a cell on 14 June 2008 at 0751 UTC	169
3.21	An RHI through a cell on 14 June 2008 at 0851 UTC	170
3.22	Vertical cross sections through a cell on 14 June 2008 at 0530 UTC	171
3.23	Time series of IWP ratio, maximum rain rate, mean drop diameter, and maximum terrain height for a cell on 14 June 2008	172
3.24	Vertical profiles of mean (a) Z_{DR} and (b) K_{DP} for part of a track on 14 June 2008	173
3.25	Percentages of Z_{DR} for all cells on 13–14 June 2008 as a function of elevation and cell type	174
3.26	Vertical profiles of (a) mean liquid water mass and (b) mean ice mass for all cells on 13–14 June as a function of elevation and cell type	175
3.27	Cumulative distribution functions of maximum rain rate for all cells on 13–14 June 2008 as a function of elevation and cell type	176

4.1	Cumulative distribution functions of maximum rain rate for isolated and organized cells during TiMREX over all elevations.....	182
4.2	Distributions of echo-top height for isolated and organized cells during TiMREX over all elevations	183
4.3	Cumulative distribution functions of maximum rain rate for isolated cells during TiMREX grouped by echo-top height	184
4.4	Distributions of maximum Z_{DR} for isolated and organized cells during TiMREX over all elevations	185
4.5	Distributions of maximum K_{DP} for isolated and organized cells during TiMREX over all elevations	186
4.6	Mean drop diameter as a function of maximum rain rate for all cells during TiMREX divided by cell type	187
4.7	Cumulative distribution functions of maximum rain rate for all cells during TiMREX grouped by cell type and elevation	188
4.8	Distributions of echo-top height for all cells during TiMREX grouped by cell type and elevation	189
4.9	Cumulative distribution functions of the maximum height of 30 dBZ for all cells during TiMREX grouped by cell type and elevation	190
4.10	Distributions of maximum Z_{DR} for all cells during TiMREX grouped by cell type and elevation	191
4.11	Mean drop diameter as a function of maximum rain rate for all cells during TiMREX grouped by cell type and elevation	192
4.12	Cumulative distribution functions of ice mass fraction for all cells during TiMREX grouped by cell type and elevation	193
4.13	Vertical profiles of (a) mean liquid water mass and (b) mean ice mass grouped by cell type and elevation	194
4.14	Contoured frequencies of K_{DP} as a function of Z_{DR} for organized cells (a) over water, (b) over the coast, and (c) over the terrain.....	195
5.1	Cumulative distribution functions of maximum rain rate for all cells during NAME and TiMREX grouped by cell type	200
5.2	Cumulative distribution functions of maximum rain rate for isolated cells during NAME and TiMREX grouped by elevation.....	201
5.3	Cumulative distribution functions of maximum rain rate for organized cells during NAME and TiMREX grouped by elevation	202
5.4	Distributions of echo-top height for all cells during NAME and TiMREX grouped by cell type.....	203
5.5	Distributions of the maximum height of 30 dBZ for all cells during NAME and TiMREX grouped by cell type	204
5.6	Cumulative distribution functions of maximum rain rate for isolated cells over all elevations during NAME and TiMREX grouped by echo-top height	205
5.7	Cumulative distribution functions of maximum rain rate for isolated cells over the slopes during NAME and TiMREX grouped by echo-top height	206
5.8	Cumulative distribution functions of hourly accumulated rain mass for all cells during NAME and TiMREX grouped by cell type	207

5.9	Cumulative distribution functions of hourly accumulated rain mass for organized cells during NAME and TiMREX grouped by elevation.....	208
5.10	Distributions of maximum Z_{DR} for all isolated cells during NAME and TiMREX grouped by elevation.....	209
5.11	Distributions of maximum Z_{DR} for all organized cells during NAME and TiMREX grouped by elevation.....	210
5.12	Vertical profiles of (a) mean water mass and (b) mean ice mass for all cells during NAME and TiMREX grouped by cell type.....	211
5.13	Cumulative distribution functions of HID-identified graupel fraction for all cells during NAME and TiMREX grouped by elevation	212
5.14	Cumulative distribution functions of water mass fraction for all cells during NAME and TiMREX grouped by elevation	213

**PART I: INVESTIGATION OF MICROPHYSICAL PROCESSES OCCURRING
IN ISOLATED CONVECTION DURING NAME**

CHAPTER 1

INTRODUCTION

The North American Monsoon (NAM) develops in the Northern Hemisphere summer in response to changes in large-scale land-sea temperature contrasts (Higgins et al. 1997). Similar to its Asian counterpart, the onset of the NAM is characterized by a reversal in the mean low-level flow (Badan-Dangon et al. 1991) and marks an abrupt transition from hot, dry weather to relatively cooler, wetter conditions in the semiarid regions of southwestern U.S. and western Mexico (e.g., Douglas et al. 1993; Adams and Comrie 1997). Areas in northwestern Mexico receive 50–80 % of annual water through monsoon rainfall (Gochis et al. 2006), highlighting the need to accurately model and forecast precipitation. The predictability of warm-season rainfall within this core monsoon region is limited by the lack of understanding of the nature of the precipitating systems, and further complicated by the presence of steep topography (e.g., Adams and Comrie 1997; Berbery 2001; Gochis et al. 2002). A comprehensive network of instrumentation was deployed for the North American Monsoon Experiment (NAME) during the summer of 2004 to characterize and understand precipitation processes influenced by the complex terrain of northwestern Mexico (Higgins et al. 2006). The Sierra Madre Occidental (SMO), extending above 3 km, dominates the local topography of this region and has been found to strongly influence the diurnal cycle of NAM precipitation (e.g., Gochis et al. 2004, 2007; Lang et al. 2007). Gochis et al. (2004),

using data from the NAME Event Rain gauge Network (NERN), described an elevation-dependent diurnal cycle in precipitation where rainfall occurs earliest and most frequently over the SMO, followed by a late evening/overnight peak across lower elevations with less frequency, but greater intensity. This diurnal trend has been mostly attributed to a westward propagation of systems off the SMO and upscale growth into mesoscale convective systems (MCSs) over the coastal plain (Lang et al. 2007).

In addition to the diurnal variability of precipitation, the vertical structure and microphysical characteristics of convection require investigation to improve the understanding and prediction of warm-season rainfall in the NAM region. In particular, knowledge of the microphysical structure is crucial for accurate estimates of precipitation from remote sensing observations (e.g., Nesbitt et al. 2008). In addition, hydrometeor identification allows for microphysical and convective parameterizations to be validated and improved, which significantly affect the model-simulated regional climate during the NAM (e.g., Gochis et al. 2002). Previous studies from NAME showed that convection over the high terrain of the SMO tended to be shallower than convection over the foothills and lower terrain (e.g., Nesbitt et al. 2008; Rowe et al. 2008). Analyses of heat sources, heating rates, and moisture sources/sinks also indicated shallow convection over the high terrain during the early afternoon hours (Johnson et al. 2010), consistent with radar and satellite observations. In addition to taller echo tops, Rowe et al. (2008) illustrated that convection over the coastal plain was characterized by greater warm-cloud depths, defined as the distance between the lifting condensation level (LCL) and the melting level (located at an average height of 5 km) and therefore a deeper layer over which accretional growth could occur. It was suggested that these elevation-dependent

trends in rainfall intensity may be explained by the observed differences in warm-cloud depth between the SMO and coastal plain (Rowe et al. 2008). Vertically intense cells containing large ice particles (i.e., graupel and small hail) were also observed in the NAME domain (Higgins et al. 2006), suggesting an additional contribution of cold-cloud processes to rainfall production. Therefore, a major remaining objective of NAME (and the basis of this study) is to describe microphysical processes, including the relative roles of warm-rain and ice-based particle growth, as a function of topography.

A recent study by Lang et al. (2010), using radar data from NAME, further suggested an increase in intensity with decreasing elevation for all convective features in the domain. Their analyses revealed an increase in ice and water masses associated with convection over the lower terrain, as well as larger rain drops, compared to over the SMO. These results further indicate differences in microphysical processes as a function of elevation, where shallower warm-cloud depths and liquid water mass imply a lesser role of warm-rain processes in convection over the higher terrain, yet increased ice mass over the coastal plain suggests an important contribution from ice-based processes in enhancing rainfall intensity.

In order to address remaining questions about the microphysical processes occurring in this region, this study used data from the National Center for Atmospheric Research (NCAR) S-band, polarimetric, Doppler radar (S-Pol) to provide information regarding the distribution and evolution of hydrometeor types in NAME convection that could be detected by S-Pol. Compared to previous NAME studies that included all precipitating features, this analysis focused on unorganized convection to evaluate the vertical structure and temporal evolution of individual convective elements through a

case study approach. The characteristics of organized (mesoscale) convection in this core monsoon region will be examined in Part II.

CHAPTER 2

DATA AND METHODOLOGY

2.1 S-Pol data

During the 2004 NAME field campaign, the S-Pol radar was located along the Gulf of California (GoC) coast about 90 km north of Mazatlan (Fig. 2.1), providing nearly continuous data from 8 July through 21 August (Higgins et al. 2006; Lang et al. 2007). Full-volume, 360° scans every 15 minutes at elevation angles of 0.8°, 1.3°, and 1.8° allowed for echo detection and rain rate estimation out to a range of 210 km (Fig. 2.1; solid circle). A less common sector-scanning mode focused on specific precipitation features, with sector widths between 90° and 120° and elevation angles selected based on the proximity of the features to the radar, to provide increased temporal (and spatial) resolution of six to seven minutes for 36 cases at a maximum range of 150 km (Fig. 2.1; dashed circle). In some cases, RHIs were included to complete the total 15-minute scanning pattern, providing fine-scale vertical resolution of convection. Due to this improved temporal information, radar volumes used in this study were selected from these microphysical cases, comprising about 95 hours of total scanning time. It is necessary to note that these scans were likely focusing on interesting, intense cases, therefore neglecting the weaker convection, especially over the SMO. For this reason, cases presented in this study represent more intense examples of convection compared to previous NAME studies.

Extensive quality control of the raw S-Pol data, discussed in detail in other NAME publications (Lang et al. 2007, 2009), involved calibration of the power-based variables, removal of non-meteorological and second-trip echo, attenuation correction, and correction for partial beam blockage. Accuracy of reflectivity was ± 1 dBZ except in blocked regions. Differential phase (ϕ_{DP}) was filtered, based on a technique described by Hubbert and Bringi (1995), in order to calculate specific differential phase (K_{DP}), which was used for rainfall analyses and hydrometeor identification. The suite of radar variables used in this study included horizontal reflectivity (Z_H), differential reflectivity (Z_{DR}), providing information about oblateness of the hydrometeors, linear depolarization ratio (L_{DR}), and the zero-lag cross-correlation coefficient ($\rho_{HV}(0)$), which both aid in distinguishing between pure rain and mixtures of hydrometeors, and K_{DP} . A more complete description of the polarimetric variables and their applications to microphysical studies can be found in Bringi and Chandrasekar (2001). The corrected radar data was then gridded to Cartesian coordinates at a 1-km horizontal and 0.5-km vertical resolution using the program REORDER (Mohr et al. 1986). Finally, using the gridded polarimetric variables (Z_H , Z_{DR} , K_{DP} , L_{DR} , $\rho_{HV}(0)$) and a mean temperature profile from Mazatlan, a hydrometeor classification algorithm, based on the methodology of Liu and Chandrasekar (2000) and described in detail by Tessendorf et al. (2005), was applied to determine the dominant hydrometeor type at each horizontal and vertical grid point.

2.2 Cell identification and tracking

In order to objectively identify convection during NAME, a tracking algorithm, similar to that described by Gauthier et al. (2010), was applied to the gridded S-Pol radar

data. This variation of the Thunderstorm Identification, Tracking, Analysis, and Nowcasting (TITAN) program (Dixon and Wiener 1993) was used to locate and track individual cells throughout the S-Pol domain over the time period of each case. This technique, referred to as the centroid identification and tracking scheme, has been applied in previous studies to develop nowcasting algorithms using radar data (e.g., Crane 1979; Rosenfeld 1987; Dixon and Wiener 1993; Johnson et al. 1998). The strength of this method over others (e.g., cross-correlation tracking; Rinehart and Garvey 1979; Li et al. 1995) is its effectiveness in tracking individual, isolated storms, thus allowing for temporal evolution of cell characteristics to be investigated (Johnson et al. 1998).

Prior to identifying cells, a composite reflectivity field was created by determining the maximum reflectivity within the vertical column at each horizontal grid point. At each time step, regions with composite reflectivity > 35 or 45 dBZ were defined as potential cells with a unique index number. Although absent in the single-threshold TITAN algorithm presented by Dixon and Wiener (1993) and the tracking methodology used by Gauthier et al. (2010), these multiple reflectivity thresholds have been implemented in the Storm Cell Identification and Tracking (SCIT) algorithm (Johnson et al. 1998), where an improvement in identification of intense, embedded cells has been shown. The choice of using the specific thresholds of 35 and 45 dBZ was somewhat arbitrary; however, these values set a reasonable limit for convection (e.g., Demott and Rutledge 1998), thereby restricting the identification to convective-only elements. Sensitivity tests on these values have shown that lowering the threshold reduces the intensity and increases the size of the cells, with a greater number of apparent cell mergers (e.g., Dixon and Wiener 1993; Gauthier et al. 2010). For this analysis, isolated

cells were compared based on terrain elevation as opposed to convective versus stratiform elements; therefore, the sensitivity of the reflectivity thresholds did not impact the results of this study.

A minimum area threshold was also implemented to prevent tracking of any noise or clutter remaining in the gridded data. For this study, a two-pixel minimum (i.e., 2 km²) was required for cells identified by the 45-dBZ threshold, and, similar to the area threshold used in SCIT (Johnson et al. 1998), potential cells with reflectivity > 35 dBZ had to contain at least ten grid points, corresponding to an area of 10 km². This differs slightly from the 8-km² threshold imposed by Gauthier et al. (2010), but both act to eliminate noisy pixels. For cells that met the reflectivity and area criteria, an ellipse-fitting method, described by Nesbitt et al. (2006), was used to estimate the major and minor dimensions of the identified cell. To focus this study on unorganized, isolated convection, a feature identification scheme was then applied to the cells. The identification of precipitating features during NAME is described in detail by Lang et al. (2007), and the classification of feature types, based on a scheme described by Rickenbach and Rutledge (1998), is explained by Pereira (2008) for this data set. Specifically, if the major axis of the cell's fitted ellipse did not exceed 100 km, and if the aspect ratio between the major and minor axes of the convective areas was less than five to one (Bluestein and Jain 1995), then the cell was considered sub-MCS, non-linear; these cells will be referred to as isolated convection in this study. This differs from previous NAME radar studies (e.g., Lang et al. 2007, 2010; Rowe et al. 2008) where convective partitioning was applied to reflectivity grids regardless of size or degree of

organization. A comparison of these observations of isolated convection to convective cells that exhibit mesoscale organization will be presented in Part II.

Initially, all individual cells in the volume scan were assigned unique track numbers, and the cell's position was recorded as the location of the reflectivity-weighted centroid. For the next radar volume, this initial location was used as the projected position of the track to be compared to the cell locations in the new volume. The distance between the projected center of the track and the center of the new cell was computed in both E-W and N-S directions, after which the angle between these distances was used to convert the projected path to Cartesian coordinates. Finally, using the axis information from the initial ellipse, the projected ellipse was determined for the track. Similarly, this method tested whether the new cell's ellipse contained the projected center of the track. The mean movement of the track was then added to the previous reflectivity-weighted centroid at each subsequent time step to relate new cells to previously defined tracks. If more than one possible track was identified for a given cell, a merger of cells was noted, and the cell with the peak reflectivity dominated the cell in the merger. An example of this can be seen in Fig. 2.2, where a merger was evident between cells 144 and 146 at time 2209 UTC, leading to the formation of cell 158 at time 2217 UTC. In this case, cell 158 was considered part of the same track as cell 144 (the cell with higher reflectivity), and the track associated with cell 146 was terminated. Also, if several new cells were associated with an existing track, this scenario was classified as a split. Any new cells that did not correspond to a previous track were given a new track number such that all cells in a volume were associated with a specific track. This method created some complications for evaluating a cell's lifetime, as it needed to be determined if the initial

(final) point of a cell's track was representative of the growing (decaying) stage of the cell or if it resulted from a split (merger); however, especially for the case studies presented in this study, it proved useful for studying the evolution of individual convective cells.

2.3 Cell properties

Each grid point contained within the tracked cells was matched with the gridded polarimetric data at that location, and vertical cross sections of polarimetric variables were created through isolated cells to investigate their vertical characteristics. Echo-top heights were estimated by the height of the 0-dBZ reflectivity contour in each individually tracked cell. To aid in the microphysical analyses, ice and liquid water masses were also calculated for each cell based on the methodology described by Carey and Rutledge (2000) and Cifelli et al. (2002). Using a modified version of the Colorado State University blended polarimetric algorithm (Cifelli et al. 2002), near-surface rain rates were computed for each cell location. Polarimetric-based equations for estimating rainfall, tailored to the NAME region, were applied to this radar data set in a similar manner as previous studies (e.g., Lang et al. 2007; Rowe et al. 2008). Applying this algorithm to all grid points within a cell allowed for mean and maximum rain rates to be calculated for each convective element. By relating these rain rates to hydrometeor characteristics within the cell, details about the microphysical processes occurring in this region could be described, as well as the specification of elevation-dependent trends and changes that occurred as the precipitating features evolved.

2.4 Topographic data

Topographic data were obtained from the National Geophysical Data Center (NGDC), available on a 0.02° grid, and was matched to the radar grids, as described in Lang et al. (2007). Maximum terrain height was determined within a 2-km area surrounding a particular grid point. This information was divided into four elevation groups: 0–1 km, 1–2 km, > 2 km, and over water, in order to remain consistent with NAME precipitation studies by Gochis et al. (2004) and Rowe et al. (2008). Of the isolated cells identified during the microphysical scans, 29% were located over the coastal plain (0–1 km), 19% over the western slopes (1–2 km), 41% over the high terrain of the SMO (> 2 km), and the remaining 11% over water. The following radar analysis is organized over these four elevation groupings.

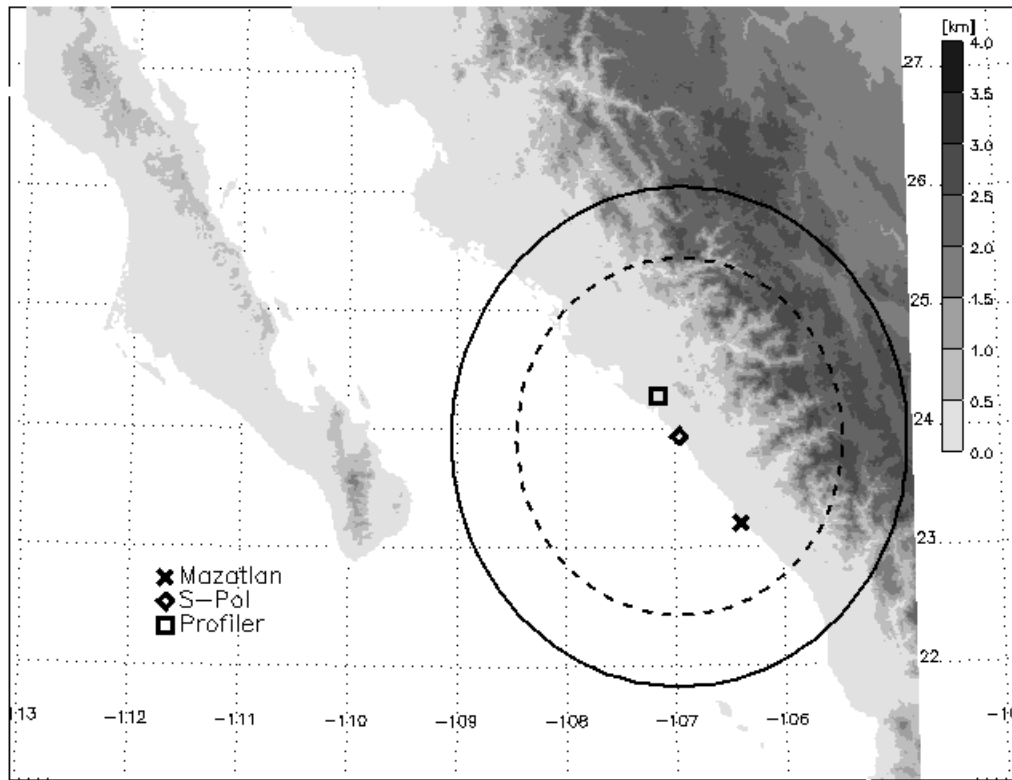


Fig. 2.1. Topography of northwestern Mexico, shaded from 0 to 4 km by increments of 0.5 km. The locations of Mazatlan, the S-Pol radar, and an S-band wind profiler are indicated by symbols along the coast. Maximum ambiguous range rings associated with S-Pol are shown as solid and dashed circles corresponding to 250 and 150 km, respectively.

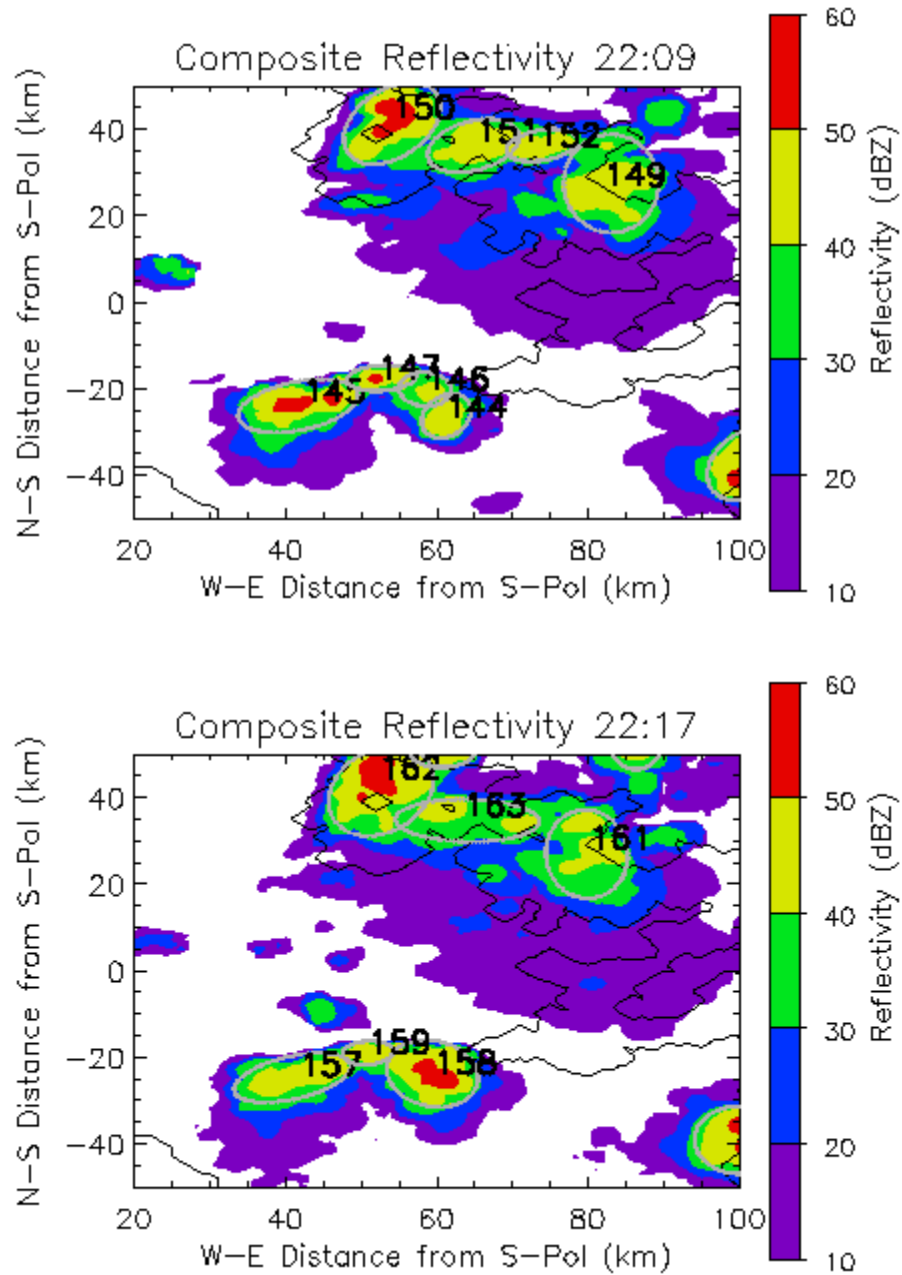


Fig. 2.2. Gridded composite reflectivity (color-filled contours) for 2209 UTC (top) and 2217 UTC (bottom) on 10 July 2004. Cell ellipses are shown in gray and the corresponding cell numbers are identified by bold, black numbers. Overlaid in black contours is the topography at 0, 1, 2, and 3 km.

CHAPTER 3

CASE STUDIES

To evaluate the relative roles of warm-rain and ice-based precipitation production, a case study approach was used to locate and describe the evolution of hydrometeors within isolated cells. Analysis of a cell in this manner required its entire lifecycle to be captured by S-Pol, and cases were chosen over a range of topography. Although previous studies have noted differences in convection occurring over the water compared to that over land (i.e., Lang et al. 2010), this study neglects cells over water due to the low frequency of isolated convection there. The majority of cells over the GoC were associated with organized systems moving off the coast during the early morning hours, while initiation of isolated convection dominated the high terrain of the SMO (Fig. 3.1). The following sub-sections describe a number of cases selected from the upper 25th percentile of tracks, in terms of echo-top height, reflectivity, and maximum rain rate, within each elevation group to compare and contrast intense convection over a range of topography. Median values of properties from all tracked isolated cells are presented in Table 3.1 to provide a basis for comparison for the following cases.

3.1 Coastal plain

An example of intense isolated convection over low terrain (0–1 km) occurred on 10 July 2004. On this day, upper-level (200 hPa) northerly flow overlaid low-level (850

hPa) southwesterly flow in an environment with approximately 2000 J kg^{-1} of CAPE. The lack of any large-scale synoptic features influencing the area allowed for diurnally forced convection to dominate the precipitation totals, with convective initiation focused over the SMO during the afternoon. Additional convection formed over the lower terrain around 2130 UTC, with the cell of interest developing over a slightly elevated area on the coastal plain due to diurnal heating. The maximum rain rate for this cell was 81 mm h^{-1} compared to a median maximum rate of 36 mm h^{-1} for isolated cells over the low terrain (Table 3.1). Peak reflectivity of 59 dBZ exceeded the median value by 12 dBZ, and echo-top heights also were above average, although the radar did not top this cell due to its proximity to the radar (approximately 50 km).

An RHI from 2159 UTC (Fig. 3.2) provides a detailed view of the vertical structure and hydrometeor characteristics of this intense, isolated cell through analysis of the polarimetric data. This image highlights a mature cell with a deep, narrow reflectivity core of 50 dBZ reaching 10 km and a narrow column of $Z_{\text{DR}} > 1.5 \text{ dB}$ extending 1–2 km above the melting level height of 5 km, commonly referred to as a positive Z_{DR} column. Such Z_{DR} columns imply the presence of supercooled raindrops lofted by strong updrafts (e.g., Illingworth et al. 1987), and have also been observed in numerous studies of convection including hail-producing, U.S. High Plains storms (e.g., Bringi et al. 1996; Tessendorf et al. 2005), isolated cells in Florida (e.g., Jameson et al. 1996; Bringi et al. 1997) and Alabama (Fulton and Heymsfield 1991), intense convection over the Tiwi Islands (e.g., Carey and Rutledge 2000; Keenan et al. 2000), and convection in Amazonia (e.g., Cifelli et al. 2002). Typically, a region of enhanced L_{DR} , known as an L_{DR} “cap,” is observed above the Z_{DR} column due to the rapid freezing of supercooled drops and

subsequent acquisition of a water coat due to collisions with liquid drops (e.g., Bringi et al. 1996). Jameson (1983) noted the presence of an L_{DR} cap atop a positive Z_{DR} column in the inflow side of a cell, indicating a process in which raindrops within the column likely grew initially via the collision-coalescence process. As updrafts increased, supercooled drops were lofted above the melting level into colder environments, allowing the drops to freeze and serve as embryos for rapid accretional growth. This process can be inferred from Fig. 3.2 with L_{DR} values increasing to -16 dB above the Z_{DR} column, likely due to the existence of wetted, aspherical ice particles (e.g., graupel). Values of $\rho_{HV}(0)$ were lower in this region compared to the surrounding areas (0.9–0.92), further indicating the coexistence of liquid and frozen drops and a dispersion of particle shapes.

A series of vertical cross sections through this cell (Fig. 3.3) from 2146 UTC through 2209 UTC places the previously described RHI in context of the cell's evolution. The initial occurrence of the Z_{DR} column (Fig. 3.3a) coincided with the presence of an L_{DR} cap (Fig. 3.4a), indicating the lofting of raindrops above the melting level and production of graupel and small hail. Subsequent melting of these large ice hydrometeors can be inferred from the local maximum in K_{DP} , with values exceeding 2° km^{-1} below the melting level (Fig. 3.3a). Indeed, the HID indicated a large region of hail at this time (Fig. 3.4b), further suggesting the riming processes implied from the Z_{DR} and L_{DR} fields. This, however, was short-lived as the large ice hydrometeors rapidly fell out, melted, and continued to produce enhanced K_{DP} below the melting level at 2152 and 2202 UTC (Fig. 3.3b,c). At 2202 UTC, the Z_{DR} column was no longer present, suggesting a weakening of the updraft, and by 2209 UTC (Fig. 3.3d), the core had descended with maximum K_{DP} near the surface, marking the decaying stage of the cell.

This particular example showed that in addition to coalescence, mixed-phase processes played a role in rainfall production in this intense convective cell via drop freezing and subsequent riming growth, similar to processes described in tropical cumulonimbi (e.g., Takahashi 1990; Takahashi and Kuhara 1993). Similar results were also presented from the Tropical Rainfall Measuring Mission Large-Scale Biosphere-Atmosphere (TRMM-LBA) experiment in Amazonia, where lofting of supercooled rain water into the mixed-phase zone and the resulting production of large amounts of precipitation ice via drop freezing was observed in convection occurring during the easterly regime (Cifelli et al. 2002). In addition, this process was noted during the Maritime Continent Thunderstorm Experiment (MCTEX) over the Tiwi Islands where an analysis of a lightning-producing cell highlighted the freezing of supercooled raindrops as an important source of precipitation-sized ice in tropical convection (Carey and Rutledge 2000).

3.2 Western slopes

Over the western slopes of the SMO (1–2 km), where rain rates tended to be maximized relative to the other terrain elevations, the vertical structure of intense, isolated convection was similar to cells over the coastal plain. One example, representing an intense cell in this elevation group, developed and persisted over the western slopes about 60 km to the northeast of S-Pol during the evening of 18 August 2004. Maximum rain rates for this cell exceeded 100 mm h^{-1} , compared to the median maximum rate over the western slopes of 33 mm h^{-1} (Table 3.1). Echo tops reached 15.5 km, compared to a median of 13 km, and the peak reflectivity of 53 dBZ exceeded the median by 7 dBZ. At

0000 UTC on 19 August, a vertical cross section through this feature (Fig. 3.5a) revealed a relatively shallow cell with weak reflectivity and low K_{DP} ($0.5^\circ \text{ km}^{-1}$). The HID at this time (Fig. 3.5b) identified graupel; however, these hydrometeors were limited to below 7 km during this early stage of development. Fifteen minutes later, the radar echo top of the cell reached 15 km (Fig. 3.5c) with graupel up to 11 km (Fig. 3.5d). This rapid growth, coincident with the increasing abundance of graupel, suggests a rapid glaciation process, where the rise in echo top results from the release of the latent heat of fusion (e.g., Cotton and Anthes 1989).

An RHI was not available for this cell; therefore, contoured frequency by altitude diagrams (CFADs) of reflectivity and Z_{DR} were employed as an additional method for evaluating microphysical processes (Yuter and Houze 1995). In particular, the key feature of the reflectivity CFAD involves the onset of diagonalization of the mode of the joint frequency distribution of reflectivity with height, which indicates the fallout of large hydrometeors and termination of accretional growth processes; in addition, the slope of the Z_{DR} CFAD reveals information about the corresponding drop sizes. Figure 3.6 shows CFADs of reflectivity and Z_{DR} for the previously described example over a selection of times throughout the cell's evolution. The abrupt vertical growth of the cell between 0000 and 0015 UTC is also seen in these figures (Fig. 3.6a,c), with reflectivities of 30 dBZ reaching 14 km at the latter time. In the lower levels, the mode of reflectivity shifted from 35 to 45 dBZ during this 15-minute interval, with a local maximum in frequency just beneath the melting level at 0015 UTC, characteristic of melting ice hydrometeors. This corresponded to the elevated maxima in reflectivity and to $K_{DP} > 1^\circ \text{ km}^{-1}$ seen in the vertical cross section (Fig. 3.5b), indicating the melting of graupel following the rapid

glaciation of the system. In addition, the occurrence of negative Z_{DR} at 0000 UTC above the melting level indicated the presence of large ice hydrometeors, which likely formed due to the freezing of drops and subsequent riming. This coincides with the rapid growth of the cell due to glaciation, as previously described, and the shift in the peak and range of Z_{DR} values in the low levels from 0000 UTC (Fig. 3.6b) to 0015 UTC (Fig. 3.6d) emphasized the contribution to rainfall from melting ice.

By 0100 UTC, the reflectivity core and largest K_{DP} values had lowered to near cloud base (Fig. 3.5e), coincident with the peak rain rate of 100 mm h^{-1} . The reflectivity CFAD at this time (Fig. 3.6e) shows a broader range of low-level reflectivity values, extending from 0 to 55 dBZ, further indicating a descending core. Similar to the analysis of Yuter and Houze (1995), this shift in the distribution indicated the mature stage of the cell, where fewer new particles were being injected into the upper levels of the cell and the large, more reflective ice particles were falling out and melting. This also corresponded to a wide range of Z_{DR} values below the melting level (Fig. 3.6f), which suggests contributions from both large and small drops to the rain water mass. At 0130 UTC, the final time of the cell's track, increased diagonalization is apparent in the reflectivity CFAD (Fig. 3.6g), indicating the demise of dominant accretional growth and suggesting a weakening of the main updraft (e.g., Zeng et al. 2001). In addition, the peak in frequency of Z_{DR} (Fig. 3.6h) at low levels shifted to lower values, signifying a reduced contribution from large, melting hydrometeors to the rainfall and likely the breakup of larger drops. Similar to the cell over the coastal plain, this case highlighted a process in which the lofting of supercooled water above the melting level, subsequent accretional growth leading to graupel formation, and the melting of the ice hydrometeors contributed

to intense rainfall. This convective microphysical scenario, discussed in other studies regarding tropical convection (i.e., MCTEX, Carey and Rutledge 2000; TRMM-LBA, Cifelli et al. 2002), was also conceptually described in the tropical-like convection of the Fort Collins flood (Petersen et al. 1999) as an “accumulation zone” model of precipitation production through a coupling of warm-rain and ice-particle accretion processes.

3.3 SMO

Reduced rainfall intensity, frequently observed over the higher terrain of the SMO (relative to lower terrain), has been related to shallower warm-cloud depths compared to the coastal plain (Rowe et al. 2008), as well as to an overall reduced amount of precipitation-sized ice aloft (Nesbitt et al. 2008; Lang et al. 2010). To further investigate these potential differences in microphysical processes, several examples of average, isolated cells over the SMO (> 2 km), located between 90 and 100 km from S-Pol were selected from 14 August 2004; a day characterized by scattered afternoon convection. Figure 3.7 shows the depth of graupel, as identified by the HID algorithm, for three cells on this day. In general, the vertical extent of graupel in these cells was limited to below 10 km, compared to 15 km over the lower elevations. This observation is consistent with the hypothesis that cells over the higher terrain had a shallower depth of precipitation-sized ice compared to the convection over the coastal plain (Nesbitt et al. 2008).

A vertical cross section through one of these isolated cells over the SMO provides further evidence for reduced warm-cloud depth and overall shallower convection over the high terrain (Fig. 3.8). Radar echo tops extended to 12 km during this mature stage of the cell, but the 30-dBZ contour remained below 8 km and K_{DP} values were $< 1^{\circ} \text{ km}^{-1}$,

possibly due to reduced precipitation water mass below the melting level or to the lack of large drops resulting from less precipitation-sized ice. This cell also lacked a Z_{DR} column; a feature observed in intense cells over lower elevations. The absence of this characteristic, along with the reduced vertical extent of graupel throughout the cell, suggests lesser emphasis on riming processes in addition to a reduced role of collision-coalescence below the melting level. Reflectivity CFADs for two mature cells on this day (Fig. 3.9a,b) show increased diagonalization compared to the mature, isolated cell over the western slopes, implying more dominant accretional processes in cells occurring over lower elevations. In addition, the mode of 35-dBZ reflectivities was lower than that over the coastal plain, indicating overall weaker convection. These observations suggest that in addition to the shallower warm-cloud cloud depth in isolated cells over the higher terrain, a reduced dependence on riming in these cells contributed to lower median rain rates.

Despite the shallower coalescence zone and reduction in graupel aloft in the examples from 14 August, the potential for higher elevations to receive brief periods of intense rainfall was also prevalent. One particularly intense example occurred on 20 July 2004 around 2015 UTC, when the cell of interest developed over the higher terrain above 2 km and gradually moved westward toward the slopes prior to dissipation. A vertical cross section through the cell at this time highlights deep convection with radar echo extending to 17 km (compared to the median height of 13 km; Table 3.1) and the 30-dBZ reflectivity contour approaching 15 km (Fig. 3.10). Values of $Z_{DR} > 2$ dB with K_{DP} exceeding 1° km^{-1} indicated heavy rainfall due to large, oblate hydrometeors; maximum rain rates for this cell were 80 mm h^{-1} , compared to the median of 29 mm h^{-1} for cells

over the SMO (Table 3.1). The reflectivity CFAD from 2015 UTC (Fig. 3.11a) shows less diagonalization compared to the previous SMO cases, suggesting a greater role of accretion similar to the lower-elevation examples. Fifteen minutes later (Fig. 3.11b), however, an increase in diagonalization occurs, beginning with a substantial increase in reflectivity aloft. Similar to the analysis by Zeng et al. (2001), this top-down movement of diagonalization suggests that accretional processes weakened first in the upper levels of the storm where available cloud water was likely depleted more quickly at these colder temperatures compared to lower levels of the cell.

The only considerable difference between this intense SMO cell and the strong cells described over the lower-elevation groups was the reduction in warm-cloud depth. Similar results were described in the Tibetan Plateau region, where convection tended to be maximized upstream of the Himalayas, yet deep convection, with 40-dBZ echo extending above 10 km, was observed over the high terrain of the Plateau despite reduced depth below the melting level (e.g., Houze et al. 2007). This NAME example also coincided with the study by Nesbitt et al. (2008), who also suggested a potentially important role for ice-based microphysical processes over the high terrain.

The examples described above, encompassing a range of topography, highlighted a dependence on mixed-phase microphysical processes in intense cells over all elevations. The previously noted reduction in warm-cloud depth and shallower convection discussed in earlier NAME studies of SMO convection was also evident in these examples, indicating a reduced role of collision-coalescence and resultant smaller rain rates compared to over the low terrain. However, an intense cell over the high terrain exhibited characteristics similar to those over the coastal plain, with deep cores and

polarimetric signatures indicative of large ice hydrometeors above the melting level. In order to place these intense examples within the larger context of all isolated cells from the NAME microphysical scans, a series of statistics based on the polarimetric radar data were computed for the elevation groups, which are described in the following chapter.

Table 3.1. Median values of specific variables for all isolated cells in a particular elevation group.

Elevation subset	# of unique tracks	Median maximum rain rate (mm h ⁻¹)	Median lifetime (min)	Median water mass (g m ⁻³)	Median ice mass (g m ⁻³)	Median maximum Z (dBZ)	Median echo top height (km)
0–1 km	488	35.7	29	0.33	0.05	47	12.5
1–2 km	366	33.2	30	0.30	0.06	45.5	13
>2 km	377	32.4	29	0.27	0.10	44	12.5

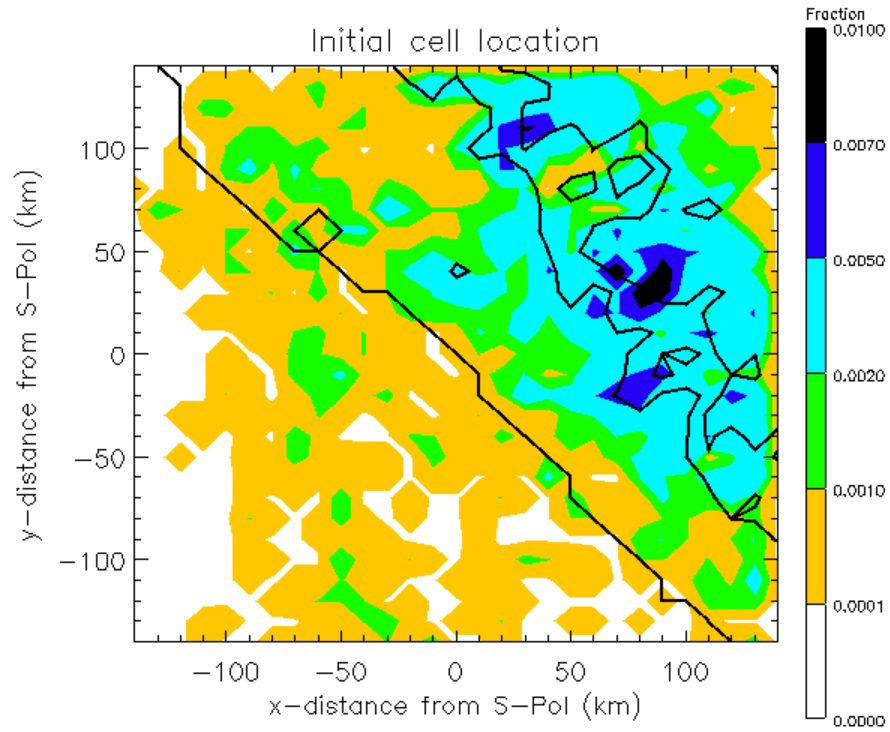


Fig. 3.1. Frequency of occurrence (color-filled contours) of the initial location of isolated cells at each grid point. Frequencies are normalized by the total number of isolated cells in the domain to obtain a fraction. Bold, black contours represent the topography contoured at 0, 1, 2, and 3 km.

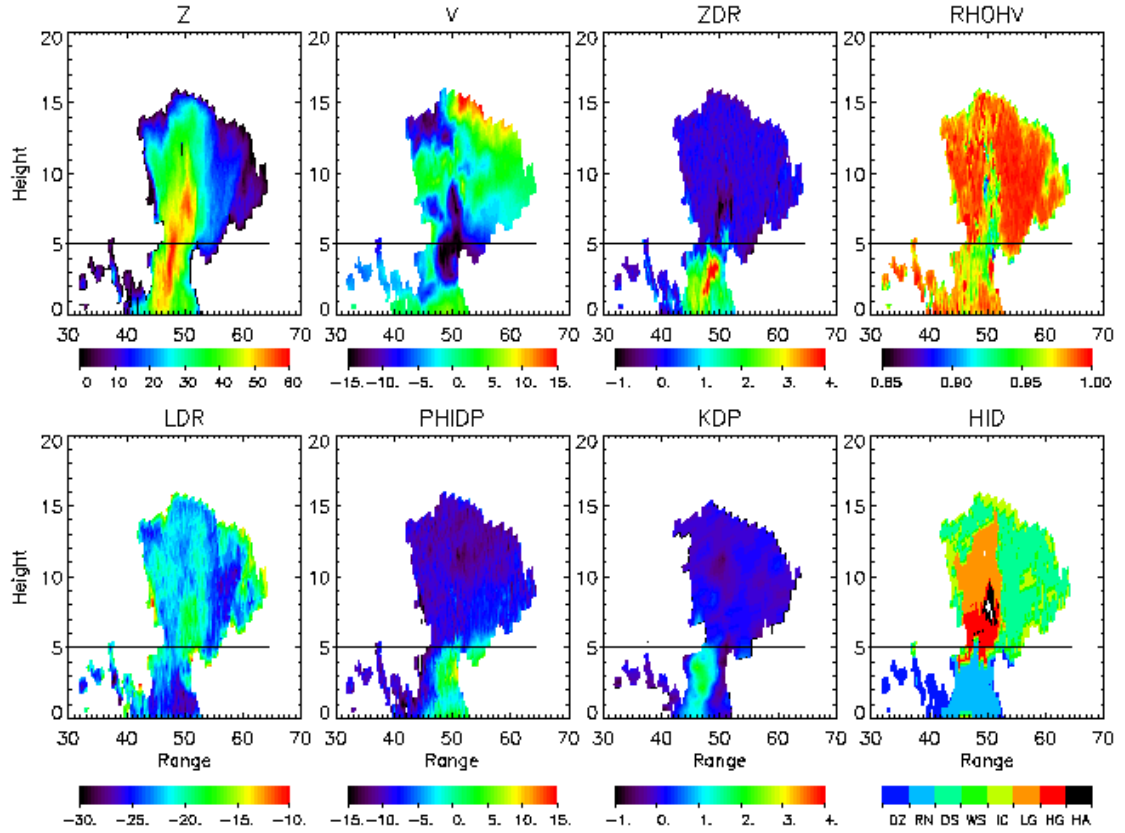


Fig. 3.2. An RHI through an isolated cell at 2159 UTC 10 July 2004 at a 118° azimuth angle. Variables from left to right, starting with the upper panel, include reflectivity (Z , dBZ), radial velocity (V_R , m s^{-1}), differential reflectivity (Z_{DR} , dB), correlation coefficient (ρ_{HV}), linear depolarization ratio (L_{DR} , dB), differential phase (Φ_{DP} , $^\circ$), specific differential phase (K_{DP} , $^\circ \text{ km}^{-1}$), and the hydrometeor identification (HID). The HID abbreviations correspond to the following: drizzle (DZ), rain (RN), dry snow (DS), wet snow (WS), ice (IC), low-density graupel (LG), high-density graupel (HG), and hail (HA). The solid black line at 5 km indicates the average height of the melting level. Height is in km, and range is distance in km from S-Pol.

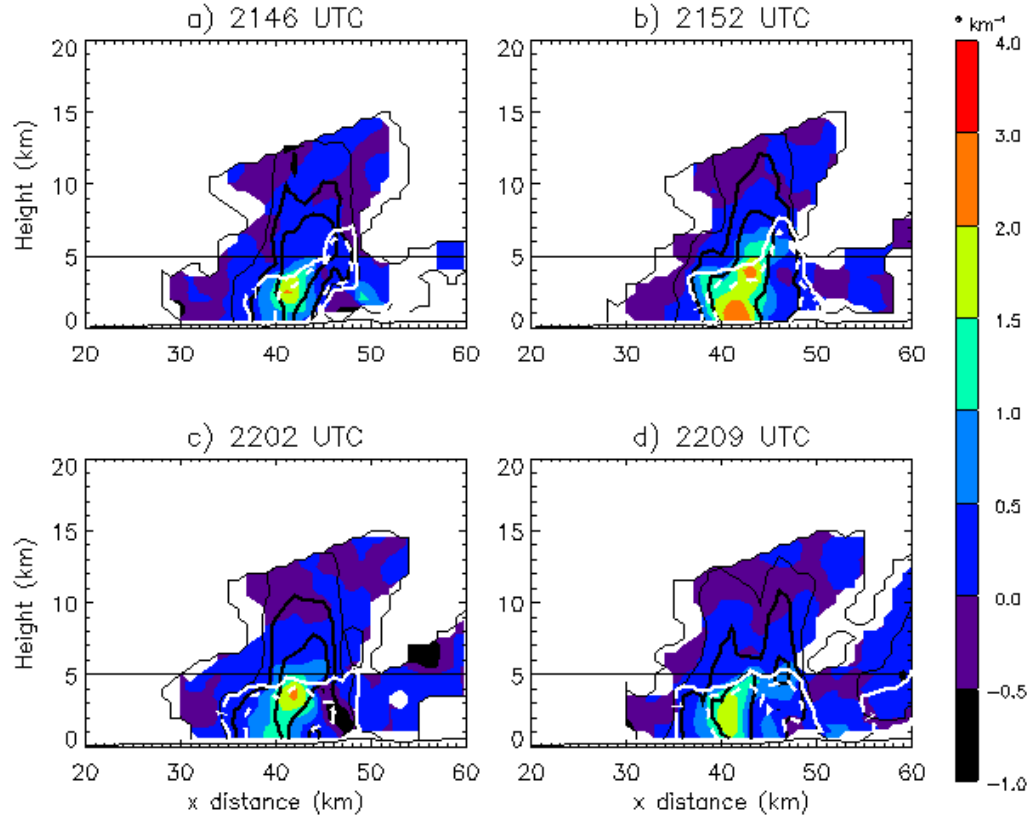


Fig. 3.3. Vertical cross sections through the point of maximum reflectivity of a cell occurring on 10 July 2004 at (a) 2146, (b) 2152, (c) 2202, and (d) 2209 UTC. Color-filled contours are associated with K_{DP} ($^{\circ} \text{km}^{-1}$). Black contours represent reflectivity at 0, 30, 40, and 50 dBZ, with the 40- and 50-dBZ contours thickened. Values of Z_{DR} are contoured in white for 1 dB (solid) and 2 dB (dashed). Terrain height is plotted at the bottom as a black contour, and the solid black line at 5 km represents the average height of the melting level during NAME. The x-axis is the distance from S-Pol in km.

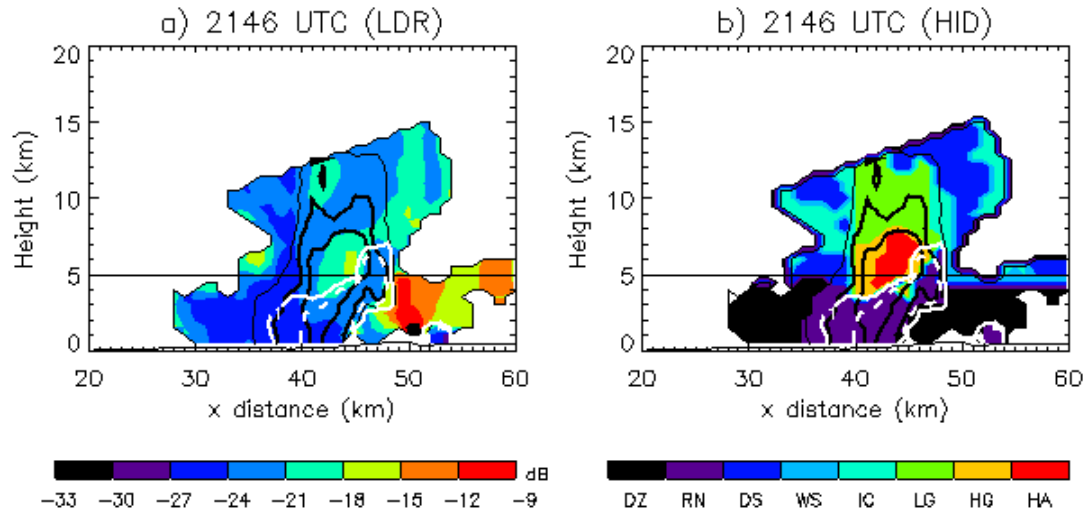


Fig. 3.4. Vertical cross sections through the point of maximum reflectivity of a cell occurring on 10 July 2004 at 2146 UTC. Color-filled contours represent (a) L_{DR} (dB) and (b) HID. The HID abbreviations are the same as in Fig. 3.2, and the remaining contours are described in Fig. 3.3.

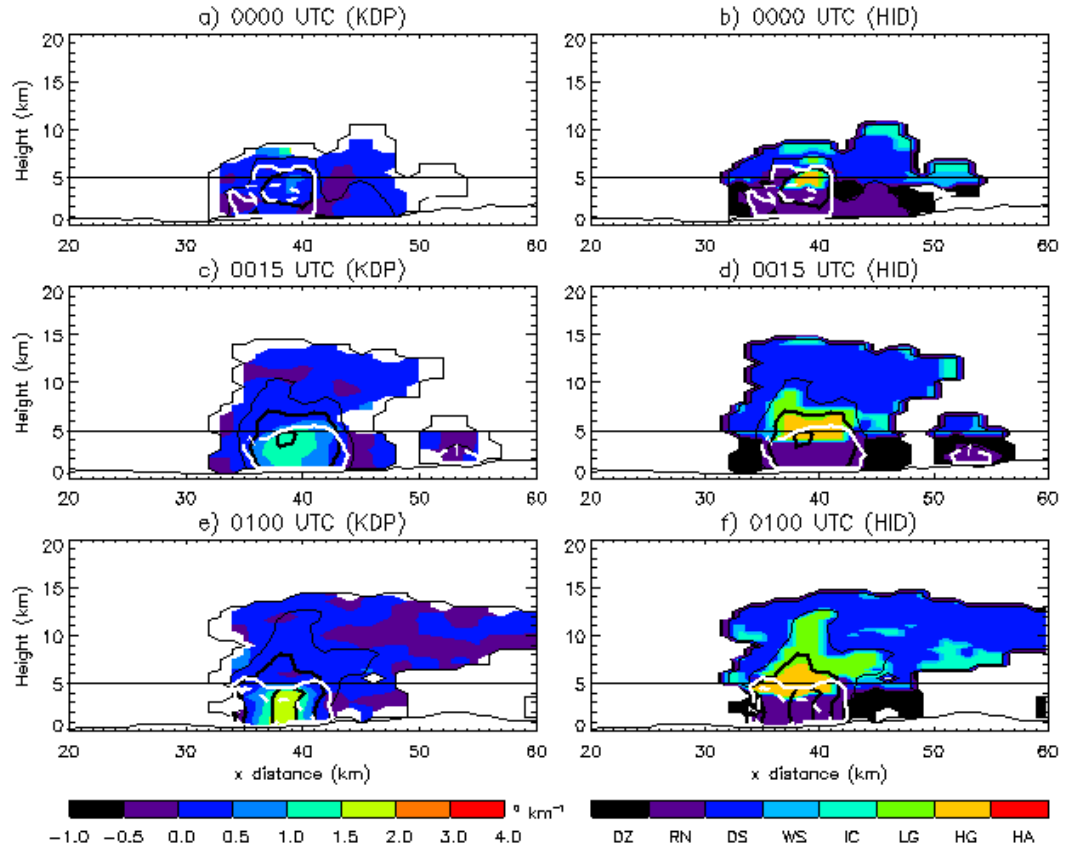


Fig. 3.5. Vertical cross sections through the maximum reflectivity of a cell at (a),(b) 0000, (c),(d) 0015, and (e),(f) 0100 UTC on 19 August 2004. Color-filled contours represent K_{DP} (left; $^{\circ} \text{ km}^{-1}$) and HID (right), with HID abbreviations explained in Fig. 3.2 and the remaining contours described in Fig. 3.3.

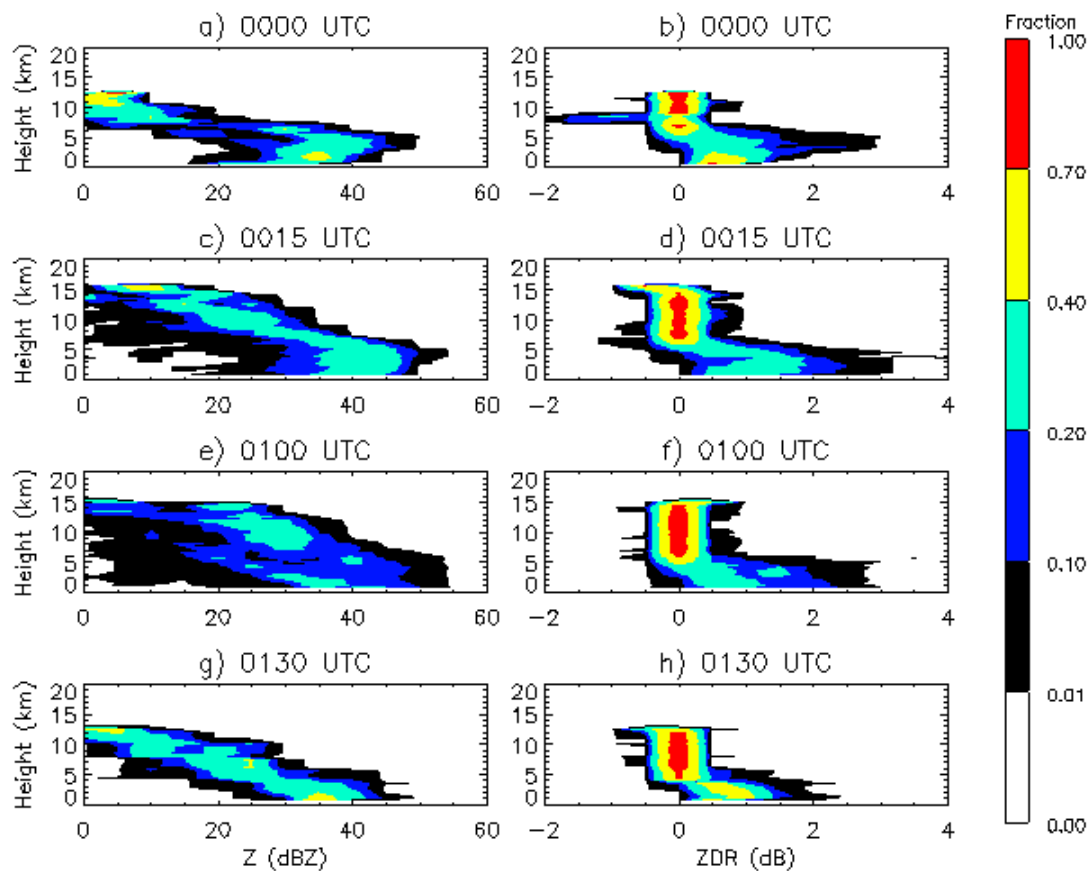


Fig. 3.6. Contoured frequency by altitude diagrams (CFADs) of reflectivity (left) and differential reflectivity (right) associated with a cell on 19 August 2004 at (a),(b) 0000, (c),(d) 0015, (e),(f) 0100, and (g),(h) 0130 UTC. Frequencies are normalized by the total frequency in each height bin.

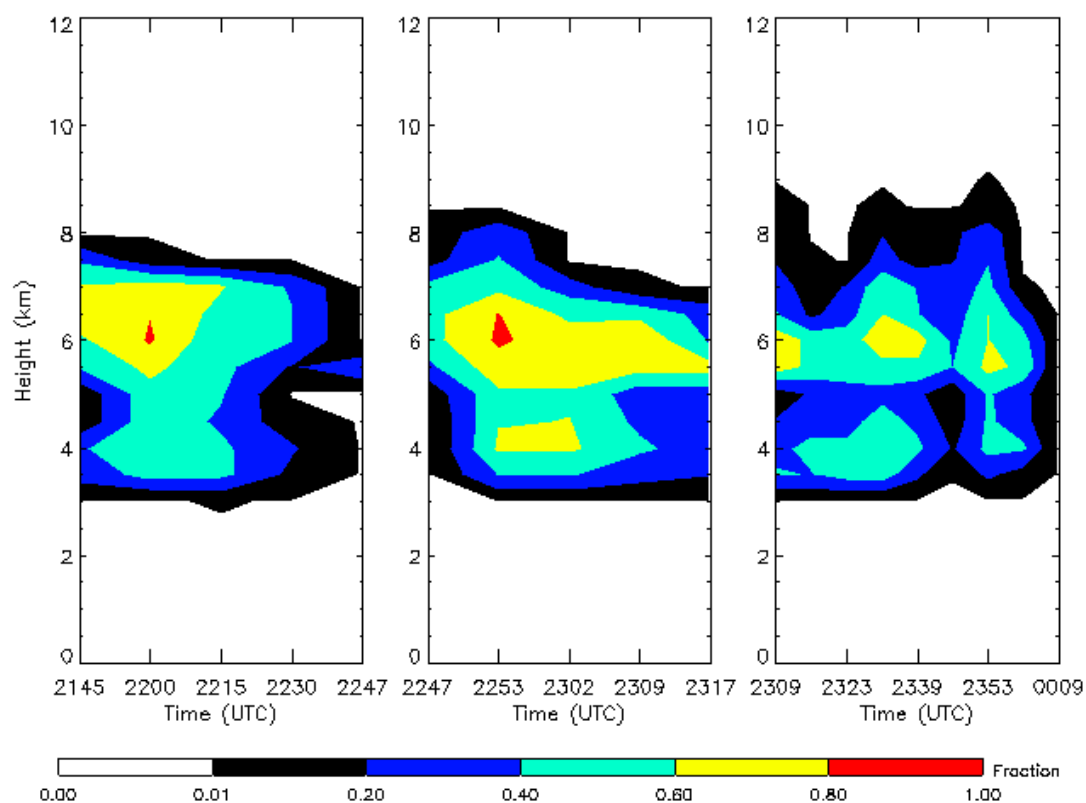


Fig. 3.7. Frequency contours of the fraction of grid points at a particular height where the HID was identified as graupel during the evolution of three isolated cells over the SMO on 14 August 2004.

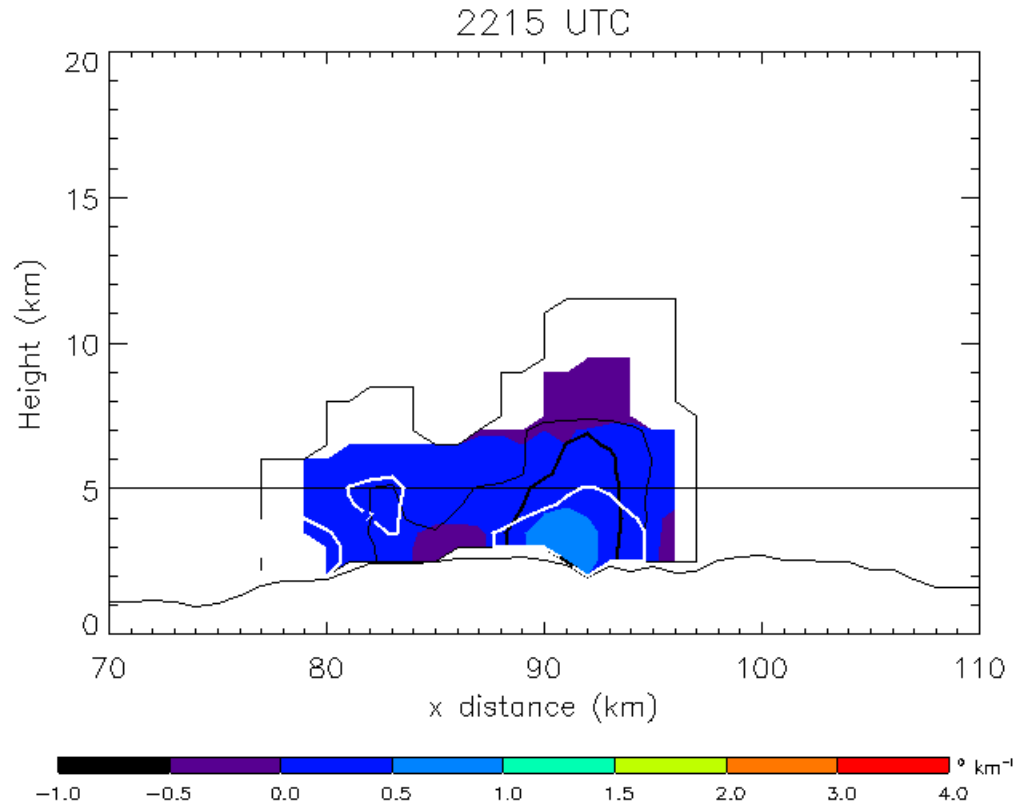


Fig. 3.8. Vertical cross section through the maximum reflectivity of a cell occurring on 14 August 2004 at 2215 UTC. Color-filled contours represent K_{DP} ($^{\circ} \text{ km}^{-1}$), and the other contours are as in Fig. 3.3.

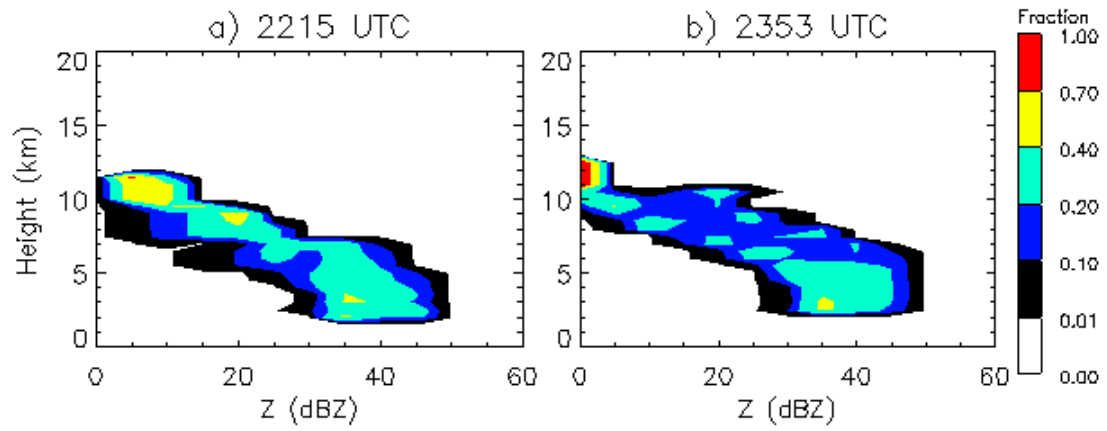


Fig. 3.9. Reflectivity CFADs for the mature stages of two separate isolated cells on 14 August 2004 occurring at (a) 2215 and (b) 2353 UTC. Frequencies are contoured as in Fig. 3.6.

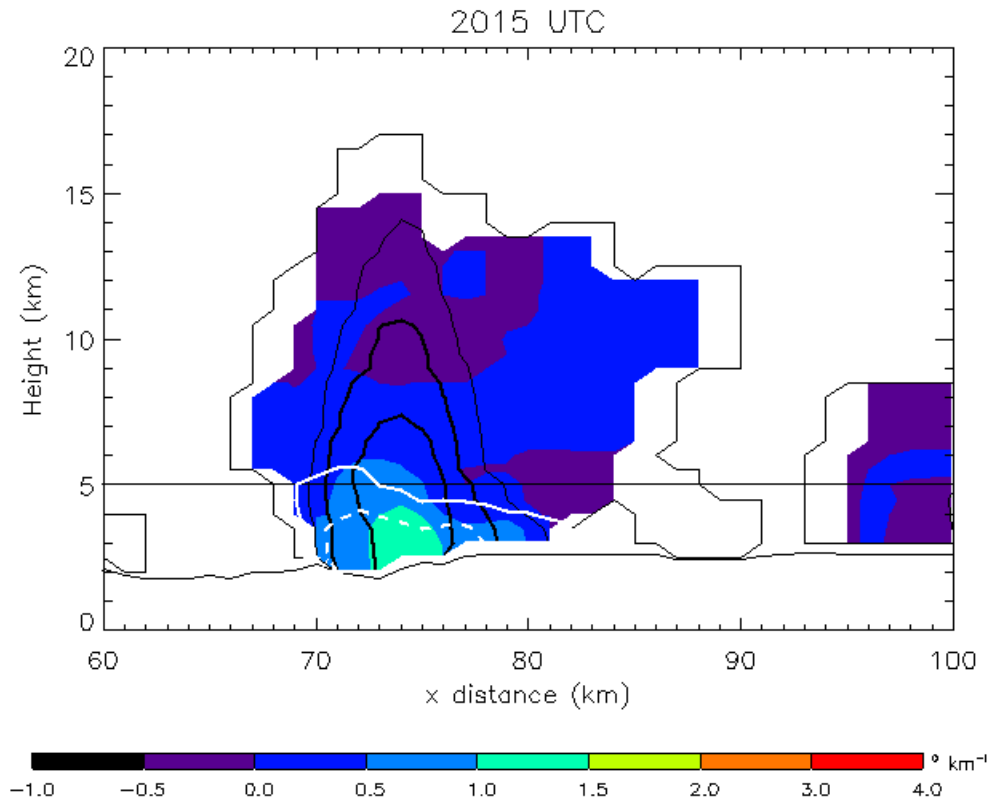


Fig. 3.10. Vertical cross section through the maximum reflectivity of a cell occurring on 20 July 2004 at 2015 UTC over the SMO. Color-filled contours represent K_{DP} ($^{\circ} \text{ km}^{-1}$), and the other contours are as in Fig. 3.3.

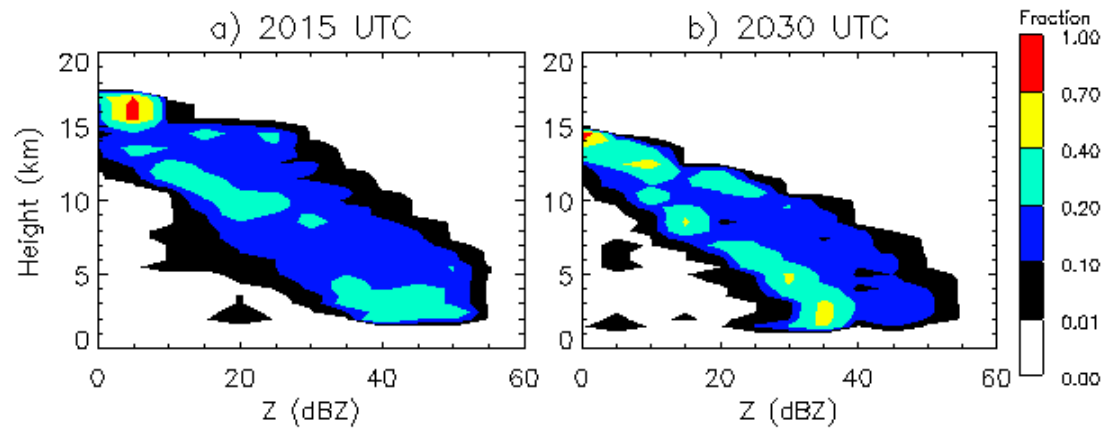


Fig. 3.11. Reflectivity CFADs for the mature stage of an isolated cell on 20 July 2004 at (a) 2015 and (b) 2030 UTC. Frequencies are contoured as in Fig. 3.6.

CHAPTER 4

GENERAL STATISTICS

4.1 Rain rate and echo-top heights

Previous results from NAME (e.g., Gochis et al. 2006, 2007; Lang et al. 2007; Nesbitt et al. 2008; Rowe et al. 2008) emphasized a peak in rainfall intensity along the coastal plain and western slopes of the SMO, hypothesized to be linked to increased warm-cloud depth (Rowe et al. 2008), deeper convection (Nesbitt et al. 2008), and increased ice mass aloft (Lang et al. 2010) over the lower terrain. However, based on comparisons between 24-h and 15-min rainfall totals, Rowe et al. (2008) suggested that rainfall may be intense both over the coast and the higher terrain of the SMO, with a longer duration of heavy precipitation over the lower elevations. The potential for the high terrain to receive brief periods of intense rainfall was also suggested from rain gauge observations (Gochis et al. 2007) and was observed in an SMO case described in Ch. 3. Cumulative distribution functions (CDFs) of maximum instantaneous rain rate as a function of maximum terrain height are shown in Fig. 4.1 for all isolated cells identified in our study. There was a tendency for the most frequent occurrence of intense rain rates ($> 50 \text{ mm h}^{-1}$) to occur within cells over the slopes (1–2 km), consistent with previous results; however, there was considerably less separation between elevation groups compared to the distributions that included all precipitating features (i.e., also including mesoscale systems and organized convection) during NAME, shown in Fig. 2 of Rowe et

al. (2008). This observation agrees with the hypothesis suggested from the case studies that isolated convection during NAME had the potential to produce brief periods of intense rainfall regardless of the underlying topography.

Previous NAME studies (e.g., Nesbitt et al. 2008; Rowe et al. 2008) suggested a relationship between reduced rainfall intensity and lower echo-top heights over the SMO for all precipitating features over the high terrain. However, when considering only isolated convection, examples of intense cells (Ch. 3), as well as general rainfall statistics (Fig. 4.1), showed that intense instantaneous rain rates also occurred over the higher elevations. The next step was to relate echo-top heights of each tracked isolated cell to those rainfall statistics and investigate any potential differences, or lack thereof, as a function of topography. First, Fig. 4.2 shows CDFs of maximum rain rates grouped by echo-top height. From this figure, it is evident that the most intense rain rates were associated with the deepest convection (i.e., echo-top heights > 11 km). The greater rain rates associated with shallow convection (0–5 km), compared to cells with echo tops in the range of 6–11 km, reflect the rain-out stage of cells and comprise only a small percentage of total cells ($< 1\%$).

The presence of deep convection over the SMO, which subsequently produced intense rainfall, was noted in Ch. 3. To investigate the frequency of occurrence of these intense cells, probability distribution functions of echo-top heights as a function of elevation were computed (Fig. 4.3). This figure shows that the most frequent echo-top height observed for isolated cells over the SMO was 11 km, whereas the greatest percentage of low-terrain isolated cells extended to 14 km, suggesting a trend of shallower convection over the higher terrain that is consistent with previous results (e.g.,

Nesbitt et al. 2008; Rowe et al. 2008). Nesbitt et al. (2008) suggested that echo-top height over the high terrain was limited by the lack of moisture available at cloud base compared to the lower elevations in proximity to the GoC. However, despite the average tendency for taller echo tops over the low terrain, it is apparent that deep convection also occurred over the SMO.

4.2 Polarimetric observations

Consistent with the presence of graupel in the examples of convection both over the coastal plain and the SMO (Ch. 3), results from Nesbitt et al. (2008) suggested that convection contained significant mixed-phase processes. Petersen et al. (1996) noted a relationship between the extension of the 30-dBZ contour above the melting level and lightning production, providing a minimum condition for the possibility of lightning. However, a number of studies have linked a higher probability of lightning production in cells to the presence of 40-dBZ echo at or above the altitude of -10°C (Laksen and Stanbury 1974; Dye et al. 1989; Buechler and Goodman 1990; Michimoto 1991, 1993; Gremillion and Orville 1999; Vincent et al. 2003). During NAME, the -10°C isotherm typically fell between 6.5 and 7 km; it is therefore assumed that lightning production, and therefore inferred large supercooled and ice water contents, was associated with the extension of the 40-dBZ contour to at least 7 km. The CDFs of the maximum height of the 40-dBZ contour (Fig. 4.4) show that roughly 15% of cells over the lower terrain reached this threshold and more than 20% of cells over the SMO had a high probability of lightning production. The larger percentage of isolated cells over the SMO with 40-dBZ reflectivities extending to 7 km compared to over lower elevations is likely a

reflection of the greater cloud-base heights over the higher elevations, although the differences between terrain bands were not significantly large. This figure does, however, highlight the likelihood of lightning production over all elevation groups.

Although mixed-phase processes were inferred in precipitating features over all elevations, Nesbitt et al. (2008) speculated that the generally shallower convection over the SMO produced a smaller depth of precipitation-sized ice particles, which was seen in several examples of average SMO cells in Ch. 3. In addition to reduced precipitation-ice mass over the SMO, Lang et al. (2010) also found a reduction in liquid water mass compared to the lower terrain for all precipitating features, consistent with the shallower warm-cloud depth over the SMO (Rowe et al. 2008). To investigate differences in the relative amounts of precipitation liquid and precipitation ice for isolated convection during the microphysical scans, ice and water masses were summed over all grid points within each identified isolated cell then normalized by the volume of the cell. Cumulative distribution functions of these variables are shown in Fig. 4.5, which highlight increased ice mass associated with cells over the SMO (Fig. 4.5b), and a greater amount of water mass contained in cells over the lower terrain (Fig. 4.5a). The increased water mass in cells over lower elevations is consistent with results from Lang et al. (2010) and suggests a greater role of warm-rain processes. The importance of collision-coalescence is also evident in the CDFs of water mass fraction (Fig. 4.6), defined as the total precipitation water mass in a cell divided by the total accumulated water and ice mass, which highlight the greater precipitation liquid water fractions over the lower terrain for isolated cells. The reduced amount and depth of rain water in isolated convection over the SMO likely resulted from smaller warm-cloud depths over the high terrain. Despite this reduced

amount of moisture flux at cloud base for cells over the SMO, due to the higher LCL, peak ice mass, which occurred just above the melting level for all elevation groups, was greater compared to cells over lower terrain (Fig. 4.7). Previous studies, such as Rosenfeld and Woodley (2003), have noted that a smaller warm-cloud depth, implied by a higher cloud-base height, reduces the depth through which warm-rain processes can occur, leaving greater relative supercooled liquid water available to the mixed-phase zone. Therefore, it is likely that the shallower warm-cloud depth in cells over the SMO limited the collision-coalescence process, allowing for a greater percentage of available moisture to be lofted above the melting level, leading to increased ice mass in convection over the higher terrain.

In agreement with these results, Williams et al. (2007), using data from a nearby profiler (see Fig. 2.1), suggested the importance of warm-rain processes at low elevations through evidence of broadening of the drop-size distribution (DSD) below the melting level. In addition, Lang et al. (2010) noted larger drop sizes over the coastal plain relative to over the SMO. To investigate this further for isolated cells, probability distributions of maximum Z_{DR} were created for portions of cells identified as rain (Fig. 4.8a). This figure not only shows a tendency for Z_{DR} to be greater in cells over the low terrain, but the distribution was also wider compared to the higher elevations, suggesting a broader DSD and larger drop sizes over the coastal plain, consistent with the previously described studies. The probability distribution for maximum K_{DP} associated with liquid water in a cell (Fig. 4.8b) shows a tendency for K_{DP} to be slightly greater over the lower terrain as well, in agreement with greater rain rates over these elevations. This trend likely reflected the larger drop sizes observed over the low terrain (Fig. 4.8a) due to melting of ice

hydrometeors and the increased precipitation water mass due to increased warm-cloud depth associated with these cells (Fig. 4.5). Both processes would tend to increase K_{DP} .

Contoured frequency diagrams of Z_{DR} and K_{DP} for rain-only points within all isolated cells (Fig. 4.9) were constructed for each elevation group to further investigate possible differences in precipitation processes. Contoured are the logarithm of normalized frequencies for the elevation group to emphasize the differences in the extremes; normalization here is in regards to total frequency. From these figures, it is evident that, in addition to the high K_{DP} values associated with large Z_{DR} , K_{DP} tends to be slightly higher over the lower elevations for lower Z_{DR} (between 1 and 2 dB; Fig. 4.9a). This suggests a significant contribution to K_{DP} , and therefore rain rates, from higher water contents of smaller drops in cells over the low terrain, and further emphasizes the importance of warm-rain processes compared to over the SMO. In addition, case studies in Ch. 3 highlighted the presence of Z_{DR} columns in cells over the low terrain, implying the lofting of liquid water above the melting level, which aided in the growth of large ice hydrometeors. Figure 4.10 shows CDFs of positive Z_{DR} for grid points located above the melting level, highlighting a slightly greater fraction of positive Z_{DR} at sub-freezing temperatures for cells over the low terrain, consistent with the more frequent observations of Z_{DR} columns compared to over the SMO in the case studies. The presence of positive Z_{DR} above the melting level over the SMO, however, is also evident in this figure, suggesting that although less common, isolated convection over the higher terrain produced significant ice mass due to riming, which subsequently melted, producing intense rain rates. This differs from the lower terrain where, in addition to ice-based

processes, warm-rain microphysical processes were important for producing copious amounts of rainfall through the addition of higher rain-water contents.

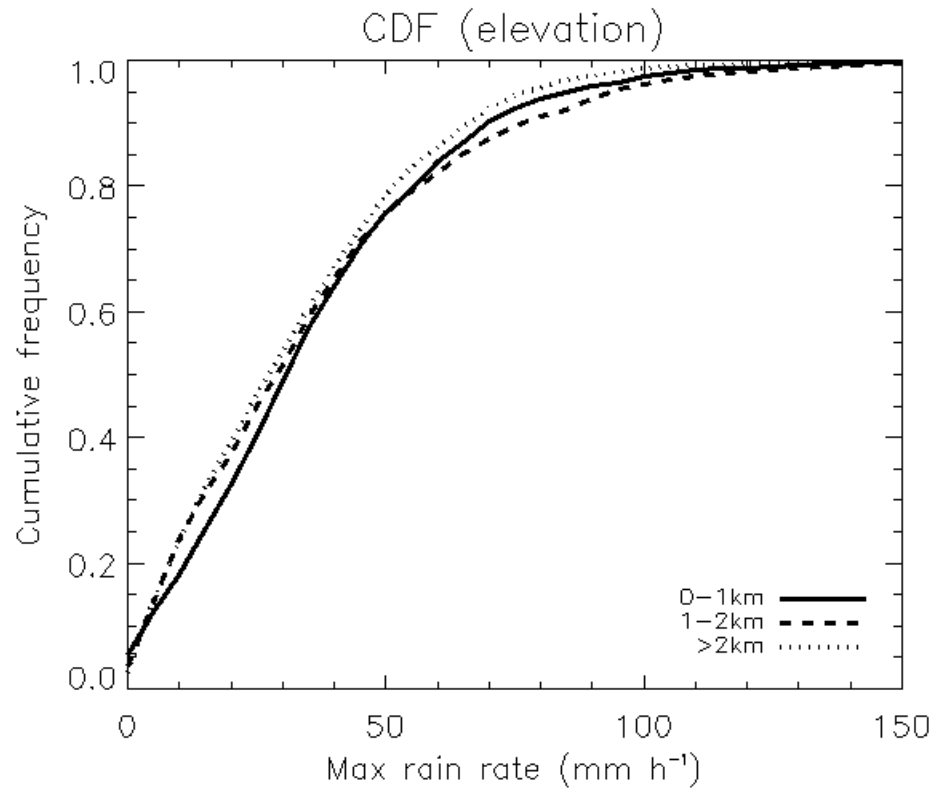


Fig. 4.1. Cumulative distribution functions of maximum rain rate (mm h⁻¹) for all isolated cells partitioned by elevation groups.

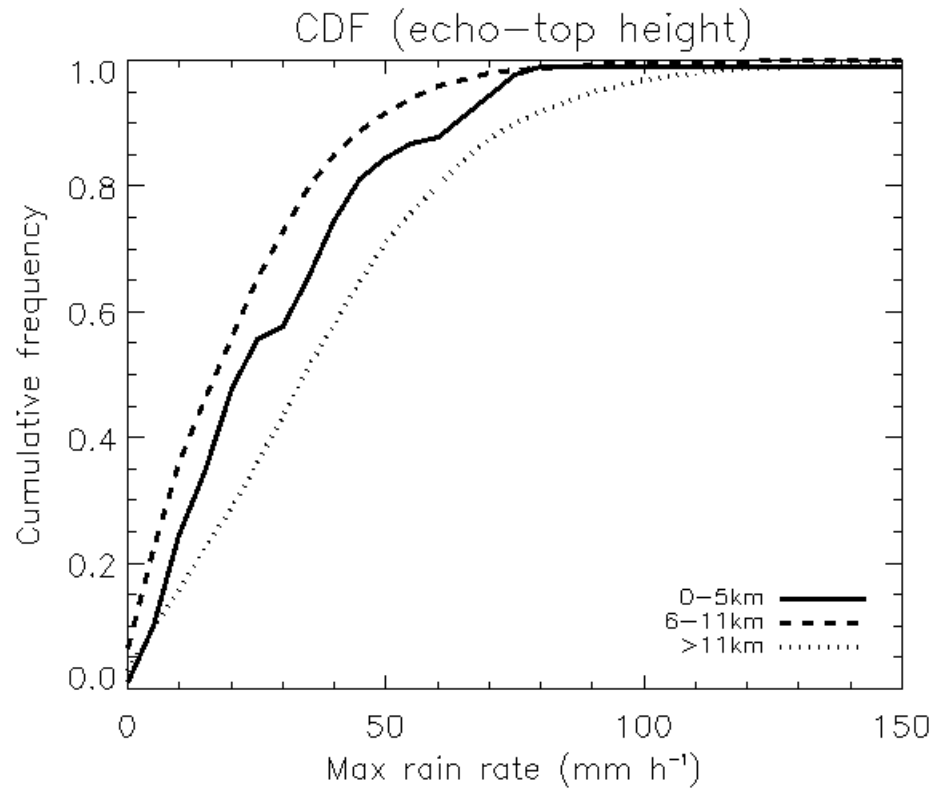


Fig. 4.2. Cumulative distribution functions of maximum rain rate (mm h⁻¹) for all isolated cells partitioned by echo-top height.

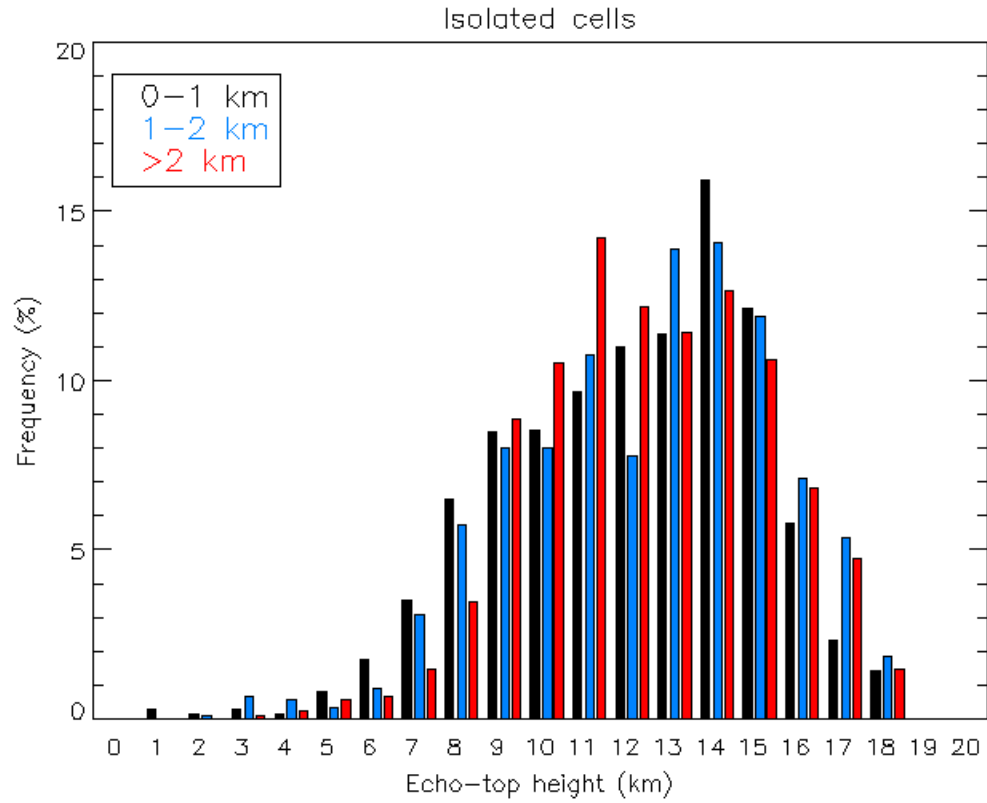


Fig. 4.3. Histogram of echo-top heights (km) for each isolated cell as a function of terrain height (km). Frequency represents the percentage of cells with the particular echo-top height for each elevation group.

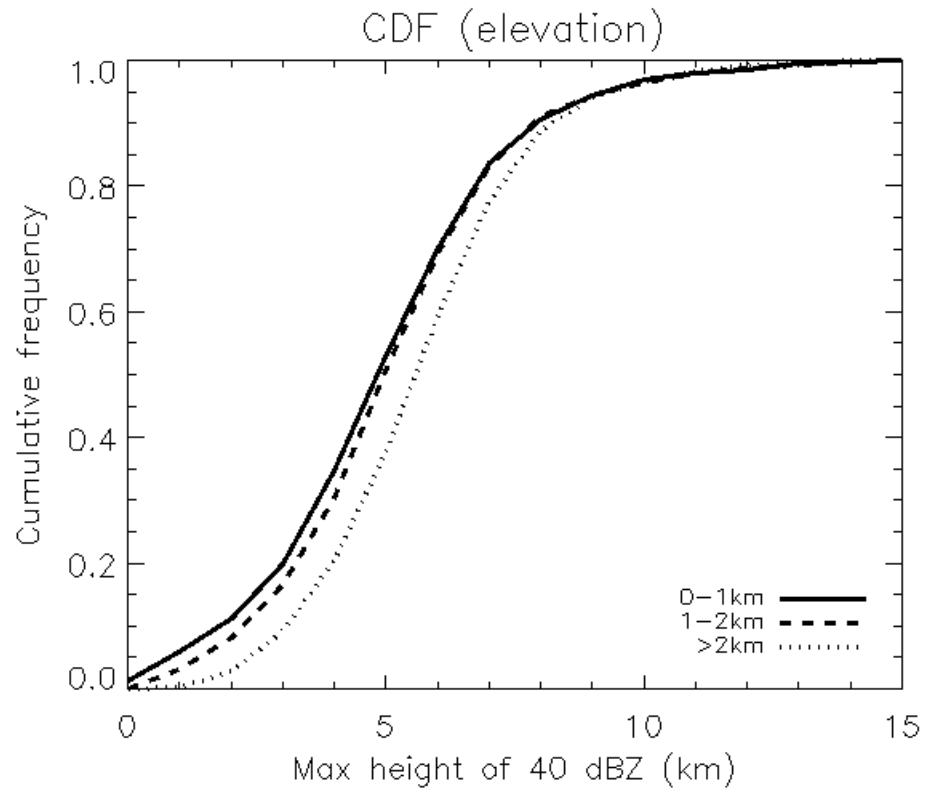


Fig. 4.4. Cumulative distribution functions of the maximum height of 40-dBZ reflectivity in isolated cells as a function of terrain height (km).

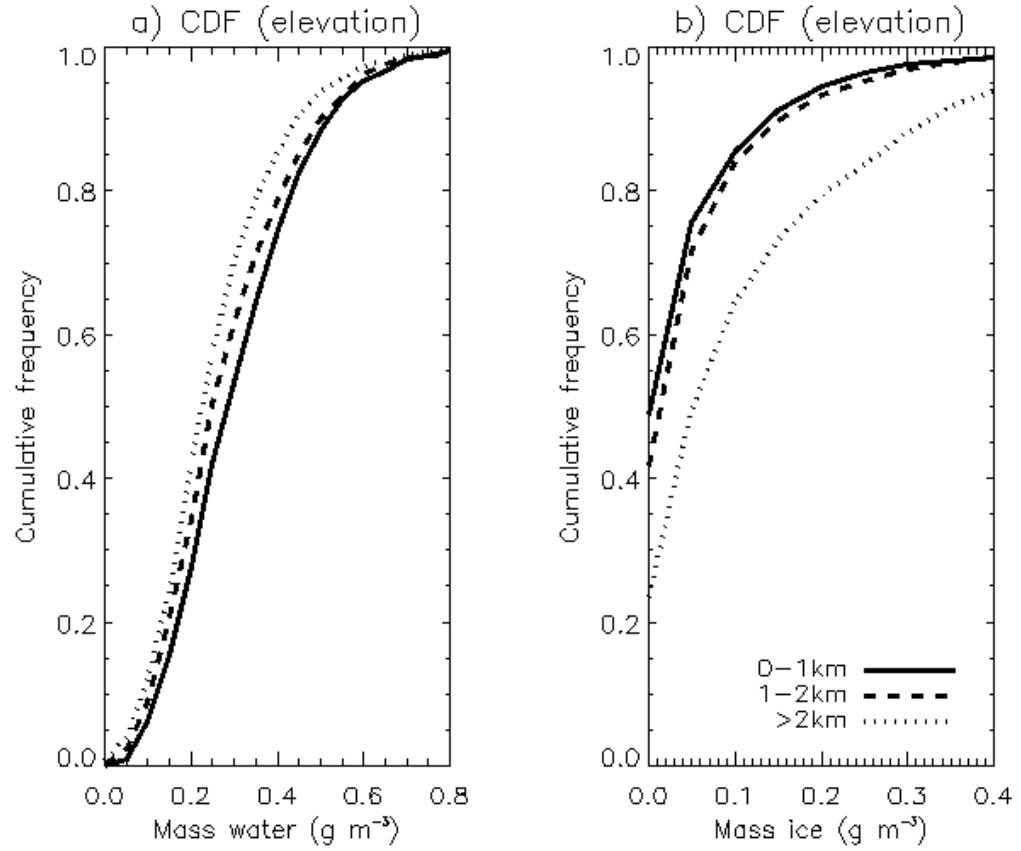


Fig. 4.5. Cumulative distribution functions of (a) liquid water mass and (b) ice mass as a function of terrain height (km). Masses (g m^{-3}) are normalized by the total number of points in each isolated cell.

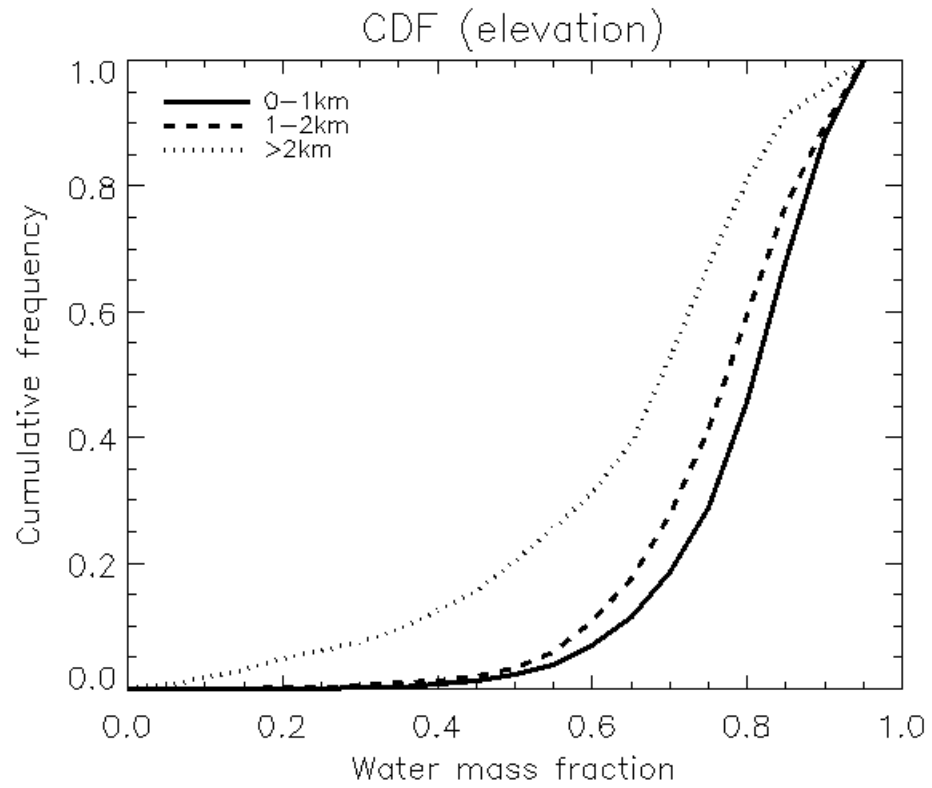


Fig. 4.6. Cumulative distribution functions of water mass fraction as a function of terrain height (km).

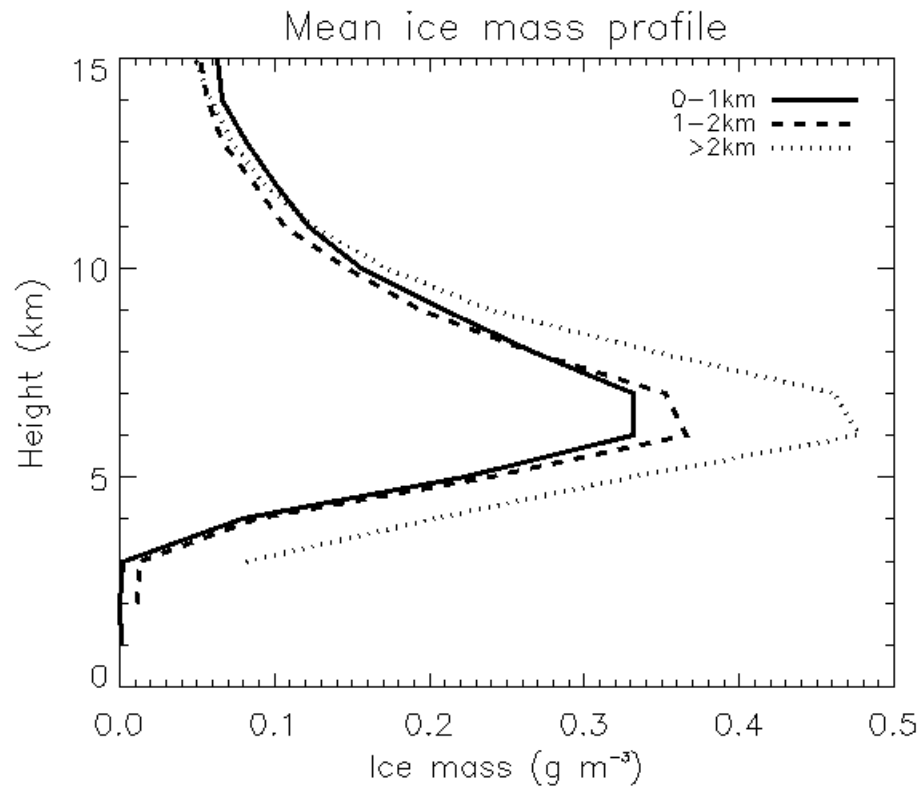


Fig. 4.7. Vertical profiles of mean ice mass as a function of terrain height (km).

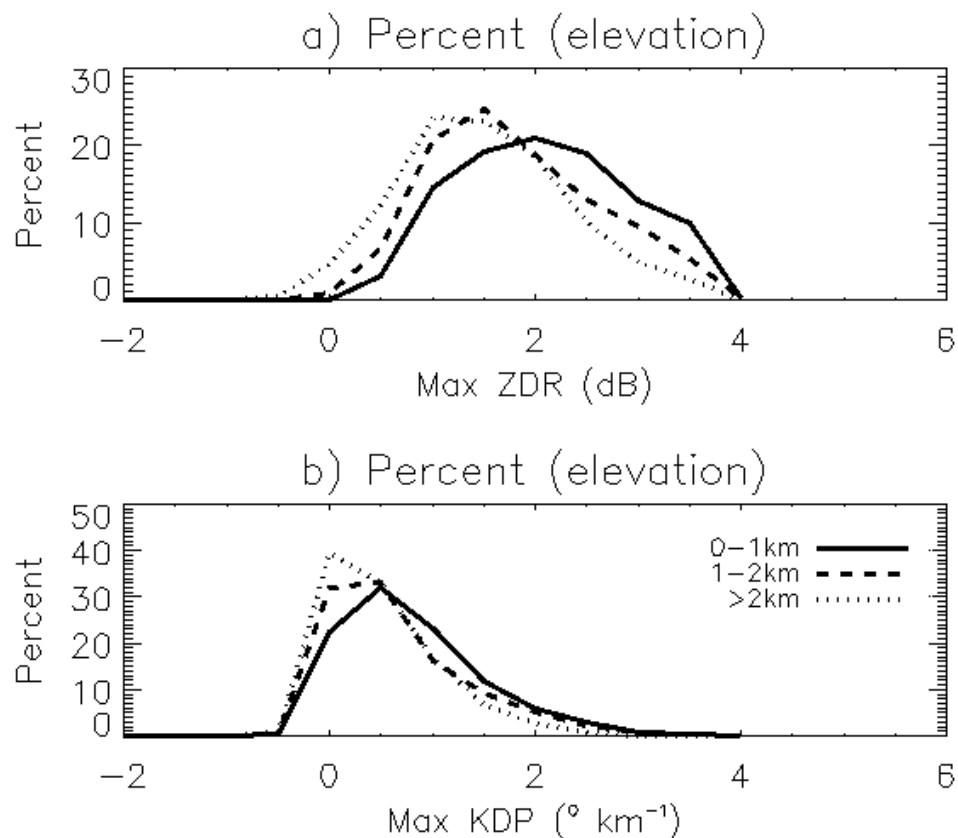


Fig. 4.8. Probability distribution functions of maximum (a) Z_{DR} (dB) and (b) K_{DP} ($^{\circ} \text{ km}^{-1}$) in isolated cells as a function of terrain height (km). Only grid points in cells identified by the HID as rain below 4 km are included to eliminate contamination by the melting level. Percent represents the percentage for the specified elevation group.

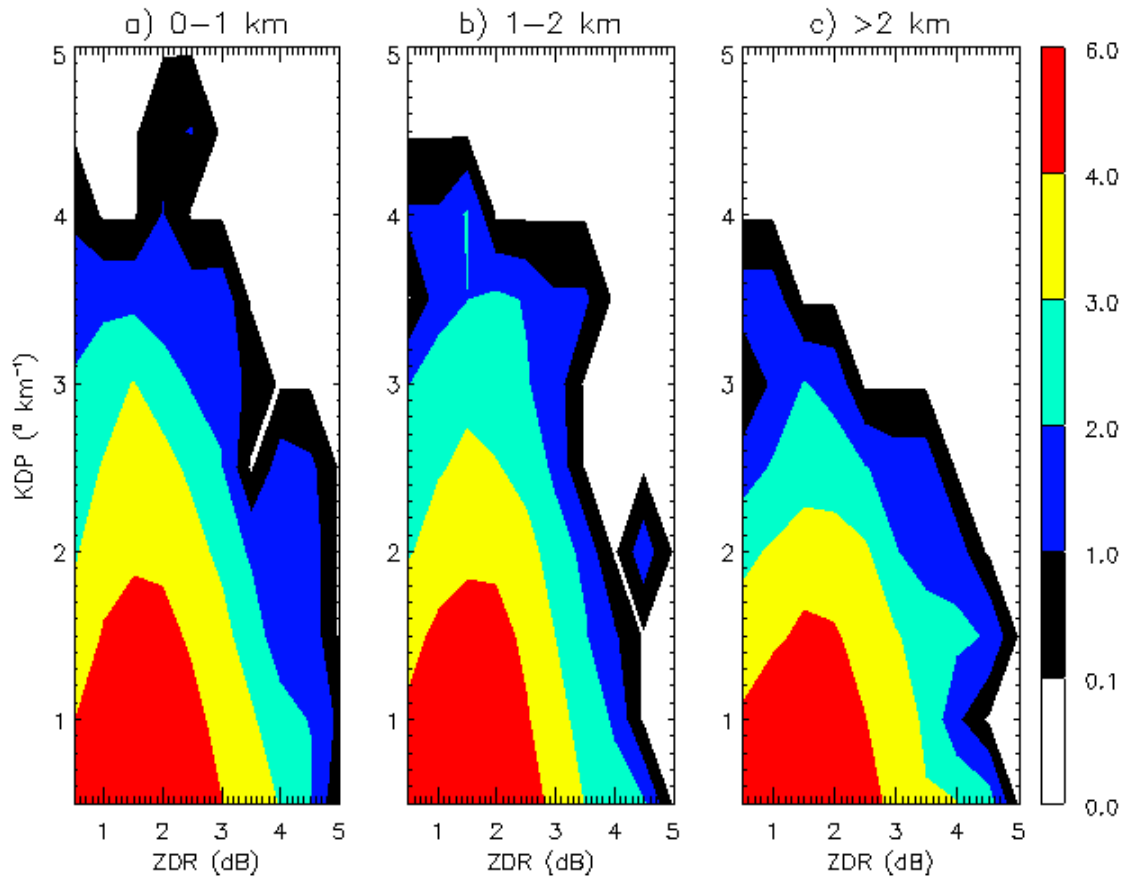


Fig. 4.9. Contoured frequency diagrams of Z_{DR} (dB) and K_{DP} (° km⁻¹) for points in isolated cells identified as rain below 4 km. Plotted are the logarithms of frequencies normalized by the total frequency for each elevation group.

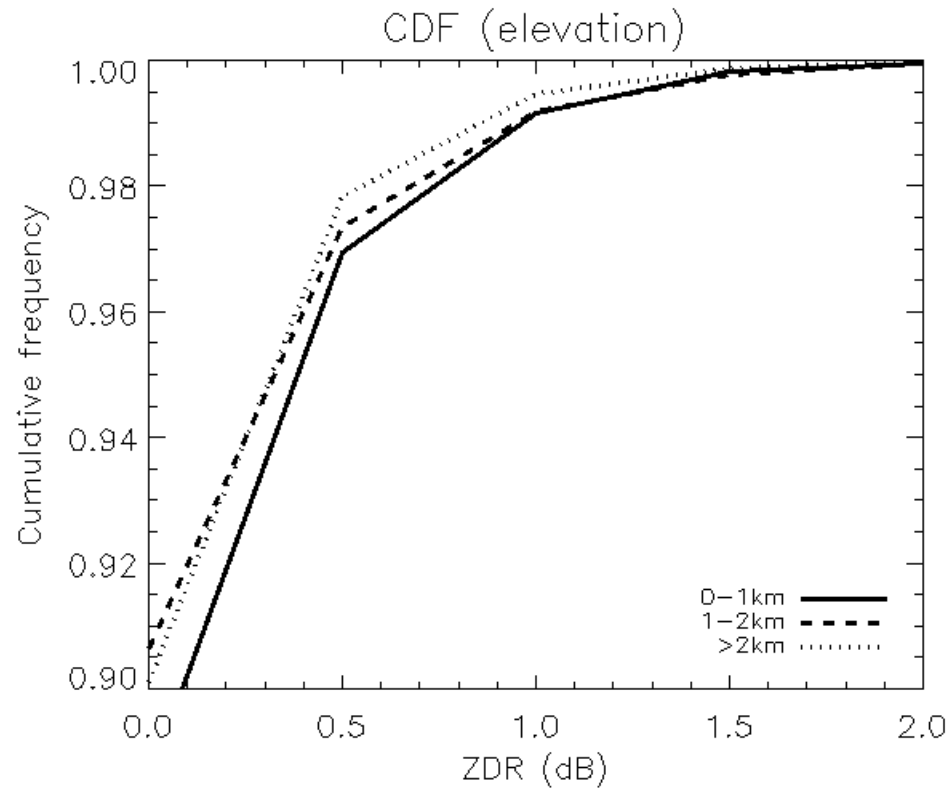


Fig. 4.10. Cumulative distribution functions of positive Z_{DR} (dB) identified above the melting level in isolated cells as a function of terrain height (km).

CHAPTER 5

DISCUSSION AND CONCLUSIONS

To address the difficulty of predicting warm-season rainfall associated with the North American monsoon, a comprehensive network of instrumentation was deployed in northwestern Mexico during the summer of 2004. The NCAR S-band, polarimetric radar (S-Pol) was used to investigate the location, size, and type of hydrometeors in individual cells during NAME. A cell identification and tracking algorithm was applied to the gridded radar fields during about 95 hours of microphysical scans to locate and track individual convective elements meeting prescribed reflectivity and area thresholds. Additional parameters were computed for the tracked cells, including ice and liquid water masses and near-surface rain rates, to relate the trends in precipitation intensity to hydrometeor characteristics within the cells. Only isolated cells over land were included for this study to begin to describe the elevation-dependent trends observed in this region.

Cases selected over the entire range of topography revealed deep cells and polarimetric signatures comparable to other studies of tropical (e.g., Carey and Rutledge 2000; Cifelli et al. 2002) and midlatitude convection (e.g., Bringi et al. 1996; Tessendorf et al. 2005). An example over the coastal plain highlighted deep, isolated convection with precipitation-sized ice extending, at times, to 15 km. The observance of Z_{DR} columns in these cells indicated the lofting of supercooled water above the melting level, where an enhanced L_{DR} “cap” above the column implied subsequent freezing to produce graupel,

which grew rapidly via riming. This process has also been shown to be an important contributor to precipitation-sized ice in tropical convection (e.g., Carey and Rutledge 2000), as well as intense convection observed in the midlatitudes (e.g., Fulton and Heymsfield 1991, Bringi et al. 1997; Zeng et al. 2001; Tessendorf et al. 2005). Contoured frequency by altitude diagrams (CFADs) of reflectivity for isolated cells over the western slopes of the SMO (1–2 km) also suggested a significant contribution from mixed-phase processes, where the observed lack of diagonalization indicated dominant accretional processes, similar to that observed over the coastal plain. Peaks in reflectivity, Z_{DR} , and K_{DP} below the melting level throughout different stages of the cells' lifetimes highlighted the combined roles of collision-coalescence and melting precipitation-sized ice in these isolated cells, where both warm-rain and ice-based processes were important for producing intense rainfall. In addition, statistics including polarimetric information for isolated cells showed a broader distribution of maximum Z_{DR} in convection over the lower terrain, with a wide range of Z_{DR} also observed for a given K_{DP} value. This further indicated the combined roles of large, melting hydrometeors and increased liquid water due to deeper warm-cloud depths for production of intense rainfall in cells over lower elevations.

A selection of average isolated cells over the SMO showed shallower warm-cloud depths, lower echo-top heights, and a lesser vertical extent of graupel throughout the column, all consistent with previous observations that included all precipitating features during NAME (e.g., Lang et al. 2007, 2010; Nesbitt et al. 2008; Rowe et al. 2008). The absence of Z_{DR} columns and increased diagonalization of reflectivity frequency distributions in these few examples suggested a reduced contribution from riming

processes, in addition to the reduction of collision-coalescence due to shallower warm-cloud depths. Despite the observation of strong upward motion over the SMO, Johnson et al. (2010) noted the absence of a moist column, where the lack of moisture available for latent heat release could likely explain the relative shallowness and reduced dependence on mixed-phase processes for rainfall production at the higher elevations. However, an example shown in Ch. 3, as well as general statistics including all isolated cells during the microphysical scans, supported the idea that short-lived, isolated convection over the SMO could produce intense rainfall despite the general tendency for weaker convection compared to over the lower elevations. One such example described an isolated cell that formed over the high terrain near the western slopes, producing heavy rainfall comparable to the low-terrain cases. Melting of large ice hydrometeors was noted, and an increased role of accretion above the melting level was inferred. The general statistics revealed increased ice mass in these particular cells over the SMO just above the melting level, further highlighting an important role of ice in producing intense rainfall during a number of cases over the higher elevations.

These characteristics associated with intense convection over the SMO were similar to convection occurring over the high terrain of the Tibetan Plateau, where despite the tendency for the most intense convective echo to be observed upstream of the Himalayas, deep, intense convection was occasionally observed over the high terrain of the Plateau, with truncated lower portions relative to cells over the lower elevations (Houze et al. 2007). The Houze et al. (2007) study also found that in some cases in this region, the intense convective echoes organized upscale to produce vertically erect, intense convective echo embedded within MCSs, which rarely occurred over the Plateau.

Previous NAME studies (e.g., Lang et al. 2007) also noted the organization of intense convection over the SMO as cells moved toward the western slopes, similar to the Tibetan Plateau region. This process suggests a likely dependence of microphysical processes on the degree of organization and will be the focus of a future study to further understand the relative roles of warm-rain and ice-based processes in convection during NAME.

REFERENCES

- Adams, D. K., and A. C. Comrie, 1997: The North American monsoon. *Bull. Amer. Meteor. Soc.*, **78**, 2197–2213.
- Badan-Dangon, A., C. E. Dorman, M. A. Merrifield, and C. D. Winant, 1991: The lower atmosphere over the Gulf of California. *J. Geophys. Res.*, **96**, 877–896.
- Berberly, E. H., 2001: Mesoscale moisture analysis of the North American monsoon. *J. Climate*, **14**, 121–137.
- Bluestein, H. B., and M. H. Jain, 1985: Formation of mesoscale lines of precipitation: Severe squall lines in Oklahoma during the spring. *J. Atmos. Sci.*, **42**, 1711–1732.
- Bringi, V. N., and V. Chandrasekar, 2001: *Polarimetric Doppler Weather Radar Principles and Applications*. Cambridge University Press, 636 pp.
- Bringi, V. N., L. Liu, P. C. Kennedy, V. Chandrasekar, and S. A. Rutledge, 1996: Dual multiparameter radar observations of intense convective storms: The 24 June 1992 case study. *Meteor. Atmos. Phys.*, **59**, 3–31.
- Bringi, V. N., K. Knupp, A. Detwiler, L. Liu, I. J. Caylor, and R. A. Black, 1997: Evolution of a Florida thunderstorm during the Convection and Precipitation/Electrification Experiment: The case of 9 August 1991. *Mon. Wea. Rev.*, **125**, 2131–2160.
- Buechler, D. E., and S. J. Goodman, 1990: Echo size and asymmetry: Impact on NEXRAD storm identification. *J. Appl. Meteor.*, **29**, 962–969.
- Carey, L. D., and S. A. Rutledge, 2000: The relationship between precipitation and lightning in tropical island convection: A C-band polarimetric radar study. *Mon. Wea. Rev.*, **128**, 2687–2710.
- Cifelli, R., W. A. Petersen, L. D. Carey, S. A. Rutledge, and M. A. F. Silva-Dias, 2002: Radar observations of the kinematic, microphysical, and precipitation characteristics of two MCSs in TRMM LBA. *J. Geophys. Res.*, **107**, 8077, doi:10.1029/2000JD000264.
- Cotton, W. R., and R. A. Anthes, 1989: *Storm and Cloud Dynamics*. Academic Press, 883 pp.

- Crane, R. K., 1979: Automatic cell detection and tracking. *IEEE Trans. Geosci. Electron.*, **GE-17**, 250–262.
- DeMott, C. A., and S. A. Rutledge, 1998: The vertical structure of TOGA COARE convection. Part I: Radar echo distributions. *J. Atmos. Sci.*, **55**, 2730–2747.
- Dixon, M., and G. Wiener, 1993: TITAN: Thunderstorm Identification, Tracking, Analysis, and Nowcasting—A radar-based methodology. *J. Atmos. Oceanic Technol.*, **10**, 785–797.
- Douglas, M., R. A. Maddox, K. Howard, and S. Reyes, 1993: The Mexican monsoon. *J. Climate*, **6**, 1665–1677.
- Dye, J. E., W. P. Winn, J. J. Jones, and D. W. Breed, 1989: The electrification of New Mexico thunderstorms. 1. Relationship between precipitation development and the onset of electrification. *J. Geophys. Res.*, **94**, 8643–8656.
- Fulton, R., and G. M. Heymsfield, 1991: Microphysical and radiative characteristics of convective clouds during COHMEX. *J. App. Meteor.*, **30**, 98–116.
- Gauthier, M. L., W. A. Petersen, and L. D. Carey, 2010: Cell mergers and their impact on cloud-to-ground lightning over the Houston area. *Atmos. Res.*, **96**, 626–632.
- Gochis, D. J., W. J. Shuttleworth, and Z. L. Yang, 2002: Sensitivity of the modeled North American monsoon regional climate to convective parameterization. *Mon. Wea. Rev.*, **130**, 1282–1298.
- Gochis, D. J., A. Jimenez, C. J. Watts, J. Garatuza-Payan, and W. J. Shuttleworth, 2004: Analysis of 2002 and 2003 warm-season precipitation from the North American Monsoon Experiment Event Rain Gauge Network. *Mon. Wea. Rev.*, **132**, 2938–2953.
- Gochis, D. J., L. Brito-Castillo, and W. J. Shuttleworth, 2006: Hydroclimatology of the North American monsoon region in northwest Mexico. *J. Hydro.*, **316**, 53–70.
- Gochis, D. J., C. J. Watts, J. Garatuza-Payan, and J. Cesar-Rodriguez, 2007: Spatial and temporal patterns of precipitation intensity as observed by the NAME Event Rain Gauge Network from 2002 to 2004. *J. Climate*, **20**, 1734–1750.
- Gremillion, M. S. and R. E. Orville, 1999: Thunderstorm characteristics of cloud-to-ground at the Kennedy Space Center, Florida: A study of lightning initiation signatures as indicated by the WSR-88D. *Wea. Forecasting*, **14**, 640–649.
- Higgins, R. W., Y. Yao, and X. L. Wang, 1997: Influence of the North American monsoon system on the U.S. summer precipitation regime. *J. Climate*, **10**, 2600–2622.

- Higgins, R. W., and Coauthors, 2006: The NAME 2004 field campaign and modeling strategy. *Bull. Amer. Meteor. Soc.*, **87**, 79–94.
- Houze, R. A., Jr., D. C. Wilton, B. F. Smull, 2007: Monsoon convection in the Himalayan region as seen by the TRMM precipitation radar. *Quart. J. Roy. Meteor. Soc.*, **133**, 1389–1411.
- Hubbert, J., and V. N. Bringi, 1995: An iterative filtering technique for the analysis of copolar differential phase and dual-frequency radar measurements. *J. Atmos. Oceanic Technol.*, **12**, 643–648.
- Illingworth A. J., J. W. F. Goddard, and S. M. Cherry, 1987: Polarization radar studies of precipitation development in convective storms. *Quart. J. Roy. Meteor. Soc.*, **113**, 469–489.
- Jameson, A., 1983: Microphysical interpretation of multi-parameter radar measurements in rain. Part I: Interpretation of polarization measurements and estimation of raindrop shapes. *J. Atmos. Sci.*, **40**, 1792–1802.
- Jameson, A., M. Murphy, and E. Krider, 1996: Multiple-parameter radar observations of isolated Florida thunderstorms during the onset of electrification. *J. Appl. Meteor.*, **35**, 343–354.
- Johnson, J. T., P. L. MacKeen, A. Witt, E. D. Mitchell, G. J. Stumpf, M. D. Eilts, and K. W. Thomas, 1998: The Storm Cell Identification and Tracking Algorithm: An enhanced WSR-88D Algorithm. *Wea. Forecasting*, **13**, 263–276.
- Johnson, R. H., P. E. Cieliecki, T. S. L’Ecuyer, and A. J. Newman, 2010: Diurnal cycle of convection during the 2004 North American Monsoon Experiment. *J. Climate*, **23**, 1060–1078.
- Keenan, T., and Coauthors, 2000: The Maritime Continent Thunderstorm Experiment (MCTEX): Overview and some results. *Bull. Amer. Meteor. Soc.*, **81**, 2433–2455.
- Lang, T. J., D. A. Ahijevych, S. W. Nesbitt, R. E. Carbone, S. A. Rutledge, and R. Cifelli, 2007: Radar-observed characteristics of precipitating systems during NAME 2004. *J. Climate*, **20**, 1713–1733.
- Lang, T. J., S. W. Nesbitt, and L. D. Carey, 2009: On the correction of partial beam blockage in polarimetric radar data. *J. Atmos. Oceanic Technol.*, **26**, 943–957.
- Lang, T. J., S. A. Rutledge, and R. Cifelli, 2010: Polarimetric radar observations of convection in northwestern Mexico during the North American Monsoon Experiment. *J. Hydrometeorol.*, **11**, 1345–1357.

- Laksen, H. R., and E. J. Stansbury, 1974: Association of lightning flashes with precipitation cores extending to height 7 km. *J. Atmos. Terr. Phys.*, **36**, 1547–1553.
- Li, L., W. Schmid, and J. Joss, 1995: Nowcasting of motion and growth of precipitation with radar over a complex orography. *J. Appl. Meteor.*, **34**, 1286–1300.
- Liu, H., and V. Chandrasekar, 2000: Classification of hydrometeors based on polarimetric radar measurements: Development of fuzzy logic and neuro-fuzzy systems, and in situ verification. *J. Atmos. Oceanic Technol.*, **17**, 140–164.
- Michimoto, K., 1991: A study of radar echoes and their relation to lightning discharge of thunderclouds in the Hokuriku District. Part I: Observation and analysis of thunderclouds in summer and winter. *J. Meteor. Soc. Japan*, **69**, 327–335.
- Michimoto, K., 1993: A study of radar echoes and their relation to lightning discharge of thunderclouds in the Hokuriku District. Part II: Observation and analysis of “single flash” thunderclouds in midwinter. *J. Meteor. Soc. Japan*, **71**, 195–204.
- Mohr, C. G., L. J. Miller, R. L. Vaughan, and H. W. Frank, 1986: The merger of mesoscale datasets into a common Cartesian format for efficient and systematic analyses. *J. Atmos. Oceanic Technol.*, **3**, 143–161.
- Nesbitt, S. W., R. Cifelli, and S. A. Rutledge, 2006: Storm morphology and rainfall characteristics of TRMM precipitation features. *Mon. Wea. Rev.*, **134**, 2702–2721.
- Nesbitt, S. W., D. J. Gochis, and T. J. Lang, 2008: The diurnal cycle of clouds and precipitation along the Sierra Madre Occidental observed during NAME-2004: Implications for warm season precipitation estimation in complex terrain. *J. Hydrometeor.*, **9**, 728–743.
- Pereira, L. G., 2008: Characteristics and organization of precipitating features during NAME 2004 and their relationship to environmental conditions. Ph.D. Dissertation, Atmospheric Science, Colorado State University, Fort Collins, Colorado, 229 pp.
- Petersen, W. A., S. A. Rutledge, and R. E. Orville, 1996: Cloud-to-ground lightning observations from TOGA COARE: Selected results and lightning location algorithms. *Mon. Wea. Rev.*, **124**, 602–620.
- Petersen, W. A., L. D. Carey, S. A. Rutledge, J. C. Knievel, N. J. Doesken, R. H. Johnson, T. B. McKee, T. Vonder Haar, and J. F. Weaver, 1999: Mesoscale and radar observations of the Fort Collins flash flood of 28 July 1997. *Bull. Amer. Meteor. Soc.*, **80**, 191–216.
- Rickenbach, T. M., and S. A. Rutledge, 1998: Convection in TOGA COARE: Horizontal scale, morphology, and rainfall production. *J. Atmos. Sci.*, **55**, 2715–2729.

Rinehart, R. E., and E. T. Garvey, 1978: Three-dimensional storm motion detection by conventional weather radar. *Nature*, **273**, 287–289.

Rosenfeld, D., 1987: Objective method for analysis and tracking of convective cells as seen by radar. *J. Atmos. Oceanic Technol.*, **4**, 422–434.

Rosenfeld, D., and W. L. Woodley, 2003: Closing the 5-year circle: From cloud seeding to space and back to climate change through precipitation physics. *Cloud Systems, Hurricanes, and the Tropical Rainfall Measuring Mission (TRMM)*, *Meteor. Monogr.*, No. 51, Amer. Meteor. Soc., 59–80.

Rowe, A. K., S. A. Rutledge, T. J. Lang, P. E. Ciesielski, and S. M. Saleeby, 2008: Elevation-dependent trends in precipitation observed during NAME. *Mon. Wea. Rev.*, **136**, 4962–4979.

Takahashi, T., 1990: Near absence of lightning in torrential rainfall producing Micronesian thunderstorms. *Geophys. Res. Lett.*, **17**, 2381–2384.

Takahashi, T., and K. Kuhara, 1993: Precipitation mechanisms of cumulonimbus clouds at Pohnpei, Micronesia. *J. Meteor. Soc.*, **71**, 21–31.

Tessendorf, S. A., L. J. Miller, K. C. Wiens, and S. A. Rutledge, 2005: The 29 June 2000 supercell observed during STEPS. Part I: Kinematics and microphysics. *J. Atmos. Sci.*, **62**, 4127–4150.

Vincent, B. R., L. D. Carey, D. Schneider, K. Keeter, and R. Gonski, 2003: Using WSR-88D reflectivity data for the prediction of cloud-to-ground lightning: A North Carolina study. *Nat. Wea. Digest*, **27**, 35–44.

Williams, C. R., A. B. White, K. S. Gage, and F. M. Ralph, 2007: Vertical structure of precipitation and related microphysics observed by NOAA profilers and TRMM during NAME 2004. *J. Climate*, **20**, 1693–1712.

Yuter, S. E., and R. A. Houze, 1995: Three-dimensional kinematic and microphysical evolution of Florida cumulonimbus. Part II: Frequency distributions of vertical velocity, reflectivity, and differential reflectivity. *Mon. Wea. Rev.*, **123**, 1941–1963.

Zeng, Z., S. E. Yuter, R. A. Houze, Jr., and D. E. Kingsmill, 2001: Microphysics of the rapid development of heavy convective precipitation. *Mon. Wea. Rev.*, **129**, 1882–1904.

**PART II: INVESTIGATION OF MICROPHYSICAL PROCESSES OCCURRING
IN ORGANIZED CONVECTION DURING NAME**

CHAPTER 1

INTRODUCTION

The arrival of strong convection in the semiarid regions of southwestern United States and northwestern Mexico during mid- to late- June marks the onset of the North American Monsoon (NAM). This atmospheric circulation develops as a result of land-sea temperature contrast, and similar to its Asian counterpart, is characterized by a reversal of the mean low-level winds, which ushers moist flow from the Gulf of California (GoC) onshore into northwestern Mexico. Within a horizontal distance of 200 km, the warm waters of the GoC transition to the steep topography of the Sierra Madre Occidental (SMO), extending more than 3 km above ground level. It is well known that diurnally forced land-sea and mountain-valley flows affect the timing and distribution of precipitation in this region (Douglas et al. 1993; Dai et al. 1999). More specifically, results from the North American Monsoon Experiment (NAME; Higgins et al. 2006) found a strong topographical influence on the diurnal variability of rainfall (e.g., Gochis et al. 2004, 2007; Lang et al. 2007), where frequent afternoon convection occurred over the SMO, dissipating by early evening. Under certain conditions, these precipitation features grew upscale to form larger, organized, mesoscale convective systems (MCSs) that propagated toward the GoC. This promoted a less frequent but more intense late-evening/early-morning peak in precipitation across lower elevations (Lang et al. 2007). In addition to this coast-perpendicular movement of organized systems (defined as Regime

A by Lang et al. 2007), periods when phase speeds were characterized by northward along-coast propagation were also identified (Regime B, same study). These systems were characterized by longer lifetimes and persisted well into the early morning hours over the GoC. On occasion, MCSs displayed both along- and cross-coast movement, leading to a combined classification referred to as Regime AB by Lang et al. (2007).

The processes by which convection initiates and organizes once off the SMO needs to be better understood to properly represent the diurnal cycle of precipitation in numerical models. In addition, the evolution of microphysical processes must be studied to provide a physical basis for estimating precipitation via remote sensing platforms (e.g., Nesbitt et al. 2008). Hydrometeor identification using polarimetric radar data provides a means to validate and improve microphysical and convective parameterizations in models, which have been found to significantly affect model simulations of the regional climate during the NAM (e.g., Gochis et al. 2002). Previous satellite- and radar-based studies of precipitating features in this region have shown that convection over the SMO tends to be shallower than over the lower elevations and coastal plain (e.g., Nesbitt et al. 2008, Rowe et al. 2008). Nesbitt et al. (2008) also described a corresponding reduction in depth of precipitation-sized ice in convection over the high terrain, suggesting a greater dependence on warm-rain processes. However, Rowe et al. (2008) found that warm-cloud depths increased toward lower elevations, suggesting an increased role of accretional growth in explaining the larger precipitation intensities over the lower terrain. In addition, the presence of graupel up to 10 km MSL in convection over the coastal plain was noted using profiler data (Lerach et al. 2010), suggesting an important contribution from ice-based processes over lower elevations as well.

Using three-dimensional gridded data from the National Center for Atmospheric Research (NCAR) S-band, polarimetric radar (S-Pol), Lang et al. (2010) found increased ice and water mass, greater rainfall intensity, and larger median drop sizes over the coastal plain relative to the high terrain. Furthermore, Rowe et al. (2011) selected NAME S-Pol volume scans with improved temporal and spatial resolution to investigate the evolution of microphysical characteristics associated with isolated convection over land. Increased water mass and rain-water depth were associated with convection over the lower elevations, consistent with the Lang et al. (2010) study, confirming the important role of warm-rain processes (Rowe et al. 2008). A selection of cases over the lower elevations illustrated a clear dependence on mixed-phase processes, in addition to coalescence, via drop freezing and subsequent riming growth. Several examples of convection over the SMO revealed shallower coalescence zones and reduced graupel depth, consistent with previous studies (e.g., Nesbitt et al. 2008); however, the potential for higher elevations to receive brief periods of intense rainfall was also noted in these cases, as well as in previous general studies of NAME precipitation (e.g., Gochis et al. 2007; Rowe et al. 2008). Peak ice mass, occurring just above the melting level for all elevation groups, was greatest over the SMO, likely due to the shallower warm-cloud depth in cells over these higher elevations. This likely limited the collision-coalescence process, allowing for a greater percentage of condensate to be lofted above the freezing level, and ultimately leading to increased ice mass and (occasional) intense rain rates over the higher terrain as these large ice hydrometeors melted. Indeed, a particularly deep cell over the SMO was characterized by rain rates $> 80 \text{ mm h}^{-1}$, despite the truncated warm-cloud depth.

Lang et al. (2007, 2010) noted increased rainfall during disturbed regimes, suggesting a likely dependence of microphysical processes on degree of organization as well. During Regime AB, a greater increase in liquid water mass compared to ice mass was observed over the GoC (Lang et al. 2010), implying an increasingly important role of warm-rain processes compared to non-disturbed periods dominated by isolated convection. A more modest change, however, was observed over the high terrain during these disturbed periods, suggesting an additional topographical dependence on microphysical processes associated with organized convection. Using S-Pol data, this study employs a case study approach to describe the evolution of *organized* convection during NAME, providing information about hydrometeor characteristics as a function of topography. In addition, cells associated with organized systems will be compared to isolated convection in the radar domain to further investigate the relationship between microphysical processes and organization.

CHAPTER 2

DATA AND METHODOLOGY

2.1 S-Pol data

During the NAME field campaign, the S-Pol radar, situated approximately 90 km north of Mazatlan along the GoC coast (see Fig. 2.1 in Part I), provided nearly continuous data from 8 July through 21 August 2004. Low-level, 360° scans to a range of 250 km were completed every 15 minutes to map rainfall. To improve temporal and spatial resolution for specific precipitating features, a sector-scanning mode, with azimuthal widths ranging from 90° to 120° and maximum range of 150 km, was employed for about 95 hours of scanning time. Similar to the isolated cases in Rowe et al. (2011), examples of organized systems for this study were selected from these microphysical scans, which allowed for improved temporal resolution to investigate the evolution of the fine-scale vertical structure of embedded convection.

Quality control of the S-Pol data is described in detail in previous NAME studies (e.g., Lang et al. 2007, 2009). Radar variables used in this study included horizontal reflectivity (Z_H), differential reflectivity (Z_{DR}), providing information about oblateness, linear depolarization ratio (L_{DR}) and zero-lag cross-correlation coefficient ($\rho_{HV}(0)$), both which allow for discrimination between pure rain and mixtures of hydrometeors, as well as specific differential phase (K_{DP}), which depends on both the mean drop size of hydrometeors and the water content within the volume. A more complete description of

these polarimetric variables can be found in Bringi and Chandrasekar (2001). Corrected S-Pol data were then interpolated to a 1-km horizontal- and 0.5-km vertical-resolution Cartesian grid using REORDER (Mohr et al. 1986). The gridded polarimetric variables, along with a mean temperature profile from nearby Mazatlan, were then incorporated into a hydrometeor classification (HID) algorithm, based on the methodology of Liu and Chandrasekar (2000) and described by Tessendorf et al. (2005), to determine dominant hydrometeors type at each grid point.

2.2 Cell identification and tracking

The gridded S-Pol data set was then subjected to a cell identification and tracking algorithm, described in detail by Rowe et al. (2011). In general, cells were identified using an ellipse-fitting method (Nesbitt et al. 2006), based on reflectivity and area thresholds, providing a means to objectively classify convective elements. Specifically, multiple reflectivity thresholds of 35 and 45 dBZ were applied to improve identification of intense convective echo embedded within larger features. An example is shown in Fig. 2.1, highlighting the effectiveness of using multiple thresholds to capture embedded convection.

In order to focus solely on MCSs, the locations of tracked cells were matched to precipitating features identified by Pereira (2008). Pereira used a feature identification algorithm, developed by Rickenbach and Rutledge (1998) and described by Lang et al. (2007) in the context of the NAME data set, to classify feature types based on the following criteria: if the major axis of the feature exceeded 100 km, it was considered an MCS; furthermore, if the ratio between the major and minor axes of the convective area

of the feature was more than five to one (Bluestein and Jain 1995), the feature was classified as linear. For this study, all cells embedded within features defined as MCS-scale, whether linear or nonlinear, were included to describe the evolution of cells within organized systems.

2.3 Cell properties

In addition to the polarimetric variables, several cell properties were computed to aid in the description of storm evolution and provide a means for comparison with isolated convection. Echo-top heights were estimated using the maximum height of the 0-dBZ reflectivity contour. Maximum Z_{DR} for each cell was computed for grid points within the cell identified as rain by the HID and located below 4 km to reduce the effects of melting; the melting level was located, on average, at 5km during NAME. Ice and liquid water masses were calculated using the same technique as the isolated cases (Rowe et al. 2011) and the Lang et al. (2010) study, following a methodology described by Carey and Rutledge (2000) and Cifelli et al. (2002). Ice (IWP) and liquid water (LWP) paths were computed by integrating ice and liquid water masses over the vertical columns associated with all grid points in individual convective cells. Ice and liquid water path ratios were then computed by dividing the IWP and LWP by the summation of the two, respectively. Near-surface rain rates, calculated using polarimetric-based equations (Cifelli et al. 2002, Cifelli et al. 2011), were also available for each cell, providing additional means for describing microphysical changes as these precipitating systems evolved.

2.4 Topographic data

Maximum terrain height was determined for each identified cell, using topographic data from the National Geophysical Data Center (NGDC). In addition to feature type, cells were grouped based on elevation using the same terrain thresholds used in previous NAME studies (Gochis et al. 2004; Rowe et al. 2008, 2011): 0–1 km, 1–2 km, > 2 km, and over water. This allowed for comparisons of cell characteristics based on organization and topography. Note that the water grouping was included for this analysis, as a greater frequency of organized convection was observed over lower elevations and over the GoC (Fig. 2.2) compared to the relatively few isolated cells over water (Fig. 3.1 in Part I). The following case studies will focus on features that organized off the SMO, as well as those which moved into the domain, allowing not only for elevation-dependent trends over land to emerge, but to analyze differences between convection over land and water.

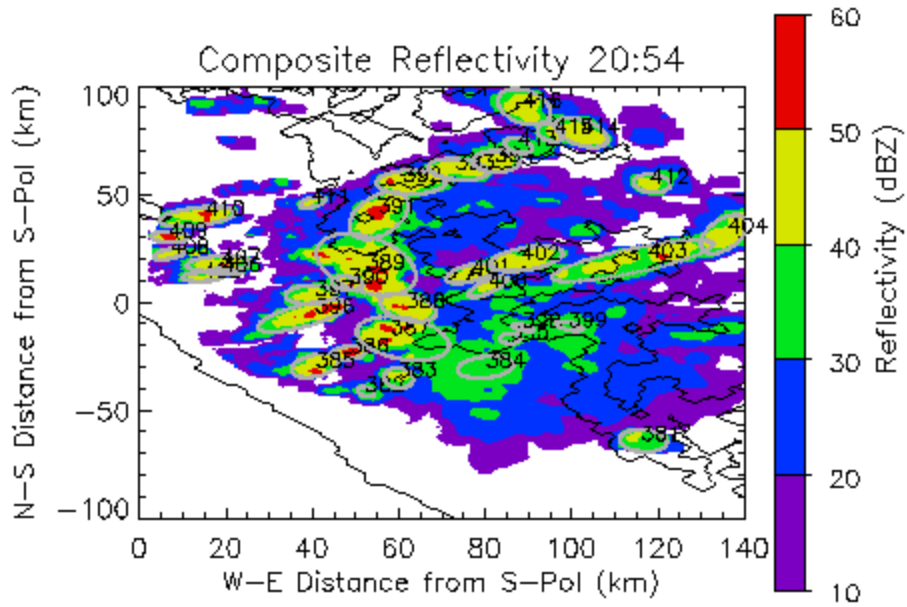


Fig. 2.1. Gridded composite reflectivity (color-filled contours) for 2054 UTC on 5 August 2004. Cell ellipses are shown in gray and associated cell numbers are identified by bold, black numbers. Overlaid in black contours is the terrain height at 0, 1, 2, and 3 km.

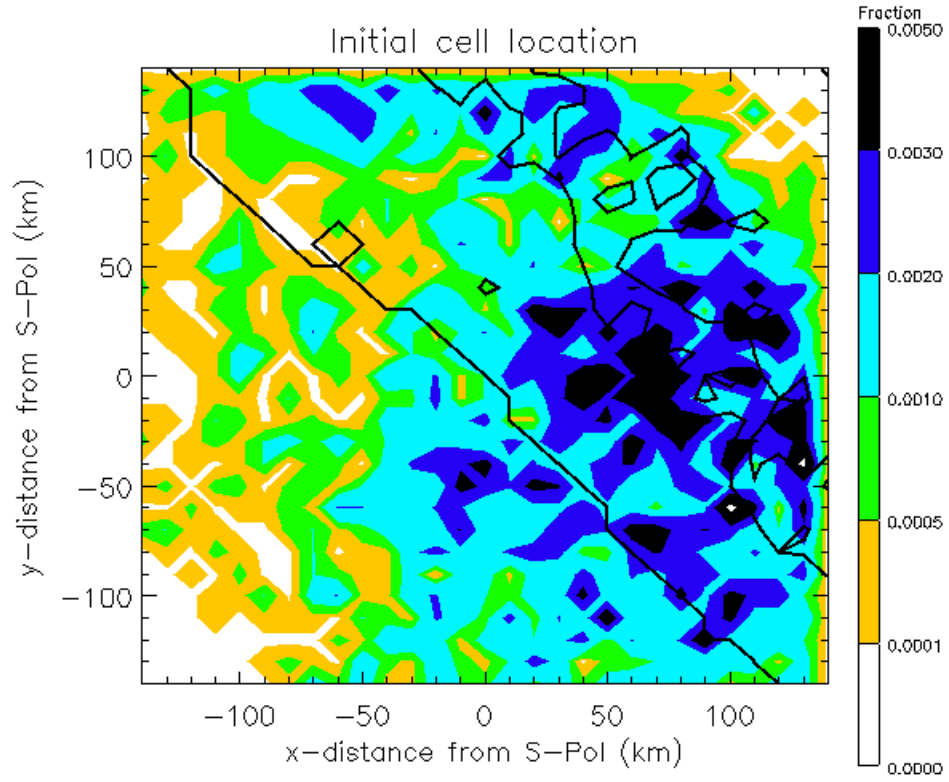


Fig. 2.2. Frequency of occurrence (color-filled contours) of the initial location of organized cells at each grid point. Frequencies are normalized by the total number of organized cells in the domain to obtain a fraction. Bold, black contours represent the topography contoured at 0, 1, 2, and 3 km.

CHAPTER 3

CASE STUDIES

3.1 12 July

The first case of interest occurred from 0700 to 1300 UTC on 12 July 2004, featuring an MCS moving into the domain from the southeast during the overnight hours. This period was characterized by moisture advection into the region as a result of enhanced east-southeast flow on the equatorward side of an anticyclone (Pereira 2008). Enhanced precipitable water led to an increase in CAPE exceeding 2000 J kg^{-1} , and the approach of an inverted trough provided dynamical support (Finch and Johnson 2010). Convection initiated over the SMO between Mazatlan and Puerto Vallarta at 0300 UTC, and by 0500 UTC, began to organize as it moved off the higher terrain. At 0700 UTC, the leading edge of this system moved into the far southeastern portion of the S-Pol domain. Of the 996 individual cells identified and tracked during this case, 78% were considered part of organized features, with 76% of those over the GoC and the remaining over land; therefore the focus of this case will be on comparing convection over land with that over water.

At 0845 UTC, the leading edge of the system extended from the coast to over the western slopes of the SMO. A vertical cross section through the northeast portion of the MCS at this time (Fig. 3.1) shows the original leading edge at 115 km over the slopes, with echo top near 17 km and K_{DP} between 2 and 3° km^{-1} at low levels (Fig. 3.1a).

Additional convective development was observed over lower elevations around 95 km, featuring a slightly deeper cell that contained a melting hail and graupel signature (Fig. 3.1b) with K_{DP} values near 2° km^{-1} and $Z_{DR} > 2 \text{ dB}$ below the melting level (associated with oblate raindrops; Fig. 3.1a). An RHI through this leading convection at 0912 UTC (Fig. 3.2) highlights a deep cell with echo top reaching 17 km and 50-dBZ reflectivities to 15 km, with high-density graupel and hail extending to 14 km. Enhanced L_{DR} , with values near -20 dB, was present in the core, indicating a mixture of ice and liquid hydrometeors. Melting graupel and hail can be inferred by K_{DP} near 4° km^{-1} and Z_{DR} of 4 dB just below the melting level, corresponding to drop sizes approaching 3 mm (estimated using $D_0 = 1.529 Z_{DR}^{0.467}$; Bringi and Chandrasekar, 2001). The velocity image shows storm-top divergence and ground-relative, low-level flow toward the radar associated with storm outflow. By 0925 UTC (Fig. 3.3), the 50-dBZ reflectivities extended only to 9 km, with peak K_{DP} values also descending to near the surface, indicating the mature phase of this cell. However, the presence of positive Z_{DR} above the melting level along the leading edge, capped by a region of slightly enhanced L_{DR} values, suggests continued production of large ice hydrometeors via drop freezing. Positive Z_{DR} values extending into the mixed-phase zone, known as Z_{DR} columns, indicate the lofting of oblate raindrops above the melting level (e.g., Illingworth et al. 1987). These drops quickly freeze and become embryos that promote rapid accretional growth. The subsequent production of wetted, aspherical ice particles (e.g., graupel) is inferred from a “cap” of enhanced L_{DR} (e.g., Jameson 1983), with values near -20 dB in this case. The increased K_{DP} above the melting level in this narrow region along the leading edge was consistent with the presence of oblate, supercooled drops. Similar features were observed

in isolated convection over the coastal plain during NAME (Rowe et al. 2011), and this role of coalescence and mixed-phase processes in rainfall production via drop freezing and subsequent accretional growth has also been described for convection in a variety of locales, including isolated convection in the U.S. (e.g., Bringi et al. 1996; Tessendorf et al. 2005; Jameson et al. 1996; Bringi et al. 1997; Fulton and Heymsfield 1991), intense cells over the Tiwi Islands (e.g., Carey and Rutledge 2000; Keenan et al. 2000), organized systems in Amazonia (e.g., Cifelli et al. 2002), and cumulonimbi in the tropics (e.g., Takahashi 1990; Takahashi and Kuhara 1993).

Also at this time, low-level flow toward the radar increased to values approaching 15 m s^{-1} (Fig. 3.3) and converged with the larger scale upslope flow (outbound velocities), as seen in the velocity PPI image at 0929 UTC (Fig. 3.4). In this region of convergence, an arc of lower reflectivities ahead of the leading line (Fig. 3.4) indicated the presence of an outflow boundary, with these features later leading to new convective development ahead over the lower terrain and adjacent water. A vertical cross section at 1008 UTC, through a portion of the system over water (Fig. 3.5), shows new convective development out ahead of the original leading line at a range of approximately 35 km. Graupel extended to only modest heights in this core (Fig. 3.5b), and maximum K_{DP} values near the surface at 60 km indicated the decaying stages of this cell (Fig. 3.5a). An elevated maximum in K_{DP} at 40 km (Fig. 3.5a) suggested a stronger leading cell where large ice hydrometeors melted to form substantial rain. Over the next ten minutes, the leading convective line intensified, resulting in deep convection along the leading edge over water at 1018 UTC, characterized by reflectivity values of 40-dBZ extending to 16 km, $K_{DP} > 2^\circ \text{ km}^{-1}$ (Fig. 3.6a), and graupel identified to near echo top (Fig. 3.6b). Older

cells continued to decay further back near the coastline, contributing to trailing stratiform precipitation. The production of stratiform rain from decaying convection is a common characteristic of MCSs in the midlatitudes (Houze 1993, 1997) and appeared to be important for the production of trailing stratiform rain in this case as well.

These features persisted through the next hour, suggesting the importance of outflow boundaries in the maintenance and propagation of the MCS as it moved offshore. The cell identification and tracking algorithm used in this study allowed for a comparison of cell characteristics between cells tracked along the leading edge over land that eventually moved offshore (0–1 km in Fig. 3.7) with those that formed along outflow boundaries over water and tracked farther away from the coast (water in Fig. 3.7). A comparison between IWP ratios (Fig. 3.7a), in particular, revealed that cells within the leading edge over the coastal plain contained more ice relative to the cells that initiated over water. Echo-top heights were also greater for the convection over land (Fig. 3.7b), possibly leading to the greater IWP ratios due to the greater depth over which ice could exist. The leading convection over the coastal plain also contained larger drops, as indicated by the wider Z_{DR} distribution compared to the track over water in Fig. 3.7c. However, despite the reduced ice mass, shallower echo tops, and smaller drops, the convection that occurred over the GoC had greater maximum instantaneous rain rates (Fig. 3.7d) and larger K_{DP} (not shown), suggesting a significant contribution from warm-rain processes. This is consistent with the Lang et al. (2010) study, where fundamental differences in drop-size distributions were described between the land and water, including a tendency for convection over the GoC to contain smaller drops and less ice mass.

3.2 5 August

A quasi-linear, asymmetric MCS developed along the SMO around 1800 UTC on 5 August, propagated toward the radar, then moved toward the northwest after 2200 UTC, paralleling the coast. Due to this combined cross- and along-coast movement, this period was classified under Regime AB (Lang et al. 2007). Environmental conditions were characterized by increased easterly zonal flow at mid and upper levels due to an approaching upper-level jet streak, resulting in 0–6 km wind shear values reaching 8.5 m s^{-1} (Pereira 2008). In addition, an upper-level inverted trough to the southeast led to moisture advection into the domain during the afternoon of 5 August. Under this moist and sheared environment, the quasi-linear MCS developed during the evening hours.

In contrast to the previous case, most convection during this time period was associated with organized features over land, with 72% of cells classified as organized and only 2% of those over water. Around 1830 UTC, scattered convection developed over the SMO ($> 2 \text{ km}$) at the far eastern edge of the radar domain. By 1933 UTC, a west-east-oriented line of convection formed to the east of the radar, with additional convection developing to the east-northeast of the radar over the high terrain. A vertical cross section at 1951 UTC through the cells to the east (Fig. 3.8a,b) shows several cells at various stages of their lifecycles, with the leading convection over an SMO peak at 90 km characterized by an echo-top height of 18 km, a Z_{DR} column, and graupel reaching 15 km MSL (Fig. 3.8b). This cell was followed by a mature cell at 110 km with maximum K_{DP} values extending to the surface (Fig. 3.8a) and a shallower extent of graupel (Fig. 3.8b). Even weaker cells were situated farther back over the terrain, although at these ranges, it

was difficult to assess the storm's true intensity. Shortly after, at 2002 UTC (Fig. 3.8c,d), the cell at 110 km had weakened, characterized by reduced K_{DP} values near the surface (Fig. 3.8c) and a slightly shallower depth of graupel (Fig. 3.8d) with less continuity between cells compared to the previous time. The cells beyond 120 km had decayed, becoming an extensive trailing stratiform region. The collapse of convective cells to form stratiform precipitation areas was similar to that observed in the previous case over the lower terrain and water; this sequence of newer convection, followed by a mature cell, and subsequently, a dissipating cell, is a signature of discrete propagation, as described for MCSs in the tropics (e.g., Houze 1977) and in the midlatitudes (e.g., Smull and Houze 1985).

The collapse of cells over the SMO, forming regions of stratiform rain, left a quasi-linear convective feature along the western slopes by 2018 UTC. A vertical cross section through this leading convection to the east-southeast at 2021 UTC (Fig. 3.9a,b) highlights a deep system, with echo tops exceeding 15 km and a Z_{DR} column evident at the leading edge. An elevated maximum in K_{DP} (Fig. 3.9a) beneath HID-identified high-density graupel (Fig. 3.9b) indicated the melting of ice hydrometeors, which appeared to be an important contributor to rainfall along the western slopes. Also at this time, new convection initiated ahead of the system over the lower elevations, likely due to outflow produced by the system. In this developing convection, the majority of echo was below the melting level, implying the importance of warm-rain processes early on. At 2042 UTC (Fig. 3.9c,d), a Z_{DR} column was again observed along the leading edge, with values exceeding 2 dB above the melting level, indicating lofting of large, liquid hydrometeors into the mixed-phase zone, manifested as a new area of HID-identified graupel along the

leading edge at 60 km (Fig. 3.9d). Values of K_{DP} remained large ($> 2^\circ \text{ km}^{-1}$; Fig. 3.9c), suggesting the continued melting of ice along the slopes. The cell ahead of this system deepened, yet remained dominated by collision-coalescence due to the large warm-cloud depth and lack of HID-identified graupel. By 2047 UTC (Fig. 3.9e,f), the Z_{DR} column was no longer present, but a new maximum in K_{DP} of $3\text{--}4^\circ \text{ km}^{-1}$ along the leading edge suggested melting of large ice hydrometeors that were produced by frozen drops and subsequent riming, similar to the processes that occurred in the leading edge of the system on 12 July. The previous K_{DP} maximum descended to near the surface at this time, further highlighting the multicellular nature of these systems.

A reflectivity PPI image at 2055 UTC (Fig. 3.10) reveals a nearly continuous leading line of convection along the western slopes, and shows the widespread, new convective development over the lower terrain. The corresponding radial velocity image at this time (Fig. 3.10) shows low-level upslope flow converging with downslope flow off the higher terrain, collocating the convergence zone with the leading convection along the western slopes. This flow pattern is consistent with a study by Johnson et al. (2010), using surface and upper-air data during NAME, that described a reversal in the daytime, upslope flow along the SMO during the evening and overnight hours. This generated a westward-propagating zone of convergence that moved downslope toward the coast by early morning. An RHI through the east-southeast portion of this feature at 2108 UTC (Fig. 3.11), corresponding to the later stages of the system shown in Fig. 3.9, shows the strong, mid-level flow in the stratiform system overlying the low-level upslope flow away from the radar. A brightband signature is evident in the trailing stratiform region, characterized by enhanced reflectivity, Z_{DR} , and L_{DR} , typical of midlatitude MCSs (e.g.,

Houze 2004). Deep convection along the leading edge consisted of higher reflectivity values above a weak echo region, defined as a reflectivity overhang that resulted from a strong upward motion in the core. Consistent with a strong updraft, positive Z_{DR} values along the leading edge extended 2 km above the melting level, with an area of enhanced L_{DR} above this Z_{DR} column, once again indicating freezing of drops and subsequent growth by riming. A maximum in K_{DP} of nearly 4° km^{-1} just below the melting level, in a region coinciding with the reflectivity overhang, where the HID abruptly transitioned from graupel to rain, suggested the melting of large ice hydrometeors, and, therefore, a strong dependence on mixed-phase processes in this system.

At 2123 UTC, an RHI through the portion of this system to the east of the radar (Fig. 3.12) reveals new leading convection following the decay of the older convective line. The combination of an elevated area of 60-dBZ reflectivity above the melting level along the back edge of the cell, with negative Z_{DR} values and $L_{DR} > -21$ dB suggested the presence of large, tumbling hail in the wet growth stage (e.g., Holler et al. 1994), similar to a preliminary analysis of this RHI by Higgins et al. (2006). Values of K_{DP} reached 4° km^{-1} below this region of hail, corresponding to rain rates of approximately 165 mm h^{-1} and indicating melting, large ice hydrometeors, consistent with the Higgins et al. (2006) study that inferred complete melting of hail into rain due to the higher Z_{DR} and lower L_{DR} values at the surface. The presence of large hail aloft in convection along the terrain in NAME was an interesting finding in the Higgins et al. (2006) study, and has been shown through this and the Rowe et al. (2011) study to have been a rather common feature in convection along the western slopes.

Similar to Fig. 3.11, Fig. 3.12 shows strong inbound velocities at midlevels converging with outbound velocities. A relatively small area of outflow (inbound velocities) at low levels promoted continued initiation of convection over the lower terrain. After this time, as this system moved toward the coast, the radar no longer topped the system; however, this MCS turned toward the northwest as it approached the water, moving parallel to the coast. The system eventually passed over the S-band profiler to the north of S-Pol (see Fig. 2.1 in Part I), where Lerach et al. (2010) noted features within the stratiform region of this system consistent with MCS-characteristics described in other regions of tropical precipitation (e.g., Williams et al. 1995; Ecklund et al. 1999; May and Keenan 2005), including a reflectivity brightband and strong Doppler velocity gradients within the melting layer.

This case, which captured the lifecycle of an MCS during NAME from its origin over the SMO to the highly organized stage over the lower elevations, provided a unique opportunity to evaluate the evolution of an organized system as a function of topography. Vertical profiles of mean ice mass for all organized cells in this case (Fig. 3.13) highlight the deep extent of ice in cells over the lower terrain, as seen in the vertical cross sections and RHIs; however, similar values existed in the mixed-phase region for all elevations. A particular track, which captured the initiation of the cell over the SMO through organization over the lower terrain, demonstrated an expected general decrease in IWP ratio as the warm-cloud depth increased over the lower terrain (Fig. 3.14). This suggests that although a similar amount of ice mass was available in the mixed-phase region, the increased warm-cloud depth over the coastal plain allowed for an additional contribution to rainfall from coalescence compared to the SMO, similar to results presented by Rowe

et al. (2008, 2011). Over the low terrain, the combined effects of ice-based and warm-rain processes led to rain rates exceeding 150 mm h^{-1} ; however, maximum rain rates were consistently $> 100 \text{ mm h}^{-1}$ for this track (Fig. 3.14), suggesting the importance of melting ice hydrometeors for rainfall production over the high terrain of the SMO.

3.3 Additional features of interest

Similar to 5 August, convection initiated over the SMO during the afternoon of 20 July, which later organized into a quasi-linear MCS along the western slopes as older convection collapsed over the higher terrain, leaving a leading convection/trailing stratiform feature typical of both the tropics and midlatitudes (Fig. 3.15). The radial velocity PPI associated with this system (Fig. 3.15) reveals mid-level rotation in the trailing stratiform region, defined as a mesoscale convective vortex (MCV; Menard and Fritsch 1989). These mid-level vortices have been observed in MCSs in both the midlatitudes (e.g., Johnston 1981; Smull and Houze 1985; Johnson et al. 1989; Scott and Rutledge 1995) and tropics (e.g., Gamache and Houze 1982; Chong and Bousquet 1999; Keenan and Rutledge 1993), and have been found to play an important role in the organization of convection (e.g., Zhang and Fritsch 1987; Bartels and Maddox 1991). More specifically, this region of positive vorticity corresponding to an MCV has been related to mid-level, storm-relative flow in the stratiform region (Houze et al. 1989). An RHI through the portion of the 20 July NAME MCS associated with the MCV (Fig. 3.16) reveals storm-relative, mid-level rear inflow of about 10 m s^{-1} in the trailing stratiform region that entered the system above the melting level between 5 and 8 km and had descended to near the surface as it approached the leading convective line. Smull and

Houze (1987) proposed that these rear inflow jets are a dynamical response to the mid-level rotation to the potential vorticity anomaly in the stratiform area. Acceleration toward the convective line is driven by a mid-level pressure gradient between the positively (negatively) buoyant anomalies above (below) the melting level, and the downward component is aided by microphysical processes, specifically cooling due to melting, sublimation, and evaporation. This rear inflow jet, in turn, amplifies the mid-level vortex by increasing horizontal vorticity, which can be tilted into the vertical within the ambient flow (Scott and Rutledge 1995).

Although surface and upper-air observations over the higher terrain were unavailable during this time, thus limiting further mesoscale analysis of this MCV, the presence of this feature in NAME convection not only emphasizes the similarities to MCSs in other regions, but also contributes to the understanding of convective organization in this region. Descending rear inflow, combined with convective downdrafts within the precipitation core, results in the development of cold pools, which propagate away from the leading convection, and, as has been suggested for MCS cases during NAME, initiates new convection. The strength of the cold pool is related to microphysical processes through convective downdrafts via precipitation loading, evaporation, and melting of ice hydrometeors (e.g., Knupp and Cotton 1985). In particular, van den Heever and Cotton (2004) noted that the melting of small hailstones resulted in strong low-level downdrafts and deeper, faster moving cold pools, further suggesting an important role of melting in the production and modification of cold pools. Ice physics also play a role in the stratiform region, controlling the latent heat distribution and thus the flow pattern resulting from temperature gradients. Numerical simulations of

tropical MCSs in western Africa noted that experiments, which did not include ice, produced less-organized systems in which the rear inflow was weaker and did not reach the ground, thereby eliminating the surface gust front (Liu et al. 1997). The generation of outflow also depended on the mass flux in convective downdrafts, with enhancement observed when the ice phase was included. Therefore, the presence of large ice hydrometeors in NAME convection was likely necessary to obtain the degree of organization observed along the western slopes and coastal plain.

Typically, MCSs became less organized and dissipated as they moved out over the GoC, especially in the northern part of the domain (Lang et al. 2010), as was the case for 20 July. Another example occurred on 16–17 August 2004, when a linear MCS formed along the western slopes in the southern portion of the radar domain and propagated out over the water as new convection initiated along outflow boundaries produced by the leading convective line. Although this system persisted a greater distance over water than on 20 July, cells that initiated along the gust front over water were relatively shallower compared to the earlier developing convection over land, with significantly less ice mass (Fig. 3.17). As a result, this newer convection quickly decayed, and new development was limited due to the lack of cold pool generation, leading to the dissipation of this system before sunrise.

3.4 Case comparison

The cases presented herein are considered to represent the spectrum of organized systems that occurred during NAME. In every case examined, afternoon convection over the high terrain of the SMO gave way to an organized system over lower elevations.

Although many systems dissipated before reaching the coast (e.g., 21 July), others persisted through the overnight hours over the GoC (e.g., 16–17 August). In some cases, these systems exhibited both cross- and along-coast movement, with the MCS on 5 August illustrating an example of a system turning parallel to the coast. Others, such as on 12 July, moved northwestward into the radar domain, thereby characterized by a longer residence time over water.

Distributions of maximum Z_{DR} within organized cells on these case days (Fig. 3.18) show that drops were, on average, smallest over the coastal plain and adjacent GoC on 12 July, suggesting more maritime characteristics (e.g., Atlas and Ulbrich 2000; Bringi et al. 2003, 2009; Rosenfeld and Ulbrich 2003; Atlas and Ulbrich 2006; Ulbrich and Atlas 2007). The largest average Z_{DR} value corresponded to cells over the low terrain on 17 August, with distributions for water-only cells on this day containing lower Z_{DR} values than cells over land. Despite containing smaller drops, liquid water mass (Fig. 3.19a) was generally greatest for cells over water on this day. Liquid water mass was also large on 12 July compared to cells on 21 July over land, consistent with the hypothesized greater dependence on warm-rain processes for systems moving over the GoC from the southeast. This is further suggested by the CDFs of ice mass (Fig. 3.19b), which show reduced ice mass over water, with the greatest amounts for cells over the SMO on 21 July and 5 August, indicating the importance of ice-phase processes in convection over the terrain.

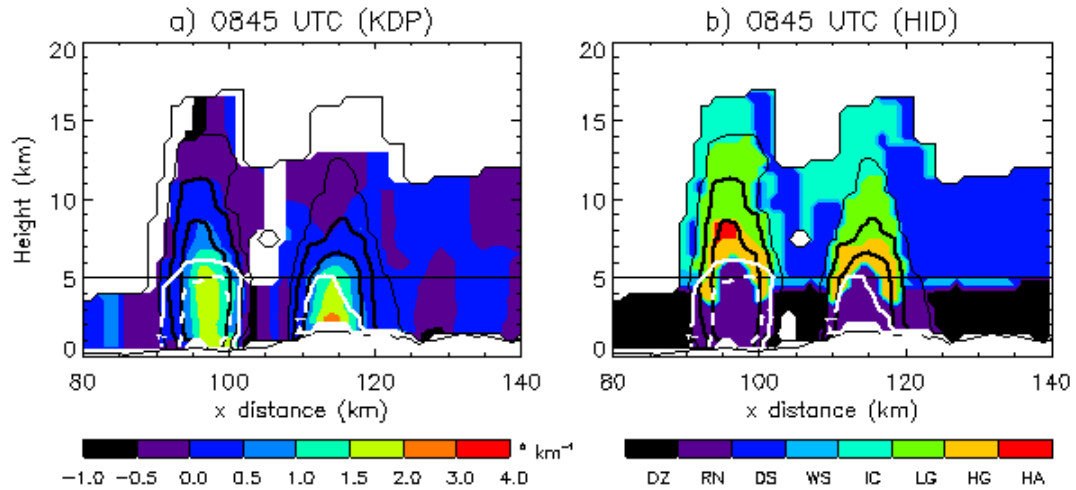


Fig. 3.1. Vertical cross sections through cells occurring at 0845 UTC on 12 July 2004. Color-filled contours are associated with (a) K_{DP} ($^{\circ} \text{ km}^{-1}$) and (b) HID. The HID abbreviations correspond to the following classifications: drizzle (DZ), rain (RN), dry snow (DS), wet snow (WS), ice (IC), low-density graupel (LG), high-density graupel (HG), and hail (HA). Black contours represent reflectivity at 0, 30, 40, and 50 dBZ, with the 40- and 50-dBZ contours thickened. Values of Z_{DR} are contoured in white for 1 dB (solid) and 2 dB (dashed). Terrain height is plotted at the bottom as a black contour, and the solid black line at 5 km represents the average height of the melting level during NAME. The x-axis is the distance from S-Pol in km.

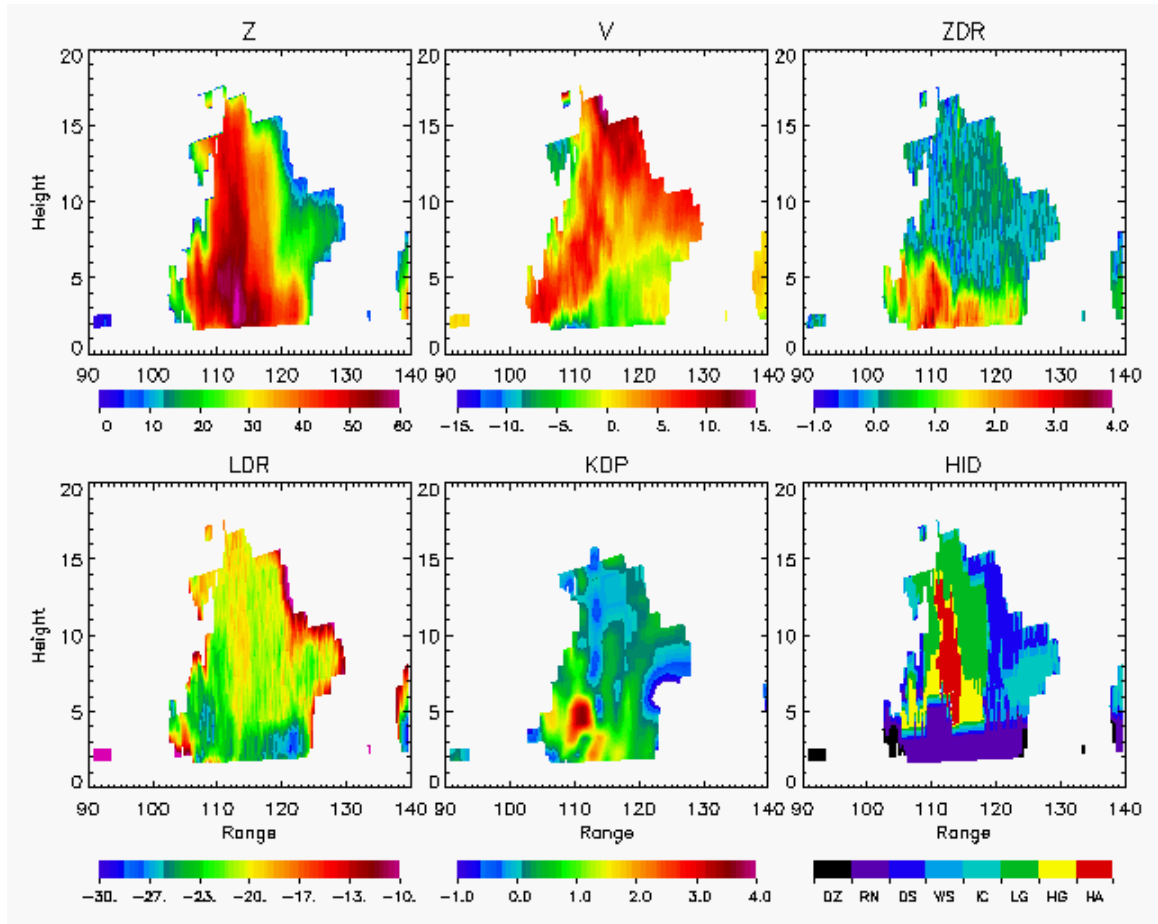


Fig. 3.2. An RHI through a cell at 0912 UTC on 12 July 2004 at an azimuth of 126° . Variables from left to right, starting with the upper panel, include reflectivity (Z , dBZ), radial velocity (V_R , m s^{-1}), differential reflectivity (Z_{DR} , dB), linear depolarization ratio (L_{DR} , dB), specific differential phase (K_{DP} , $^\circ \text{ km}^{-1}$), and the hydrometeor identification (HID). The HID abbreviations are the same as in Fig. 3.1. Range and height are in km.

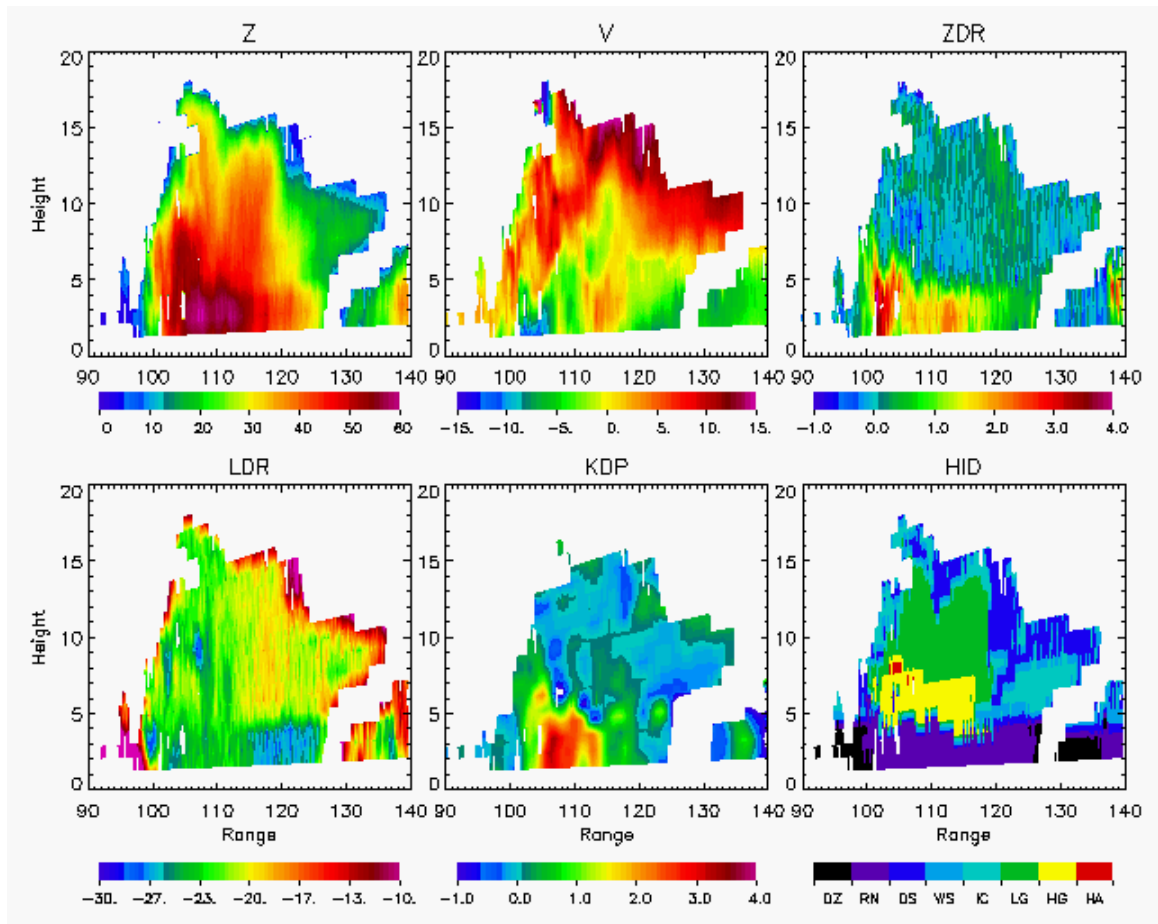


Fig. 3.3. An RHI through a cell at 0925 UTC on 12 July 2004 at an azimuth of 126°. Variables are as described in Fig. 3.2.

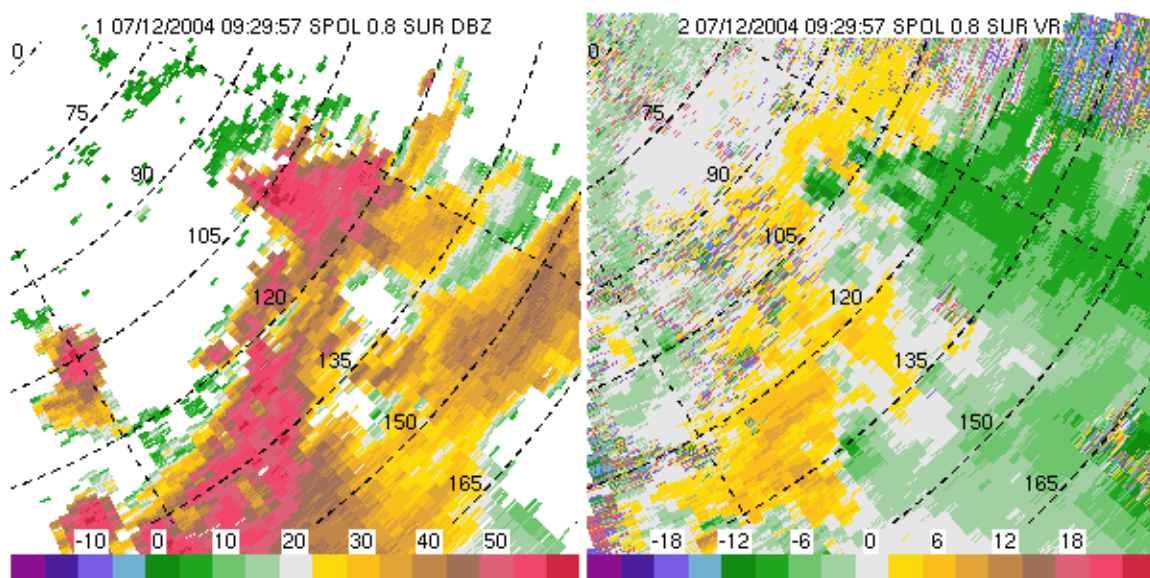


Fig. 3.4. PPI images of reflectivity (dBZ; left) and radial velocity (m s^{-1} ; right) at 0929 UTC on 12 July 2004 at a 0.8° elevation angle. Range rings are in km and azimuths are in degrees.

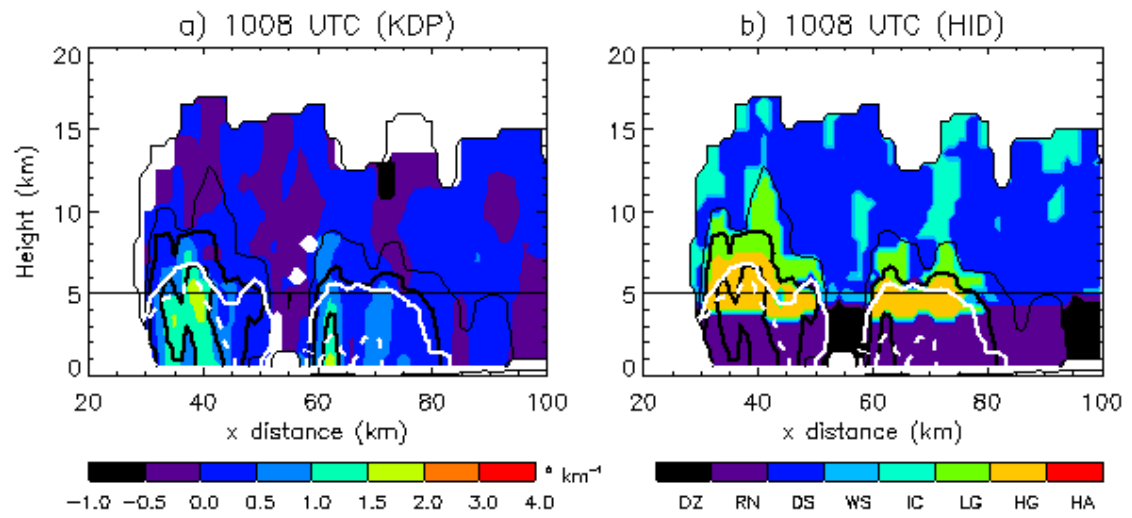


Fig. 3.5. Vertical cross sections through cells at 1008 UTC on 12 July 2004. See Fig. 3.1 for a description of the variables and contours.

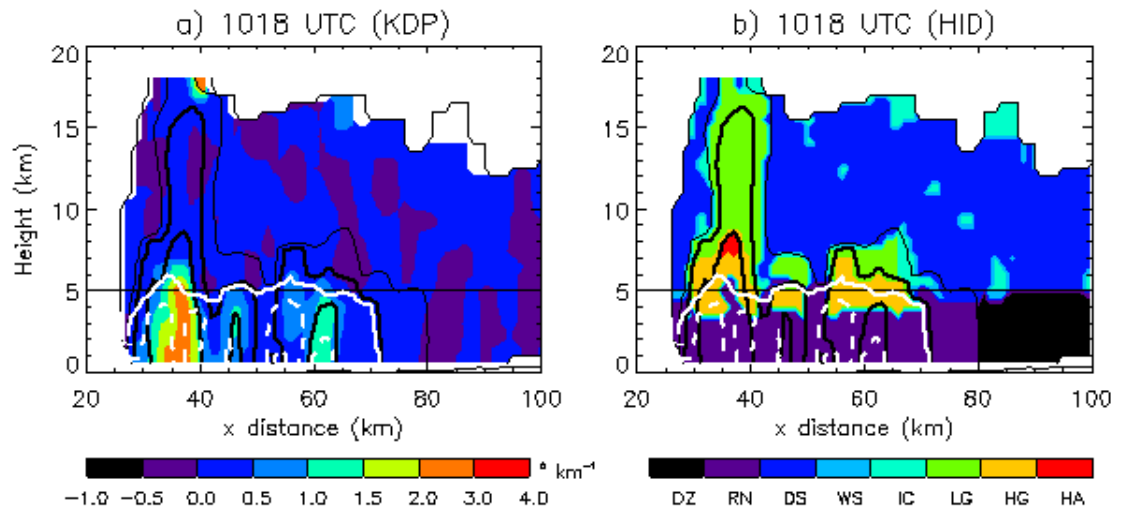


Fig. 3.6. Vertical cross sections through cells at 1018 UTC on 12 July 2004. See Fig. 3.1 for a description of the variables and contours.

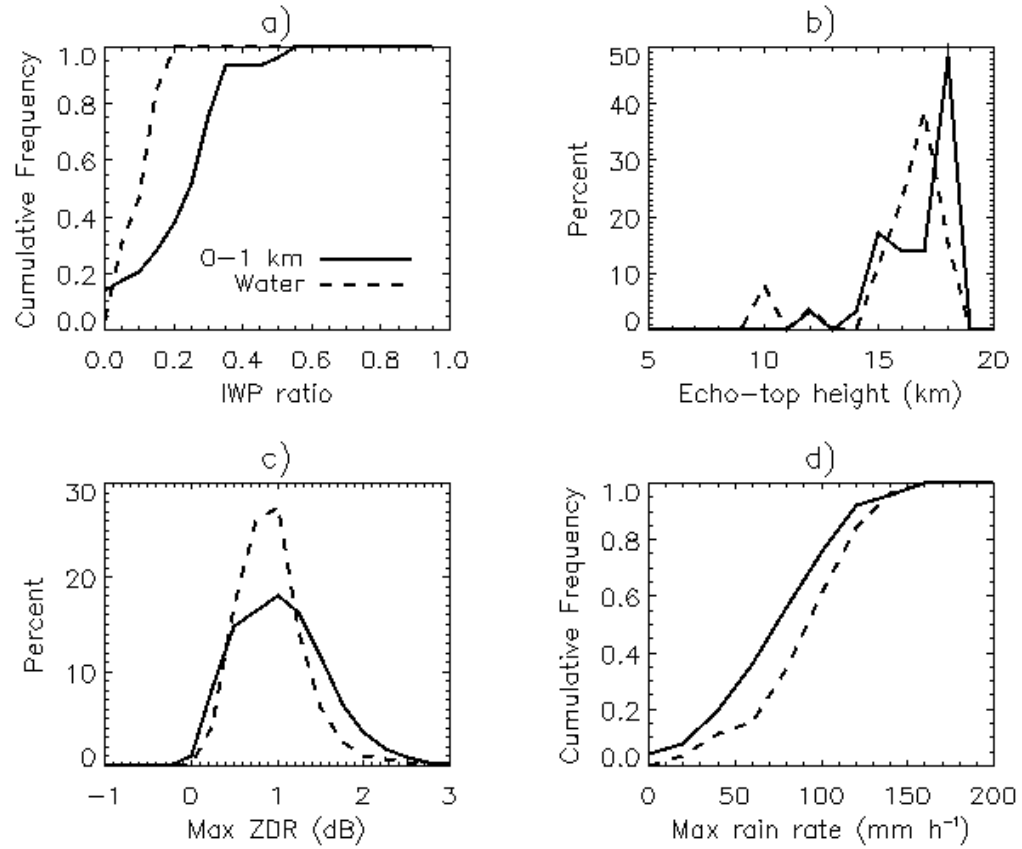


Fig. 3.7. Distributions associated with a track over the low terrain (0–1 km) and with a track over water (water) on 12 July 2004. Plotted variables represent values associated only with cells in these particular tracks.

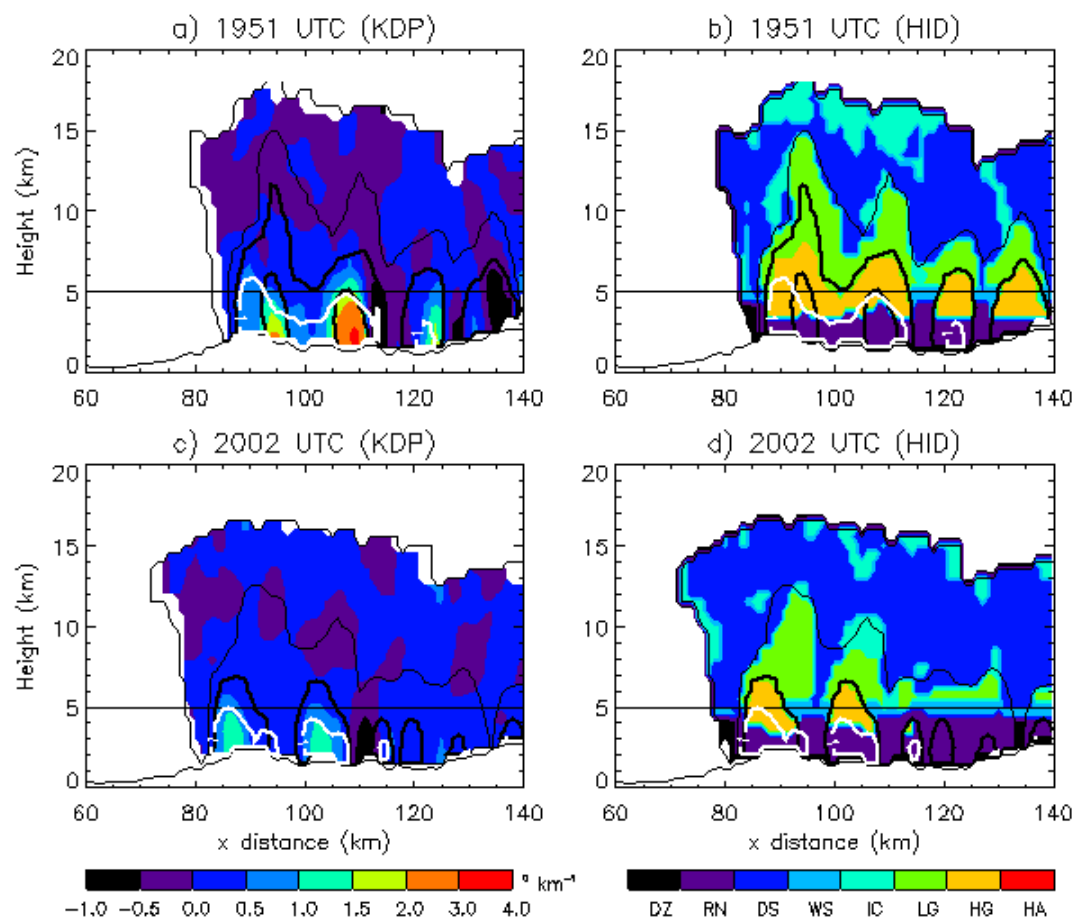


Fig. 3.8. Vertical cross sections through cells at (a),(b) 1951 and (c),(d) 2002 UTC on 5 August 2004. See Fig. 3.1 for a description of the variables and contours.

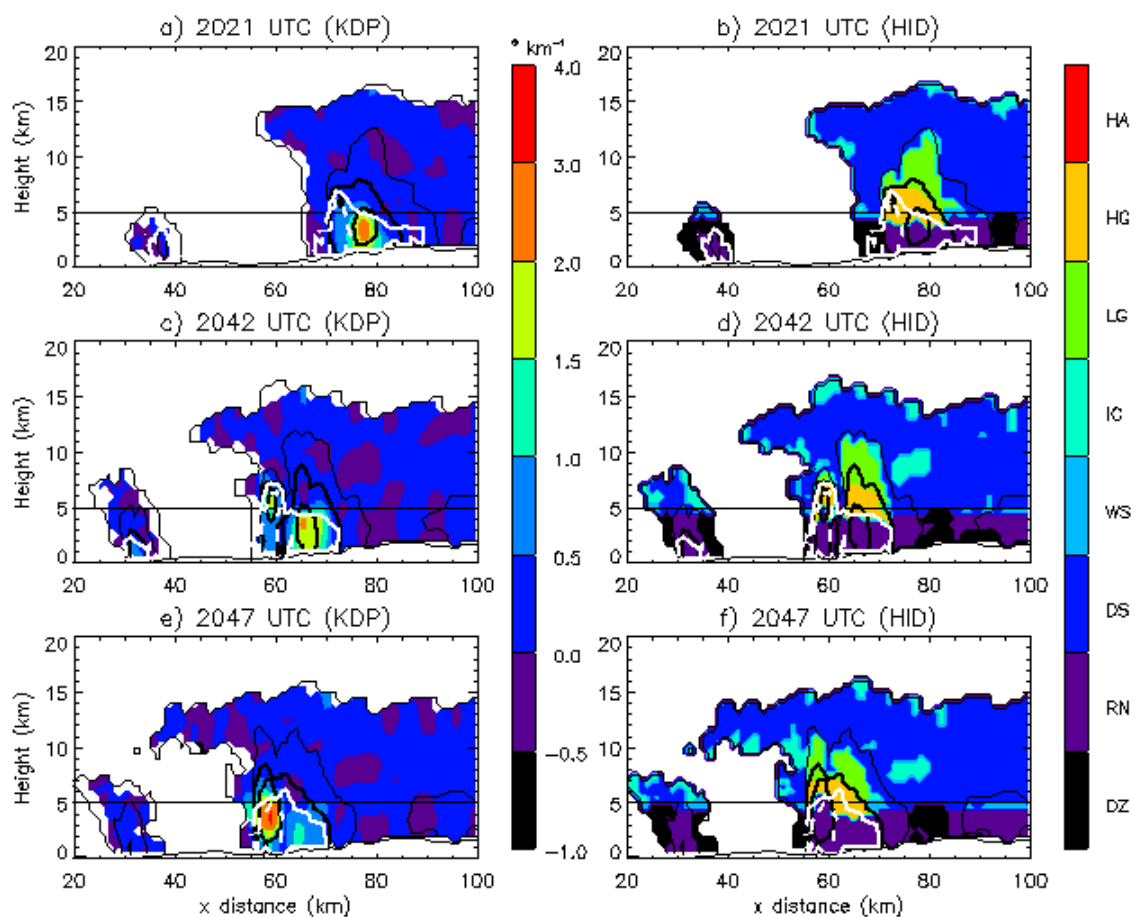


Fig. 3.9. Vertical cross sections through cells at (a),(b) 2021, (c),(d) 2042, and (e),(f) 2047 UTC on 5 August 2004. See Fig. 3.1 for a description of the variables and contours.

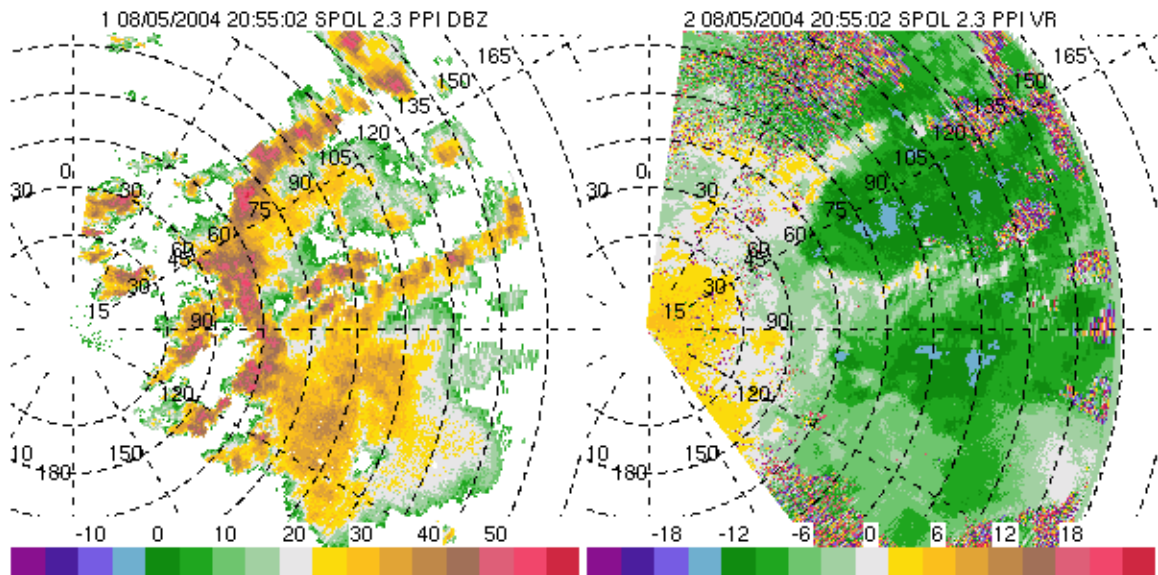


Fig. 3.10. PPI images of reflectivity (dBZ; left) and radial velocity (m s^{-1} ; right) at 2055 UTC on 5 August 2004 at a 2.3° elevation angle.

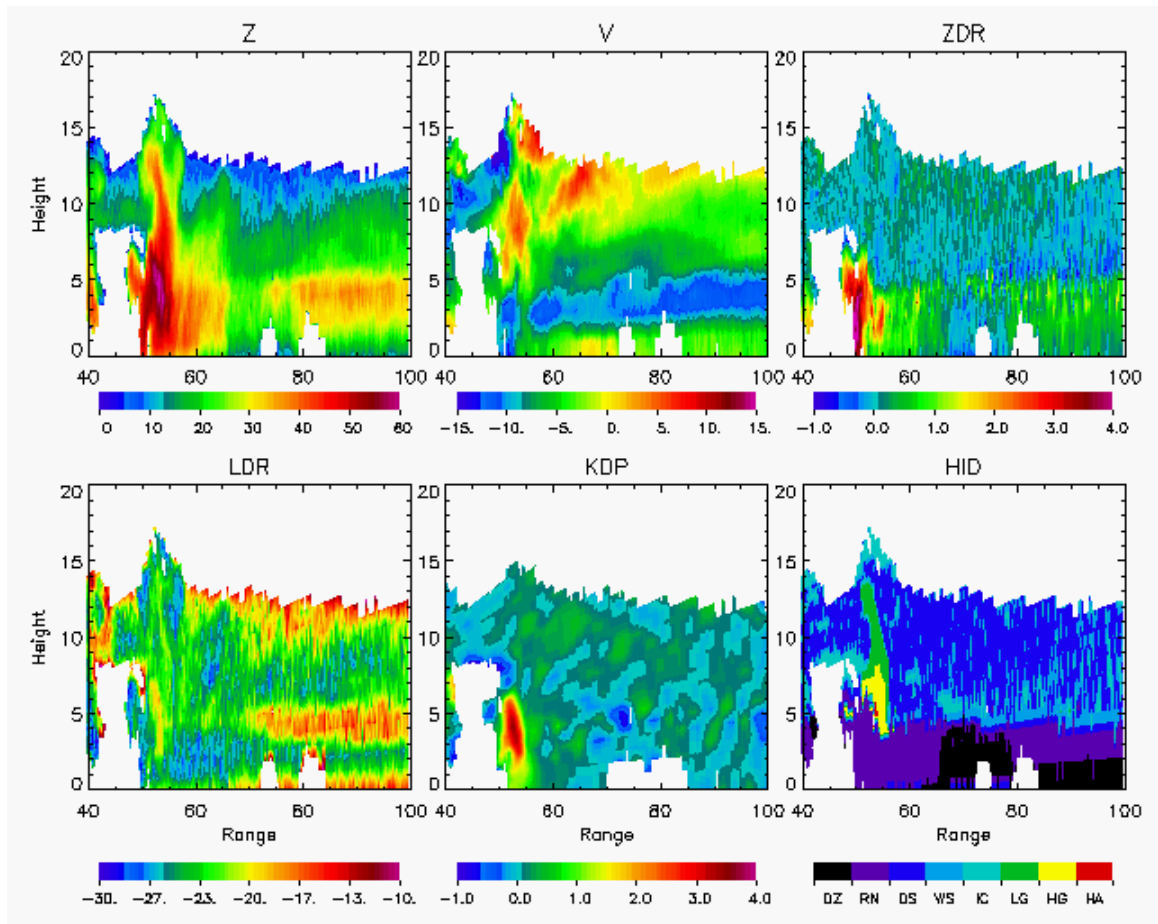


Fig. 3.11. An RHI through a cell at 2108 UTC on 5 August 2004 at an azimuth of 101.9°. Variables are as described in Fig. 3.2.

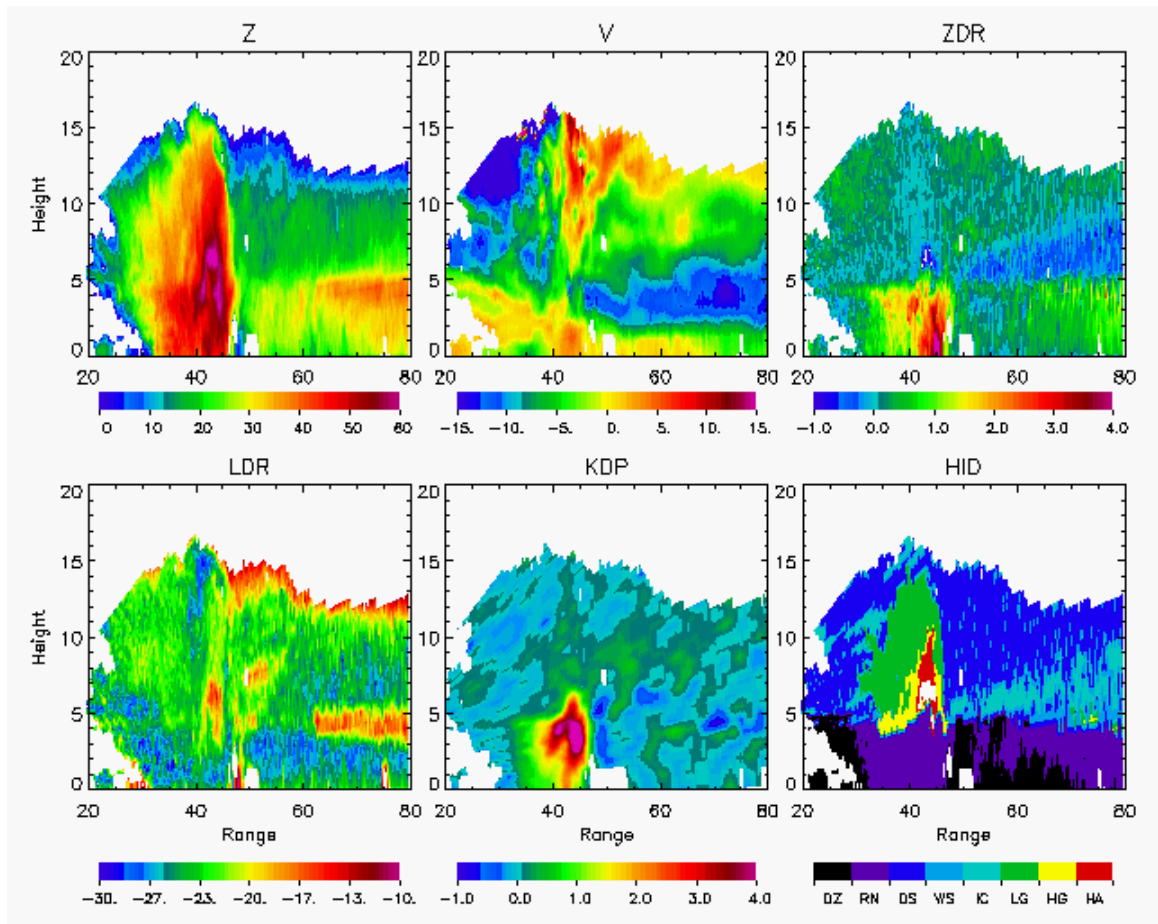


Fig. 3.12. An RHI through a cell at 2123 UTC on 5 August 2004 at an azimuth of 96.6° . Variables are as described in Fig. 3.2.

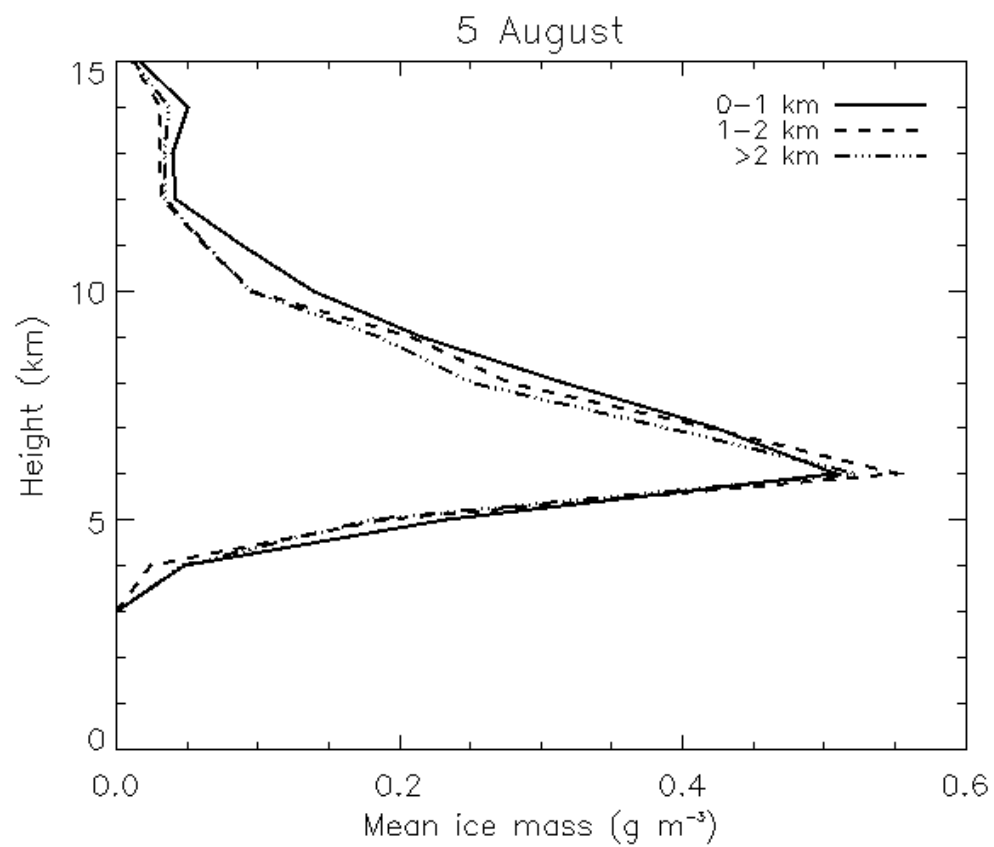


Fig. 3.13. Vertical profiles of mean ice mass as a function of terrain height (km) for all organized cells identified on 5 August.

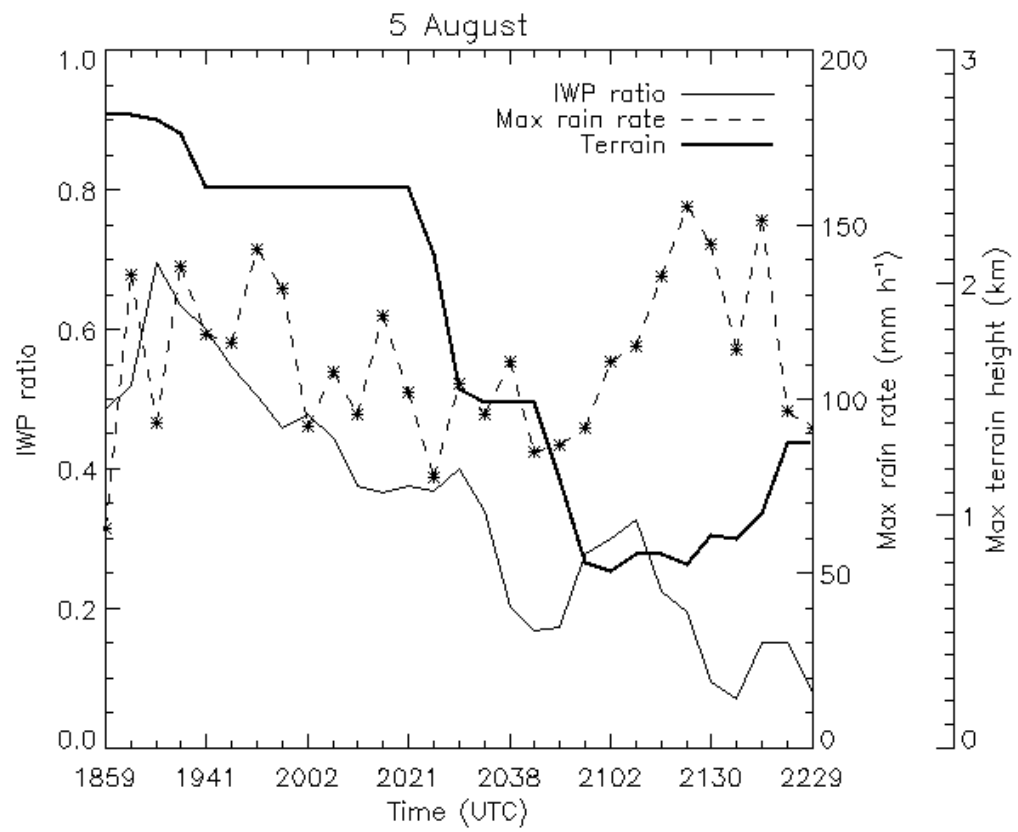


Fig. 3.14. A time series of IWP ratio and maximum rain rate (mm h^{-1}) associated with a particular track on 5 August 2004 as a function of terrain height (km).

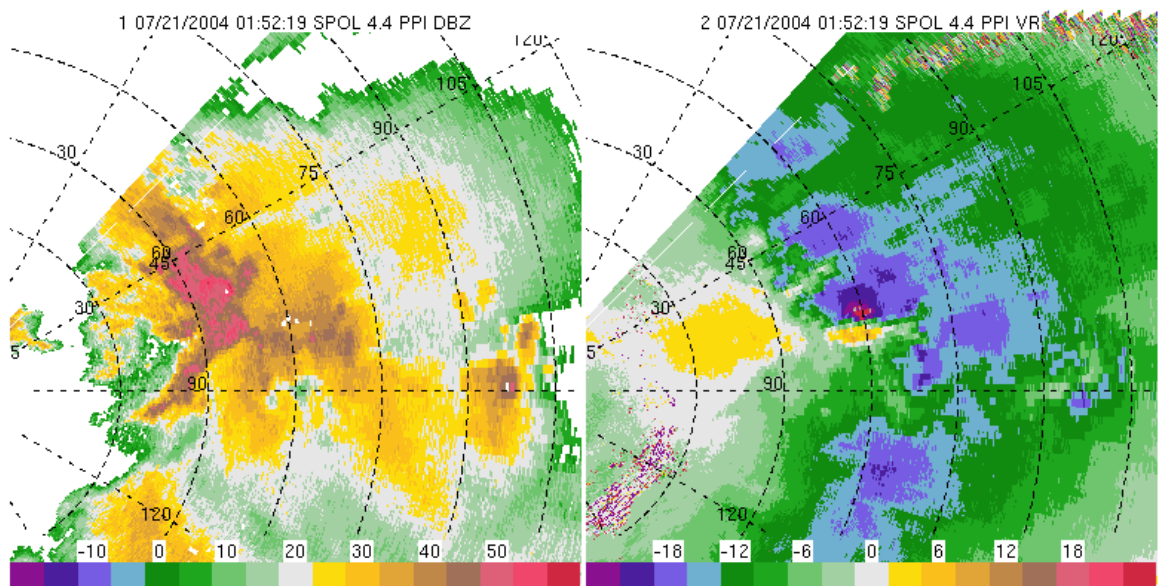


Fig. 3.15. PPI images of reflectivity (dBZ; left) and radial velocity (m s^{-1} ; right) at 0152 UTC on 21 July 2004 at a 4.4° elevation angle.

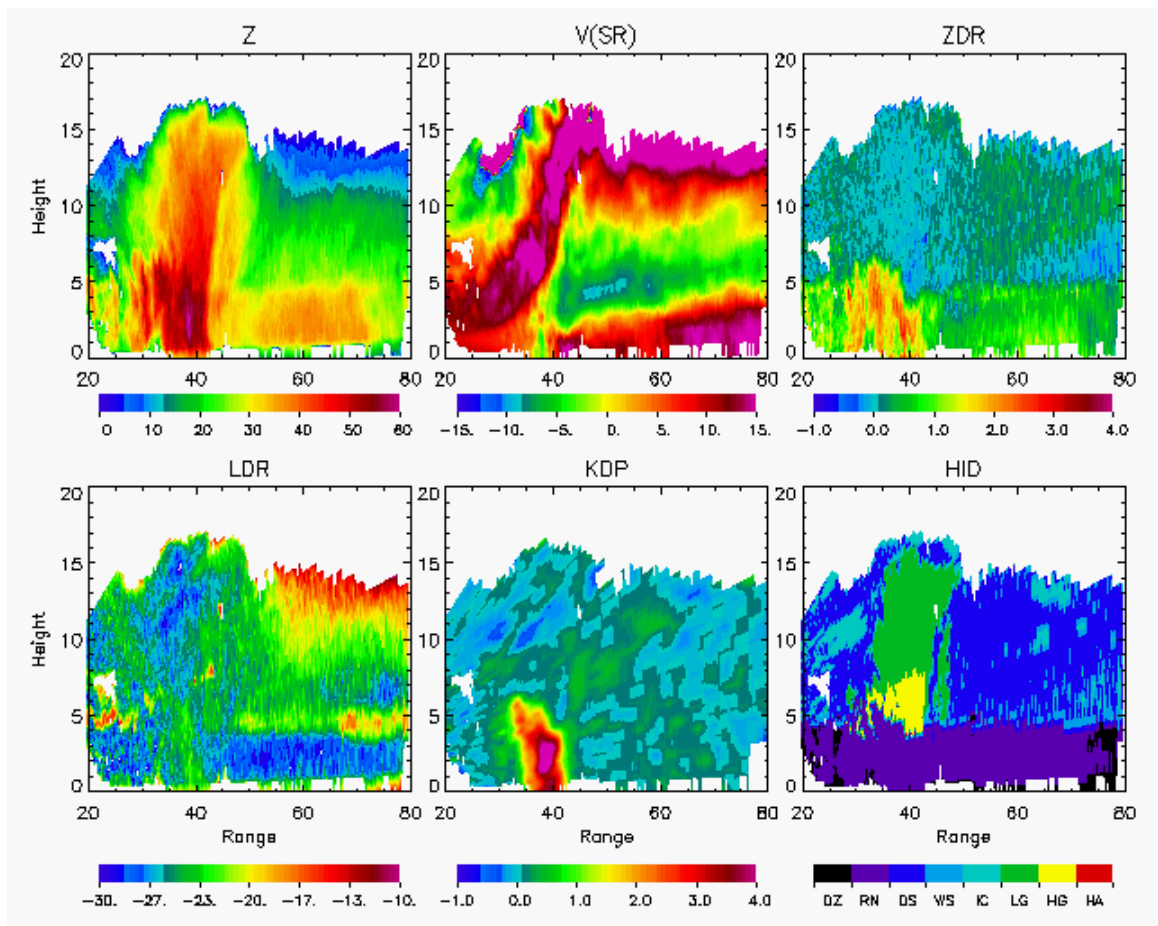


Fig. 3.16. An RHI through a cell at 0208 UTC on 21 July 2004 at an azimuth of 75°. Variables are as described in Fig. 3.2.

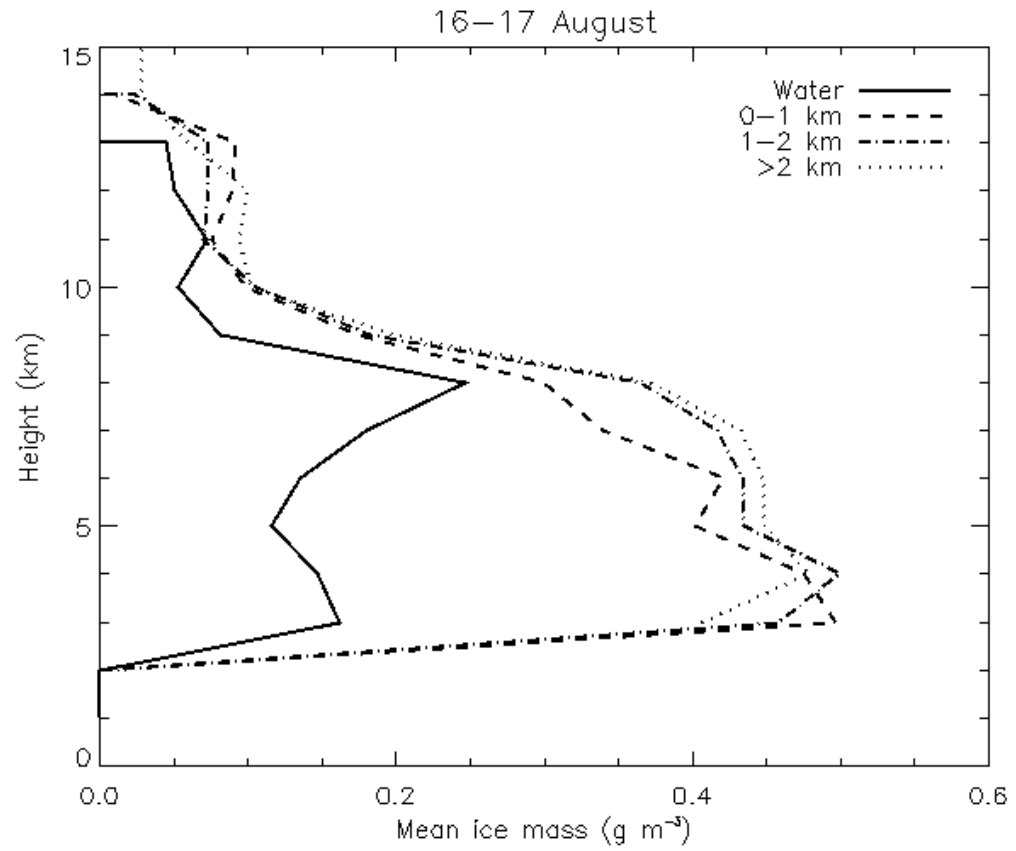


Fig. 3.17. Vertical profiles of mean ice mass as a function of terrain height (km) for organized cells identified during the 16–17 August case.

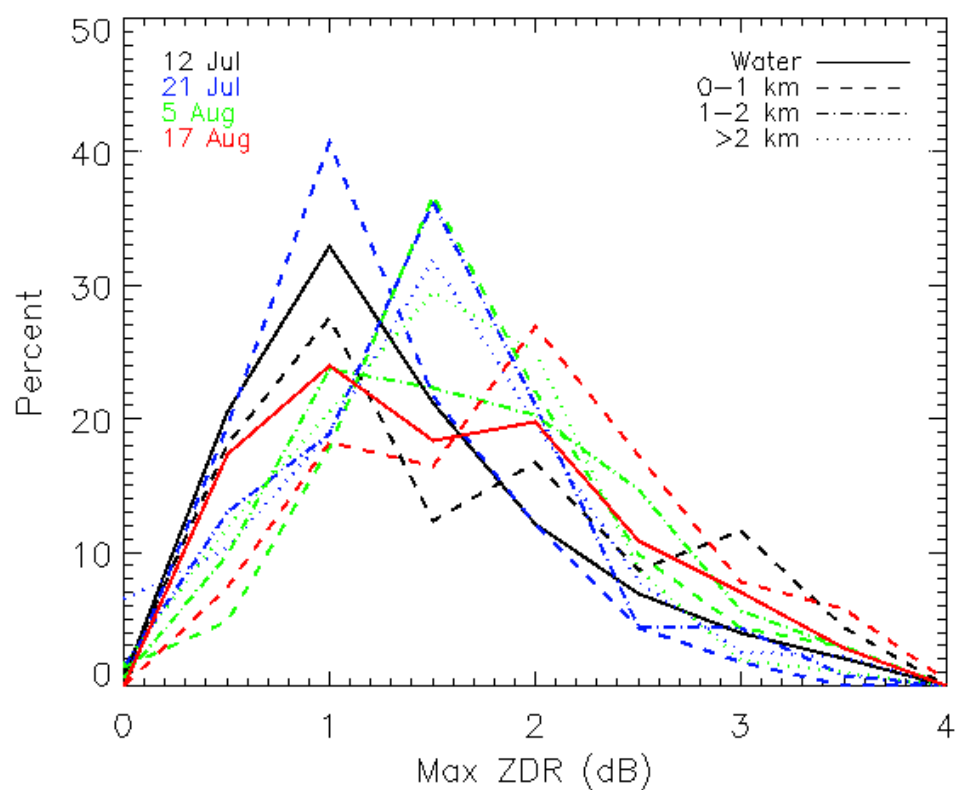


Fig. 3.18. Distributions of maximum Z_{DR} (dB), identified by the HID as rain below 4 km, for all organized cells identified during the specific case days as a function of terrain height (km).

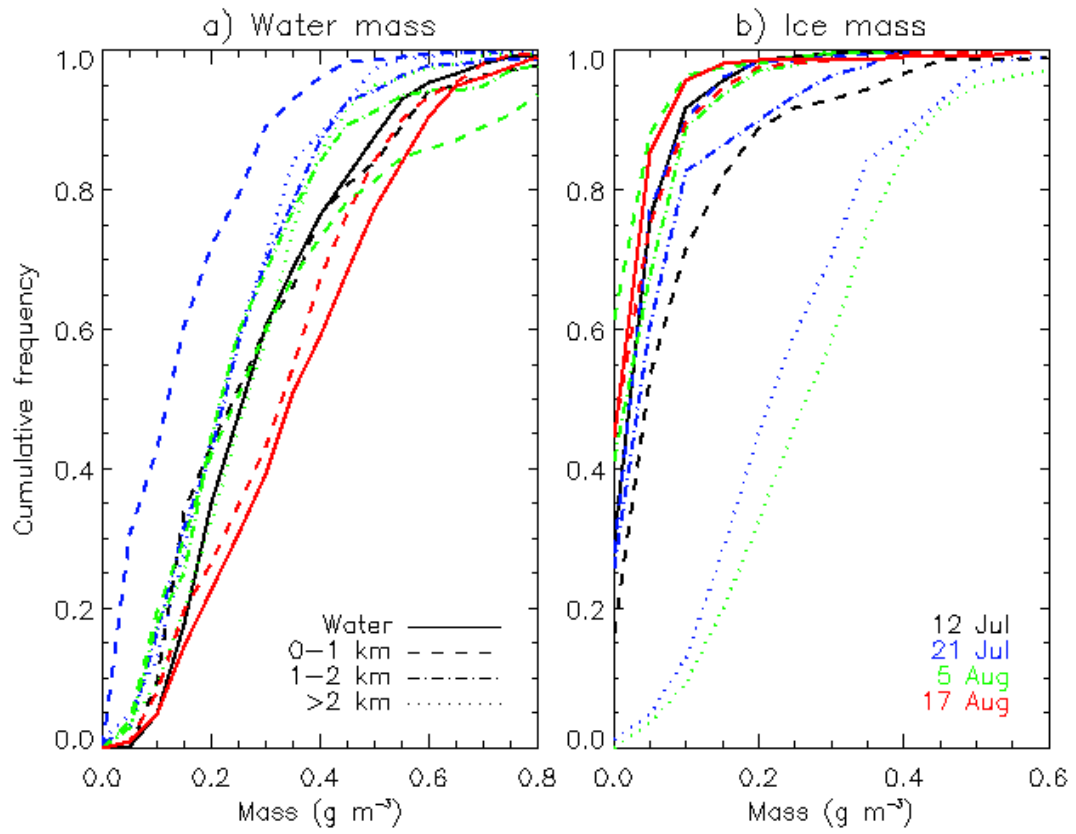


Fig. 3.19. Cumulative distribution functions of (a) liquid water mass and (b) ice mass for all organized cells identified during the specific case days as a function of terrain height (km). Liquid water and ice masses are normalized by cell volume.

CHAPTER 4

COMPARISON WITH ISOLATED CONVECTION

The previous case studies described the evolution of MCSs during NAME, comprised of deep, intense convective elements that exhibited similar characteristics to isolated convection in this region, including Z_{DR} columns, large ice hydrometeors aloft, and an elevation-dependence in microphysical processes (Rowe et al. 2011). Despite the greater frequency of isolated convection during NAME (Pereira 2008; Rowe et al. 2011), the long-lived, organized features were responsible for 75% of the total rainfall in the radar domain (Lang et al. 2007; Pereira 2008). The number of tracked, individual isolated cells was similar to the number of organized cells collected during the microphysical scans (6568 and 6908, respectively), allowing for meaningful comparisons; however, the majority of isolated cells occurred over the high terrain of the SMO (> 2 km), whereas most organized convection was located over the western slopes (1–2 km) and coastal plain (0–1 km). Therefore, to provide a more accurate comparison, isolated and organized cells were further subdivided based on elevation.

4.1 Rain rate and echo-top heights

Cumulative distribution functions of maximum rain rate (Fig. 4.1), grouped by feature type and elevation, show greater rainfall intensity for cells contained within MCSs, suggesting that both increased intensity and duration were responsible for greater

precipitation totals associated with organized systems in this region. Maximum instantaneous rain rates were associated with organized convection along the western slopes, consistent with the tendency for the most intense rainfall to occur during the late afternoon/early evening over the lower elevations of the SMO (Rowe et al. 2008). A direct relationship existed between storm depth and rain rates for isolated cells (Rowe et al. 2011); therefore, it is no surprise that echo-top heights were generally higher for organized convection regardless of terrain height (Fig. 4.2). Similar results were found using radar data from the Tropical Ocean Global Atmosphere Coupled Ocean Atmosphere Response Experiment (TOGA COARE), where mesoscale systems contributed 80% of rainfall during the experiment despite a greater frequency of isolated cells, and the tallest echo-top heights were associated with organized features (Rickenbach and Rutledge 1998). In addition, organized systems during NAME tended to be shallower over the water compared to land for both organized and isolated convection, consistent with observations presented in the previous cases (Ch. 3).

4.2 Polarimetric observations

To further investigate differences in microphysical characteristics, distributions of maximum Z_{DR} within cells were grouped by feature type and topography (Fig. 4.3). Overall, for a given range of topography, there was a tendency for isolated cells to have larger Z_{DR} values compared to organized convection. In particular, isolated cells over the lower elevations contained the largest drops and exhibited the widest distribution compared to the cells embedded within organized systems over the coastal plain and adjacent water. To further investigate this difference, frequency contours of K_{DP} as a

function of Z_{DR} are presented for isolated and organized convection over the 0–1-km elevation range (Fig. 4.4a,b). Both figures show the occurrence of Z_{DR} values exceeding 4 dB, corresponding to drop sizes near 3 mm, yet for lower Z_{DR} (i.e., 1–2 dB, drop sizes ranging from roughly 1.5–2 mm), K_{DP} values were greater for cells embedded within organized systems, indicating higher liquid water contents. This suggests that although large drops were present in both isolated and organized convection, the presence of high concentrations of smaller liquid hydrometeors in organized convection contributed significantly to the intense rainfall associated with these systems. The shift in the drop-size distribution is even more evident when comparing these frequencies between organized convection over land compared to that over water (Fig. 4.4c,d), which show that large K_{DP} was limited to lower Z_{DR} values over the GoC, characteristic of more maritime-like systems (e.g., 12 July).

In addition, organized convection over the slopes, where rain rates were maximized, was also characterized by greater Z_{DR} values compared to cells embedded within MCSs over the lower terrain and GoC (Fig. 4.3), further indicating a greater contribution from melting ice hydrometeors to rainfall. Vertical profiles of mean ice mass within a cell (Fig. 4.5a) show greater ice mass in organized cells over the higher elevations compared to the coastal plain, further implying an important role of ice-based processes in producing heavy rainfall over the terrain. In addition, MCS-convection generally contained higher ice mass compared to isolated convection over land, especially within the mixed-phase region near 6 km, with the exception of the intense, isolated cells over the SMO captured by the microphysical scans. This smaller change in ice mass with organization over the SMO, compared to other elevations, is consistent

with previous NAME studies that found little day-to-day variability in precipitating systems over the high terrain (Lang et al. 2007, 2010; Nesbitt et al. 2008). Furthermore, there was a clear trend of reduced ice mass over water compared to that over land, similar to results found for all S-Pol-detected cells during NAME (Lang et al. 2010), and as described in the case studies (Ch. 3). Vertical profiles of mean liquid water mass (Fig. 4.5b) reveal a trend in which greater amounts of liquid water were present below the melting level for MCS-convection compared to isolated cells, consistent with the abundance of small drops in organized convection, as shown in Fig. 4.4. The smallest values at low levels were associated with isolated convection over lower elevations, likely due to a larger concentration of big drops, as well as increased sub-cloud evaporation over the coastal plain (Nesbitt et al. 2008).

In general, convection in organized systems was characterized by greater rain rates, in addition to longer life spans, with taller echo-top heights and greater water and ice masses compared to isolated cells. Isolated cells had larger Z_{DR} values, especially compared to organized systems over the lower terrain and GoC, which contained larger concentrations of smaller drops and led to increased liquid water mass in the low levels. This apparent dependence of intensity on organization may also be due, at least in part, to the more favorable environmental conditions under which MCSs develop. Intense rainfall associated with organized systems occurred during periods of both enhanced CAPE and shear (Pereira 2008), consistent with studies of heavy rain-producing MCSs during TOGA-COARE (Lucas and Zipser 2000). The relative contributions of the mesoscale dynamics inherent to the organized systems and the favorable large-scale environmental

conditions to the enhanced rainfall associated with MCSs during NAME are topics beyond the scope of this study.

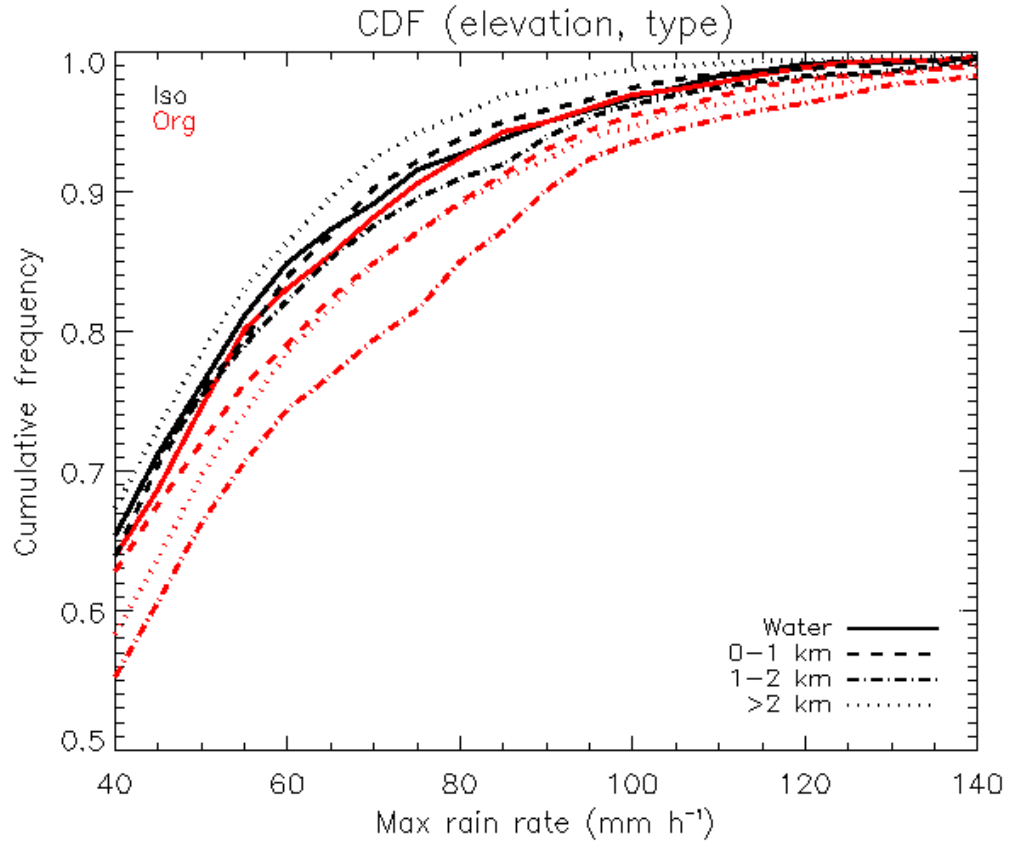


Fig. 4.1. Cumulative distribution functions of maximum rain rate (mm h^{-1}) for all cells identified within the S-Pol microphysical scans, divided by terrain height (km) and cell type, where “Iso” and “Org” represent isolated and organized cells, respectively. Note that rain rates are truncated to focus on intense values.

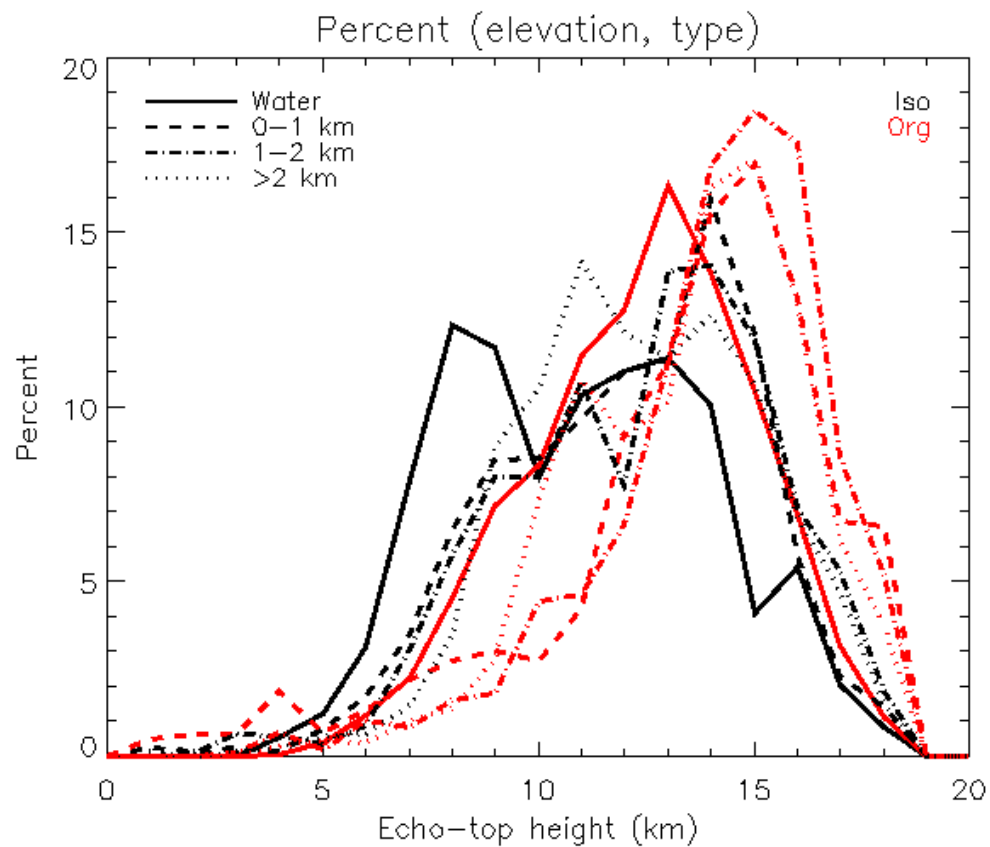


Fig. 4.2. Distributions of echo-top height (km) for all cells identified within the S-Pol microphysical scans, divided by terrain height (km) and cell type. Percent represents the percentage for the specified type and elevation group.

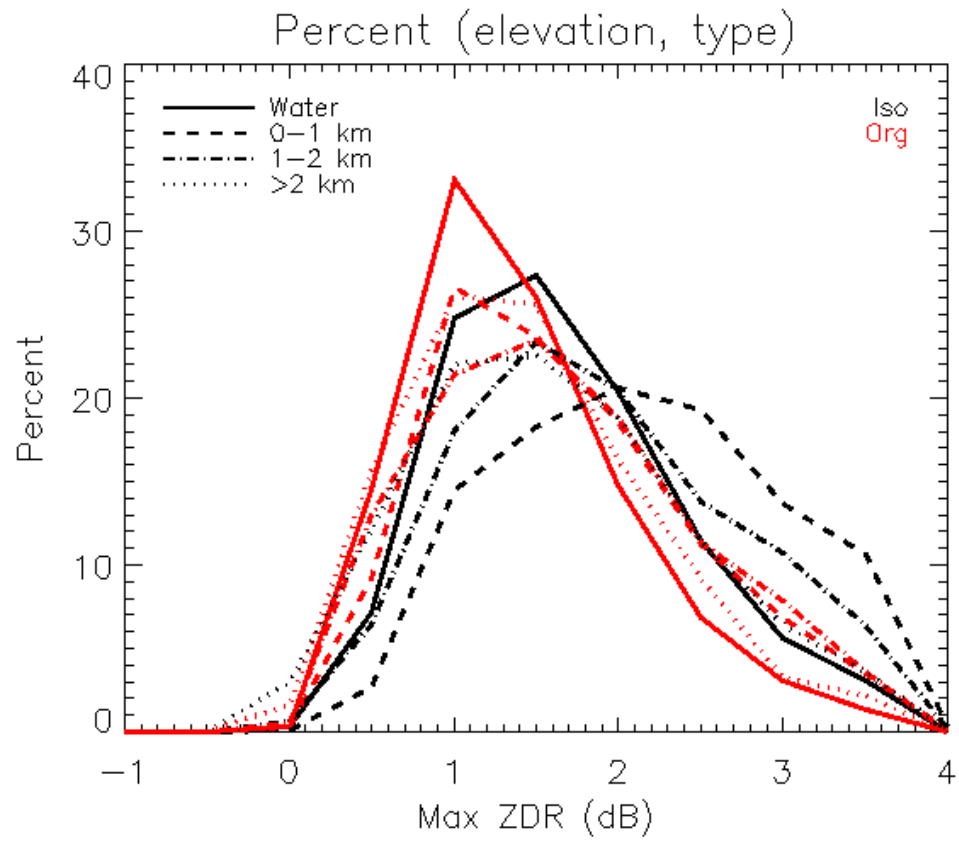


Fig. 4.3. Distributions of maximum Z_{DR} (dB) for all cells identified within the S-Pol microphysical scans, divided by terrain height (km) and cell type.

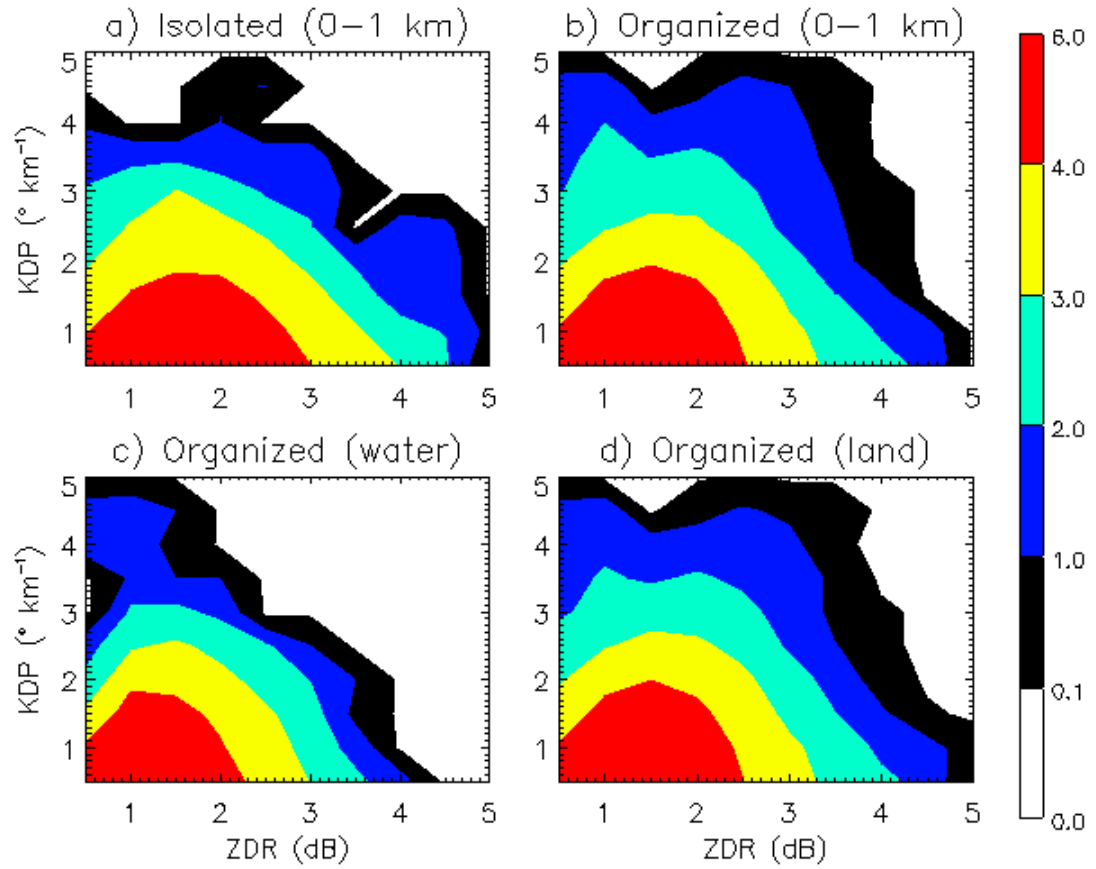


Fig. 4.4. Contoured frequencies of K_{DP} as a function of Z_{DR} for (a) isolated cells over low terrain (0–1 km), (b) organized cells over low terrain (0–1 km), (c) organized cells over water, and (d) organized cells over land (all elevations). Only values identified as rain by the HID and located below 4 km are considered. Plotted are the logarithms of frequencies normalized by the total frequency in the specified elevation grouping.

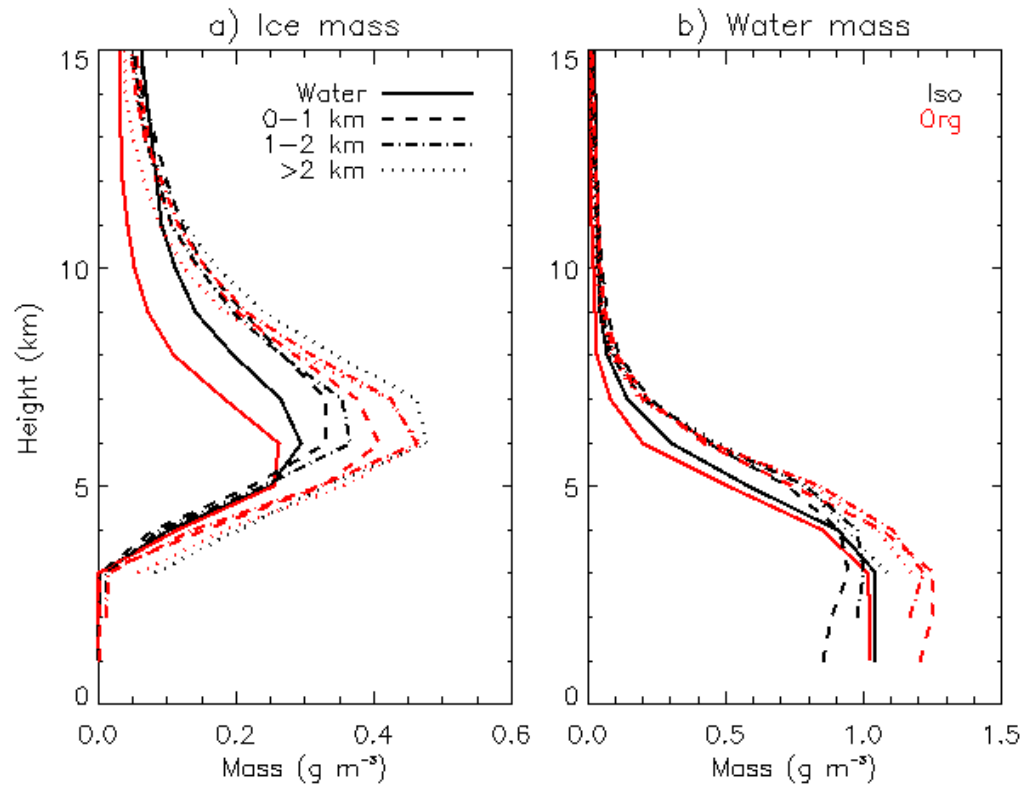


Fig. 4.5. Profiles of (a) mean ice mass and (b) mean liquid water mass for all cells within the microphysical scans, categorized by terrain height (km) and cell type.

CHAPTER 5

DISCUSSION AND CONCLUSIONS

A major objective of NAME was to describe the microphysical processes of convection relative to the significant terrain variations in that region. Previous work by Rowe et al. (2011) compared examples of isolated convection and found increased water mass and rain-water depth in convection over the coastal plain, consistent with Lang et al. (2010), suggesting an important role of warm-rain processes due to increased warm-cloud depth compared to over the high terrain of the SMO (Rowe et al. 2008). Cases over the western slopes revealed an additional dependence on ice-based processes via drop freezing and subsequent riming growth, with an even greater role of melting ice in rainfall production over the highest elevations. Despite the high frequency of occurrence of these isolated cells during NAME, MCSs were responsible for 75% of rainfall in the radar domain, suggesting an additional dependence on organization.

Compared to isolated convection, cells embedded within MCSs were characterized by larger rain rates, taller echo-top heights, and greater liquid and ice mass. The larger drops associated with isolated convection over the coastal plain were subjected to increased evaporation below cloud base (e.g., Nesbitt et al. 2008), leading to lower liquid water contents and rain rates compared to the organized systems moving off the higher terrain, which contained larger concentrations of smaller drops. This maritime-like drop-size distribution was more evident within organized convection over the GoC,

where systems moving into the domain from the south had longer residence times over water. One such example occurred on 12 July 2004, when a mature system entered the domain from the southeast, characterized by deep leading convection with trailing stratiform rain. A comparison of the portion of this system over the coastal plain with cells over water revealed reduced ice mass, shallower echo tops, and smaller drops in convection over the GoC; however, this part of the system produced greater liquid water contents and, therefore, greater maximum rain rates, suggesting a significant contribution from warm-rain processes. Greater warm-cloud depths further indicated the importance of warm-rain processes, but the significance of mixed-phase processes was also inferred due to the presence of Z_{DR} columns along the leading edge. Along with this characteristic, the collapse of older cells to maintain the trailing stratiform region highlighted similarities to MCSs in the midlatitudes (Houze 1993, 1997).

An example of convection organizing along the western slopes was presented from 5 August 2004, yielding differences in microphysical processes based on elevation. Deep cells with comparable amounts of ice mass to lower elevations developed over the SMO during the afternoon hours. As time progressed, a quasi-linear MCS organized, with a leading convective line along the western slopes and a trailing stratiform region comprised of older, dissipating cells, as was the case on 12 July. Elevated maxima in K_{DP} and Z_{DR} indicated the important role of melting ice in producing rainfall along the western slopes. This led to the development of outflow boundaries, which initiated new convection along the lower terrain, resulting in the propagation of this system toward the coast. Amounts of ice mass above the melting level remained consistent as the system

moved over the coastal plain, yet increased warm-cloud depth allowed for an additional contribution from warm-rain processes, leading to the most intense rain rates.

A similar case on 20 July 2004, characterized by convection initiating over the high terrain and organizing along the western slopes, revealed mid-level rotation in the trailing stratiform region. This MCV was related to the presence of strong rear inflow in the system, which later descended and contributed to the production of outflow ahead of the convective line. On 17 August 2004, the propagation of outflow over the GoC led to additional convective development ahead of the leading convective line along the coast, resulting in offshore propagation of the system. The new convection over water, however, was shallower, with less ice, thereby limiting further production of strong outflow and leading to the dissipation of the system by morning.

This study emphasized the importance of both warm-rain and ice processes in producing intense rainfall in organized systems during NAME. Although similarities existed between isolated cells and convection embedded within MCSs, organized cells were typically deeper and contained more ice, which then melted and contributed to the development of mesoscale convective outflow boundaries. Similar to a hypothesis presented by Nesbitt et al. (2008), it is suggested that organization was limited over the higher elevations due to the complicated interaction between cold pools and the topography. As convection organized along the western slopes, outflow boundaries spread over the lower elevations, converging with larger scale upslope flow, leading to new convective development and propagation of systems toward the GoC. Once over lower elevations, additional warm-cloud depth aided in the production of intense, long-lasting rainfall, and allowed for continued ice production, which, along with the

development of rear inflow in the trailing stratiform region, maintained and strengthened the convective outflow. Keenan and Carbone (1992) noted the importance of a spreading cold pool, produced by descending rear inflow and downdrafts, for initiation of new convection in tropical systems in northern Australia, comparable to classic tropical squall lines described during GATE (e.g., Houze 1977; Barnes and Sieckman 1984; Zipser and LeMone 1989). Similar processes were also described for systems in the Himalayas where Medina et al. (2010) noted the triggering of strong convection along the foothills, which began to grow, merge, and develop cold pools that moved toward the plains. New convection initiated over the plains as these outflow boundaries converged with the monsoonal flow, allowing for propagation of these systems off the terrain. This occurrence of organized systems upstream of and over lower elevations, observed in the Himalayas and during NAME, has also been noted in other mountainous regions, including the western Ghats (Grossman and Durran 1984), the European Alps (Houze et al. 2001), and the Pyrenees (Romero et al. 2001).

Due to the apparent importance of cold pools in the organization and propagation of MCSs in the NAME region, the understanding of hydrometeor characteristics described in this study should prove useful not only for convective parameterization schemes, but also for the simulation of cold pool dynamics. The interaction of these cold pools with environmental conditions will likely affect the organization of these systems (e.g., RKW theory, Rotunno et al. 1988), which are crucial to properly simulate the MCS lifecycle and feedbacks to the monsoon circulation via heat and momentum transport.

REFERENCES

- Atlas, D., and C. R. Williams, 2000: An observationally based conceptual model of warm oceanic convective rain in the Tropics. *J. Appl. Meteor.*, **39**, 2165–2181.
- Atlas, D., and C. W. Ulbrich, 2006: Drop size spectra and integral remote sensing parameters in the transition from convective to stratiform rain. *Geophys. Res. Lett.*, **33**, L16803, doi:10.1029/2006GL026824.
- Barnes, G. M., and K. Sieckman, 1984: The environment of fast- and slow-moving tropical mesoscale convective cloud lines. *Mon. Wea. Rev.*, **112**, 1782–1794.
- Bartels, D. L., and R. A. Maddox, 1991: Midlevel cyclonic vortices generated by mesoscale convective systems. *Mon. Wea. Rev.*, **119**, 104–118.
- Bluestein, H. B., and M. H. Jain, 1985: Formation of mesoscale lines of precipitation: Severe squall lines in Oklahoma during the spring. *J. Atmos. Sci.*, **59**, 1711–1732.
- Bringi, V. N., and V. Chandrasekar, 2001: *Polarimetric Doppler Weather Radar Principles and Applications*. Cambridge University Press, 636 pp.
- Bringi, V. N., L. Liu, P. C. Kennedy, V. Chandrasekar, and S. A. Rutledge, 1996: Dual multiparameter radar observations of intense convective storms: The 24 June 1992 case study. *Meteor. Atmos. Phys.*, **59**, 3–31.
- Bringi, V. N., K. Knupp, A. Detwiler, L. Liu, I. J. Caylor, and R. A. Black, 1997: Evolution of a Florida thunderstorm during the Convection and Precipitation/Electrification Experiment: The case of 9 August 1991. *Mon. Wea. Rev.*, **125**, 2131–2160.
- Bringi, V. N., V. Chandrasekar, J. Hubbert, E. Gorgucci, W. L. Randeu, and M. Schoenhuber, 2003: Raindrop size distribution in different climatic regimes with disdrometer and dual-polarized radar analysis. *J. Atmos. Sci.*, **60**, 354–365.
- Bringi, V. N., C. R. Williams, M. Thurai, and P. T. May, 2009: Using dual-polarized radar and dual-frequency profiler for DSD characteristics: A case study from Darwin, Australia. *J. Atmos. Oceanic Technol.*, **26**, 2107–2122.

Carey, L. D., and S. A. Rutledge, 2000: The relationship between precipitation and lightning in tropical island convection: A C-band polarimetric radar study. *Mon. Wea. Rev.*, **128**, 2687–2710.

Chong, M., and O. Bousquet, 1999: A mesovortex within a near-equatorial mesoscale convective system during TOGA COARE. *Mon. Wea. Rev.*, **127**, 1145–1156.

Cifelli, R., W. A. Petersen, L. D. Carey, S. A. Rutledge, and M. A. F. Silva-Dias, 2002: Radar observations of the kinematic, microphysical, and precipitation characteristics of two MCSs in TRMM LBA. *J. Geophys. Res.*, **107**, 8077, doi:10.1029/2000JD000264.

Cifelli, R., V. Chandrasekar, S. Lim, P. C. Kennedy, Y. Wang, and S. A. Rutledge, 2011: A new dual-polarization radar rainfall algorithm: Application in Colorado precipitation events. *J. Atmos. Oceanic Technol.*, **28**, 352–364.

Dai, A. G., F. Giorgi, and K. E. Trenberth, 1999: Observed and model-simulated diurnal cycles of precipitation over the contiguous United States. *J. Geophys. Res.*, **104**, 6377–6402.

Douglas, M., R. A. Maddox, K. Howard, and S. Reyes, 1993: The Mexican monsoon. *J. Climate*, **6**, 1665–1677.

Ecklund, W. L., C. R. Williams, P. E. Johnston, and K. S. Gage, 1999: A 3-GHz profiler for precipitating cloud studies. *J. Atmos. Oceanic Technol.*, **16**, 309–322.

Finch, Z. O., and R. H. Johnson, 2010: Observational analysis of an upper-level inverted trough during the 2004 North American Monsoon Experiment. *Mon. Wea. Rev.*, **138**, 3540–3555.

Fulton, R., and G. M. Heymsfield, 1991: Microphysical and radiative characteristics of convective clouds during COHMEX. *J. App. Meteor.*, **30**, 98–116.

Gamache, J. F., and R. A. Houze, Jr., 1982: Mesoscale air motions associated with a tropical squall line. *Mon. Wea. Rev.*, **110**, 118–135.

Gochis, D. J., W. J. Shuttleworth, and Z. L. Yang, 2002: Sensitivity of the modeled North American monsoon regional climate to convective parameterization. *Mon. Wea. Rev.*, **130**, 1282–1298.

Gochis, D. J., A. Jimenez, C. J. Watts, J. Garatuza-Payan, and W. J. Shuttleworth, 2004: Analysis of 2002 and 2003 warm-season precipitation from the North American Monsoon Experiment Event Rain Gauge Network. *Mon. Wea. Rev.*, **132**, 2938–2953.

Gochis, D. J., C. J. Watts, J. Garatuza-Payan, and J. Cesar-Rodriguez, 2007: Spatial and temporal patterns of precipitation intensity as observed by the NAME Event Rain Gauge Network from 2002 to 2004. *J. Climate*, **20**, 1734–1750.

Grossman, R. L., and D. R. Durran, 1984: Interaction of low-level flow with the Western Ghat mountains and offshore convection in the summer monsoon. *Mon. Wea. Rev.*, **112**, 652–672.

Higgins, R. W., and Coauthors, 2006: The NAME 2004 field campaign and modeling strategy. *Bull. Amer. Meteor. Soc.*, **87**, 79–94.

Holler, H., M. Hagen, P. F. Meischner, V. N. Bringi, and J. Hubbert, 1994: Life cycle and precipitation formation in a hybrid-type hailstorm revealed by polarimetric and Doppler radar measurements. *J. Atmos. Sci.*, **51**, 2500–2522.

Houze, R. A., Jr., 1977: Structure and dynamics of a tropical squall-line system. *Mon. Wea. Rev.*, **105**, 1540–1567.

Houze, R. A., Jr., 1993: *Cloud Dynamics*. Academic Press, 573 pp.

Houze, R. A., Jr., 1997: Stratiform precipitation in regions of convection: A meteorological paradox? *Bull. Amer. Meteor. Soc.*, **78**, 2179–2196.

Houze, R. A., Jr., 2004: Mesoscale convective systems. *Rev. Geophys.*, **42**, 0.1029/2004RG000150, 43 pp.

Houze, R. A., Jr., M. I. Biggerstaff, S. A. Rutledge, and B. F. Smull, 1989: Interpretation of Doppler weather radar displays of midlatitude mesoscale convective systems. *Bull. Amer. Meteor. Soc.*, **70**, 608–619.

Houze, R. A., Jr., C. N. James, and S. Medina, 2001: Radar observations of precipitation and airflow on the Mediterranean side of the Alps: Autumn 1998 and 1999. *Q. J. R. Meteorol. Soc.*, **127**, 2537–2558.

Illingworth A. J., J. W. F. Goddard, and S. M. Cherry, 1987: Polarization radar studies of precipitation development in convective storms. *Quart. J. Roy. Meteor. Soc.*, **113**, 469–489.

Jameson, A., 1983: Microphysical interpretation of multi-parameter radar measurements in rain. Part I: Interpretation of polarization measurements and estimation of raindrop shapes. *J. Atmos. Sci.*, **40**, 1792–1802.

Jameson, A., M. Murphy, and E. Krider, 1996: Multiple-parameter radar observations of isolated Florida thunderstorms during the onset of electrification. *J. Appl. Meteor.*, **35**, 343–354.

Johnson, R. H., S. Chen, and J. J. Toth, 1989: Circulations associated with a mature-to-decaying midlatitude mesoscale convective systems. Part I: Surface features—heat bursts and mesolow development. *Mon. Wea. Rev.*, **117**, 942–959.

Johnson, R. H., P. E. Cielsielski, T. S. L'Ecuyer, and A. J. Newman, 2010: Diurnal cycle of convection during the 2004 North American Monsoon Experiment. *J. Climate*, **23**, 1060–1078.

Johnston, E. C., 1981: Mesoscale vorticity centers induced by mesoscale convective complexes. Master's Thesis, University of Wisconsin, 54 pp.

Keenan, T. D., and R. E. Carbone, 1992: A preliminary morphology of precipitation systems in tropical northern Australia. *Q. J. R. Meteorol. Soc.*, **118**, 283–326.

Keenan, T. D., and S. A. Rutledge, 1993: Mesoscale characteristics of monsoonal convection and associated stratiform precipitation. *Mon. Wea. Rev.*, **121**, 352–374.

Keenan, T. D., and Coauthors, 2000: The Maritime Continent Thunderstorm Experiment (MCTEX): Overview and some results. *Bull. Amer. Meteor. Soc.*, **81**, 2433–2455.

Knupp, K. R., and W. R. Cotton, 1985: Convective cloud downdraft structure: An interpretive survey. *Rev. Geophys.*, **23**, 183–215.

Lang, T. J., D. A. Ahijevych, S. W. Nesbitt, R. E. Carbone, S. A. Rutledge, and R. Cifelli, 2007: Radar-observed characteristics of precipitating systems during NAME 2004. *J. Climate*, **20**, 1713–1733.

Lang, T. J., S. W. Nesbitt, and L. D. Carey, 2009: On the Correction of Partial Beam Blockage in Polarimetric Radar Data. *J. Atmos. Oceanic Technol.*, **26**, 943–957.

Lang, T. J., S. A. Rutledge, and R. Cifelli, 2010: Polarimetric radar observations of convection in northwestern Mexico during the North American Monsoon Experiment. *J. Hydrometeor.*, **11**, 1345–1357.

Lerach, D. G., S. A. Rutledge, C. R. Williams, and R. Cifelli, 2010: Vertical structure of convective systems during NAME 2004. *Mon. Wea. Rev.*, **138**, 1695–1714.

Liu, C., M. W. Moncrieff, and E. J. Zipser, 1997: Dynamical influence of microphysics in tropical squall lines: A numerical study. *Mon. Wea. Rev.*, **125**, 2193–2210.

Liu, H., and V. Chandrasekar, 2000: Classification of hydrometeors based on polarimetric radar measurements: Development of fuzzy logic and neuro-fuzzy systems, and in situ verification. *J. Atmos. Oceanic Technol.*, **17**, 140–164.

Lucas, C., and E. J. Zipser, 2000: Environmental variability during TOGA COARE. *J. Atmos. Sci.*, **57**, 2333–2350.

May, P. T., and T. D. Keenan, 2005: Evaluation of microphysical retrievals from polarimetric radar with wind profiler data. *J. Appl. Meteor.*, **44**, 827–838.

- Medina, S., R. A. Houze, Jr., A. Kumar, and D. Niyogi, 2010: Summer monsoon convection in the Himalayan region: Terrain and land cover effects. *Q. J. R. Meteorol. Soc.*, **136**, 593–616.
- Menard, R. D., and J. M. Fritsch, 1989: A mesoscale convective complex-generated inertially stable warm core vortex. *Mon. Wea. Rev.*, **117**, 1237–1261.
- Mohr, C. G., L. J. Miller, R. L. Vaughan, and H. W. Frank, 1986: The merger of mesoscale datasets into a common Cartesian format for efficient and systematic analyses. *J. Atmos. Oceanic Technol.*, **3**, 143–161.
- Nesbitt, S. W., R. Cifelli, and S. A. Rutledge, 2006: Storm morphology and rainfall characteristics of TRMM precipitation features. *Mon. Wea. Rev.*, **134**, 2702–2721.
- Nesbitt, S. W., D. J. Gochis, and T. J. Lang, 2008: The diurnal cycle of clouds and precipitation along the Sierra Madre Occidental observed during NAME-2004: Implications for warm season precipitation estimation in complex terrain. *J. Hydrometeor.*, **9**, 728–743.
- Pereira, L. G., 2008: Characteristics and organization of precipitating features during NAME 2004 and their relationship to environmental conditions. Ph.D. dissertation, Atmospheric Science, Colorado State University, Fort Collins, Colorado, 229 pp.
- Rickenbach, T. M., and S. A. Rutledge, 1998: Convection in TOGA COARE: Horizontal scale, morphology, and rainfall production. *J. Atmos. Sci.*, **55**, 2715–2729.
- Romero, R., C. A. Doswell III, and R. Riosalido, 2001: Observations and fine-grid simulations of a convective outbreak in northeastern Spain: Importance of diurnal forcing and convective cold pools. *Mon. Wea. Rev.*, **129**, 2157–2182.
- Rosenfeld, D., and C. W. Ulbrich, 2003: Cloud microphysical properties, processes, and rainfall estimation opportunities. *Meteor. Monogr.*, **30**, 237.
- Rotunno, R., J. B. Klemp, and M. L. Weisman, 1988: A theory for strong, long-lived squall lines. *J. Atmos. Sci.*, **45**, 463–485.
- Rowe, A. K., S. A. Rutledge, T. J. Lang, P. E. Ciesielski, and S. M. Saleeby, 2008: Elevation-dependent trends in precipitation observed during NAME. *Mon. Wea. Rev.*, **136**, 4962–4979.
- Rowe, A. K., S. A. Rutledge, and T. J. Lang, 2011: Investigation of microphysical processes occurring in isolated convection during NAME. *Mon. Wea. Rev.*, **139**, 424–443.

- Scott, J. D., and S. A. Rutledge, 1995: Doppler radar observations of an asymmetric mesoscale convective system and associated vortex couplet. *Mon. Wea. Rev.*, **123**, 3437–3457.
- Smull, B. F., and R. A. Houze, Jr., 1985: A midlatitude squall line with a trailing region of stratiform rain: Radar and satellite observations. *Mon. Wea. Rev.*, **113**, 117–133.
- Smull, B. F., and R. A. Houze, Jr., 1987: Rear inflow in squall lines with trailing stratiform precipitation. *Mon. Wea. Rev.*, **115**, 2869–2889.
- Takahashi, T., 1990: Near absence of lightning in torrential rainfall producing Micronesian thunderstorms. *Geophys. Res. Lett.*, **17**, 2381–2384.
- Takahashi, T., and K. Kuhara, 1993: Precipitation mechanisms of cumulonimbus clouds at Pohnpei, Micronesia. *J. Meteor. Soc.*, **71**, 21–31.
- Tessendorf, S. A., L. J. Miller, K. C. Wiens, and S. A. Rutledge, 2005: The 29 June 2000 supercell observed during STEPS. Part I: Kinematics and microphysics. *J. Atmos. Sci.*, **62**, 4127–4150.
- Ulbrich, C. W., and D. Atlas, 2007: Microphysics of raindrop size spectra: Tropical continental and maritime storms. *J. Appl. Meteor. Climatol.*, **46**, 1777–1791.
- Williams, C. R., W. L. Ecklund, and K. S. Gage, 1995: Classification of precipitating clouds in the Tropics using 915-MHz wind profilers. *J. Atmos. Oceanic Technol.*, **12**, 996–1012.
- Zhang, D.-L., and J. M. Fritsch, 1987: Numerical simulations of the meso- β scale structure and evolution of the 1977 Johnstown flood. Part II: Inertially stable warm-core vortex and the mesoscale convective complex. *J. Atmos. Sci.*, **44**, 2593–2612.
- Zipser, E. J., and M. A. LeMone, 1989: Cumulonimbus vertical velocity events in GATE. Part II: Synthesis and model core structure. *J. Atmos. Sci.*, **37**, 2458–2469.

**PART III: A STUDY OF MICROPHYSICAL PROCESSES OCCURRING IN
CONVECTION DURING TIMREX**

CHAPTER 1

INTRODUCTION

During May and June, episodic, extreme rainfall events occur with a high degree of regularity in Taiwan. Typically, these events are associated with mesoscale convective systems (MCSs) embedded within the quasi-stationary Mei-Yu front, an important component of the East Asian monsoon (e.g., Chen and Chang 1980; Tao 1980). During the Mei-Yu season, these disturbances are steered onshore by the southwesterly environmental flow (e.g., Kuo and Chen 1990), producing localized heavy rainfall exceeding 200 mm d^{-1} along the southwestern slopes of the Central Mountain Range (CMR) resulting in loss of life, property, and agricultural production due to flooding and landslides (Chen et al. 1999). These events are dependent on the annual variability in the onset and strength of the monsoon system, and prediction of the distribution and intensity of precipitation is further complicated by the complex local topography. For example, Johnson and Bresch (1991) found an early afternoon rainfall maximum occurring along the western foothills of the CMR. Furthermore, Yeh and Chen (1998) showed that more than 50% of the rainfall during one particular monsoon season was associated with diurnally driven orographic precipitation, suggesting that local mountain circulations, in addition to large-scale flow impinging on the terrain, can influence the timing, location, and intensity of the monsoon precipitation. An understanding of the upstream environmental conditions, diurnal variability, and local microphysical processes is

therefore required to better predict heavy rainfall in Taiwan, especially in vulnerable areas along the steep topography.

To improve understanding of the mesoscale-dynamic and microphysical processes responsible for heavy monsoonal precipitation, the Taiwan Area Mesoscale Experiment (TAMEX) was conducted in northern Taiwan during May and June of 1987 (Kuo and Chen 1990). Results from this experiment have provided insights into the characteristics of squall lines (e.g., Wang et al. 1990; Chen and Chou 1993), the influence of the low-level jet on precipitation (e.g., Jou and Deng 1992), and large-scale environmental conditions leading to heavy rainfall (e.g., Chen and Li 1995). Despite the valuable information gained from this project, TAMEX was necessarily limited to the northern part of Taiwan, whereas it has been shown that extreme rainfall events occur most frequently in southwestern Taiwan (e.g., Chen et al. 1999). The Terrain-influenced Monsoon Rainfall Experiment (TiMREX), during May and June of 2008, sought to study monsoon precipitation in this region using a suite of ground-based and upper-air instrumentation, including rain gauges, wind profilers, rawinsondes, aircraft, a research vessel, and a network of radars. This project was partly designed to address scientific questions left unanswered from previous field experiments (e.g., TAMEX), in order to improve understanding of terrain-influenced, warm-season precipitation in this region.

Previous studies from other monsoon regions have described a strong influence of topography on precipitation. For example, during the South Asian monsoon, moisture from the Arabian Sea and Bay of Bengal flows over the hot continent, leading to some of the deepest and most intense convection in the world (Zipser et al. 2006). Recent studies have related the location of this extreme rainfall to the regional topography (e.g., Hirose

and Nakamura 2002; Xie et al. 2006; Houze et al. 2007; Romatschke et al. 2010). More specifically, Houze et al. (2007) suggested that convective systems containing the most intense radar echo tend to occur in the western indentation of the Himalayas due to orographic lifting of potentially unstable, low-level flow. Similarly, moist, onshore flow during the North American Monsoon (NAM) in northwestern Mexico leads to frequent convective initiation over the high terrain of the Sierra Madre Occidental (SMO) during the afternoon (e.g., Gochis et al. 2004, 2007; Lang et al. 2007). In contrast to the mesoscale systems moving onshore and impinging on the mountain barrier in Taiwan, convection over the SMO organizes while propagating westward toward the coastal plain, leading to a maximum in rain rates over the lower elevations during the late-evening/early-morning hours (Lang et al. 2007). Using data from the National Center for Atmospheric Research's (NCAR) S-band, polarimetric radar (S-Pol), deployed during the North American Monsoon Experiment (NAME; Higgins et al. 2006) in 2004, Rowe et al. (2011a,b) assessed the relative importance of warm-rain versus ice-based microphysical processes as a function of topography, highlighting the combined roles of collision-coalescence and melting (precipitation-sized) ice for producing intense rainfall in this region. In addition, a comparison between isolated cells and those embedded within organized systems in those studies revealed that MCSs were characterized by deeper convection with more ice and greater rain rates, suggesting a dependence on degree of organization (Rowe et al. 2011b).

A remaining objective from TiMREX involves an understanding of the microphysical processes leading to heavy rainfall. Similar to the analysis from NAME, polarimetric data from the S-Pol radar, also deployed during TiMREX, provided a means

to evaluate hydrometeor characteristics of convection in southwestern Taiwan and to investigate the relative contributions of ice-based and warm-rain processes as a function of terrain. It is hypothesized that although warm-rain processes likely played a dominant role in rainfall production, the growth and subsequent melting of ice hydrometeors became more important along the western slopes of the CMR as environmental southwesterly flow lifted along the terrain. This analysis uses a case study approach to describe the microphysical characteristics and evolution of convection, especially along the foothills of the CMR. In addition, general statistics of TiMREX convection will be presented, allowing for comparisons between isolated cells and those embedded within organized systems, similar to the analysis conducted in northwestern Mexico; therefore, providing a basis for comparison with results from NAME.

CHAPTER 2

DATA AND METHODOLOGY

Field observations during TiMREX occurred from 15 May through 26 June 2008, including nine intense observation periods (IOPs) that focused on upstream environmental conditions, initiation of diurnally forced convection, the interaction between environmental flow and precipitating systems with terrain, and microphysical studies, among others. Operational facilities included six Doppler radars, a boundary layer wind profiler, 418 automatic rain gauges, 57 GPS integrated water vapor sensors, and geostationary and polar orbiting satellite imagery. Additional research facilities were placed in the observation domain (115-123°E, 20-26°N), including an integrated sounding system (ISS), a tethered sonde, four mobile rawinsondes, a VHF wind profiler, several rain profilers, a dropsonde aircraft, a research vessel, National Central University's X-band, polarimetric, mobile radar (TEAM-R), and the NCAR S-Pol radar.

2.1 S-Pol data

The S-Pol radar was located along the southwestern coast of Taiwan, providing adequate baselines for dual-Doppler analyses with TEAM-R and the operational KCCG radar (Fig. 2.1). Full-volume, 360° surveillance scans provided data to a maximum unambiguous range of 150 km. Low-level surveillance scans were available every 15 minutes for coordinated rain-mapping and boundary-layer analyses with nearby radars.

To maintain this cycle, elevation angles were adjusted according to the proximity of the precipitating features to the radar and for inclusion of sector scans and vertical cross sections. Quality control of the raw S-Pol data followed the methodology described by Lang et al. (2007) and involved calibration of the power-based variables, removal of clutter, insects, and second trip contamination, and attenuation corrections. Rainfall attenuation correction was based on the methodology described by Carey et al. (2000), with additional correction for gaseous attenuation following Battan (1973, Ch. 6). Thresholds were applied for removal of non-meteorological echo, similar to those used for the NAME data set (Lang et al. 2007). Differential phase (Φ_{DP}) was filtered based on a technique described by Hubbert and Bringi (1995) in order to calculate specific differential phase (K_{DP}) by fitting a line to the filtered Φ_{DP} field. The full suite of polarimetric radar variables included the following:

- reflectivity at horizontal polarization (Z_H): provides information about the size and concentration of the hydrometeors, weighted heavily by the largest hydrometeors in the volume;
- differential reflectivity (Z_{DR}): the ratio of the returned power in the horizontal to that in the vertical, providing information about the shape, orientation, and thermodynamic state (i.e., liquid or solid) of hydrometeors and mean drop size;
- specific differential phase (K_{DP}): allows for the amount of precipitating liquid water to be estimated, even in the presence of ice;
- linear depolarization ratio (L_{DR}): the ratio of cross-polar to copolar power, providing information about the orientation and canting of hydrometeors, as well as detecting mixtures of hydrometeors within the volume; and

- zero-lag cross-correlation coefficient (ρ_{HV}): the correlation between copolar horizontally and vertically polarized echo signals, allowing for detection of pure rain versus a mixture of hydrometeors.

A more complete description of these variables and their applications to microphysical studies can be found in Bringi and Chandrasekar (2001). Corrected radar data was gridded to Cartesian coordinates at a 1-km horizontal and 0.5-km vertical resolution using the program REORDER (Mohr et al. 1986). Next, using the gridded polarimetric variables, as well as temperature profiles from nearby soundings at Pingtung (Fig. 2.1), a hydrometeor classification algorithm was applied to determine the dominant hydrometeor type at each horizontal and vertical grid point. The HID algorithm is based on the fuzzy logic methodology of Liu and Chandrasekar (2000), described in detail by Tessendorf et al. (2005), yielding nine hydrometeor types, including drizzle, rain, dry snow, wet snow, vertical ice, low-density graupel, high-density graupel, small hail, and large hail.

2.2 Cell identification and tracking

A cell identification algorithm, described by Rowe et al. (2011a), was then applied to the gridded files to objectively locate and track individual convective elements. To determine whether a cell was isolated or part of an organized convection, a modified version of Rickenbach and Rutledge's (1998) classification scheme was used, differentiating only between sub-MCS and MCS features instead of using the four separate feature types described in that study. For this study, a threshold of 15 dBZ was applied to the composite reflectivity field to eliminate weak pixels, similar to the methodology used for NAME (Pereira 2008; Rowe et al. 2011a,b). An ellipse-fitting

method, similar to that applied in the cell identification algorithm, was then used to determine the dimensions of the features. To further eliminate weak pixels, features had to have an area of at least 20 km^2 to be included in the analysis. All remaining features were classified based on the length of the ellipse's major axis, where values $< 100 \text{ km}$ and $> 100 \text{ km}$ were labeled sub-MCS and MCS, respectively. The location of each cell was then matched with the features to classify each identified cell as either isolated or organized.

2.3 Cell properties

Numerous parameters were computed to provide a means to describe and compare individual convective cells. Instantaneous rain rates were calculated using a modified version of the Colorado State University blended polarimetric algorithm (Cifelli et al. 2002). Polarimetric-based equations for estimating rainfall were applied to all horizontal and vertical grid points within an echo using similar thresholds as in NAME studies (e.g., Lang et al. 2007; Rowe et al. 2008, 2011a,b), allowing for mean and maximum rain rates to be calculated for each identified cell. In addition, using the rain rates at each grid point within a cell, mass fluxes were computed for each cell, leading to calculations of hourly rain masses for selected groupings of cells.

Ice and liquid water masses were computed based on the methodology described by Carey and Rutledge (2000) and Cifelli et al. (2002), and ice (IWP) and liquid water (LWP) paths were determined by integrating the ice and liquid water masses over vertical columns within individual cells. Ice and liquid water path ratios were then calculated by dividing the IWP and LWP, respectively, by the summation of the two. In addition, ice

and liquid water mass fractions were determined by dividing the sum of ice and water mass, respectively, by the total mass (both ice and liquid) within the cell. Echo-top heights were estimated using the maximum height of the 0-dBZ reflectivity contour. Maximum Z_{DR} and K_{DP} were computed for grid points within the cell identified as rain by the HID and located below 4 km to reduce the effects of melting; the average melting level during TiMREX was 5 km. Finally, using only the Z_{DR} values classified as rain below 4 km, mean drop diameters of cells were computed using the following relationship: $D_0 = 1.529Z_{DR}^{0.467}$ (Bringi and Chandrasekar 2001).

2.4 Topographic data

Topographic data from the National Geophysical Data Center (NGDC) were used to compute the maximum terrain height associated with each identified cell. To remain consistent with the NAME analyses (e.g., Gochis et al. 2004; Rowe et al. 2008, 2011a,b), cells were grouped based on the following elevation ranges: 0–1 km, 1–2 km, > 2 km, and over water. Furthermore, to compare cells along the coast with those occurring along the foothills and higher elevations of the CMR, cells were also divided based on an elevation threshold of 0.5 km. These subdivisions allowed for comparisons of cells as a function of terrain, as well as between land and water.

2.5 Case selection

During NAME, approximately 95 hours of radar data were selected for the analyses by Rowe et al. (2011a,b) due to increased temporal and spatial resolution of those scans. Those so-called “microphysical scans” focused on intense precipitating

features, thereby neglecting periods of short-lived, weak mountain convection. A selection of time periods characterized by intense rainfall during TiMREX was therefore required to allow for accurate comparisons between the two regions. As a result, cases for this study were selected from 19 May through 26 June 2008, corresponding to the beginning of IOP1 and the end of IOP9, respectively. This 39-day time period captured a variety of precipitation features, ranging from isolated, diurnally forced convection along the foothills to widespread mesoscale convective systems moving onshore within strong southwesterly flow. Individual rain events were selected to be consistent with the Tong (2009) study, in which 3-h-averaged rain rates were subjected to an arbitrary threshold of 5.5 mm h^{-1} , thereby excluding weakly precipitating, short-lived systems. In order to further classify these periods into mutually exclusive events, Hovmöller diagrams of radar-derived rain rate were also analyzed, leading to 40 individual rain events, comprising about 40% of the total time of the entire study (Tong 2009). These 40 events served as the basis for computation of general statistics during TiMREX, presented in later chapters.

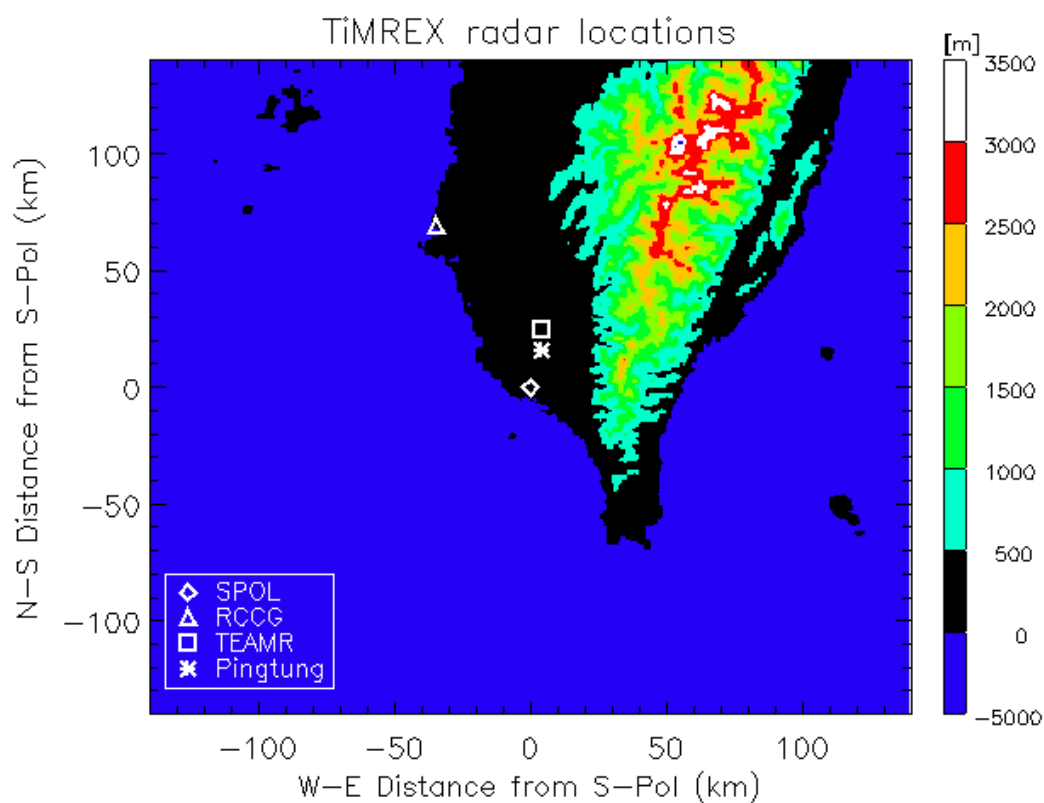


Fig. 2.1. Locations of the S-Pol, RCCG, and TEAM-R radars during the TiMREX field experiment during 2008. The location of Pingtung, the nearest sounding location to S-Pol, is indicated by an asterisk. Topography (m) is shaded, with water indicated in dark blue as negative values.

CHAPTER 3

CASE STUDIES

Due to the contribution to heavy rainfall from isolated convection and large-scale systems during TiMREX, examples from both types were selected from the 40 rainfall events to describe the evolution of precipitating features in this region. Particular attention was given to days when convection initiated or moved toward the western slopes of the CMR in order to investigate the effect of topography on precipitation. Prior to 2 June 2008, precipitation was concentrated over land (71% of cells identified during this period), as diurnally driven convection dominated the region under weak synoptic conditions. For this analysis, three days were chosen to focus on this intense, diurnally forced convection along the foothills of the CMR: 30 May, 31 May, and 1 June 2008. The remainder of the experiment was under the influence of the Mei-Yu front, with 85% of precipitating cells considered part of organized features and 57% of those over water, reflecting the high frequency of organized systems moving from water to land within the strong, southwesterly environmental flow. An example of an MCS moving onshore and interacting with the terrain on 13–14 June 2008 was selected to investigate elevation-dependent trends in microphysical properties associated with an organized system.

3.1 30 May 0100–1200 UTC

The large-scale environment on 30 May 2008 was characterized by weak westerly flow with moist low levels and dry air aloft. There was a lack of synoptic-scale systems influencing the domain, and the Mei-Yu front was situated to the north of Taiwan, allowing for diurnally driven convection to dominate precipitation on this day. Of all cells identified throughout this time period, 77% were classified as isolated, with the vast majority (85%) of those occurring over land. In addition, of these isolated cells over land, 75% were associated with terrain > 0.5 km, allowing for an analysis of isolated convection along the western slopes of the CMR.

An example of isolated convection along the terrain is shown in Figs. 3.1 and 3.2. During these selected times, the vertical depth of radar echo was limited to below 12 km, likely due to the prominent midlevel drying on this day. The HID-classified graupel existed within only a shallow layer near the melting level (Fig. 3.2), suggesting a dominance of warm-rain processes during this cell's lifetime. Collision and coalescence led to reflectivity exceeding 50 dBZ at 0422 (Fig. 3.1a) and 0430 UTC (Fig. 3.1b), with K_{DP} values approaching $3^{\circ} \text{ km}^{-1}$ and $Z_{DR} > 3$ dB during these times (Fig. 3.1a,b), equivalent to drop sizes of at least 2.5 mm in diameter. The corresponding vertical profile of mean Z_{DR} at 0430 UTC (Fig. 3.3) revealed a peak near 1 km, associated with large drops near the surface due to collision and coalescence, and an additional peak near the melting level (5 km), where the melting of graupel also likely contributed to production of large drops at this time. As the cell evolved, Z_{DR} decreased aloft, leaving peak mean values of 1 dB near the surface (Fig. 3.3). This is consistent with the vertical cross sections, which show smaller Z_{DR} values at 0437 UTC (Fig. 3.1c) and reduced K_{DP} at

0445 UTC (Fig. 3.1d), corresponding to decreasing areas of graupel (Fig. 3.2), as the core descended. Peak rainfall rates during this time reached 90 mm h^{-1} , suggesting the efficient role of warm-rain processes in producing precipitation in isolated cells along the terrain. An RHI at 0450 UTC (Fig. 3.4) further highlights the production of large drops in these shallow cells due to collision and coalescence near the slopes of the CMR. Both cells existed primarily below the melting level with only nominal amounts of precipitation-sized ice, yet Z_{DR} values of 4 dB and an elevated K_{DP} maximum near 4° km^{-1} further suggested an important contribution of large drops to the liquid water content and subsequent intense rainfall in these isolated cells.

Another common characteristic of convection on this day was an upslope tilt along the terrain. Two examples, shown in Fig. 3.5, both exhibited this tilt, with maximum reflectivity and Z_{DR} elongated upslope in a region of slight westerly speed shear (indicated by the 0600 UTC sounding). Updrafts of shallow convection forming along steep topography in Hawaii have been shown to also exhibit an upslope tilt in an environment of weak mid-level shear, resulting in quasi-stationary systems with convective cores extending along the crest of the terrain (Murphy and Businger 2010). In that study, similar to that hypothesized for southwestern Taiwan (see Ch. 1), the vertical advection of hydrometeors, originating below the melting level, by terrain-influenced updrafts was crucial to the formation and growth of graupel via the freezing of raindrops and subsequent growth by riming. However, due to the dry midlevels during the TiMREX case on 30 May 2008, the depth of convection along the CMR was limited, thereby reducing the vertical extent of precipitation-sized ice. Despite this limitation, cumulative distribution functions (CDFs) of maximum rain rate (Fig. 3.6) show greater

values for cells over elevations > 0.5 km compared to the few along the coast and over water, suggesting a topographical dependence on microphysical processes.

3.2 31 May 0100–1100 UTC

Compared to the previous day, sounding data highlighted a moister environment, with relatively weak midlevel flow and a slight southwesterly component at low levels ahead of the approaching Mei-Yu front. Developing precipitating features were associated with this quasi-stationary boundary, including gradually organizing linear convective systems in the northern portion of the domain, as well as diurnally forced convective initiation over the foothills and adjacent plain (e.g., Tong 2009). Only 24% of the cells during this time period were considered isolated, with 53% of those occurring over land, of which half were along the western slopes and higher terrain. This relatively even distribution of cells with elevation allowed for investigation of differences between cells over the coast compared to those over the CMR.

A vertical cross section through an isolated cell along the terrain at 0422 UTC (Fig. 3.7) reveals a similar upslope tilt as was observed in the previous case, with reflectivity and Z_{DR} increasing toward the higher terrain. The HID-classified graupel was displaced from the peak reflectivity and Z_{DR} (Fig. 3.7b), suggesting that hydrometeors were lofted upslope above the melting level, allowing for production of precipitation-sized ice over the terrain. The layer of graupel, however, was shallow, further suggesting the dominance of warm-rain processes in these isolated cells, as Z_{DR} values exceeded 3 dB near the surface and peak rain rates reached 200 mm h^{-1} at this time.

The evolution of another example of tilted cells along the slopes is shown in Fig. 3.8, where hydrometeors were advected upslope, leading to the propagation of the K_{DP} and Z_{DR} maxima from low elevations to over the slopes, thereby shifting the location of maximum rainfall. More specifically, initially at 0815 UTC (Fig. 3.8a), the K_{DP} maximum was near the surface within the upslope-tilted cell, located at a range of 20 km. By 0822 UTC (Fig. 3.8b), this maximum had shifted to over higher elevations and upward to 4 km, consistent with our hypothesis that hydrometeors were lofted above the melting level along the slopes, where they froze, then subsequently melted to produce intense rainfall over the slopes. The tendency for precipitation to be maximized along the slopes has been shown to be a common characteristic in regions of steep topography, where upward air motion and microphysical growth processes on the windward side of these barriers are most robust at lower levels (e.g., Houze 2011). A descending core was observed in the TiMREX case at 0830 UTC (Fig. 3.8c) when maximum values of K_{DP} began to decrease and lower, followed by further weakening and descent to the surface by 0837 UTC (Fig. 3.8d), suggesting lower rain rates as the cell moved over yet higher terrain.

Time series of maximum rain rate, IWP ratio, and mean drop diameter as a function of maximum terrain height (Fig. 3.9), associated with another cell on this day, highlight an initial increase, albeit slight, in the ratio of ice to total water mass as the cell approached the terrain. Ratios, however, remained below 0.2 signifying the continued dominance of warm-rain processes, similar to processes observed in the previous example (Fig. 3.8). Rain rates initially increased with increasing terrain, reaching a maximum of 170 mm h^{-1} at 0615 UTC. A slight decrease in mean drop size diameter (D_0)

was also noted during this time, although the difference between 0600 and 0622 UTC was only approximately 0.3 mm. The relatively constant mean diameter during this time period, corresponding to rain rates increasing above 80 mm h^{-1} , suggested the establishment of an equilibrium drop-size distribution (DSD; e.g., Atlas and Ulbrich 2000), resulting from a balance between collision, coalescence, and drop breakup (e.g., Hu and Srivastava 1995). Therefore, for the example presented in Fig. 3.9, melting hydrometeors likely broke up and contributed to high liquid water contents at the time of maximum rainfall intensity along the slopes.

Convection exhibiting a greater degree of organization also occurred near the terrain on this day. A series of RHIs through one particular feature at an azimuth of 45.5° are shown in Figs. 3.10–3.13. Figure 3.10 highlights a multicellular system at 0758 UTC, characterized by several cores along the lower terrain that contained graupel extending above 10 km, Z_{DR} approaching 4 dB, and elevated K_{DP} maxima $> 3^\circ \text{ km}^{-1}$, suggesting the melting of ice hydrometeors. Low-level outflow was present below the convective cores, with a large area of outbound velocities over the slopes, reflecting the west-southwesterly flow impinging on the terrain at this time. Over the CMR, stratiform precipitation was characterized by a reflectivity brightband, with corresponding enhancements of Z_{DR} and L_{DR} near the melting level. Both the low-level outflow and brightband features remained at 0806 UTC (Fig. 3.11), although the vertical extent of graupel had lowered along the back edge as convection within the leading portion of the system intensified. Maximum Z_{DR} and K_{DP} values exceeded 3 dB and 3° km^{-1} , respectively, both at the surface and just below the melting level, as ice continued to melt near the slopes. By 0813 UTC (Fig. 3.12), the leading convective cell became dominant as convection along the slopes

decayed. The adjacent stratiform region maintained a brightband feature, manifested as an enhancement in reflectivity, Z_{DR} , and L_{DR} as relatively smaller ice particles fell below the melting level. The presence of upslope flow in midlevels, seen as strong outbound velocities in this and the previous RHIs, suggests that ice forming in the convective portion of the system was advected horizontally at these upper levels toward higher terrain, similar to the “particle fountain” process described by Yuter and Houze (1995). The remnant cellular structure near the leading convective line, however, also suggests a contribution to stratiform rain from weakening convection, a common characteristic of MCSs in the midlatitudes (Houze 1993, 1997).

A narrow region of graupel persisted along the leading edge through 0821 UTC (Fig. 3.13), with a continued contribution to maxima in Z_{DR} and K_{DP} from the melting of large ice hydrometeors. New convection initiated adjacent to the main leading edge over lower elevations, likely due to converging of the low-level outflow, seen in the velocity image at this and previous times, with the environmental upslope flow. This process of upstream regeneration likely contributed to a longer lifetime of this feature compared to isolated convection and has been shown to influence the longevity of organized systems in other mountainous regions, including the Himalayas (Medina et al. 2010), the European Alps (Houze et al. 2001), and the steep topography of northwestern Mexico (Rowe et al. 2011b). This process led to extended periods of enhanced rain rates over the sloping terrain, as suggested by CDFs of maximum rain rate (Fig. 3.14), which show greater instantaneous rain rates, compared to over the coast on this day, and increased hourly rain mass totals, as 57% of this cumulative rain mass corresponded to elevations > 0.5 km.

3.3 1 June 0300–1100 UTC

Additional examples of isolated, diurnally driven convection near the slopes occurred on 1 June due to the continued lack of synoptic-scale disturbances and a northward retreat of the Mei-Yu front, leaving weak westerly flow both at low levels and aloft. The majority of cells on this day (76%) were considered isolated, with all of those over land and 95% concentrated along the terrain. Vertical cross sections, displaying the evolution of one of these isolated features, are presented in Figs. 3.15 and 3.16. Initially, at 0530 UTC, there was almost no graupel identified (Fig. 3.16a), K_{DP} values were $< 1^\circ \text{ km}^{-1}$ (Fig. 3.15a), and Z_{DR} ranged from 1 to 2 dB. At the next time step, 0537 UTC, high-density graupel vertically extended several kilometers about the melting level (Fig. 3.16b), likely due to the freezing of lofted supercooled drops, which grew initially by warm-rain processes at lower levels; the melting of these large ice hydrometeors led to elevated maxima in Z_{DR} and K_{DP} along the slopes at this time (Fig. 3.15b).

This contribution from melting hydrometeors to rainfall was also inferred at 0545 UTC, where Z_{DR} values exceeded 3 dB near the melting level, corresponding to an elevated maximum in $K_{DP} > 3^\circ \text{ km}^{-1}$ (Fig. 3.15c). In addition, graupel extended farther upslope (Fig. 3.16c), as was also seen in previous examples of isolated convection along the terrain. Maximum rain rates in this cell increased from near 100 mm h^{-1} at 0530 UTC to 160 mm h^{-1} at 0545 UTC, during the period when large ice hydrometeors were melting along the slopes. An RHI at an azimuth of 64° at this time (Fig. 3.17) reveals more detailed characteristics of this feature, including a reflectivity overhang in a region of HID-classified graupel and enhanced L_{DR} , indicating a mixture of hydrometeors and

implying growth of large ice particles via freezing and subsequent riming. This feature overlaid Z_{DR} values of 4 dB and K_{DP} reaching 4° km^{-1} , further suggesting the growth and melting of large ice hydrometeors. Later, at 0552 UTC, continued melting and possible drop breakup were suggested due to enhanced K_{DP} coincident with reduced Z_{DR} values (Figs. 3.15d, 3.16d). Distributions of Z_{DR} and K_{DP} (Fig. 3.18), corresponding to the times of these vertical cross sections (Figs. 3.15, 3.16), highlight the initial shift to higher percentages of large drops with time. Greater percentages of large K_{DP} also coincided with an increase in larger Z_{DR} , suggesting a contribution from large drops to water mass, and therefore, rainfall, in these cells. At the final time, the distribution shifted to smaller drops, reflecting the breakup of drops suggested in the vertical cross section at 0552 UTC.

This case study revealed yet another example of enhancement of rainfall along the slopes as hydrometeors were lofted above the melting level, where they froze, grew by riming, then melted, leading to contributions from both warm-rain and ice-based microphysical processes, as has been observed in other mountainous regions (e.g., Rowe et al. 2011a; Houze 2011). Elongated, elevated regions of maximum reflectivity indicated upslope advection of hydrometeors, which was further emphasized by tilting of convection toward the foothills of the CMR, observed in examples of both isolated and organized convection described in this study. Next, microphysical characteristics associated with MCSs are further investigated to determine if similar processes occurred as these organized systems moved onshore and interacted with the steep terrain.

3.4 13 June 1700 UTC – 14 June 1800 UTC

In contrast to the previous cases, deepening of an upper-level trough provided favorable synoptic conditions for the development of organized systems. Southwesterly winds ahead of the advancing Mei-Yu front brought moist flow into southwestern Taiwan, and several north-south-oriented convective lines with associated trailing stratiform moved successively onshore on 13 and 14 June within this southwesterly environmental flow. Although the majority of cells (87%) were associated with these organized systems, isolated convection was also observed ahead of the first convective line, providing an interesting case for analysis of both isolated and organized features. Of the isolated cells, 59% were observed over water, with 30% of the remaining cells over land occurring along the terrain, leaving 11% along the coastal plain. For organized cells, 44% were over water, with 42% of the cells over land identified at elevations > 0.5 km. This spread of cells over a range of topography allowed for further comparisons between cells over water with those along the coast and adjacent terrain.

A vertical cross section through a portion of the system over water at 0507 UTC shows a convective core embedded within a large region of deep stratiform echo (Fig. 3.19). Values of Z_{DR} remained below 1 dB within a region of reflectivity between 40 and 50 dBZ and K_{DP} near 2° km^{-1} (Fig. 3.19a) with only a shallow layer of graupel observed near the melting level (Fig. 3.19b). An RHI at 0751 UTC (Fig. 3.20), through a portion of the system near the coast, also shows an almost complete lack of precipitation-sized ice with low values of Z_{DR} and K_{DP} near the surface, indicating weaker precipitation. Studies of wide echo near the Himalayas (e.g., Houze et al. 2007; Romatschke et al. 2010) also showed that organized systems over nearby water lacked the deep intense cores that

occurred over the adjacent land near the mountains. Medina et al. (2010), however, noted that the stratiform precipitation regions of these MCSs were strengthened after moving inland over the mountain ranges. Similarly, as the broad region of stratiform precipitation moved onshore and began to interact with the topography during the 13–14 June case at 0851 UTC, graupel production increased (Fig. 3.21). This shallow depth of graupel along the leading edge of the system contributed to enhanced reflectivity, Z_{DR} , and K_{DP} below the melting level as the large ice hydrometeors melted. Widespread stratiform precipitation extended over the higher topography, with a brightband feature evident in a region of strong outbound velocities indicating upslope flow. After zooming in on an earlier example of an organized cell as it impinged on the CMR, a prominent upslope tilt was evident in the reflectivity and K_{DP} fields (Fig. 3.22a), as was described in previous cases of diurnally driven convection during TiMREX. This further suggests lofting of hydrometeors above the melting level over higher elevations, as indicated by the HID-identified graupel at this time (Fig. 3.22b). Strong precipitation particle growth via coalescence below the melting level and riming aloft has been shown to also be favored in convective cells embedded within preexisting widespread cloud systems in other regions characterized by locally enhanced upward motion along windward slopes of steep terrain (e.g., Medina and Houze 2003; Georgis et al. 2003; Yuter and Houze 2003).

One particular track, following a segment of an MCS as it moved from the water toward the CMR, exhibited a gradual increase in rain rate with terrain to values exceeding 100 mm h^{-1} (Fig. 3.23). An enhancement in the IWP ratio was observed along the coast, although mean drop diameters remained nearly constant throughout the track, suggesting that breakup of melting ice hydrometeors led to equilibrium DSDs along the

slopes. Vertical profiles of Z_{DR} and K_{DP} (Fig. 3.24), for the period of time when this track moved toward higher elevations, show an enhancement of both variables near the melting level at the time when the system was located over the terrain (1122 UTC), likely due to melting. The increase in Z_{DR} at this level, in particular, suggests production of large drops over the terrain, yet distributions of maximum Z_{DR} (Fig. 3.25) and vertical profiles of mean water and ice mass (Fig. 3.26) do not show an increase in these values with increasing elevation. Smaller values of Z_{DR} at lower levels at 1122 UTC (Fig. 3.24) indicated a loss of large drops near the surface, possibly, in part, via breakup mechanisms (Kirankumar et al. 2008), as was described for Fig. 3.23, suggesting that smaller drops contributed to high liquid water mass regardless of elevation.

Also in agreement with that example, CDFs of maximum rain rate (Fig. 3.27) reveal slightly greater intensities over the terrain for organized cells, with the weakest rates over water. Similar trends are noted for isolated convection, with greater rates over the terrain compared to organized convection, as well as greater liquid water masses (Fig. 3.26a) and greater percentages of large drops (Fig. 3.25) over all elevations. This suggests that large drops may have been an important contributor to rainfall in isolated convection due to melting of precipitation-size ice hydrometeors, which is further implied by greater ice masses in isolated cells near the melting level (Fig. 3.26b). However, long-lived organized convection contributed 90% of accumulated hourly rain mass on this day, with 57% of that corresponding to elevations > 0.5 km, emphasizing the need to understand the processes leading to the enhancement and persistence of these systems as they interact with the topography.

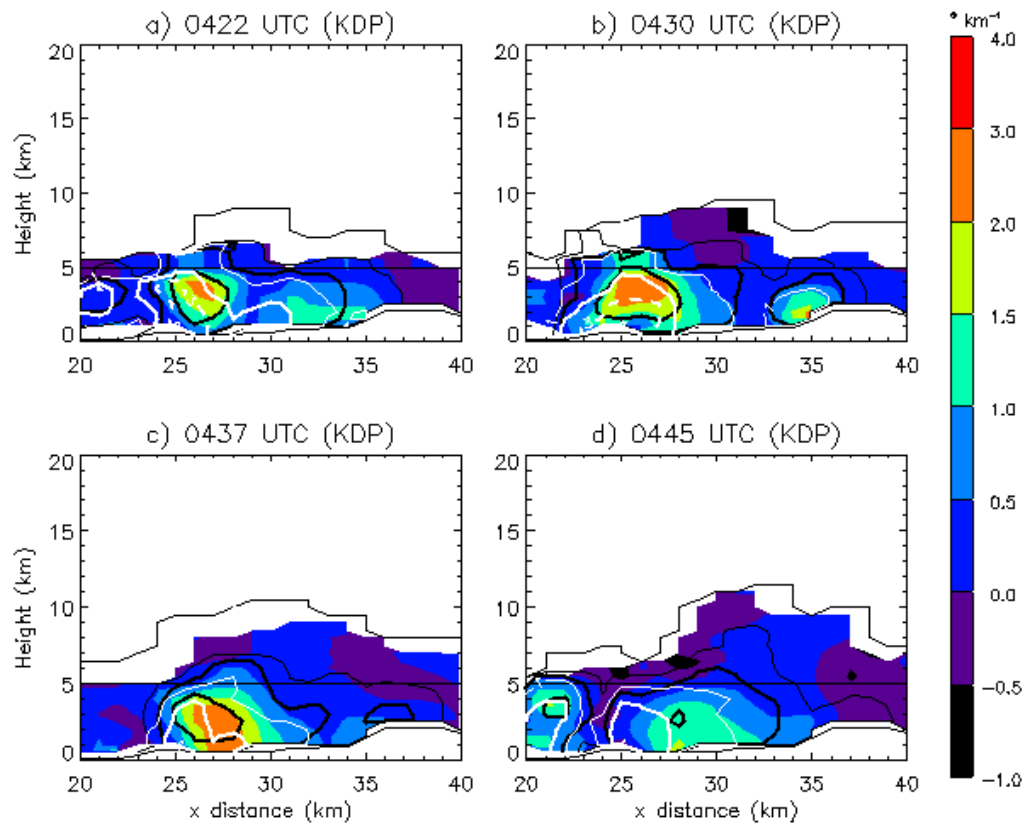


Fig. 3.1. Vertical cross sections through a cell at (a) 0422, (b) 0430, (c) 0437, and (d) 0445 UTC on 30 May 2008. Color-filled contours are associated with K_{DP} ($^{\circ} \text{km}^{-1}$). Black contours represent reflectivity at 0, 30, 40, and 50 dBZ, with the 40- and 50-dBZ contours thickened. Values of Z_{DR} are contoured in white for 1 dB (solid), 2 dB (thick solid), and 3 dB (dashed). Terrain height is plotted at the bottom as a black contour, and the solid black line at 5 km represents the average height of the melting level during TiMREX. The x-axis is the distance from S-Pol in km.

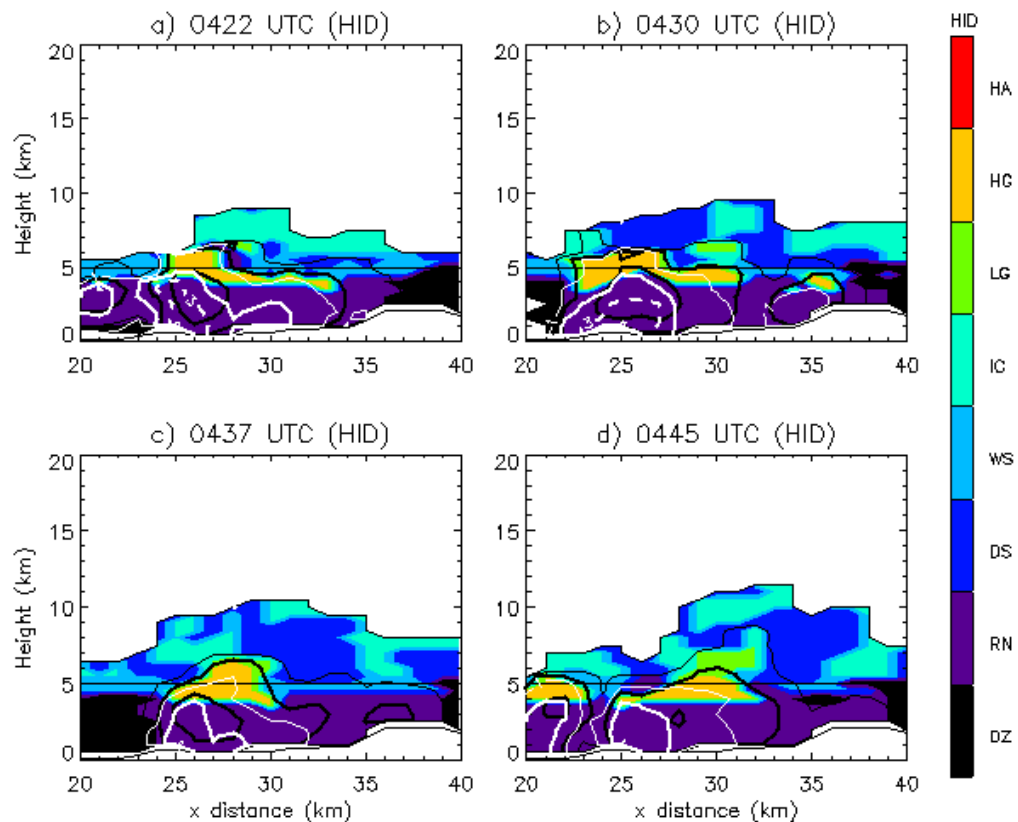


Fig. 3.2. Vertical cross sections through a cell at (a) 0422, (b) 0430, (c) 0437, and (d) 0445 UTC on 30 May 2008. Color-filled contours are associated with HID. The HID abbreviations correspond to the following classifications: drizzle (DZ), rain (RN), dry snow (DS), wet snow (WS), ice (IC), low-density graupel (LG), high-density graupel (HG), and hail (HA). Black contours represent reflectivity at 0, 30, 40, and 50 dBZ, with the 40- and 50-dBZ contours thickened. Values of Z_{DR} are contoured in white for 1 dB (solid), 2 dB (thick solid), and 3 dB (dashed). Terrain height is plotted at the bottom as a black contour, and the solid black line at 5 km represents the average height of the melting level during TiMREX. The x-axis is the distance from S-Pol in km.

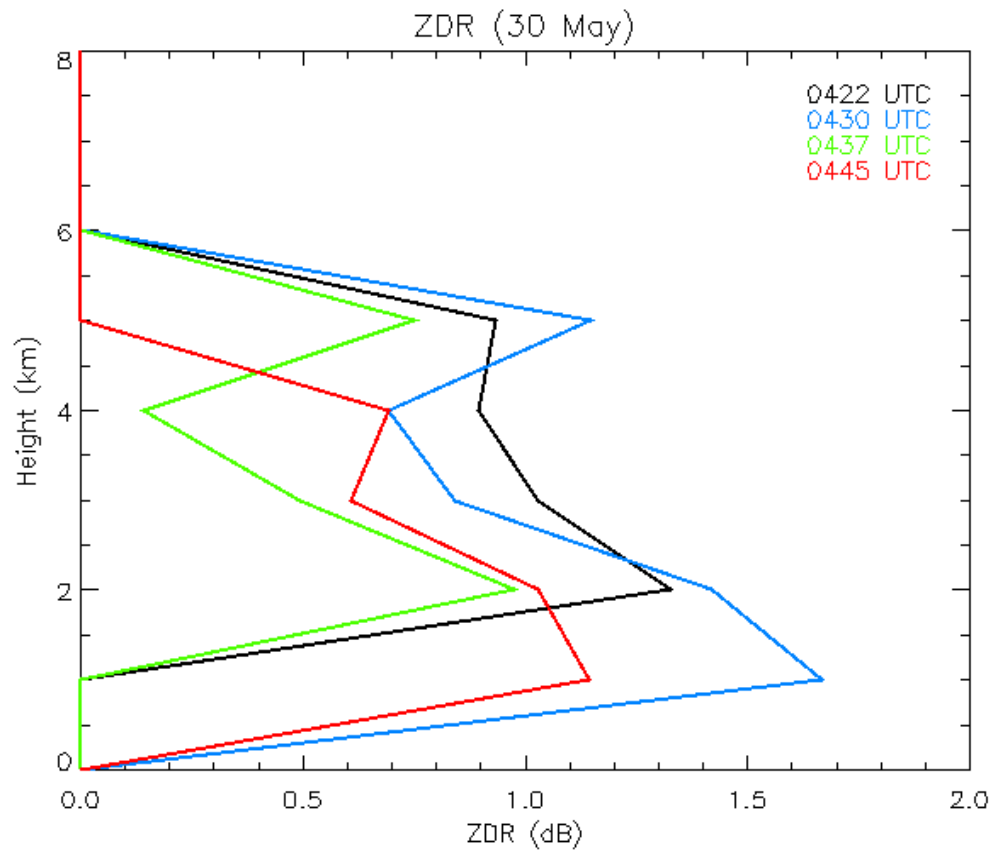


Fig. 3.3. Vertical profiles of mean Z_{DR} associated with the evolution of the cell presented in the vertical cross sections of Figs. 3.1 and 3.2.

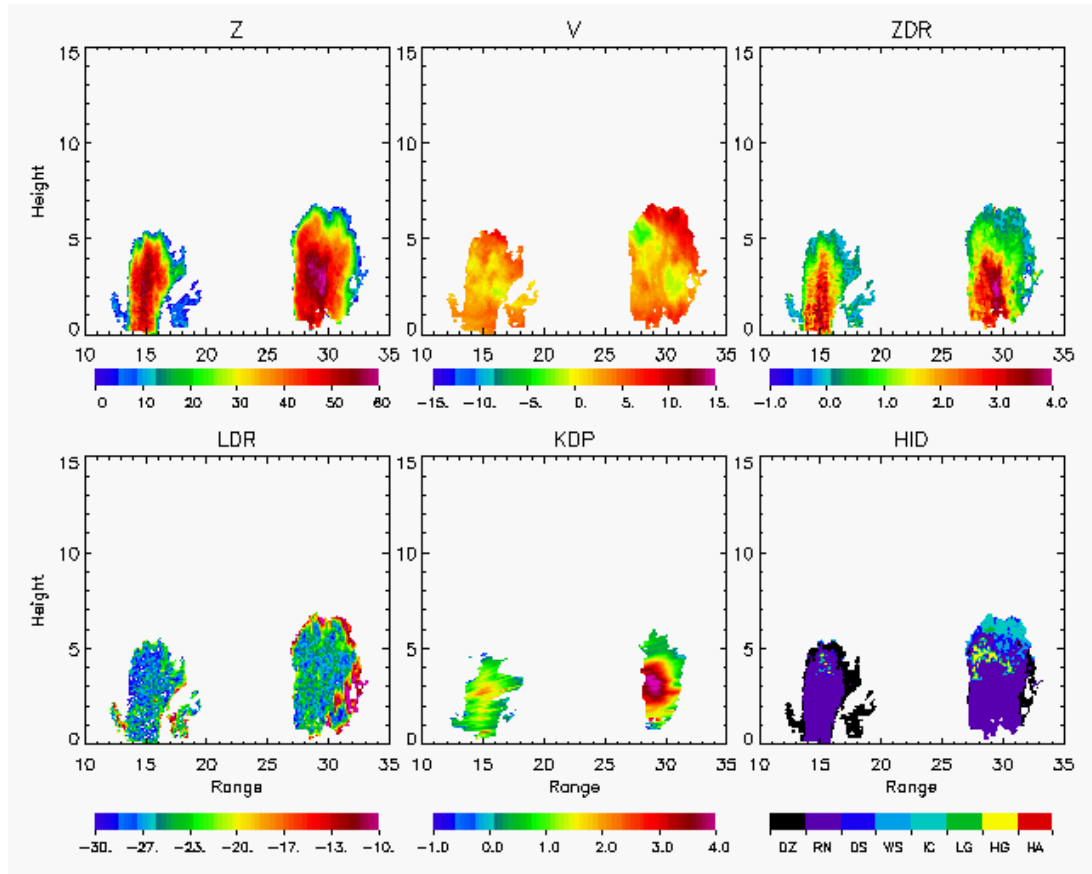


Fig. 3.4. An RHI through cells at 0450 UTC on 30 May 2008 at an azimuth of 51° . Variables from left to right, starting with the upper panel, include reflectivity (Z , dBZ), radial velocity (V_R , m s^{-1}), differential reflectivity (Z_{DR} , dB), linear depolarization ratio (L_{DR} , dB), specific differential phase (K_{DP} , $^\circ \text{ km}^{-1}$), and the hydrometeor identification (HID). The HID abbreviations are the same as in Fig. 3.2. Range and height are in km.

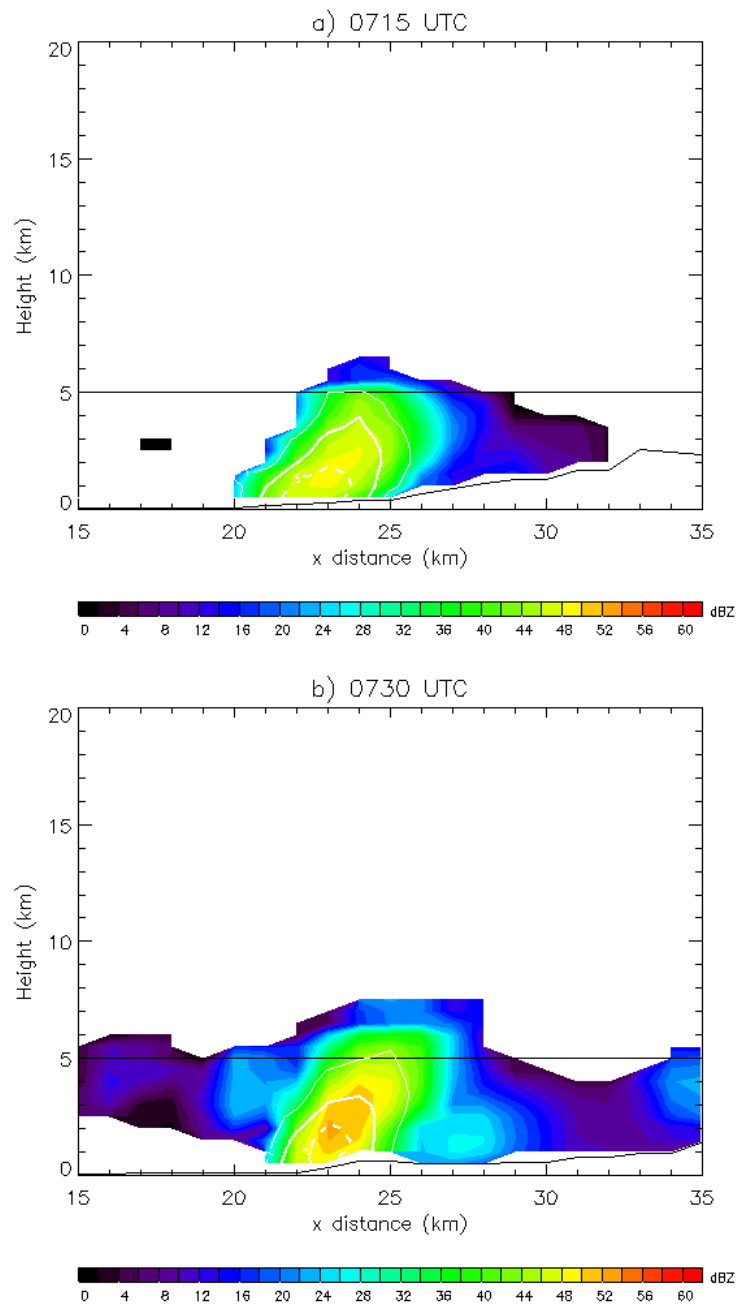


Fig. 3.5. Vertical cross sections of cells on 30 May 2008 at (a) 0715 and (b) 0730 UTC. Color-filled contours represent reflectivity, and the remaining contours and lines are as described in Fig. 3.1.

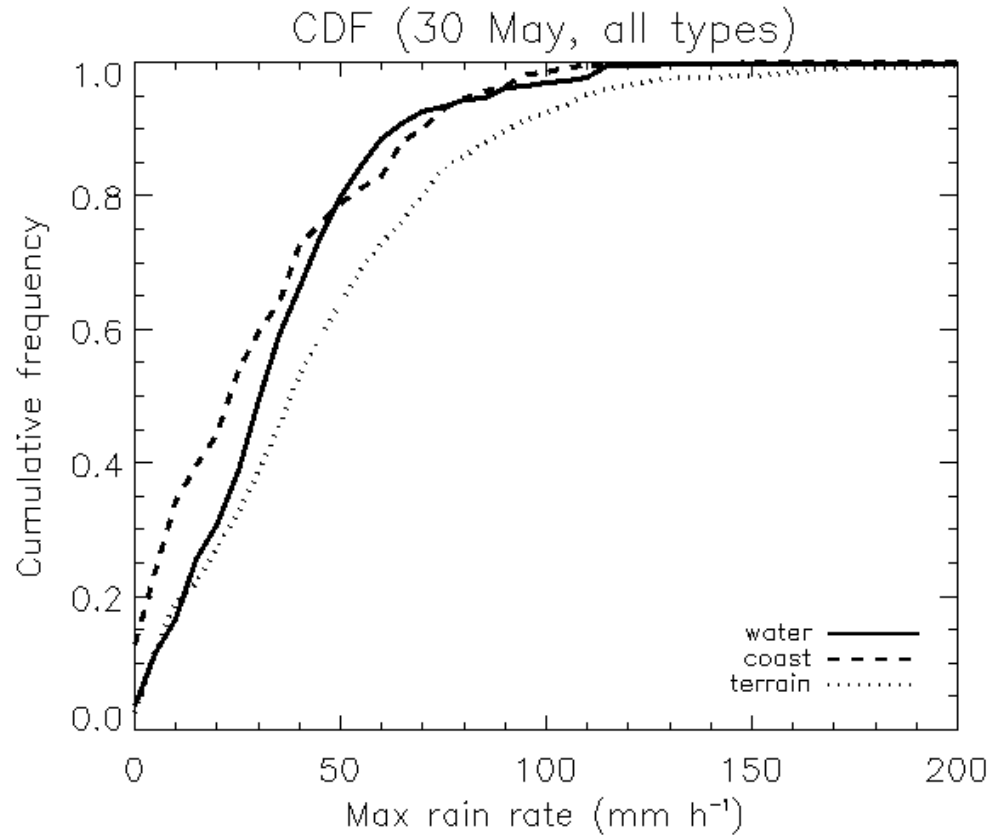


Fig. 3.6. Cumulative distribution functions of maximum rain rate (mm h^{-1}) as a function of elevation for all cells, regardless of type (i.e., isolated and organized), identified on 30 May 2008. The coast (terrain) elevation group represents maximum terrain height < 0.5 km (≥ 0.5 km).

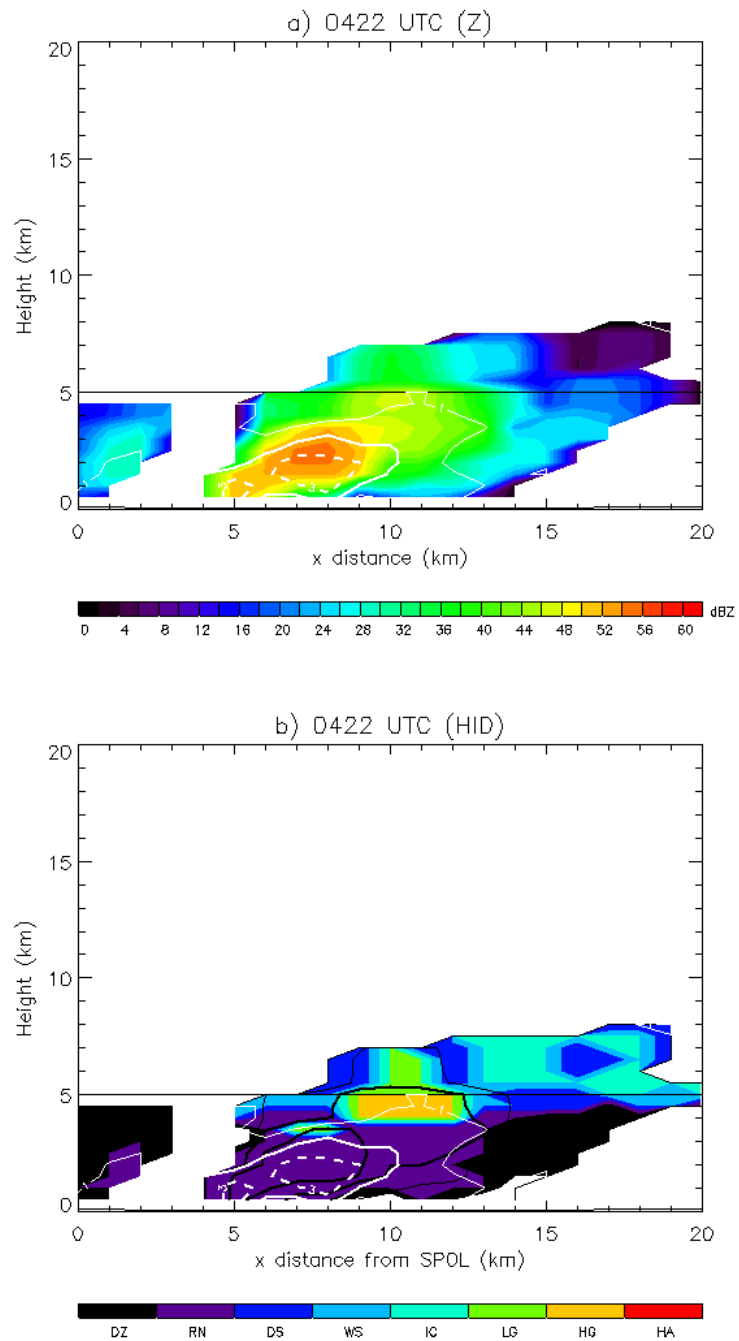


Fig. 3.7. Vertical cross sections through a cell at 0422 UTC on 31 May 2008. Color-filled contours represent (a) reflectivity and (b) HID. The HID abbreviations are described in Fig. 3.2. Remaining contours are as described in Fig. 3.1.

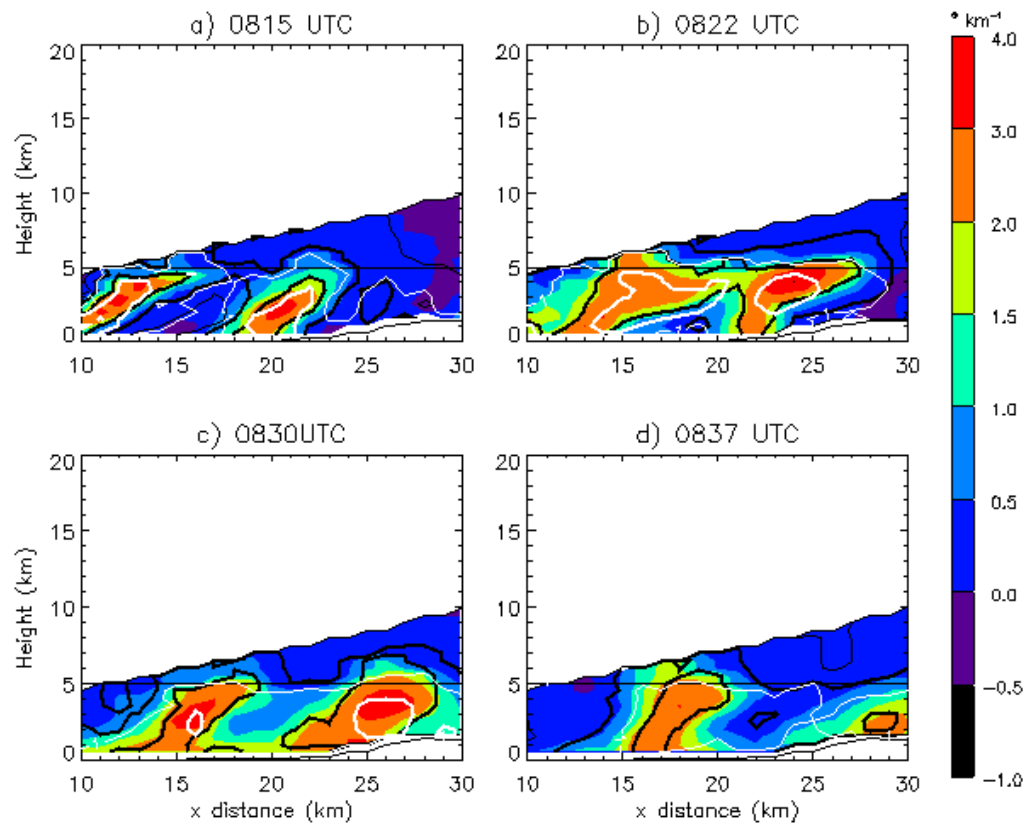


Fig. 3.8. Vertical cross sections of K_{DP} (color-filled contours) for the time evolution of a cell on 31 May 2008 at (a) 0815, (b) 0822, (c) 0830, and (d) 0837 UTC. Other contours are as in Fig. 3.1.

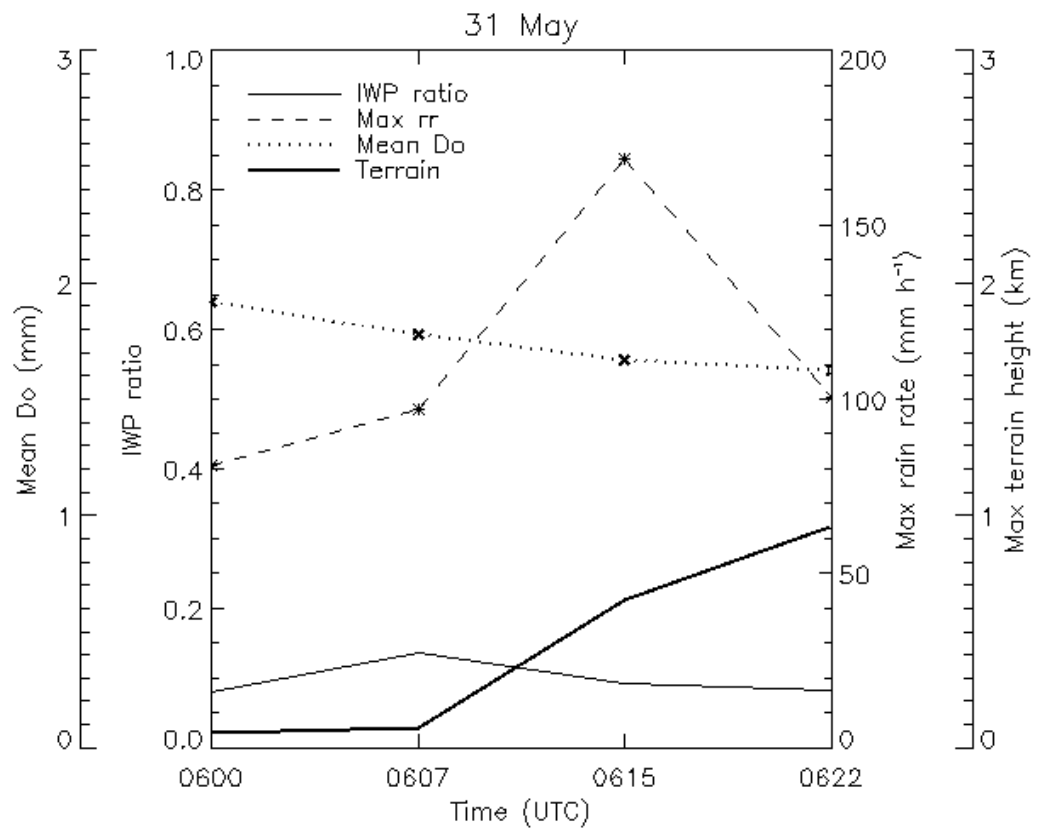


Fig. 3.9. Time series of IWP ratio, maximum rain rate (mm h^{-1}), mean drop diameter (mm), and maximum terrain height (km) for a cell on 31 May 2008.

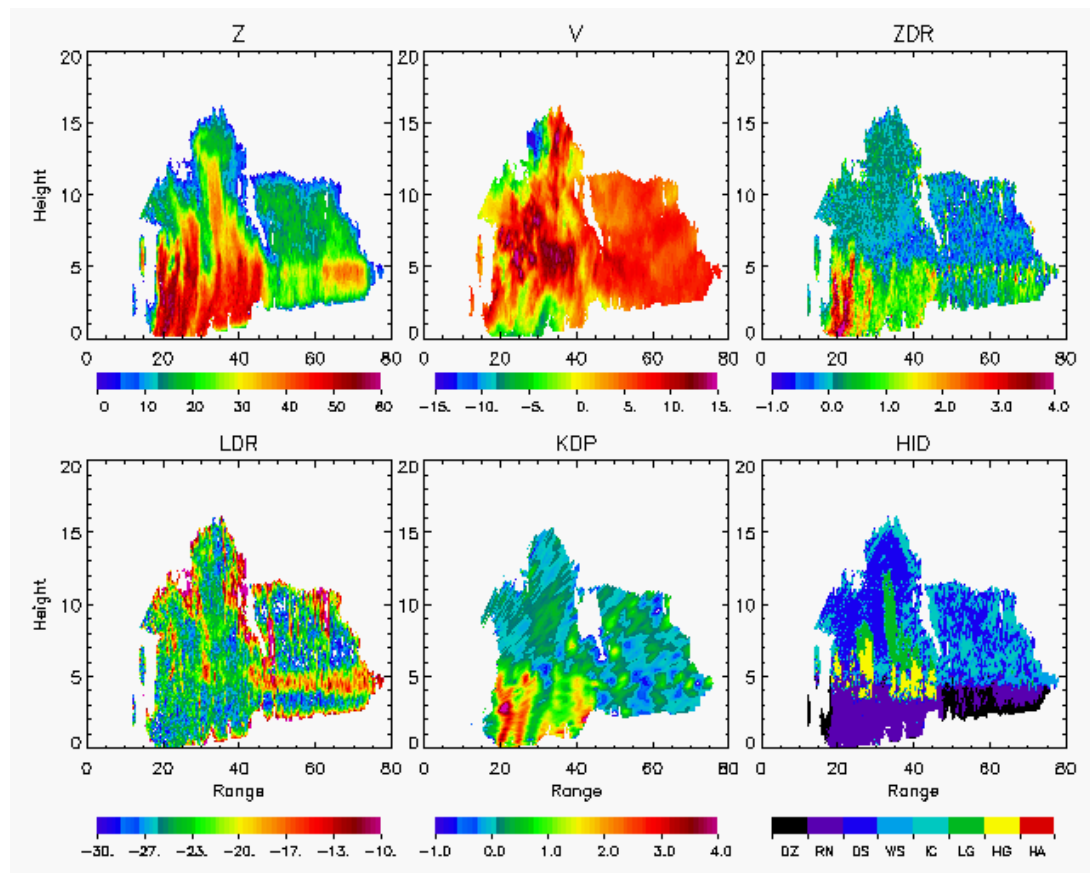


Fig. 3.10. An RHI through a cell at 0758 UTC on 31 May 2008 at an azimuth of 45.5°. Variables are described in Fig. 3.4.

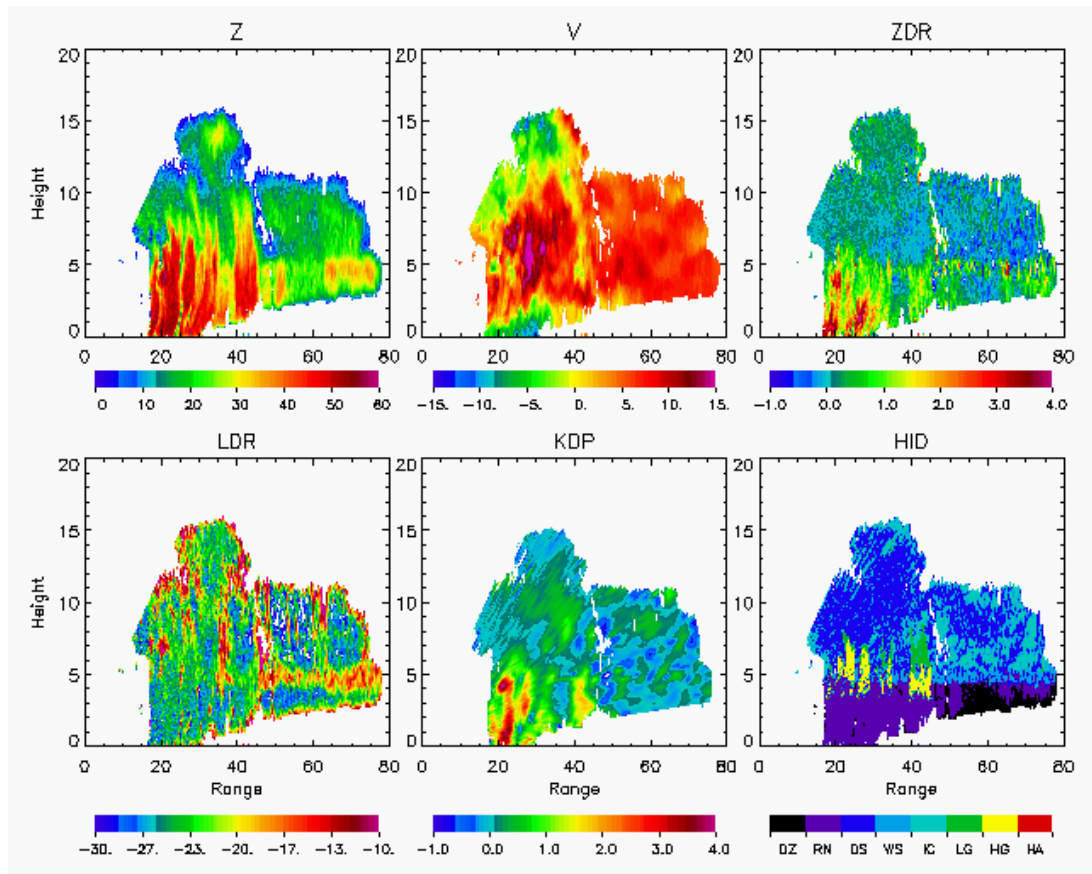


Fig. 3.11. An RHI through a cell at 0806 UTC on 31 May 2008 at an azimuth of 45.5° . Variables are described in Fig. 3.4.

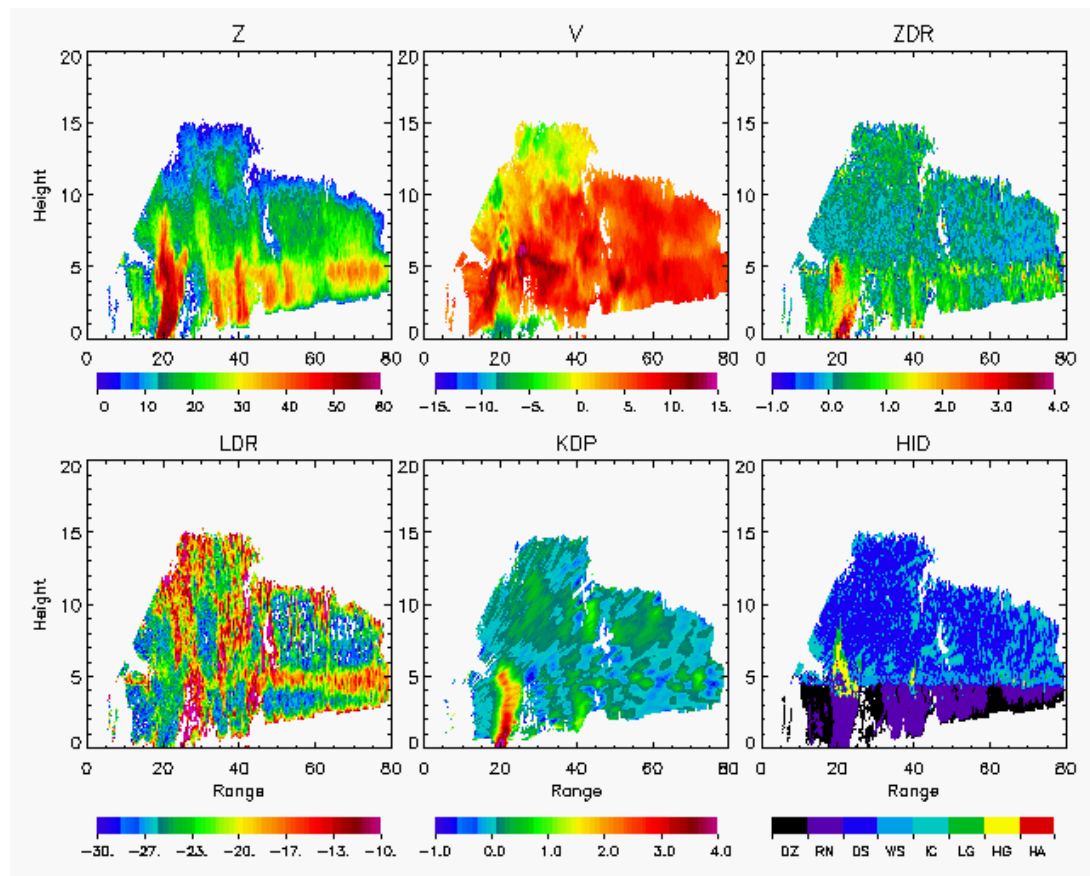


Fig. 3.12. An RHI through a cell at 0813 UTC on 31 May 2008 at an azimuth of 45.5° . Variables are described in Fig. 3.4.

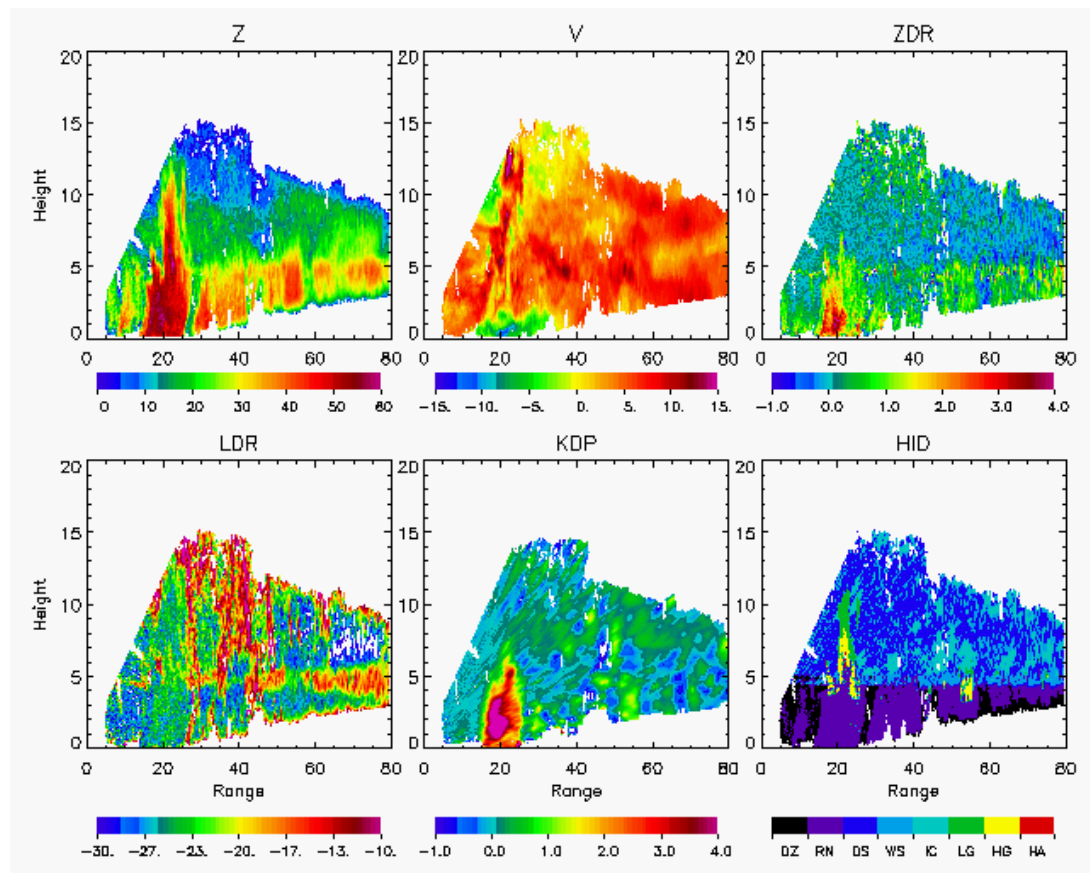


Fig. 3.13. An RHI through a cell at 0821 UTC on 31 May 2008 at an azimuth of 45.5° . Variables are described in Fig. 3.4.

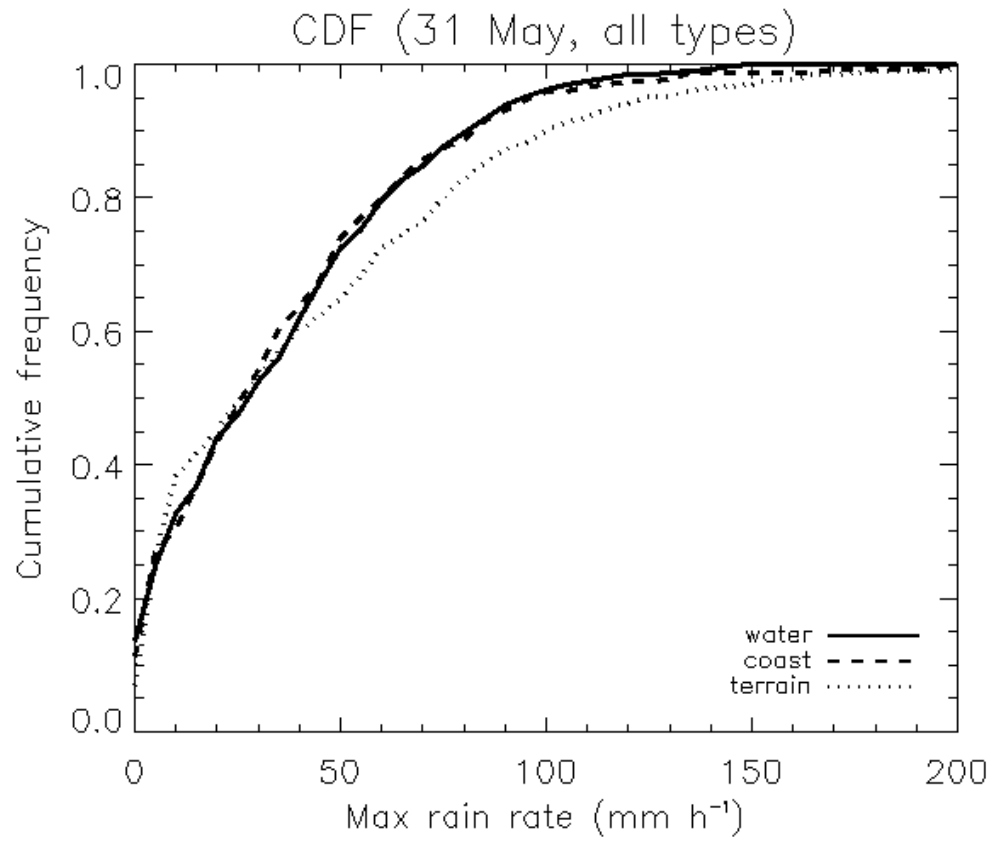


Fig. 3.14. Cumulative distribution functions of maximum rain rate (mm h⁻¹) as a function of elevation for all cells, regardless of type (i.e., isolated and organized), identified on 31 May 2008.

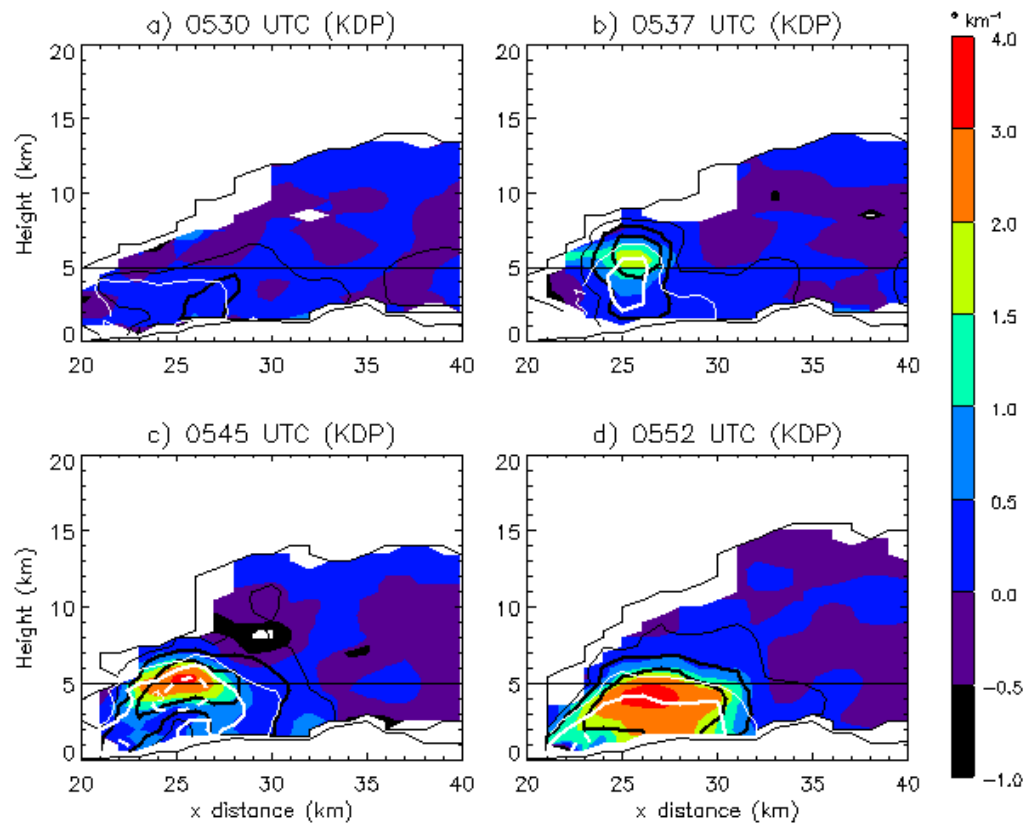


Fig. 3.15. Vertical cross sections of K_{DP} (color-filled contours) for the time evolution of a cell on 1 June 2008 at (a) 0530, (b) 0537, (c) 0545, and (d) 0552 UTC. Other contours are as in Fig. 3.1.

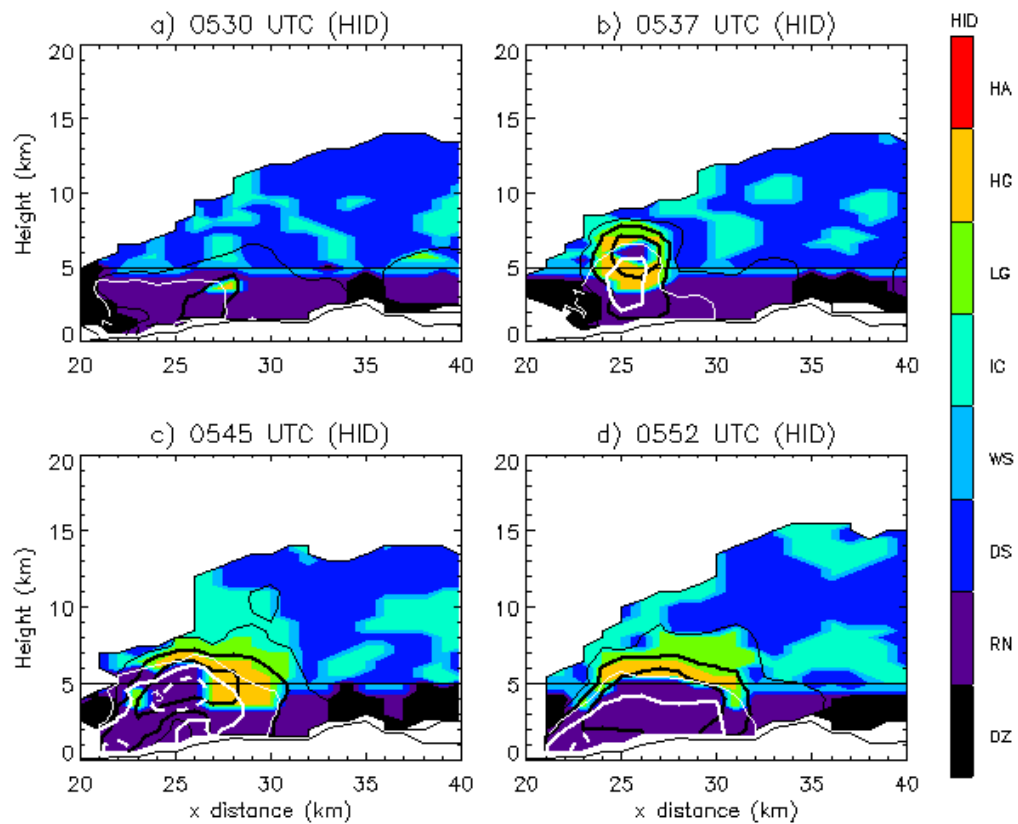


Fig. 3.16. Vertical cross sections of HID (color-filled contours) for the time evolution of a cell on 1 June 2008 at (a) 0530, (b) 0537, (c) 0545, and (d) 0552 UTC. The HID abbreviations and other contours are as described in Fig. 3.2.

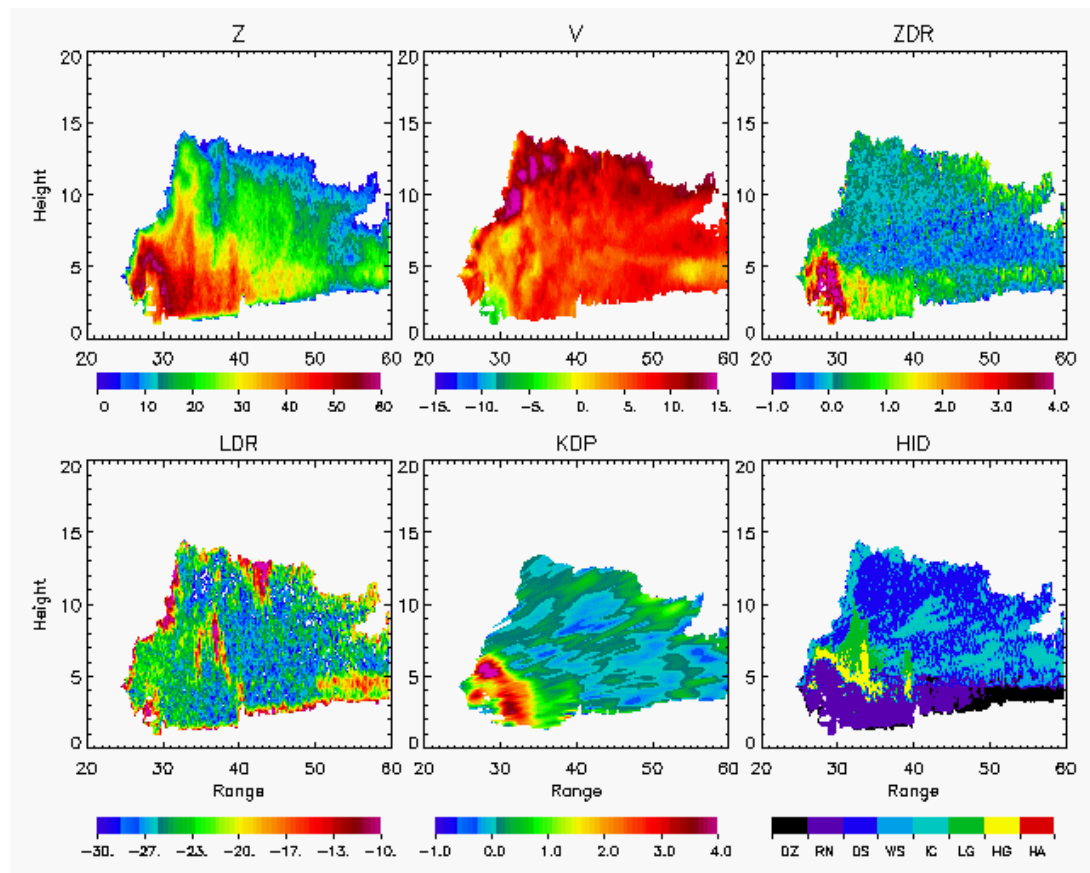


Fig. 3.17. An RHI through a cell at 0551 UTC on 1 June 2008 at an azimuth of 64° . Variables are described in Fig. 3.4.

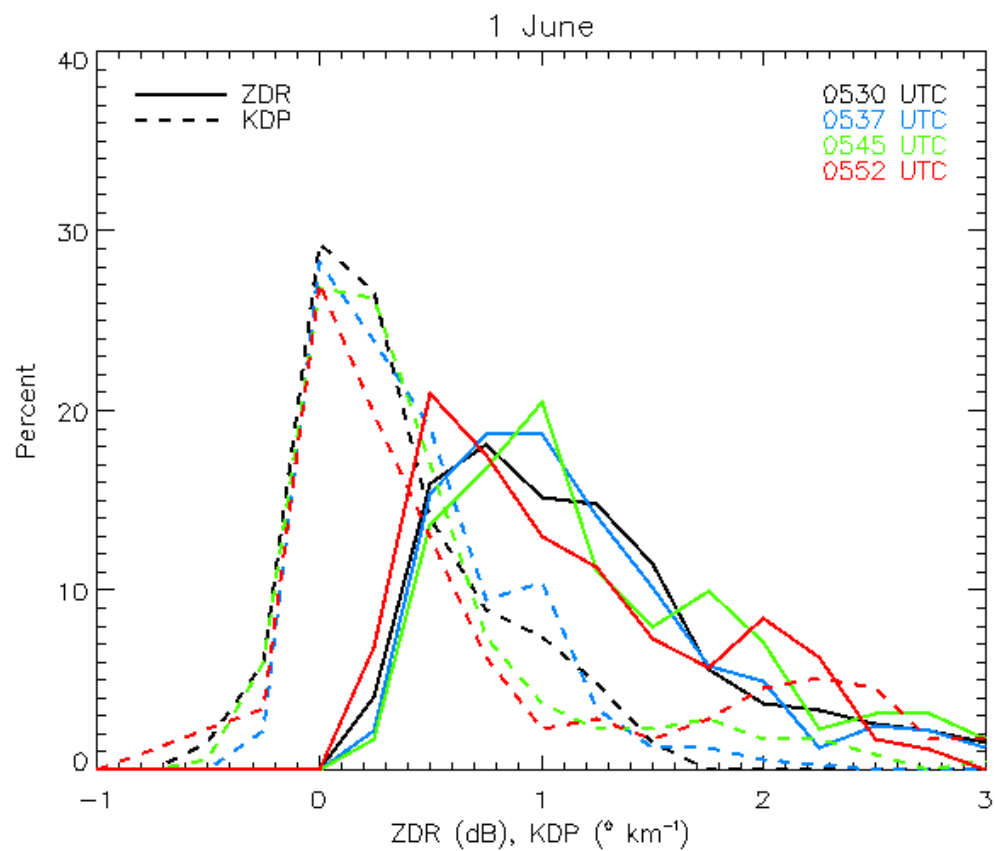


Fig. 3.18. Distributions of Z_{DR} (dB) and K_{DP} ($^{\circ} \text{ km}^{-1}$) values for points identified as rain below 4 km within the cells shown in Figs. 3.15 and 3.16.

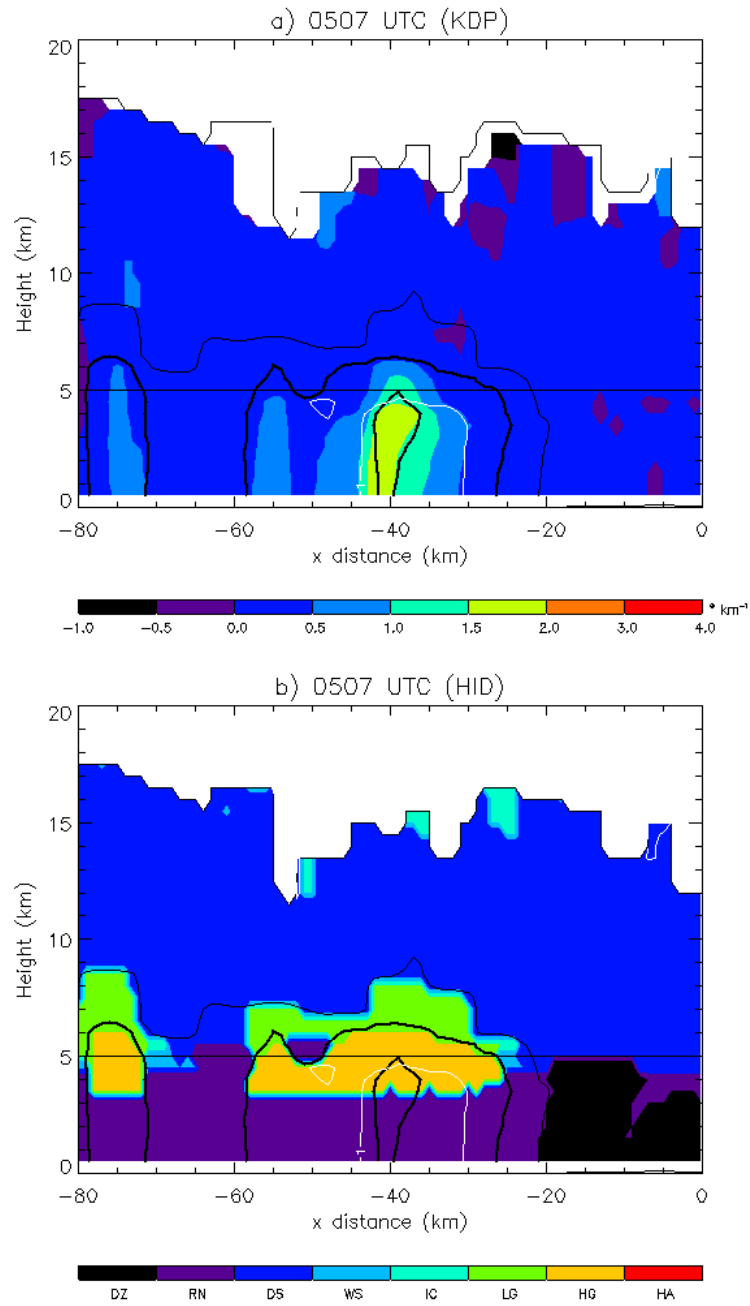


Fig. 3.19. Vertical cross sections through a cell at 0507 UTC on 14 June 2008. Color-filled contours represent (a) K_{DP} and (b) HID. The HID abbreviations are described in Fig. 3.2. Remaining contours are as described in Fig. 3.1.

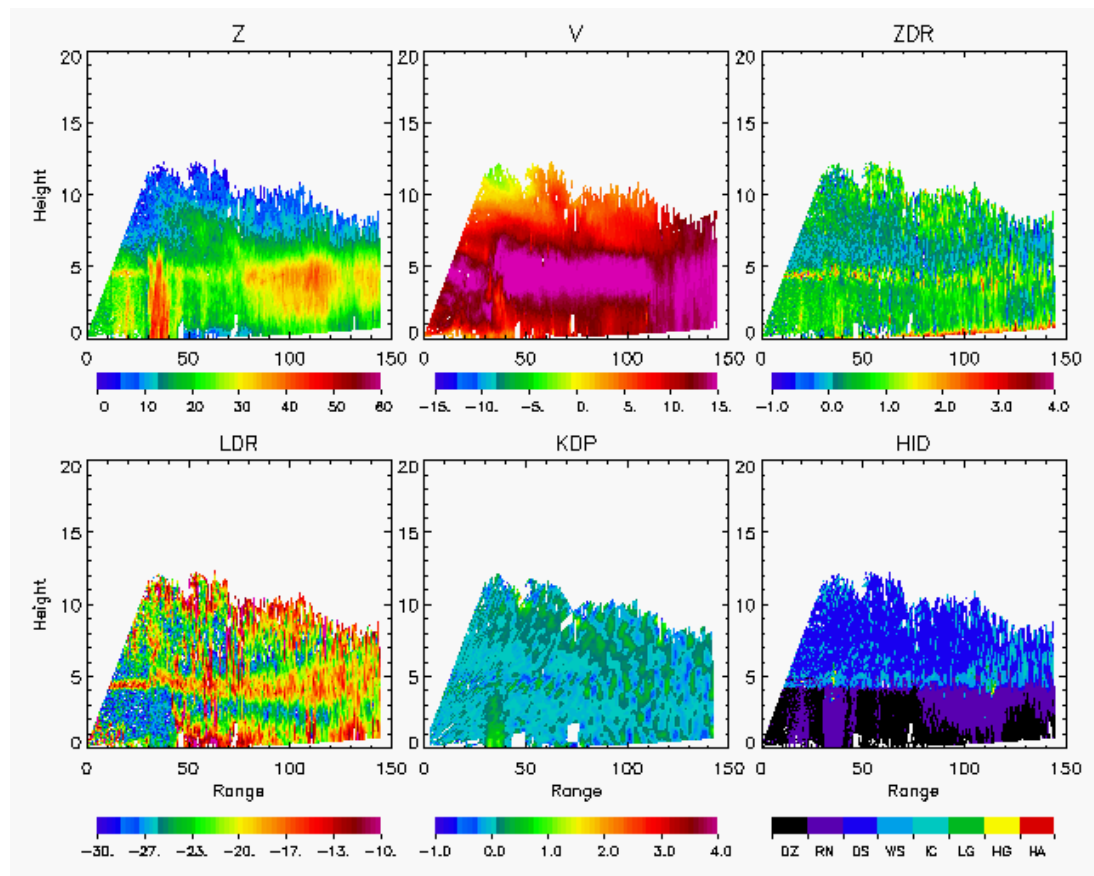


Fig. 3.20. An RHI through a cell at 0751 UTC on 14 June 2008 at an azimuth of 15°. Variables are described in Fig. 3.4.

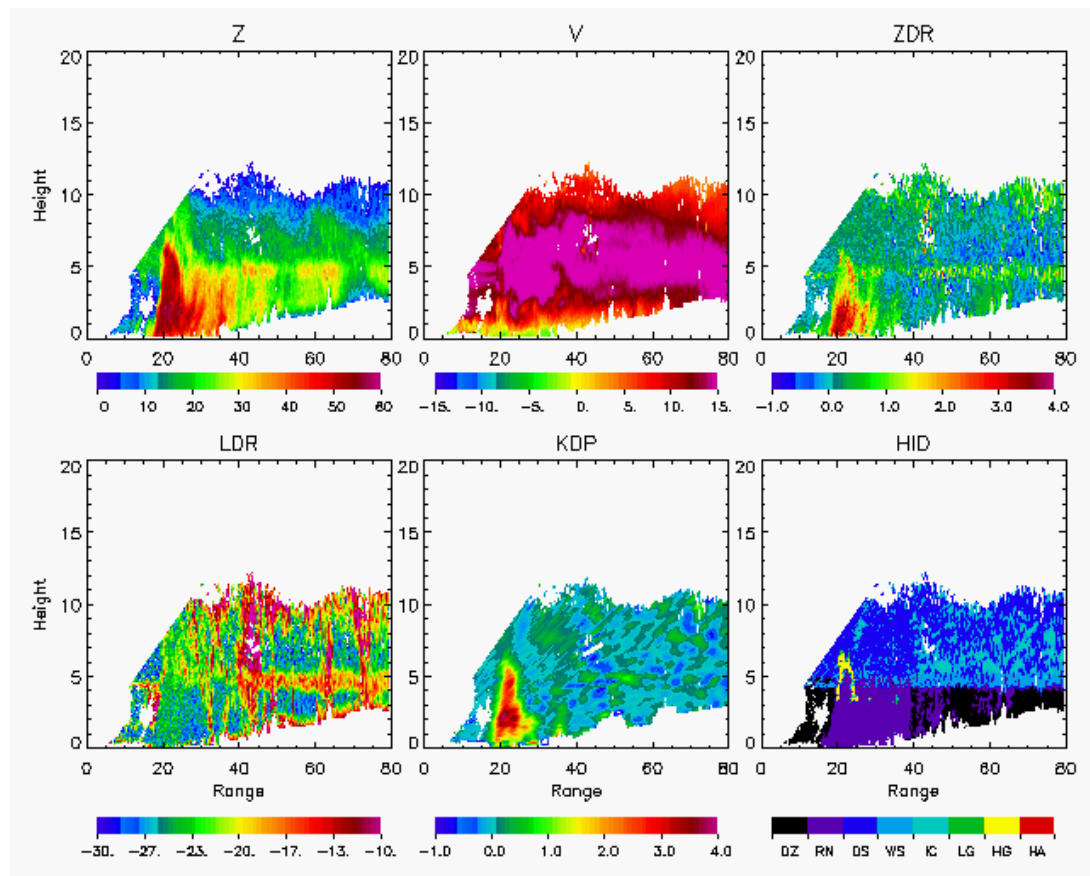


Fig. 3.21. An RHI through a cell at 0851 UTC on 14 June 2008 at an azimuth of 40°. Variables are described in Fig. 3.4.

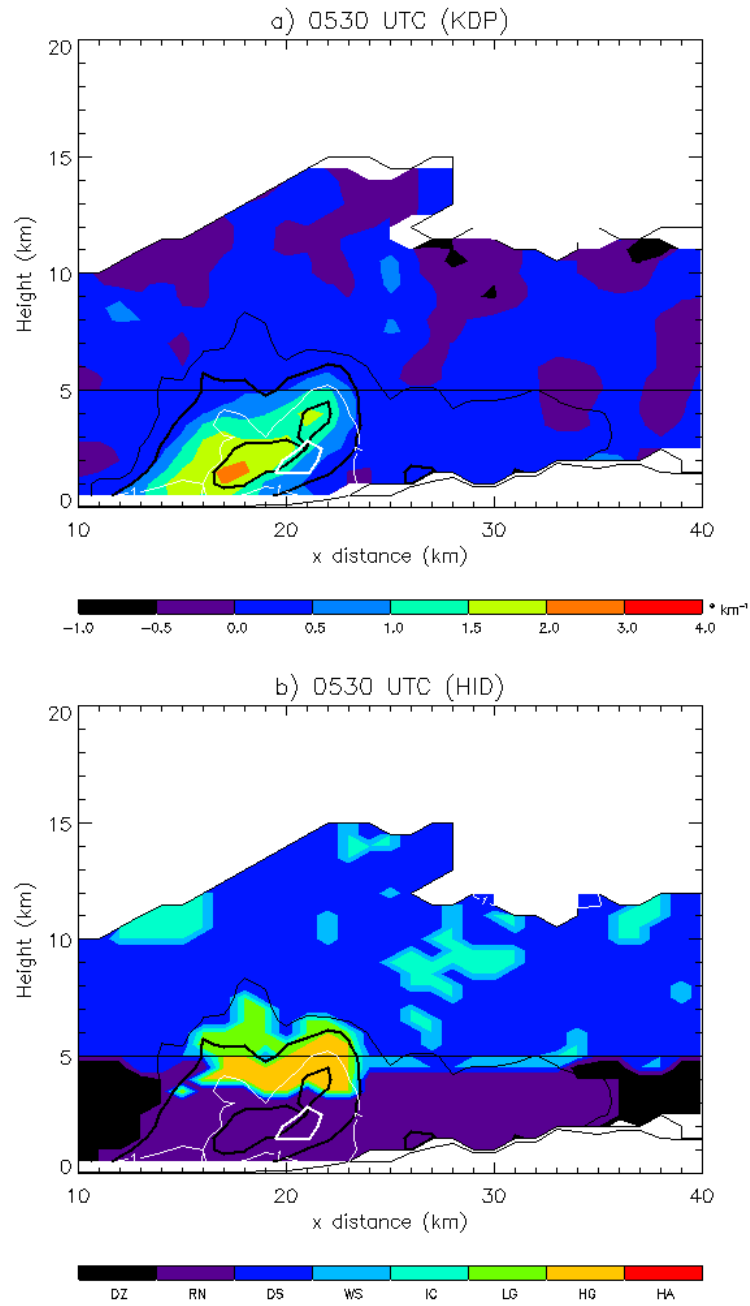


Fig. 3.22. Vertical cross sections through a cell at 0530 UTC on 14 June 2008. Color-filled contours represent (a) K_{DP} and (b) HID. The HID abbreviations are described in Fig. 3.2. Remaining contours are as described in Fig. 3.1.

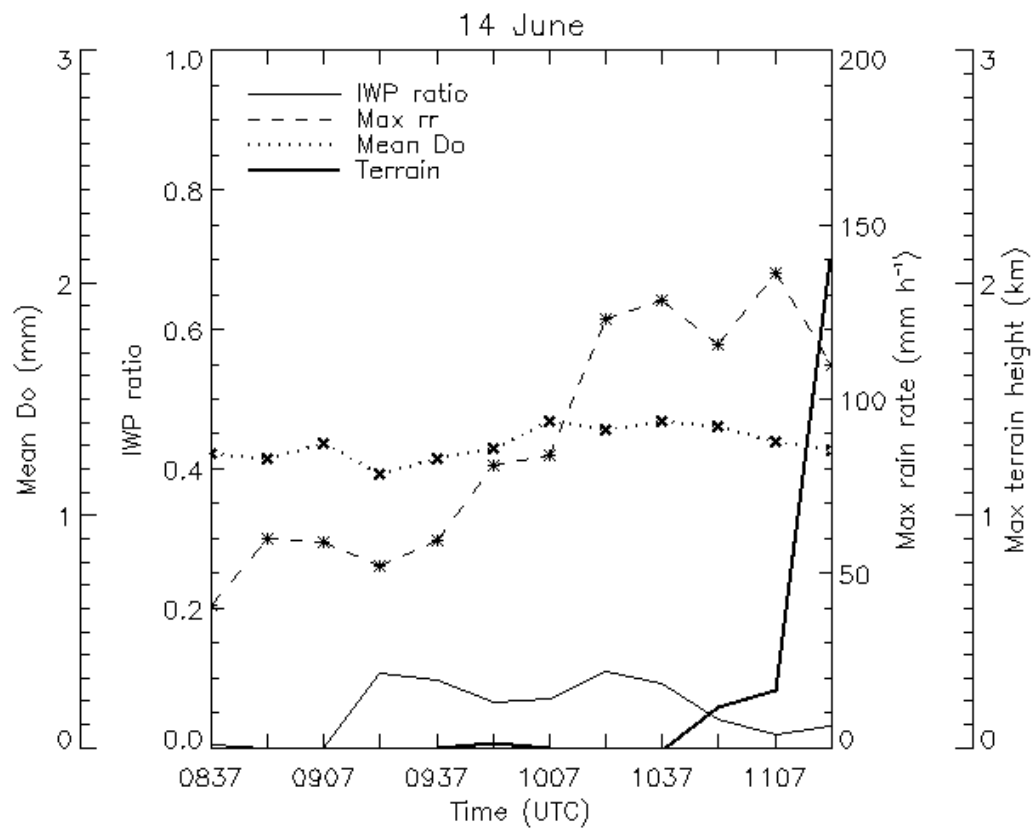


Fig. 3.23. Time series of IWP ratio, maximum rain rate (mm h^{-1}), mean drop diameter (mm), and maximum terrain height (km) for a cell on 14 June 2008.

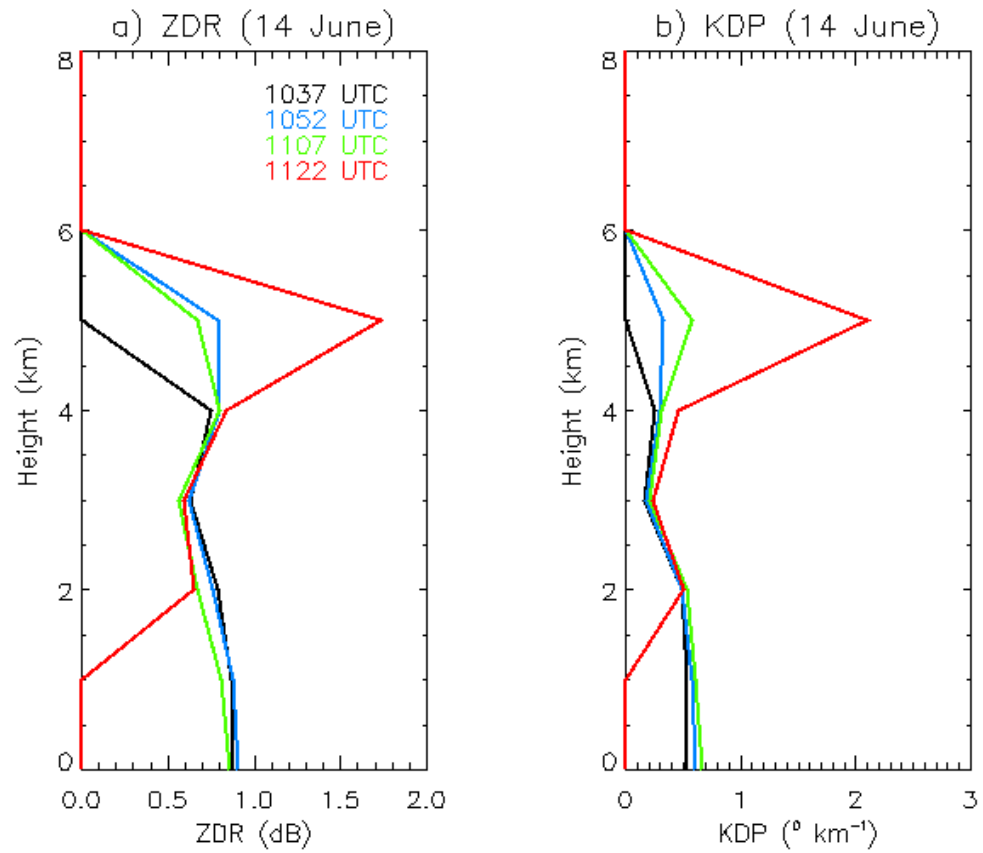


Fig. 3.24. Vertical profiles of mean (a) Z_{DR} (dB) and (b) K_{DP} ($^{\circ} \text{km}^{-1}$) for the period of a track on 14 June when the cells were located at terrain elevations > 0.5 km.

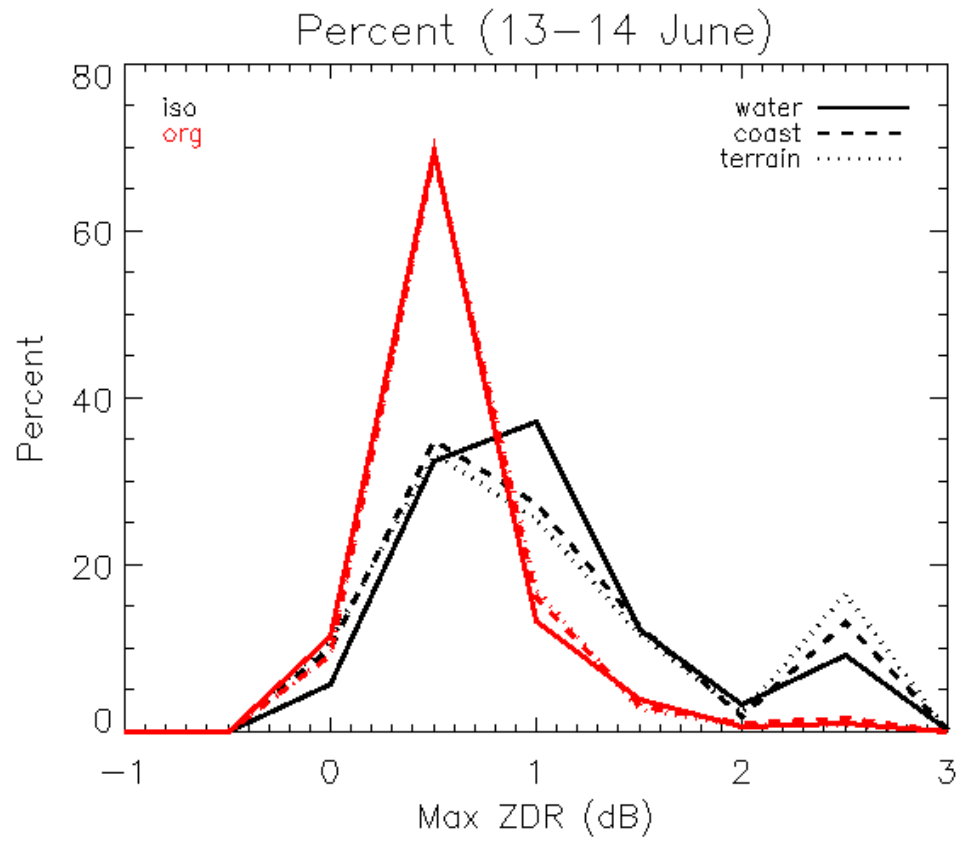


Fig. 3.25. Percentages of Z_{DR} (for points identified as rain below 4 km) as a function of elevation and cell type (i.e., isolated or organized) for all cells identified on 13–14 June 2008.

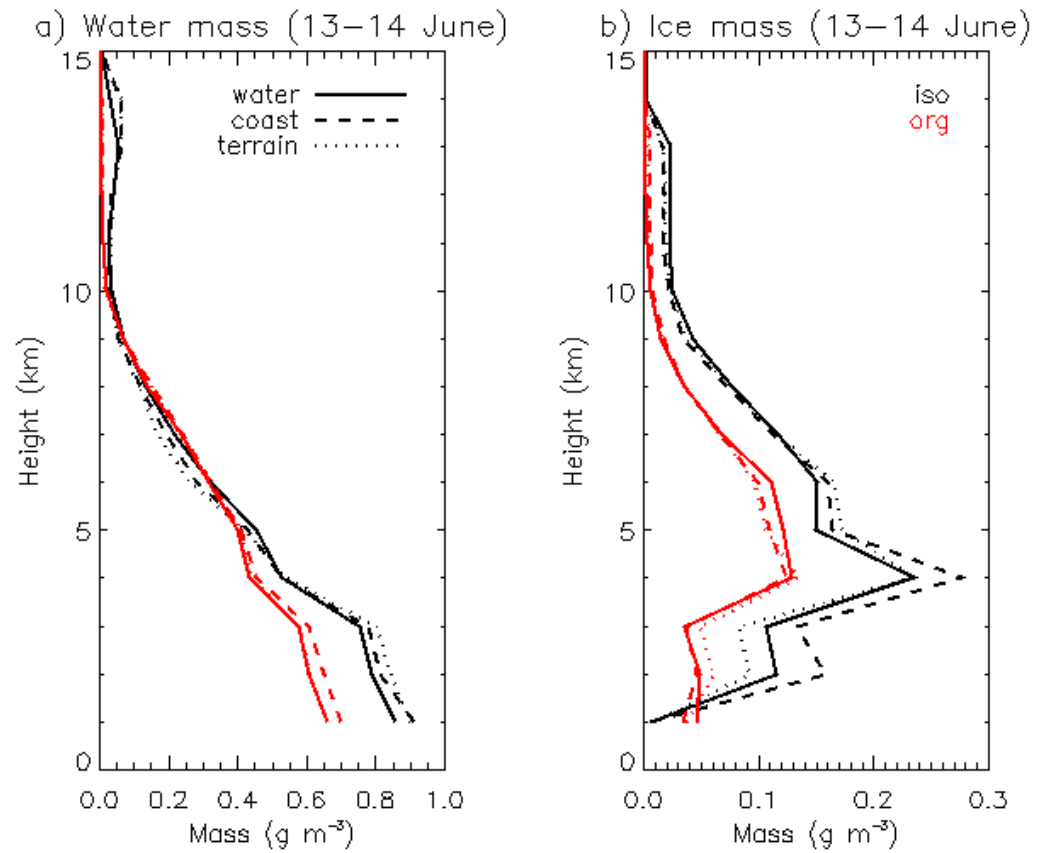


Fig. 3.26. Vertical profiles of (a) mean liquid water mass and (b) mean ice mass as a function of elevation and cell type.

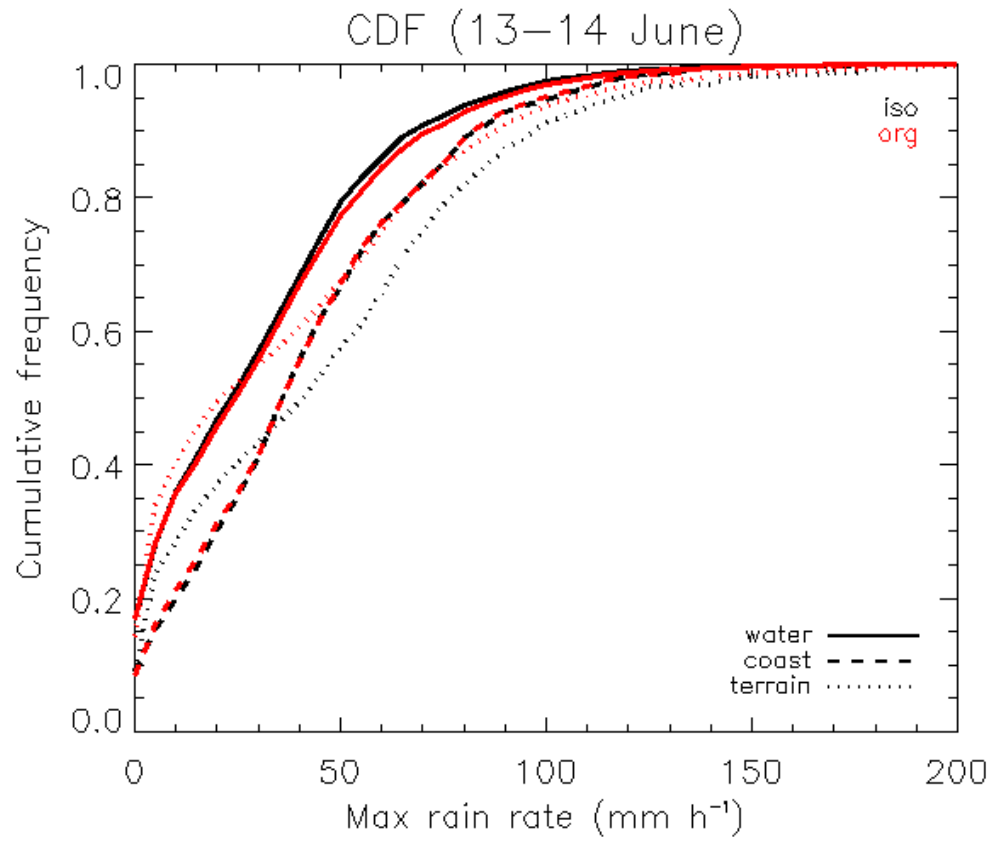


Fig. 3.27. Cumulative distribution functions of maximum rain rate (mm h^{-1}) as a function of elevation and cell type (i.e., isolated or organized) for all cells identified on 13–14 June 2008.

CHAPTER 4

GENERAL STATISTICS

Although case studies provided valuable information regarding the evolution of individual cells, general statistics can place these examples within the broader spectrum of convection that occurred during TiMREX, as well as afford an opportunity for comparisons with other locations. Of all cells identified within the 40 cases, 80% were classified as organized, reflecting the high frequency of MCSs that influence southwestern Taiwan during the Mei-Yu season. Over half (55%) of the organized cells occurred over water, with 60% of the remaining cells over land identified along the coast as systems moved onshore. Isolated convection occurred primarily over land, with approximately half of those cells located along the higher terrain; only 36% of isolated cells were located over water.

4.1 Type comparison

Differences in maximum instantaneous rain rates between features types for all cells were investigated, as the example from 13-14 June revealed greater rates for isolated cells. Cumulative distribution functions of maximum rain rate for all elevations (Fig. 4.1) show only a small difference between isolated and organized cells, with isolated convection characterized by slightly greater instantaneous rain rates. Distributions of echo-top height (Fig. 4.2), however, reveal that organized cells tended to be deeper than

isolated cells, with respective means of 12 and 9 km. When rain rates of isolated convection are sub-divided by these echo-top heights (Fig. 4.3), it is apparent that shallow convection produced similar, if not slightly greater, rates as the deep cells. This is consistent with the cells described in Ch. 3 that were characterized by intense rainfall despite the majority of echo existing below the melting level, implying efficient warm-rain processes.

In addition, distributions of maximum Z_{DR} (Fig. 4.4) show greater percentages of values > 1 dB for isolated cells and a wider range of values, implying a broader DSD compared to organized convection. Distributions of maximum K_{DP} (Fig. 4.5) also show a higher frequency of large values for isolated cells suggesting greater liquid water contents and/or greater concentrations of highly oblate drops compared to organized cells.

Slightly larger mean drop sizes in isolated convection were observed regardless of rain rate (Fig. 4.6), with organized cells producing similar rain rates via smaller drops, further suggesting differences in DSDs. Despite intense instantaneous rain rates, likely associated with larger drops compared to organized convection, shallow isolated cells, dominated by warm-rain processes, were responsible for only $< 1\%$ of the total hourly rain mass, indicating a relatively insignificant contribution to total rainfall despite intense instantaneous rain rates.

4.2 Elevation comparison

Examples from the case studies indicated increased ice production along the western slopes, for both isolated and organized cells, leading to enhanced rain rates. To further investigate these potential elevation-dependent trends, organized and isolated cells

were further subdivided based on topography: water, coast (0–0.5 km), and terrain (> 0.5 km). The CDFs of maximum rain rate (Fig. 4.7), divided by type and elevation, show the greatest rates along the terrain for both isolated and organized cells, consistent with cell tracks described in Ch. 3. Cells over water, regardless of type, produced the weakest rain rates, with generally greater intensities for organized cells compared to isolated convection. Over the higher terrain, isolated cells had greater rainfall intensities compared to organized cells. Distributions of echo-top height (Fig. 4.8) indicate that organized convection was deeper than isolated convection regardless of elevation, although peaks at 16 km suggest that isolated convection, on occasion, reached greater heights. Also, organized convection exhibited little difference between elevation groups, whereas isolated cells over water were slightly deeper than those over land.

Previous studies have used the maximum height of the 30-dBZ contour as another means to describe the vertical intensity of storms (e.g., Zipser 1994; Petersen et al. 1996), linking the production of lightning to the extension of 30-dBZ reflectivities above the melting level. Cumulative distributions of the maximum height of 30 dBZ (Fig. 4.9) reveal that at least 70% of cells, regardless of type or elevation, had 30-dBZ contours extending above this level, suggesting the coexistence of ice and supercooled water in the majority of cells during TiMREX. In addition, maximum 30-dBZ heights were comparable for all elevation groups and types during TiMREX (Fig. 4.9) despite major differences in echo-top height, similar to results from NAME (Rowe et al. 2008) and TOGA COARE (DeMott and Rutledge 1998), suggesting that convection, whether shallow or deep, was just as intense in terms of rain rates. However, as was previously mentioned, shallow, isolated convection contributed negligibly to hourly accumulated

rain mass, whereas 93% of total rain was associated with deep organized systems extending above 11 km. In addition, of the total hourly rain mass associated with these organized cells, 46% occurred over the terrain, with 34% over the coast, and the remaining 20% over water, further suggesting elevation-dependent trends in microphysicals.

Distributions of maximum Z_{DR} below the melting level (Fig. 4.10) show the most frequent occurrence of $Z_{DR} > 1$ dB over the terrain for both cell types, with greater percentages of large drops in isolated convection. An analysis of mean drop-size diameter as a function of maximum rain rate (Fig. 4.11) suggests that a given rain rate was produced by smaller drops, on average, for organized convection, with the smallest mean diameters corresponding to organized systems over water, implying more maritime-like DSDs (e.g., Atlas and Ulbrich 2000, 2006; Bringi et al. 2003, 2009; Rosenfeld and Ulbrich 2003; Ulbrich and Atlas 2007). Recall, however, that cells over water during the 13–14 June period contained slightly greater amounts of ice mass than over the terrain despite smaller mean drop diameters. Cumulative distribution functions of ice mass fraction (Fig. 4.12) reveal that for a given degree of organization, cells over the terrain had the greatest fractions, due, in part, to the truncated depth below the melting level. It is interesting to note that cells over water had a greater ice mass fraction than those over the coast, consistent with the greater ice masses in organized convection off the coast shown in the case study. Vertical profiles of mean ice mass for these groupings (Fig. 4.13b) also show that cells over water, for both types, contained larger amounts of ice mass near the melting level, although greater ice mass at higher altitudes was associated with the occasional deep, isolated cells along the CMR (e.g., 31 May, 1 June).

Profiles of mean liquid water mass (Fig. 4.13a) indicate greater masses below the melting level for isolated convection with previously implied broader DSDs compared to organized cells. Consistent with the suggestion that organized cells contained relatively small drops, especially over water, frequency contours of maximum Z_{DR} versus maximum K_{DP} for organized cells (Fig. 4.14) show larger K_{DP} for values of $Z_{DR} < 3$ dB over the water (Fig. 4.14a), indicating a greater contribution from small drops to high liquid water mass. Larger drops contributed to greater K_{DP} over the higher elevations (Fig. 4.14c), further suggesting a role of ice-based processes in producing intense rainfall as those systems impinged on the terrain.

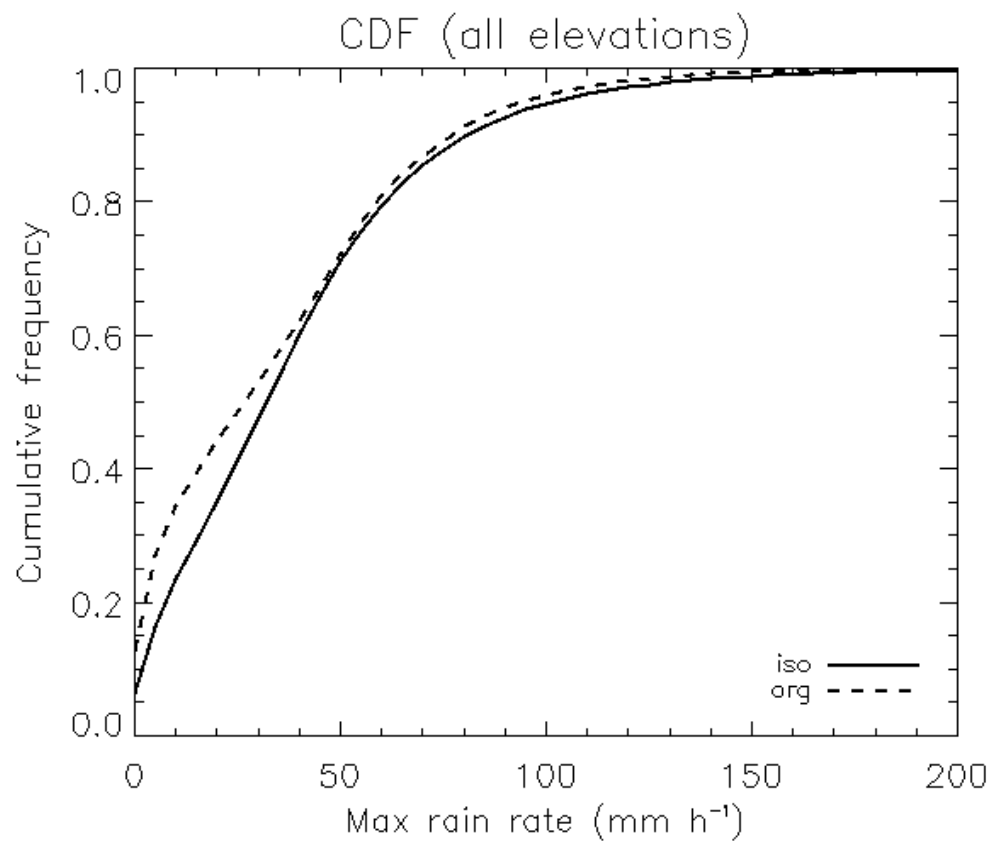


Fig. 4.1. Cumulative distribution functions of maximum rain rate (mm h^{-1}) for isolated (iso) and organized (org) cells during TiMREX over all elevations.

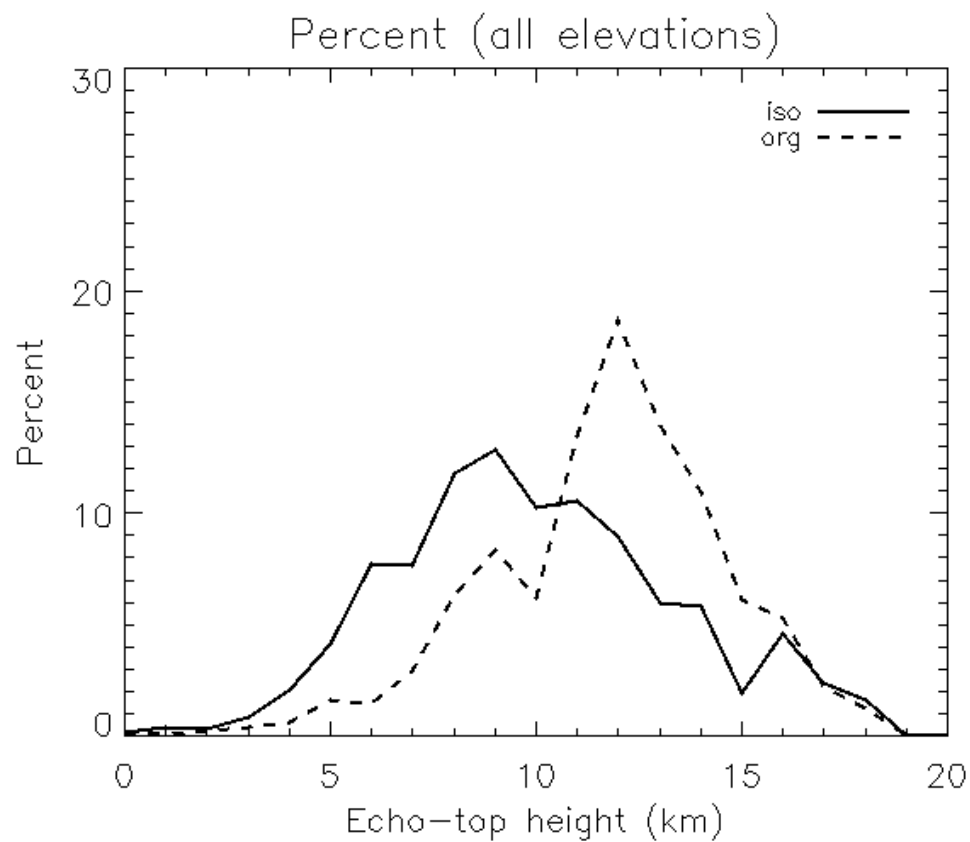


Fig. 4.2. Distributions of echo-top height (km) for isolated (iso) and organized (org) cells during TiMREX over all elevations.

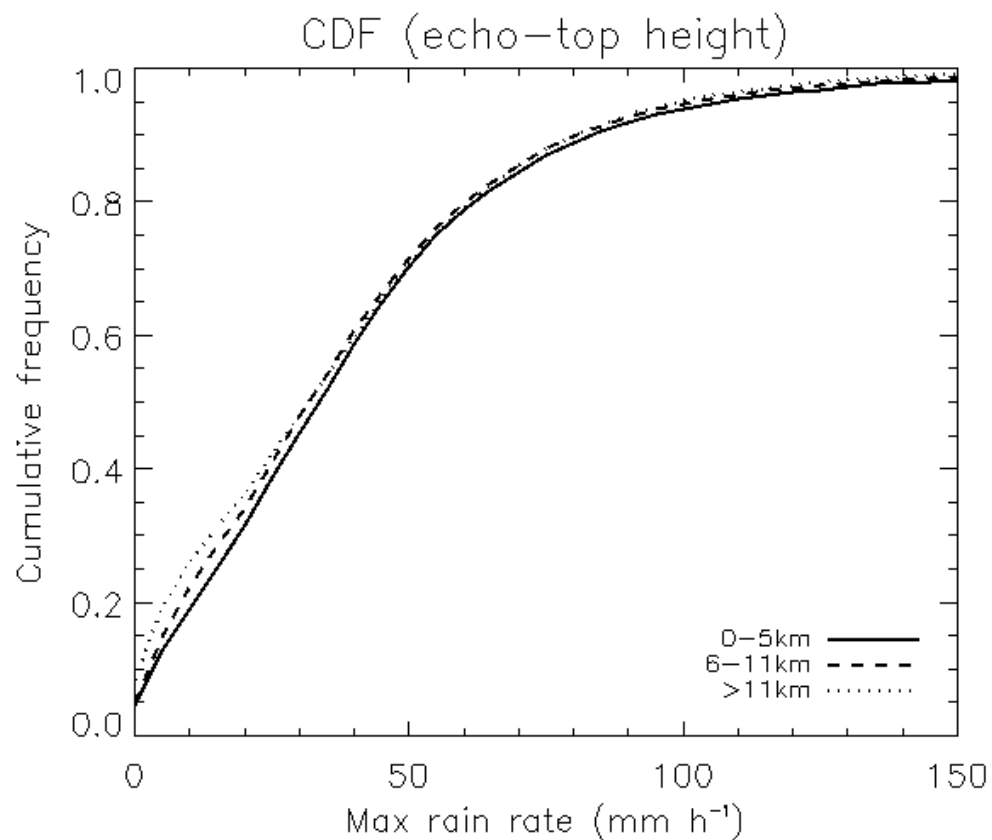


Fig. 4.3. Cumulative distribution functions of maximum rain rate (mm h⁻¹) for all isolated cells during TiMREX, grouped by echo-top height.

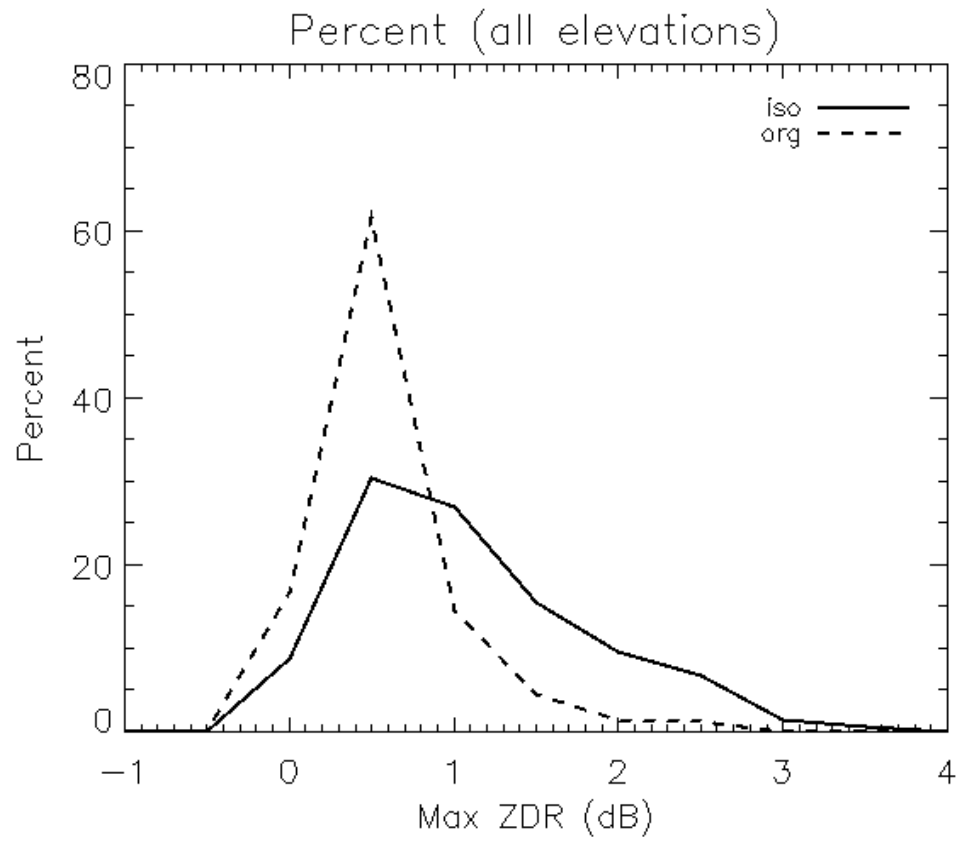


Fig. 4.4. Distributions of maximum Z_{DR} (dB) below 4 km for isolated (iso) and organized (org) cells during TiMREX over all elevations.

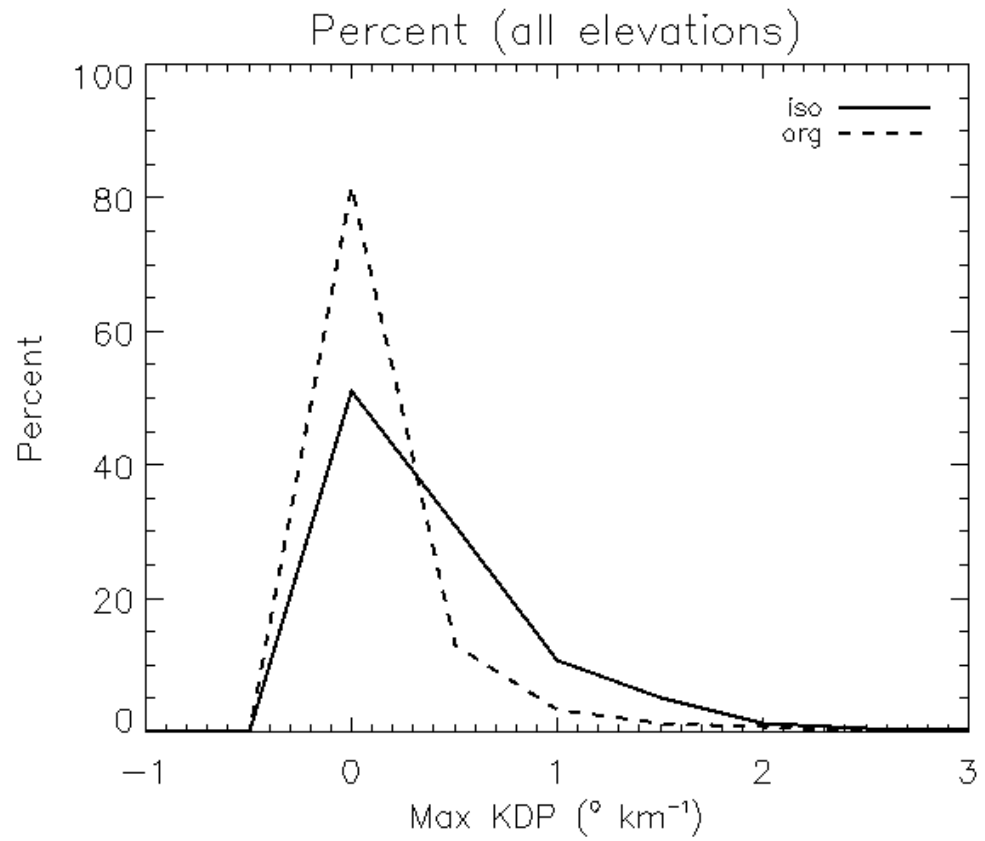


Fig. 4.5. Distributions of maximum K_{DP} ($^{\circ} \text{ km}^{-1}$) below 4 km for isolated (iso) and organized (org) cells during TiMREX over all elevations.

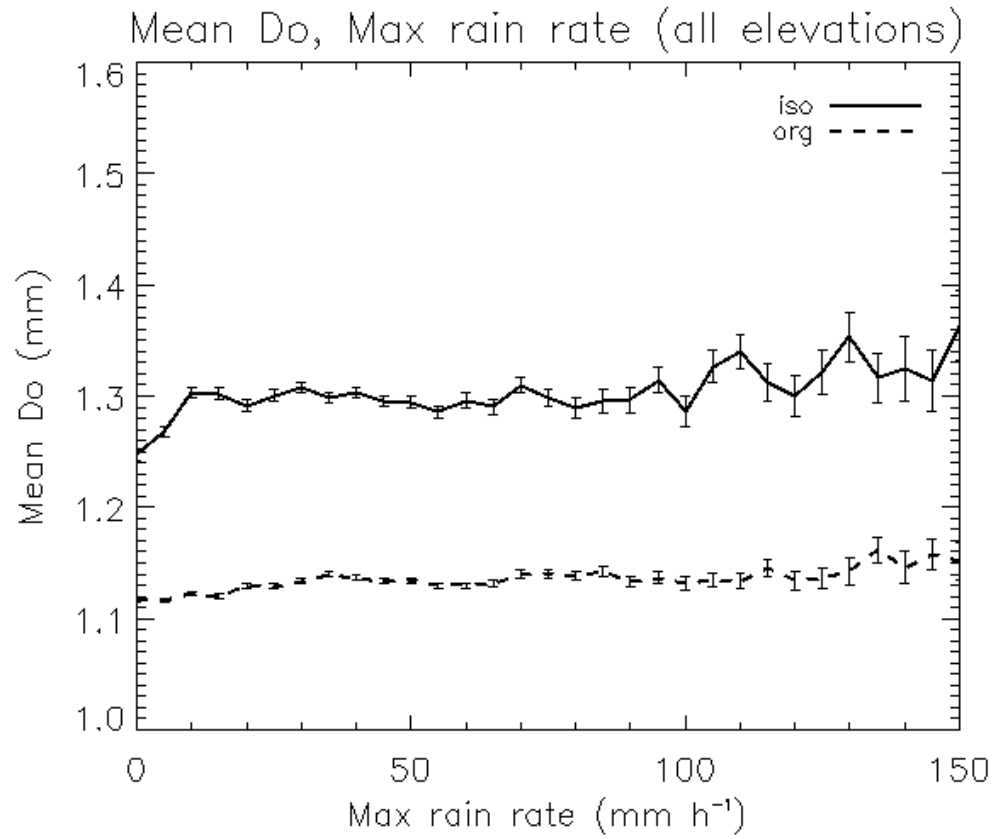


Fig. 4.6. Mean drop diameter (mm) as a function of maximum rain rate (mm h^{-1}) for all cells during TiMREX, divided by cell type (i.e., isolated and organized). Error bars are associated with 95% confidence intervals for each grouping.

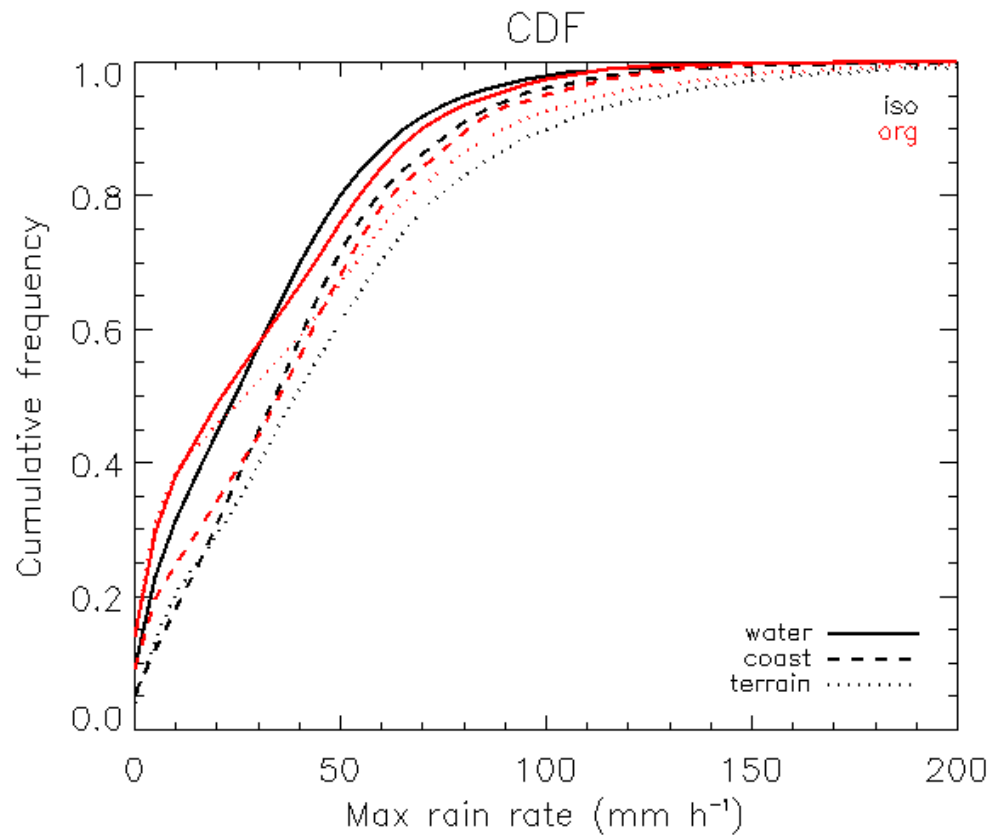


Fig. 4.7. Cumulative distribution functions of maximum rain rate (mm h⁻¹) for all cells during TiMREX, grouped by cell type and elevation.

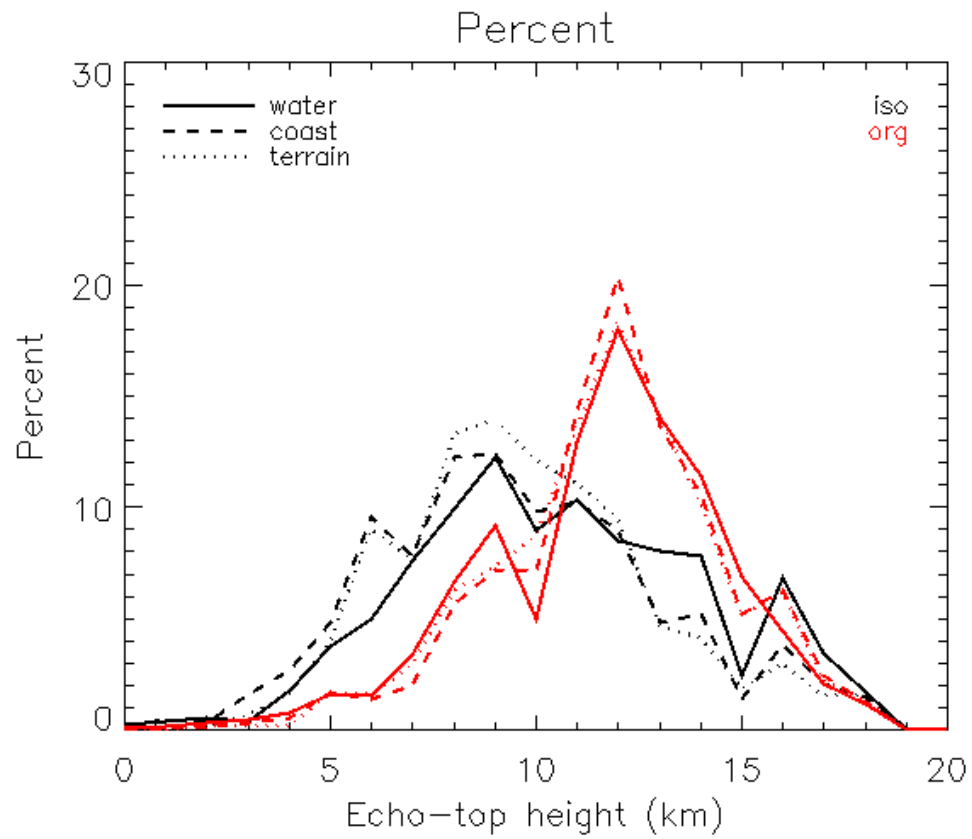


Fig. 4.8. Distributions of echo-top height (km) for all cells during TiMREX, grouped by cell type and elevation.

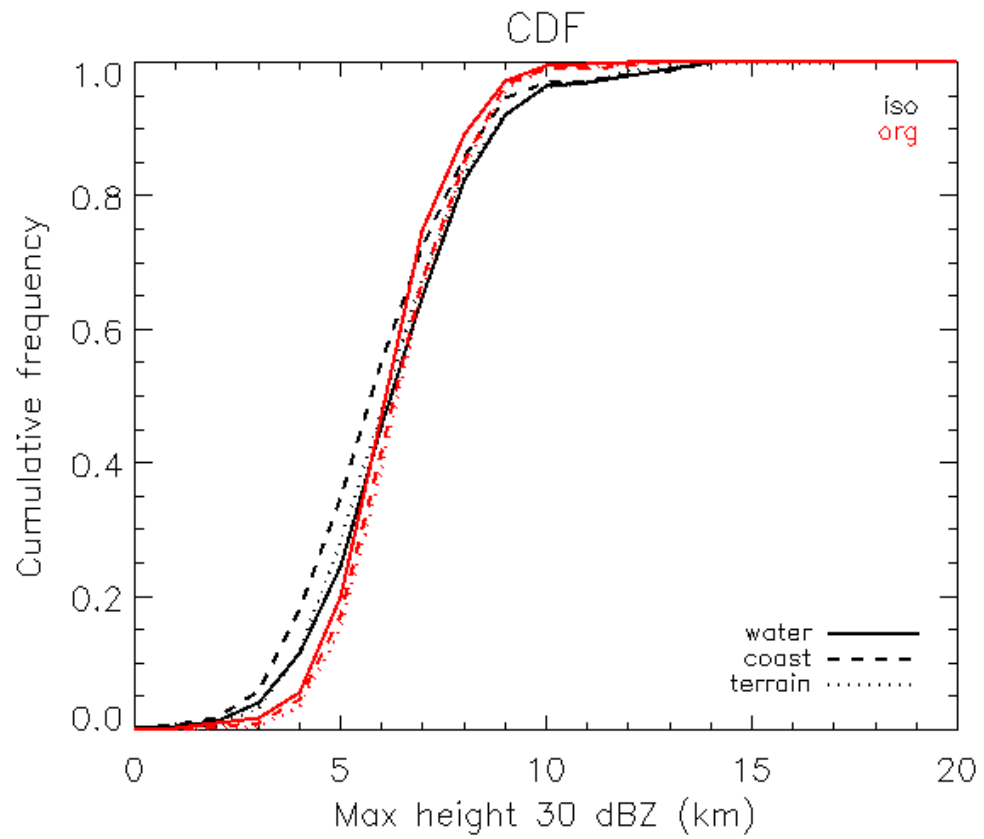


Fig. 4.9. Cumulative distribution functions of the maximum height of 30 dBZ (km) for all cells during TiMREX, grouped by cell type and elevation.

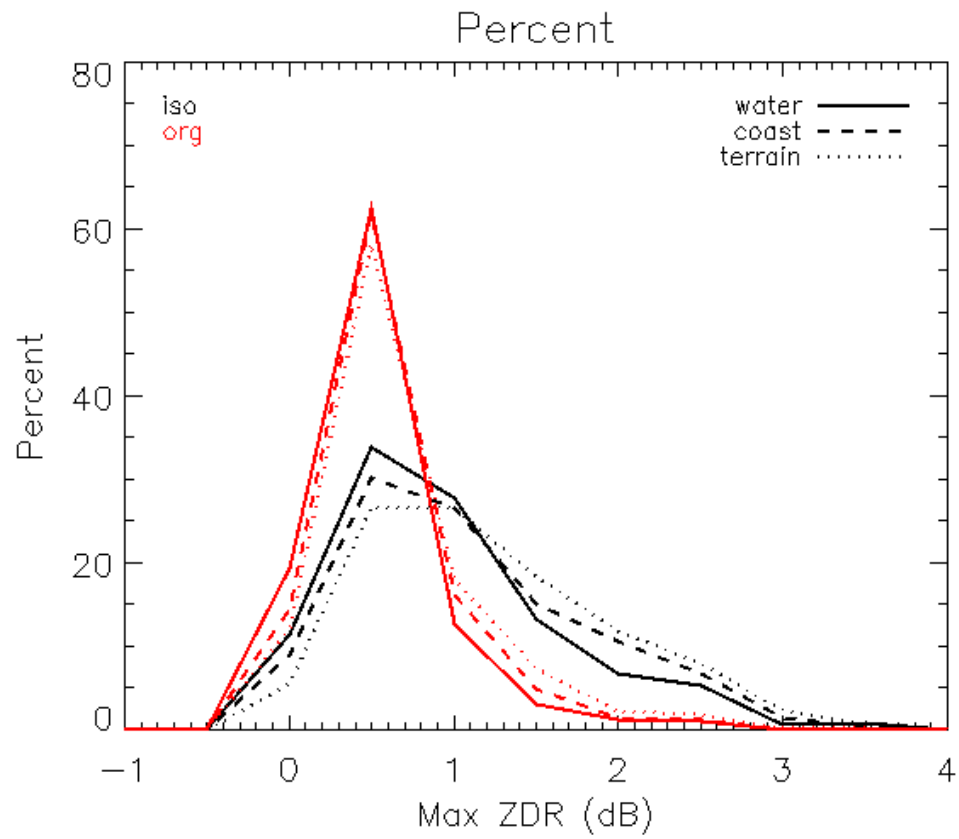


Fig. 4.10. Distributions of maximum Z_{DR} (dB) below 4 km for all cells during TiMREX, grouped by cell type and elevation.

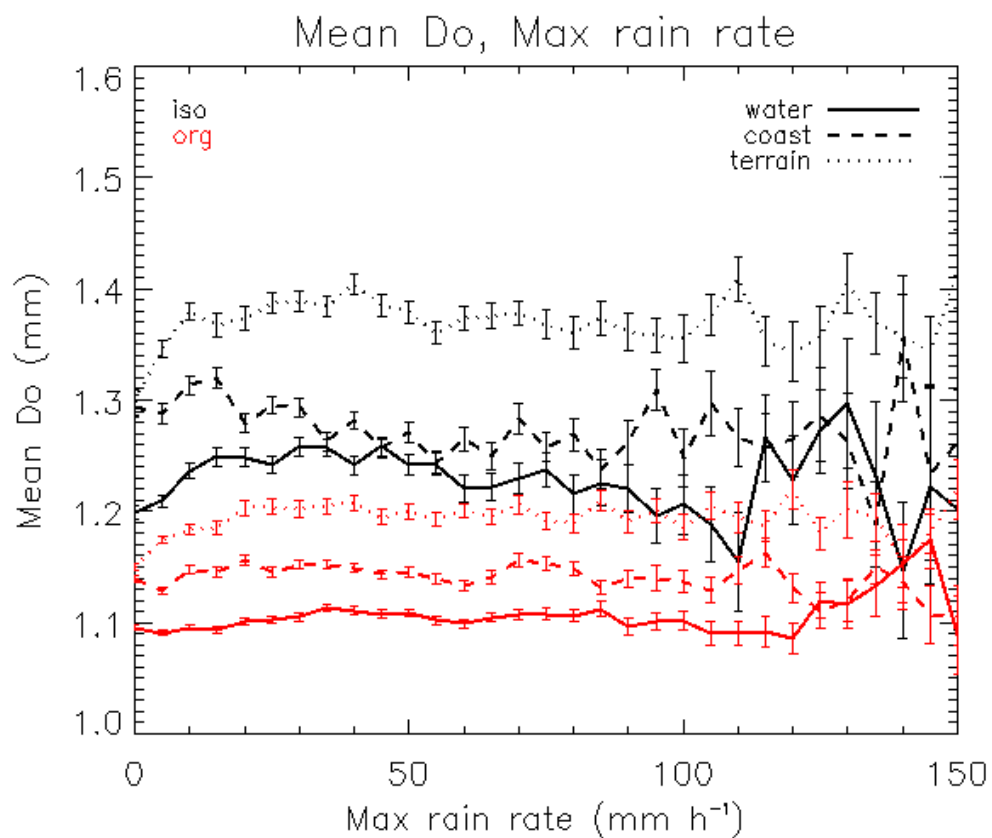


Fig. 4.11. Mean drop diameter (mm) as a function of maximum rain rate (mm h⁻¹) for all cells during TiMREX, grouped by cell type and elevation. Error bars are associated with 95% confidence intervals for each grouping.

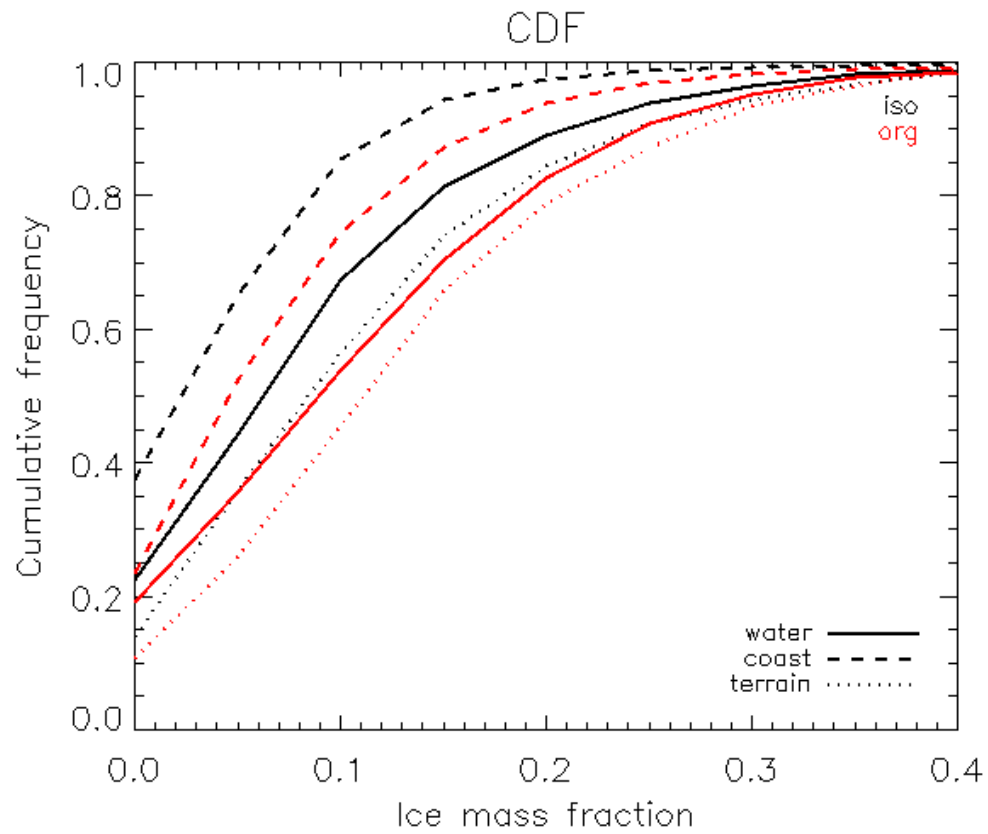


Fig. 4.12. Cumulative distribution functions of ice mass fraction for all cells during TiMREX, grouped by cell type and elevation.

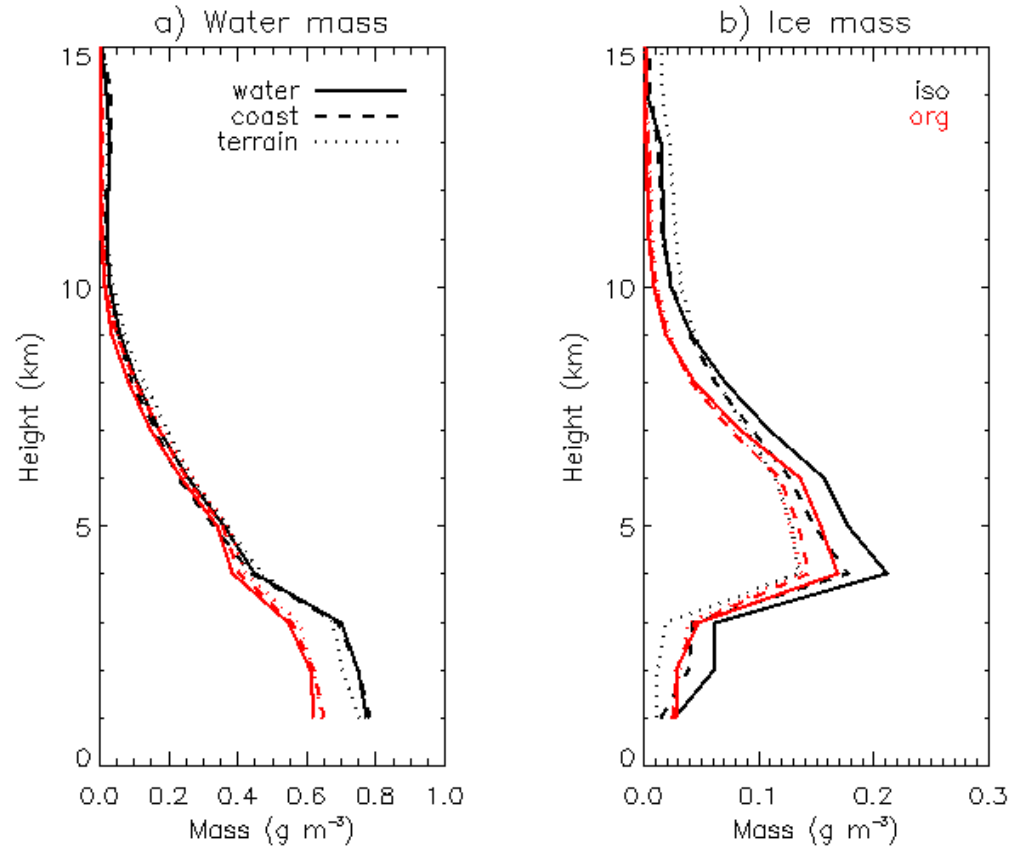


Fig. 4.13. Vertical profiles of (a) mean liquid water mass and (b) mean ice mass, grouped by cell type and elevation.

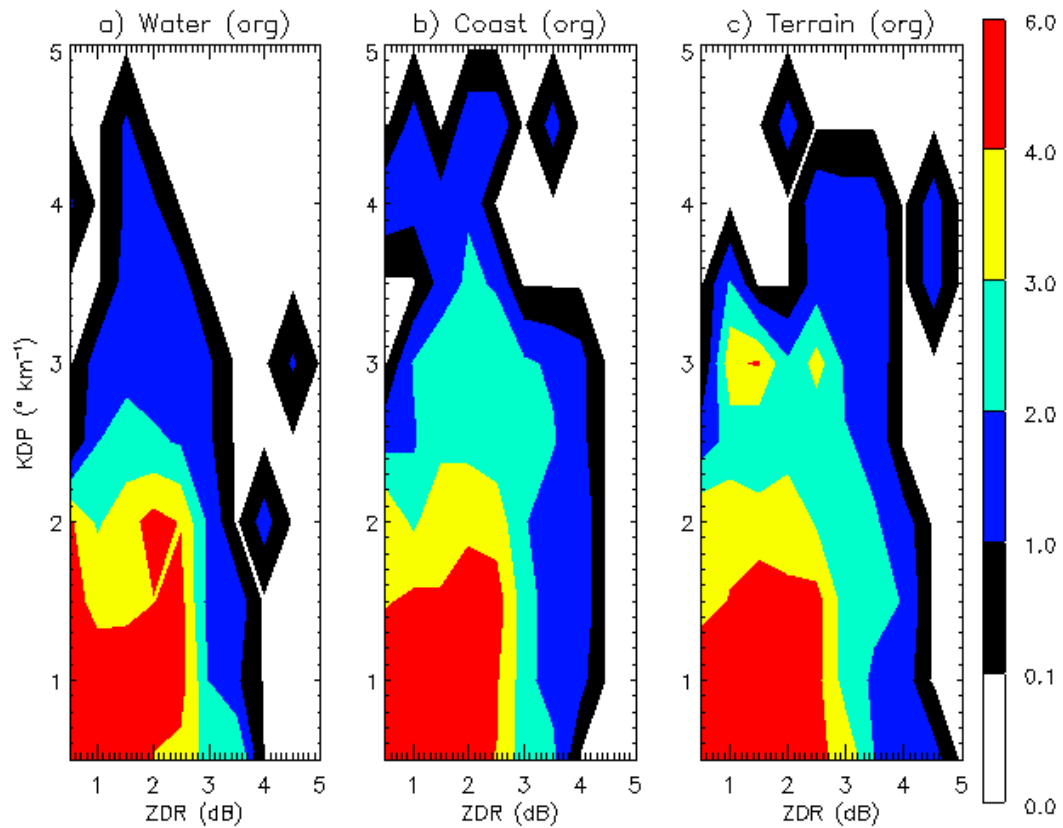


Fig. 4.14. Contoured frequencies of K_{DP} as a function of Z_{DR} for organized cells (a) over water, (b) over the coast (0–0.5 km), and (c) over the terrain (> 0.5 km). Only values identified as rain by the HID located below 4 km were included. Plotted are the logarithms of frequencies normalized by the total frequency in the specified elevation grouping.

CHAPTER 5

COMPARISON WITH NAME

The analyses presented in Ch. 4 allowed for a comparison with studies of isolated and organized convection observed during NAME (Rowe et al. 2011a,b). While both southwestern Taiwan and northwestern Mexico are influenced by diurnally driven convection, the latter region is characterized by weaker, short-lived cells over the highest peaks of the local topography. In Taiwan, however, as shown in this study, the western slopes of the CMR receive the most intense rainfall primarily from isolated cells, although organized systems produce the greatest rain totals in this region. In contrast to the systems organizing along the Sierra Madre Occidental (SMO) and moving toward the coast within easterly flow during the overnight hours during NAME, the TiMREX domain was characterized by large systems moving onshore within southwesterly flow during the late-evening hours, interacting with the Central Mountain Range (CMR) by morning and leading to enhanced rainfall.

Case-based and statistical studies of NAME convection, presented in Parts I and II, revealed a significant contribution from ice-based processes, especially over the higher elevations. While ice has been shown to play a role in producing intense rainfall along the western slopes of the CMR (Ch. 3), shallow convection dominated by warm-rain processes led to as intense rain rates as deeper convection associated with organized systems during TiMREX. On the other hand, organized systems produced the greatest

accumulated hourly rainfall during TiMREX, highlighting an interest for comparisons between both instantaneous and cumulative rainfall for all cells during both experiments, in addition to assessing potential differences in microphysical processes.

Cumulative distribution functions of maximum rainfall rate (Fig. 5.1) show similar rates when considering all cells. Isolated convection during TiMREX appears to have been slightly more intense in terms of rainfall rates compared to NAME, after dividing the isolated cells further by elevation (Fig. 5.2), it is apparent that these rates were greatest in cells over land. There was less distinction between elevation groups for organized convection (Fig. 5.3), especially along the sloped terrain where rain rates tended to be maximized in both regions. Distributions of echo-top height for all elevations (Fig. 5.4) show a trend of deeper convection during NAME for both isolated and organized cells. Convection during NAME also had a greater percentage of 30-dBZ reflectivities extending above 9 km compared to isolated and organized cells from TiMREX (Fig. 5.5), suggesting greater intensities and, therefore, vertical motions. However, recall that rain rates were similar between the two regions, with isolated convection in Taiwan producing slightly greater rates than those in northwestern Mexico despite shallower echo tops and reduced 30-dBZ heights. The division of rain rates in terms of echo-top height for isolated cells (Fig. 5.6) shows that these shallower cells produced greater instantaneous rain rates than the deep convection during NAME, with an even clearer distinction when focusing only on isolated convection along the slopes (1–2 km) in both regions (Fig. 5.7), implying an emphasis on warm-rain processes during TiMREX.

In addition to these differences in instantaneous rain rates, CDFs of hourly accumulated rain mass (Fig. 5.8) show greater values during TiMREX, with the greatest distinction observed for organized systems. After further dividing the organized cells by terrain, it is clear that this trend is independent of elevation (Fig. 5.9). It was shown that shallow, isolated cells during TiMREX produced as intense rain rates as deep isolated convection during NAME; however, the former only contributed < 1% of hourly accumulated rain during the project, while the latter provided 42% of total rain mass during NAME. Regardless of the region, deep, long-lived organized systems contributed significantly to cumulative rainfall, especially along the sloping terrain, as was described in previous case studies. Differences in rain totals from the two projects, however, also suggest varying microphysical processes between these regions that lead to this intense rainfall, which were further examined using the polarimetric data from the experiments.

Distributions of maximum Z_{DR} for isolated cells (Fig. 5.10) suggest broader DSDs during TiMREX with smaller means and higher percentages of $Z_{DR} > 2$ dB compared to NAME for all land elevations. When considering organized convection (Fig. 5.11), generally larger drops were observed during NAME for all land elevations, consistent with results from Rowe et al. (2011a,b) that suggested a significant contribution from melting ice for production of intense rainfall as systems organized along the western slopes of the SMO. While it was previously suggested that ice become increasingly important along the terrain of the CMR in Taiwan, vertical profiles of mean ice mass (Fig. 5.12b) clearly show a greater amount of ice associated with NAME convection for both cell types. The deep vertical extent of greater ice mass in NAME cells also reflects the stronger vertical motions suggested by higher 30-dBZ heights compared to TiMREX.

Interestingly, mean profiles of water mass (Fig. 5.12a) also show greater values for NAME for both organized and isolated cells compared to TiMREX, possibly due to production of large drops due to melting at all elevations. Cumulative distribution functions of graupel fraction (Fig. 5.13), calculated as the number of HID-classified graupel divided by all HID points in the cell, show a greater fraction of graupel in over-land NAME convection (for a given elevation group) compared to over-land convection in TiMREX. Over water, TiMREX convection was deeper and had greater amounts of ice compared to the NAME cells over water, likely reflecting differences in organized convection as few isolated cells were observed over the GoC during NAME (Rowe et al. 2011a). This trend of greater graupel fractions over land during NAME suggests a more important role of melting precipitation-sized ice in rainfall production compared to TiMREX. Although mean liquid water mass was also greater for NAME cells, larger water mass fractions (Fig. 5.14) occurred in cells during TiMREX, especially for organized convection over lower elevations, further suggesting efficient warm-rain processes in convection during this experiment.

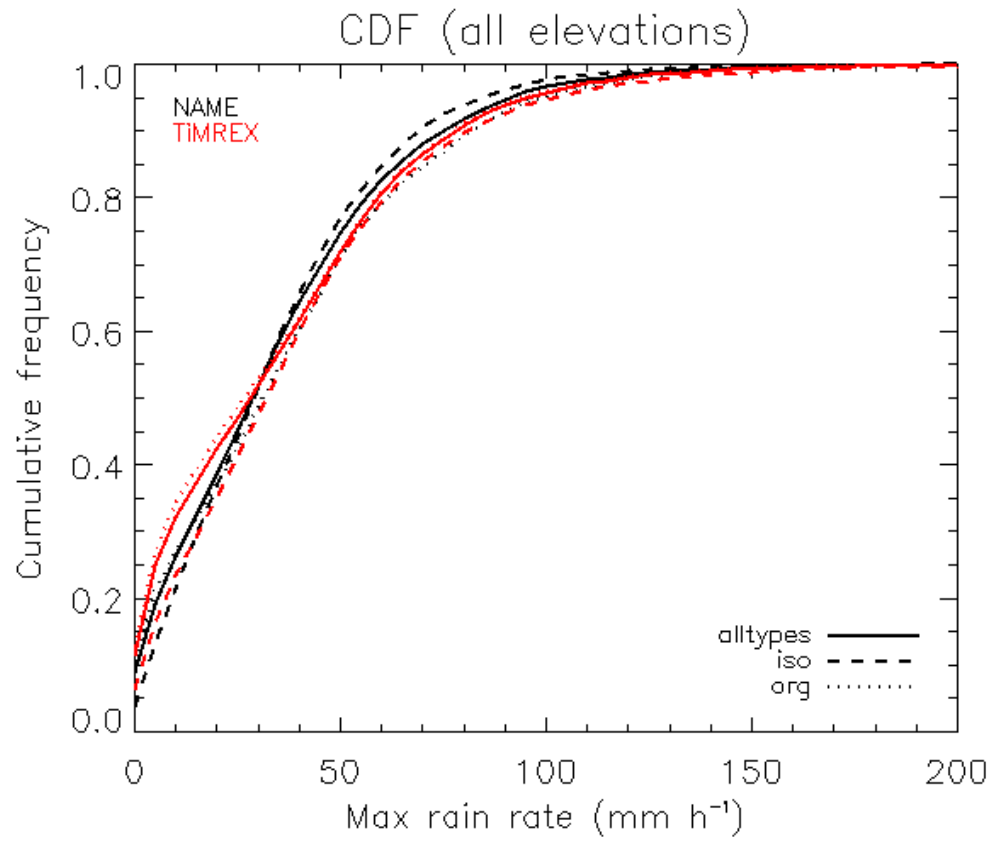


Fig. 5.1. Cumulative distribution functions of maximum rain rate (mm h⁻¹) for all cells identified during NAME and TiMREX, grouped by cell type (i.e., isolated and organized).

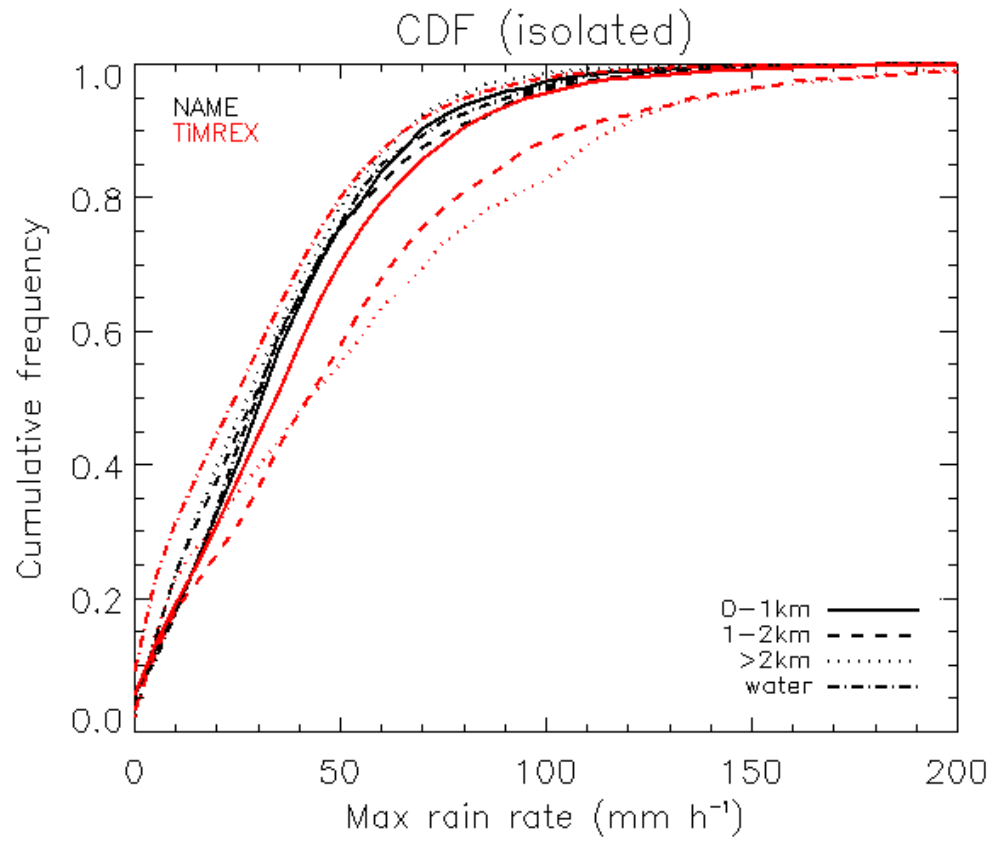


Fig. 5.2. Cumulative distribution functions of maximum rain rate (mm h⁻¹) for all isolated cells identified during NAME and TiMREX, grouped by elevation.

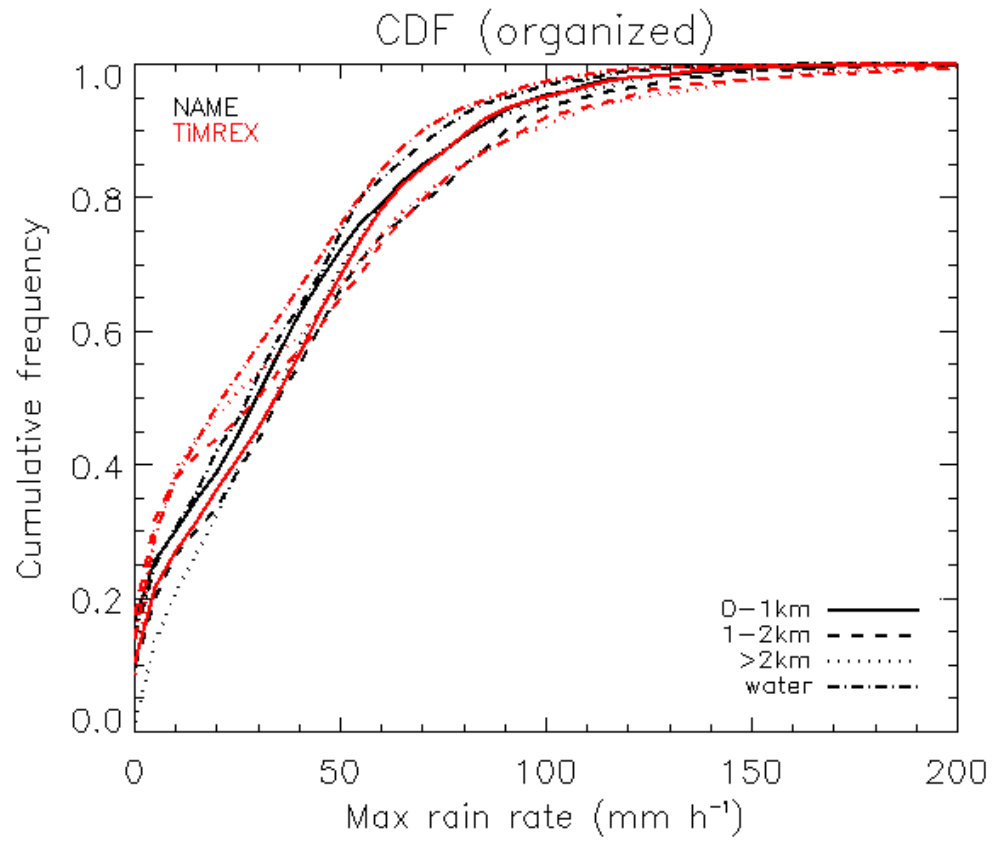


Fig. 5.3. Cumulative distribution functions of maximum rain rate (mm h^{-1}) for all organized cells identified during NAME and TiMREX, grouped by elevation.

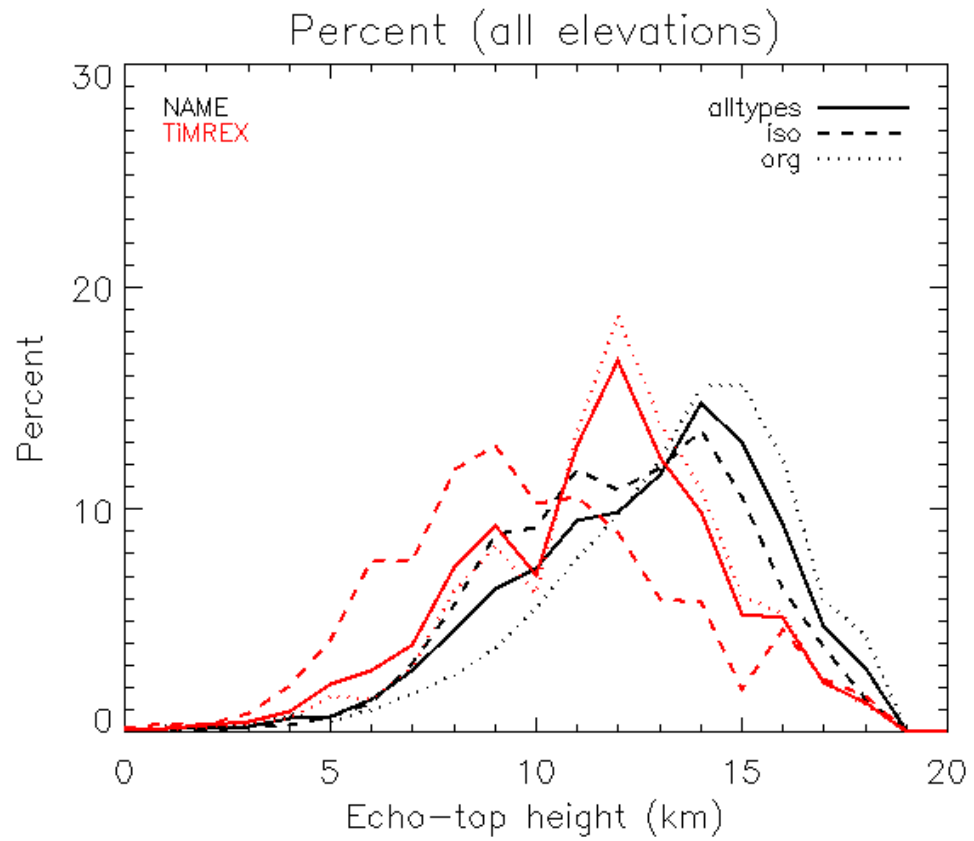


Fig. 5.4. Distributions of echo-top height (km) for all cells identified during NAME and TiMREX, grouped by cell type.

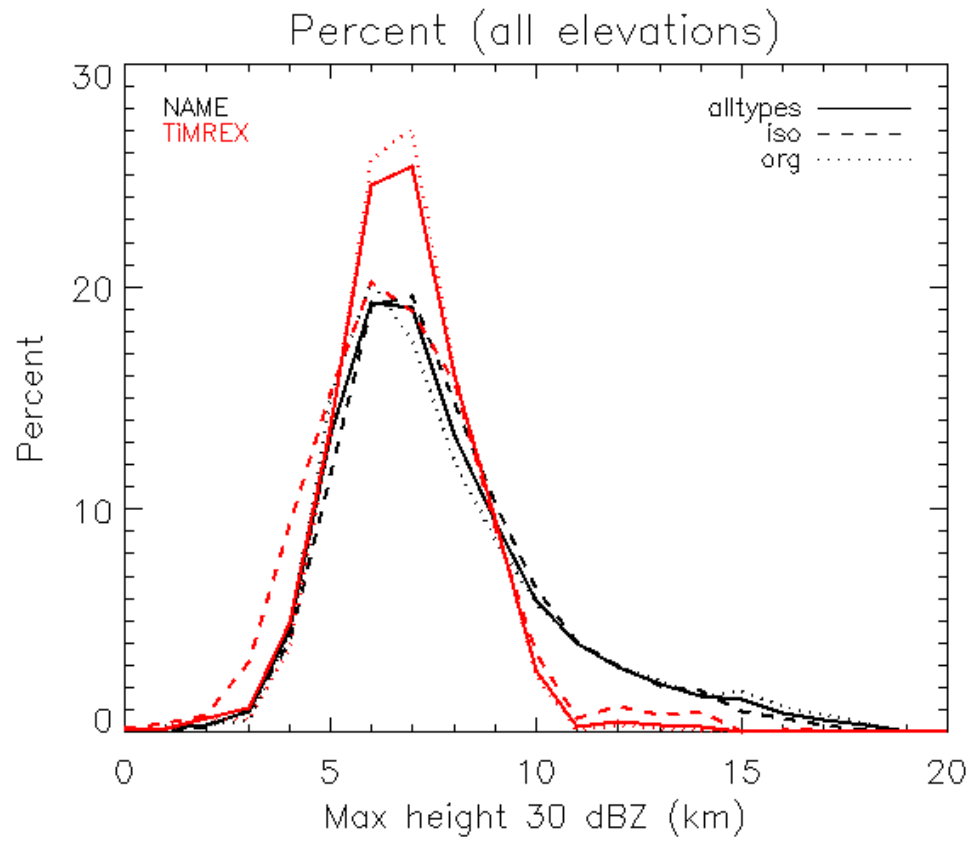


Fig. 5.5. Distributions of maximum height of 30 dBZ for all cells identified during NAME and TiMREX, grouped by cell type.

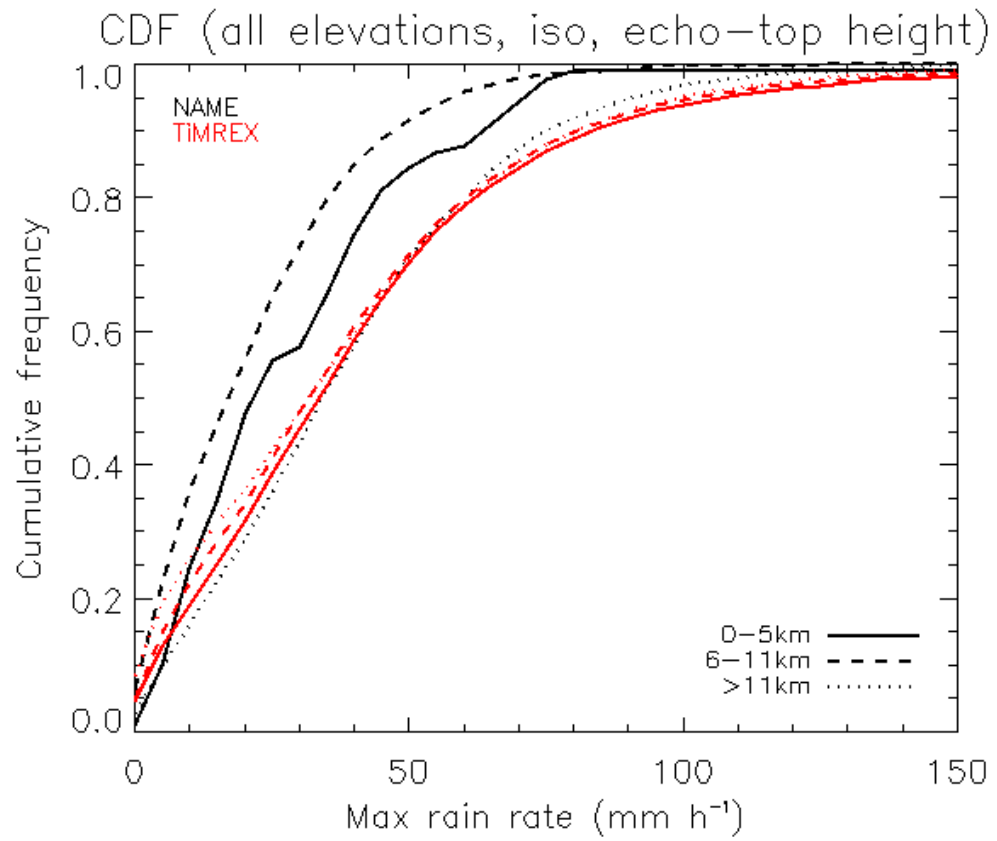


Fig. 5.6. Cumulative distribution functions of maximum rain rate (mm h⁻¹) for isolated cells over all elevations during NAME and TiMREX, grouped by echo-top height.

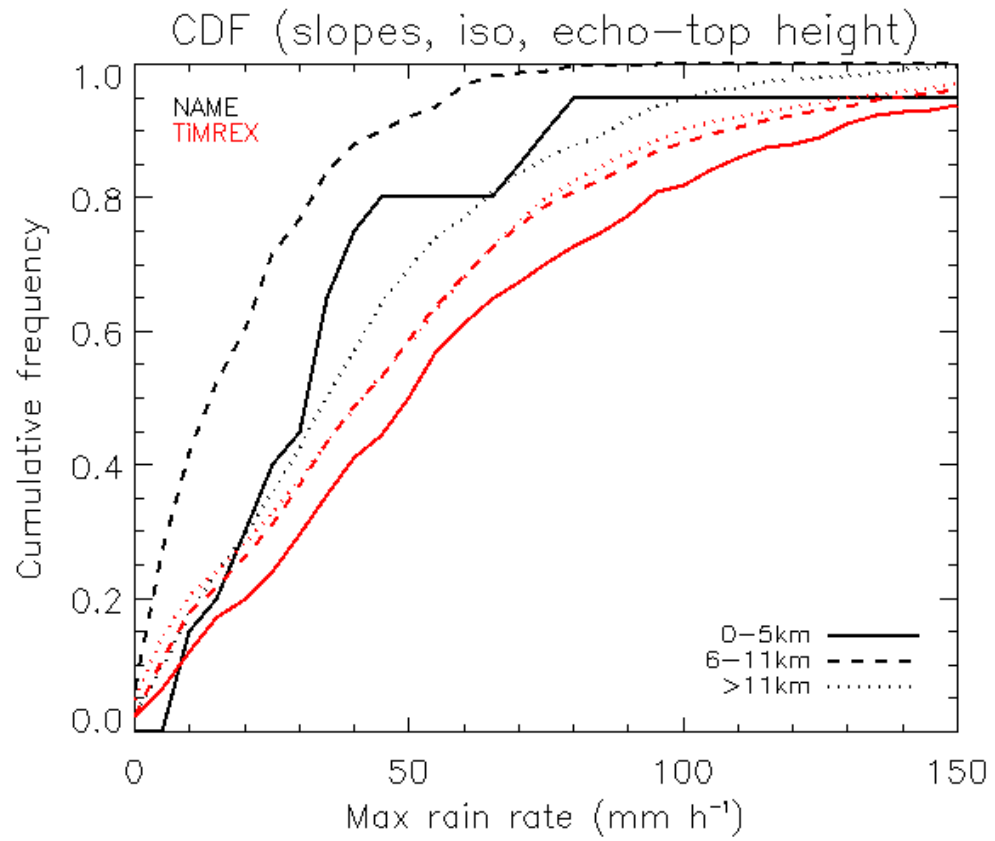


Fig. 5.7. Cumulative distribution functions of maximum rain rate (mm h⁻¹) for isolated cells over the slopes (1–2 km) during NAME and TiMREX, grouped by echo-top height.

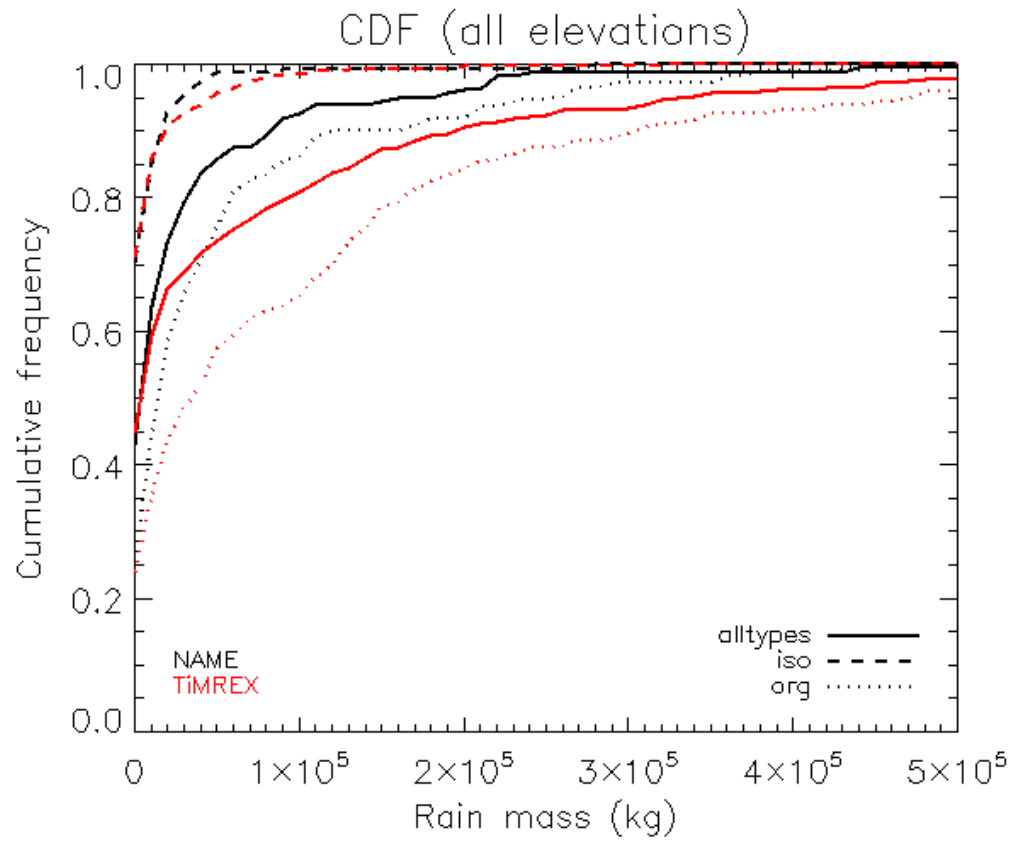


Fig. 5.8. Cumulative distribution functions of hourly accumulated rain mass (kg) for all cells during NAME and TiMREX, grouped by cell type.

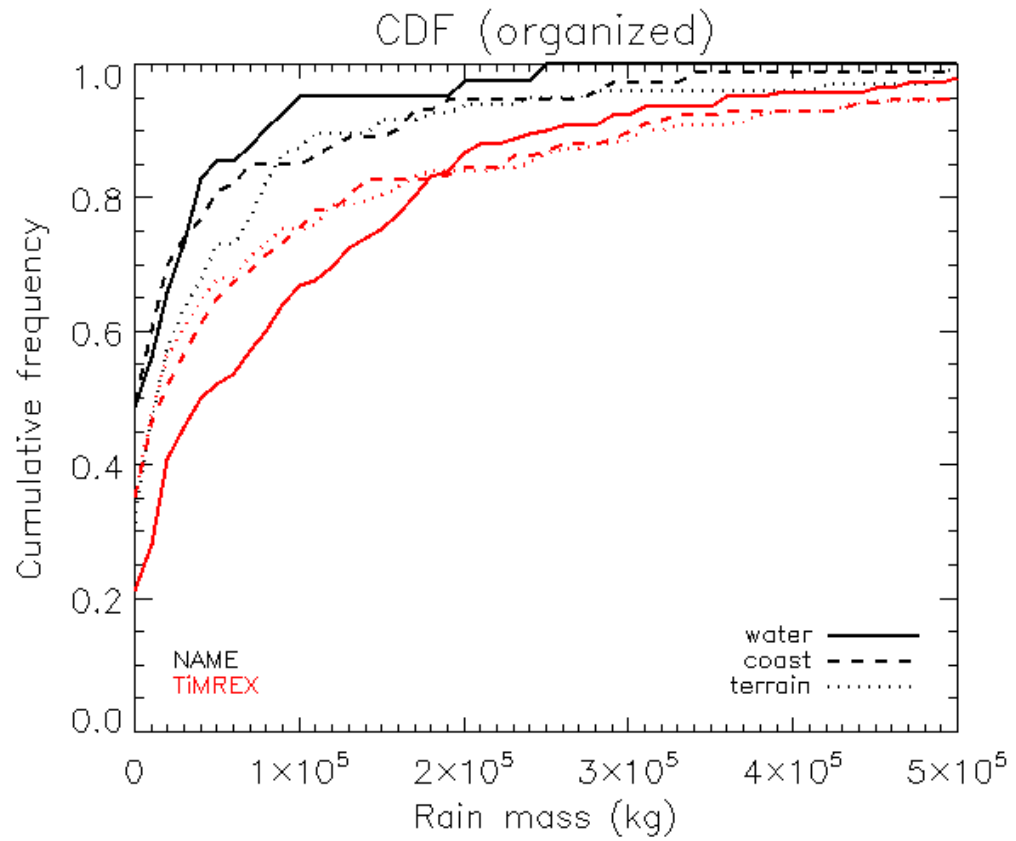


Fig. 5.9. Cumulative distribution functions of hourly accumulated rain mass (kg) for organized cells during NAME and TiMREX, grouped by elevation.

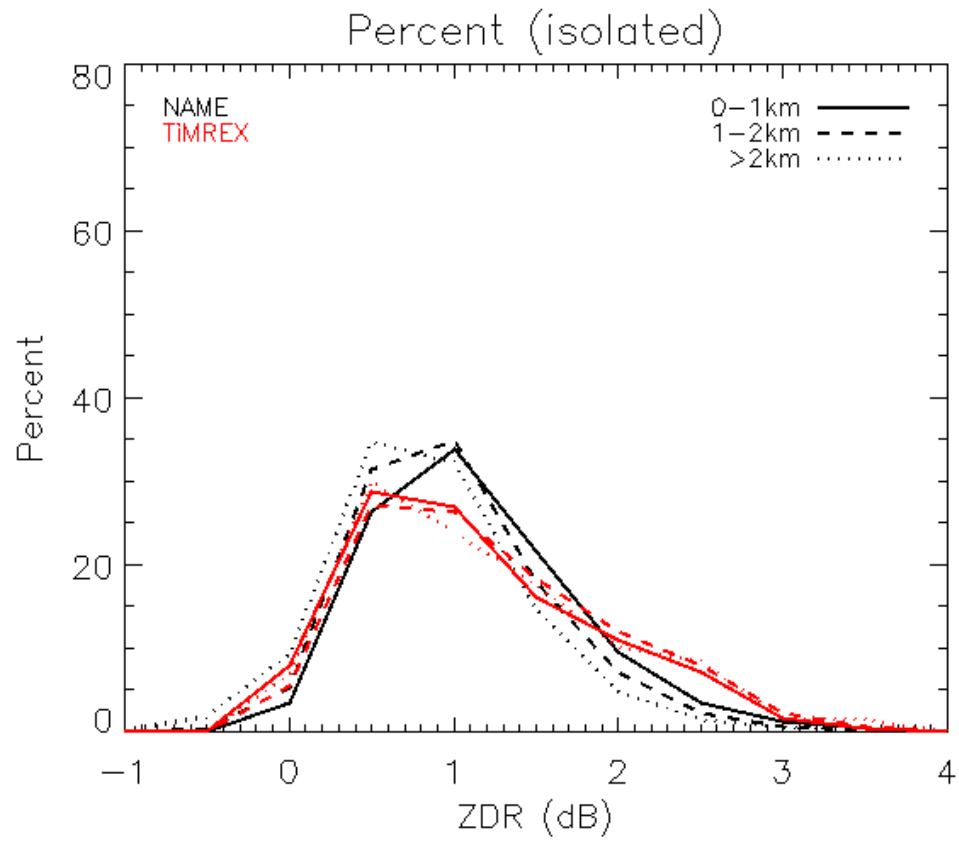


Fig. 5.10. Distributions of maximum Z_{DR} (dB) below 4 km for all isolated cells identified during NAME and TiMREX, grouped by elevation.

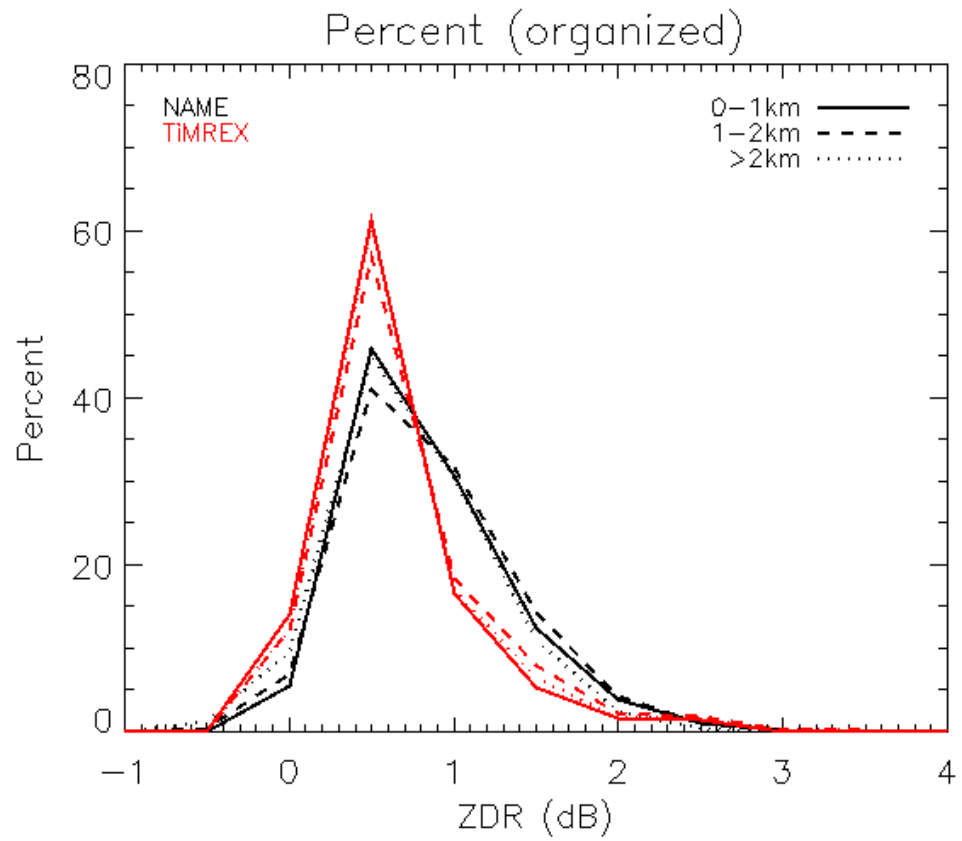


Fig. 5.11. Distributions of maximum Z_{DR} (dB) below 4 km for all organized cells identified during NAME and TiMREX, grouped by elevation.

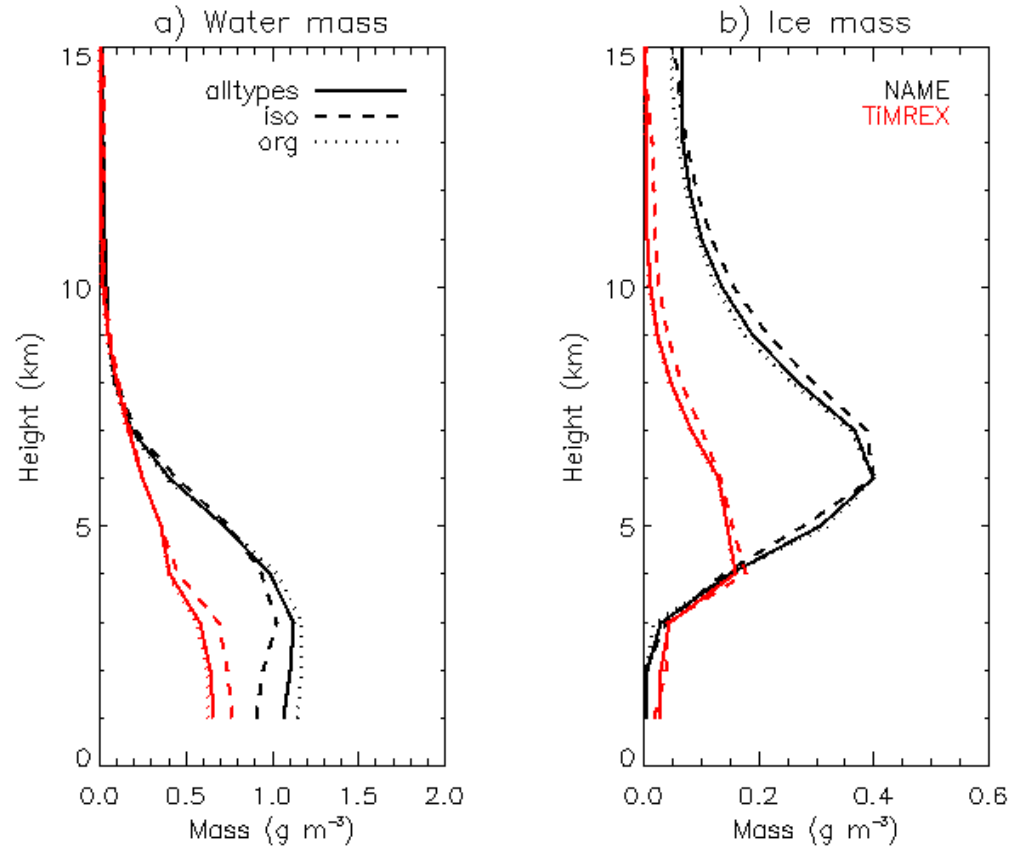


Fig. 5.12. Vertical profiles of (a) mean water mass and (b) mean ice mass for all cells identified during NAME and TiMREX, grouped by cell type.

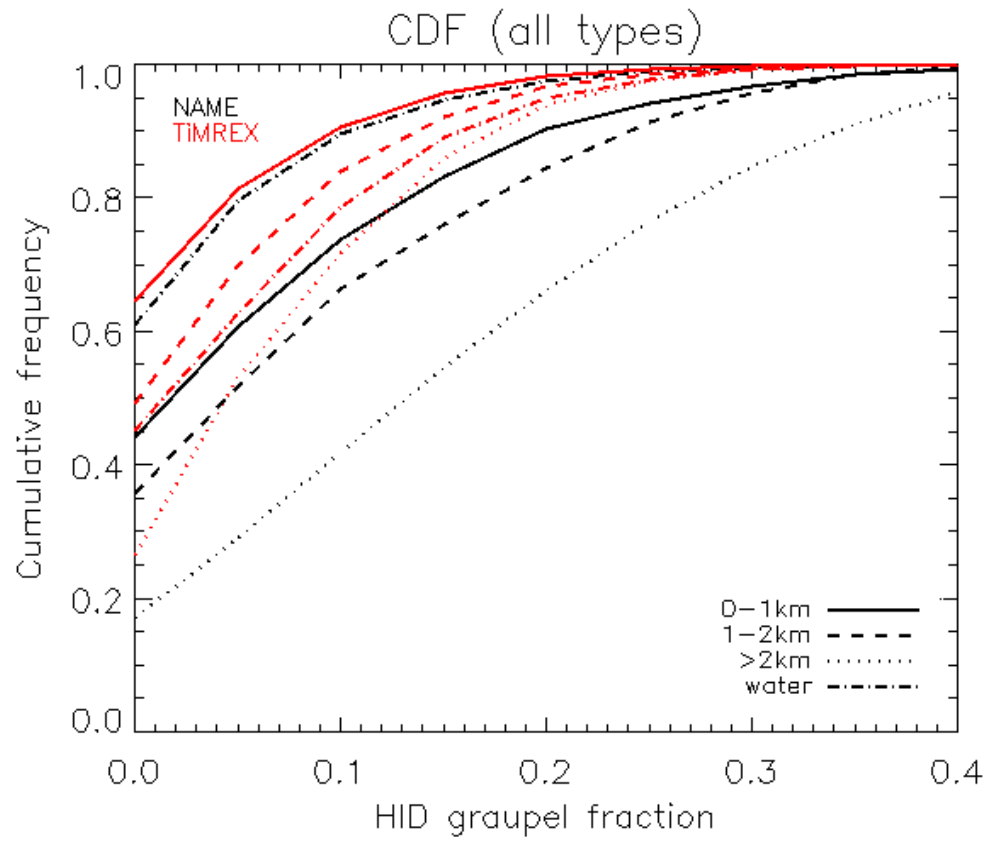


Fig. 5.13. Cumulative distribution functions of HID graupel fraction for all cells, regardless of type, identified during NAME and TiMREX, grouped by elevation.

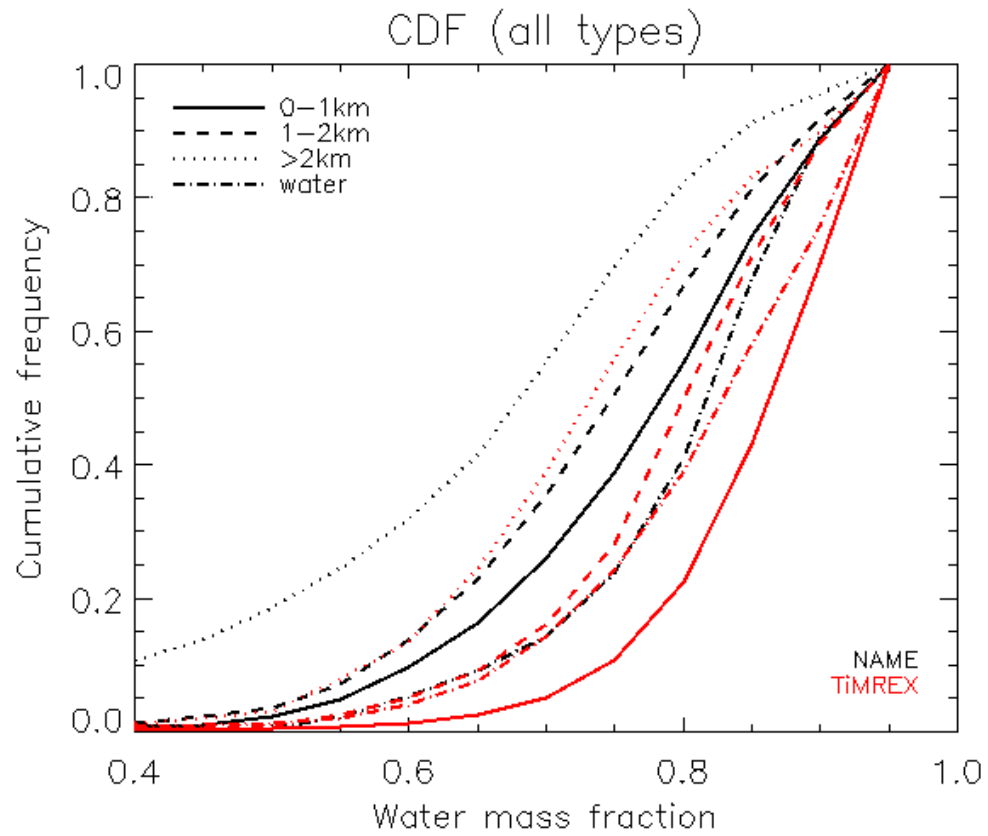


Fig. 5.14. Cumulative distribution functions of water mass fraction for all cells, regardless of type, identified during NAME and TiMREX, grouped by elevation.

CHAPTER 6

DISCUSSION AND CONCLUSIONS

Polarimetric data from NCAR's S-Pol radar, deployed during TiMREX, provided a means to evaluate hydrometeor characteristics of convection in southwestern Taiwan and furthermore, for an investigation of microphysical processes as a function of terrain in this region. Due to moist, onshore flow impinging on the terrain during the Mei-Yu season, it was hypothesized that despite the likely dominance of warm-rain processes, ice-based mechanisms for rainfall production became more important along the western slopes of the Central Mountain Range (CMR) as hydrometeors were lofted above the melting level, where they froze and grew via riming. Using a cell identification and tracking algorithm, similar to that applied to the NAME data set (Rowe et al. 2011a,b), characteristics of isolated cells were compared to convection embedded within organized systems moving onshore to determine trends based on both elevation and degree of organization.

Case studies of diurnally forced isolated convection, selected from 40 mutually exclusive rainfall events during TiMREX, revealed relatively shallow isolated cells with HID-identified graupel confined to within a few kilometers above the melting level, suggesting dominant warm-rain processes. Initially, localized maxima in reflectivity, Z_{DR} , and K_{DP} indicated the presence of large drops and high liquid water contents within the

warm-cloud portion of these cells. In several examples, these maxima in radar variables were elongated upslope due to a prominent tilt of these cells, with precipitation-sized ice displaced over higher terrain. This upslope tilt has also been observed in shallow convection along steep topography in Hawaii, where vertical advection of hydrometeors above the melting level within locally enhanced updrafts led to the growth of graupel via the freezing of raindrops and subsequent accretional processes (Murphy and Businger 2010). This process was also described for isolated cells along the CMR during TiMREX, which likely lead to enhanced rain rates over the western foothills. This tendency for precipitation to be maximized along the slopes has been shown to be a common feature in regions characterized by steep topography, where upward motion and microphysical growth processes on the windward side of these barriers are most robust at lower levels (e.g., Houze 2011).

A specific example of convection near the terrain exhibiting a greater degree of organization also displayed features similar to precipitating systems in other locations. For example, elevated maxima in polarimetric variables within the leading convective line along the lower slopes suggested the contribution of melting ice to intense rainfall at these elevations, as was described in cases of organized convection along the slopes of the SMO during NAME (Rowe et al. 2011b). Deep upslope flow at midlevels resulted in horizontal advection of smaller ice particles toward higher terrain, leading to the production of widespread stratiform precipitation regions characterized by a radar brightband. These radar-echo features were similar to observations of MCSs in other regions of tropical precipitation (e.g., Williams et al. 1995; Ecklund et al. 1999; May and Keenan 2005), as well as in the midlatitudes (e.g., Houze et al. 1989; Houze 2004). In

addition, upstream regeneration of convection in these TiMREX examples due to convergence between the low-level outflow and the environmental upslope flow led to extended periods of intense rainfall over the steep slopes. This specific convergence process has also been shown to influence the longevity of organized systems in other mountainous regions, including the Himalayas (Medina et al. 2010), the European Alps (Houze et al. 2001), and the Sierra Madre Occidental in northwestern Mexico (Rowe et al. 2011b).

In order to determine if similar processes occurred in organized systems moving onshore within the environmental southwesterly flow, one particular case, characterized by successive convective lines with trailing stratiform rain, was analyzed to compare differences in microphysical processes from when these systems were located over water compared to later when they interacted with the terrain. Both individual tracks and statistics including all cells identified during this case revealed smaller mean drop diameters over water, characteristic of maritime DSDs, followed by an upslope tilt leading to enhanced production of graupel along the western slopes of the CMR. Despite melting-ice signatures and an observed enhancement of rain rates along the terrain, an analysis of mean drop diameters showed almost no elevation-dependent trend for these organized systems. A possible explanation suggested for this observation was the breakup of melting hydrometeors produced equilibrium DSDs (e.g., Hu and Srivastava 1995) that may have been comprised of smaller drops in the mean, comparable to distributions observed over the adjacent water.

The suggested differences in DSDs for this particular case were also described in an analysis of all organized cells during TiMREX. In addition to smaller mean drop

sizes, organized cells over all elevations were deeper than isolated convection, yet shallow isolated cells, with broader Z_{DR} distributions and, therefore, DSD spectra, produced slightly greater rain rates. This suggested efficient warm-rain processes during TiMREX, although hourly accumulated rain masses were significantly greater for deep, organized cells. Despite differences in total rainfall, an enhancement of rain rates along the topography was observed for all cell types, likely due to melting hydrometeors.

A similar analysis of isolated and organized convection during the North American Monsoon Experiment (NAME) in northwestern Mexico (Rowe et al. 2011a,b) also revealed enhanced rain rates along the western slopes of the local topography, where ice-based processes played a greater role. As opposed to the organized systems moving onshore during TiMREX, characterized by smaller mean drop diameters, upscale growth of convection during NAME originated over the topography, later propagating offshore and allowing for a greater contribution from melting ice to intense rain rates compared to in southwestern Taiwan. This was reflected in comparisons between vertical profiles of mean ice mass for both regions, where convection during NAME contained greater amounts of ice mass throughout the entire vertical depth of convection regardless of organization. Higher percentages of large Z_{DR} further emphasized the greater role of warm-rain processes in convection during TiMREX, highlighting differences in microphysical processes between these two regions characterized by warm-season, orographic rainfall.

Despite these differences, enhanced rainfall, as well as regeneration of convection, over the sloping terrain in both the NAME and TiMREX can lead to flash flooding and subsequent landslides, and thus emphasizes the need for accurate prediction

and understanding of processes that lead to intense rain in these vulnerable regions. The polarimetric information available from S-Pol provided a unique opportunity to study hydrometeor characteristics within convection in these areas. Placing the analysis of these features within a dynamical framework (e.g., dual-Doppler analysis, modeling studies) would provide additional information regarding mass fluxes within intense convection. Also, modeling studies relating the longevity of these systems to the environmental conditions and topography would be beneficial for forecasting these extreme rainfall events, given the significant contribution to accumulated rainfall from long-lived organized systems along the sloping terrain.

REFERENCES

- Atlas, D., and C. W. Ulbrich, 2000: An observationally based conceptual model of warm oceanic convective rain in the tropics. *J. Appl. Meteor.*, **39**, 2165–2181.
- Atlas, D., and C. W. Ulbrich, 2006: Drop size spectra and integral remote sensing parameters in the transition from convective to stratiform rain. *Geophys. Res. Lett.*, **33**, L16803, doi:10.1029/2006GL026824.
- Battan, L. J., 1973: *Radar Observation of the Atmosphere*. University of Chicago Press, 324 pp.
- Bringi, V. N., and V. Chandrasekar, 2001: *Polarimetric Doppler Weather Radar Principles and Applications*. Cambridge University Press, 636 pp.
- Bringi, V. N., V. Chandrasekar, J. Hubbert, E. Gorgucci, W. L. Randeu, and M. Schoenhuber, 2003: Raindrop size distribution in different climatic regimes with disdrometer and dual-polarized radar analysis. *J. Atmos. Sci.*, **60**, 354–365.
- Carey, L. D., and S. A. Rutledge, 2000: The relationship between precipitation and lightning in tropical island convection: A C-band polarimetric radar study. *Mon. Wea. Rev.*, **128**, 2687–2710.
- Chen, G. T.-J., and C.-P. Chang, 1980: The structure and vorticity budget of an early summer monsoon trough (Mei-Yu) over southeastern China and Japan. *Mon. Wea. Rev.*, **108**, 942–953.
- Chen, G. T.-J., and H. Chou, 1993: General characteristics of squall lines observed in TAMEX. *Mon. Wea. Rev.*, **121**, 726–733.
- Chen, T.-C., M.-C. Yen, J.-C. Hsieh, and R. W. Arritt, 1999: Diurnal and seasonal variations in the rainfall measured by the automatic rainfall and meteorological telemetry system in Taiwan. *Bull. Amer. Meteor. Soc.*, **80**, 2299–2312.
- Chen, Y.-L., and J. Li, 1995: Large-scale conditions favorable for the development of heavy rainfall during TAMEX IOP 3. *Mon. Wea. Rev.*, **123**, 2978–3002.

- Cifelli, R., W. A. Petersen, L. D. Carey, S. A. Rutledge, and M. A. F. Silva-Dias, 2002: Radar observations of the kinematic, microphysical, and precipitation characteristics of two MCSs in TRMM LBA. *J. Geophys. Res.*, **107**, 8077, doi:10.1029/2000JD000264.
- DeMott, C. A., and S. A. Rutledge, 1998: The vertical structure of TOGA COARE convection. Part I: Radar echo distributions. *J. Atmos. Sci.*, **55**, 2730–2747.
- Ecklund, W. L., C. R. Williams, P. E. Johnston, and K. S. Gage, 1999: A 3-GHz profiler for precipitating cloud studies. *J. Atmos. Oceanic Technol.*, **16**, 309–322.
- Georgis, J.-F., F. Roux, M. Chong, and S. Pradier, 2003: Triple-Doppler radar analysis of the heavy rain event observed in the Lago Maggiore region during MAP IOP2b. *Q. J. R. Meteorol. Soc.*, **129**, 495–522.
- Gochis, D. J., A. Jimenez, C. J. Watts, J. Garatuza-Payan, and W. J. Shuttleworth, 2004: Analysis of 2002 and 2003 warm-season precipitation from the North American Monsoon Experiment Event Rain Gauge Network. *Mon. Wea. Rev.*, **132**, 2938–2953.
- Gochis, D. J., C. J. Watts, J. Garatuza-Payan, and J. Cesar-Rodriguez, 2007: Spatial and temporal patterns of precipitation intensity as observed by the NAME Event Rain Gauge Network from 2002 to 2004. *J. Climate*, **20**, 1734–1750.
- Higgins, R. W., and Coauthors, 2006: The NAME 2004 field campaign and modeling strategy. *Bull. Amer. Meteor. Soc.*, **87**, 79–94.
- Hirose, M., and K. Nakamura, 2002: Spatial and seasonal variation of rain profiles over Asia observed by spaceborne precipitation radar. *J. Climate*, **15**, 3443–3458.
- Houze, R. A., Jr., 1993: *Cloud Dynamics*. Academic Press, 573 pp.
- Houze, R. A., Jr., 1997: Stratiform precipitation in regions of convection: A meteorological paradox? *Bull. Amer. Meteor. Soc.*, **78**, 2179–2196.
- Houze, R. A., Jr., 2004: Mesoscale convective systems. *Rev. Geophys.*, **42**, 0.1029/2004RG000150, 43 pp.
- Houze, R. A., Jr., 2011: Orographic effects on precipitating clouds. Submitted to *Rev. Geophys.*
- Houze, R. A., Jr., M. I. Biggerstaff, S. A. Rutledge, and B. F. Smull, 1989: Interpretation of Doppler weather radar displays of midlatitude mesoscale convective systems. *Bull. Amer. Meteor. Soc.*, **70**, 608–619.

- Houze, R. A., Jr., C. N. James, and S. Medina, 2001: Radar observations of precipitation and airflow on the Mediterranean side of the Alps: Autumn 1998 and 1999. *Q. J. R. Meteorol. Soc.*, **127**, 2537–2558.
- Houze, R. A., Jr., D. C. Wilton, B. F. Smull, 2007: Monsoon convection in the Himalayan region as seen by the TRMM precipitation radar. *Quart. J. Roy. Meteor. Soc.*, **133**, 1389–1411.
- Hu, Z., and R. C. Srivastava, 1995: Evolution of raindrop size distribution by coalescence, breakup, and evaporation: Theory and observations. *J. Atmos. Sci.*, **52**, 1761–1783.
- Hubbert, J., and V. N. Bringi, 1995: An iterative filtering technique for the analysis of copolar differential phase and dual-frequency radar measurements. *J. Atmos. Oceanic Technol.*, **12**, 643–648.
- Johnson, R. H., and J. F. Bresch, 1991: Diagnosed characteristics of precipitation systems over Taiwan during the May-June 1987 TAMEX. *Mon. Wea. Rev.*, **119**, 2540–2557.
- Jou, B. J., and S. M. Deng, 1992: Structure of a low-level jet and its role in triggering and organizing moist convection over Taiwan: A TAMEX case study. *Terr. Atmos. Oceanic Sci.*, **3**, 35–58.
- Kirankumar, N. V. P., T. Narayana Rao, B. Radhakrishna, and D. Narayana Rao, 2008: Statistical characteristics of raindrop size distribution in southwest monsoon season. *J. Appl. Meteor. Climatol.*, **47**, 576–590.
- Kuo, Y. H., and G. T.-J. Chen, 1990: The Taiwan Area Mesoscale Experiment (TAMEX): An overview. *Bull. Amer. Meteor. Soc.*, **71**, 488–503.
- Lang, T. J., D. A. Ahijevych, S. W. Nesbitt, R. E. Carbone, S. A. Rutledge, and R. Cifelli, 2007: Radar-observed characteristics of precipitating systems during NAME 2004. *J. Climate*, **20**, 1713–1733.
- Liu, H., and V. Chandrasekar, 2000: Classification of hydrometeors based on polarimetric radar measurements: Development of fuzzy logic and neuro-fuzzy systems, and in situ verification. *J. Atmos. Oceanic Technol.*, **17**, 140–164.
- May, P. T., and T. D. Keenan, 2005: Evaluation of microphysical retrievals from polarimetric radar with wind profiler data. *J. Appl. Meteor.*, **44**, 827–838.
- Medina, S., and R. A. Houze, Jr., 2003: Air motions and precipitation growth in alpine storms. *Q. J. R. Meteorol. Soc.*, **129**, 345–371.

- Medina, S., R. A. Houze, Jr., A. Kumar, and D. Niyogi, 2010: Summer monsoon convection in the Himalayan region: Terrain and land cover effects. *Q. J. R. Meteorol. Soc.*, **136**, 593–616.
- Mohr, C. G., L. J. Miller, R. L. Vaughan, and H. W. Frank, 1986: The merger of mesoscale datasets into a common Cartesian format for efficient and systematic analyses. *J. Atmos. Oceanic Technol.*, **3**, 143–161.
- Murphy, M. J., Jr., and S. Businger, 2010: Orographic influences on an Oahu flood. *Mon. Wea. Rev.*, in press.
- Pereira, L. G., 2008: Characteristics and organization of precipitating features during NAME 2004 and their relationship to environmental conditions. Ph.D. Dissertation, Atmospheric Science, Colorado State University, Fort Collins, Colorado, 229 pp.
- Petersen, W. A., S. A. Rutledge, and R. E. Orville, 1996: Cloud-to-ground lightning observations from TOGA COARE: Selected results and lightning location algorithms. *Mon. Wea. Rev.*, **124**, 602–620.
- Rickenbach, T. M., and S. A. Rutledge, 1998: Convection in TOGA COARE: Horizontal scale, morphology, and rainfall production. *J. Atmos. Sci.*, **55**, 2715–2729.
- Romatschke, U., S. Medina, and R. A. Houze, Jr., 2010: Regional, seasonal, and diurnal variations of extreme convection in the South Asian region. *J. Climate*, **23**, 419–439.
- Rosenfeld, D., and C. W. Ulbrich, 2003: Cloud microphysical properties, processes, and rainfall estimation opportunities. *Meteor. Monogr.*, **30**, 237.
- Rowe, A. K., S. A. Rutledge, T. J. Lang, P. E. Ciesielski, and S. M. Saleeby, 2008: Elevation-dependent trends in precipitation observed during NAME. *Mon. Wea. Rev.*, **136**, 4962–4979.
- Rowe, A. K., S. A. Rutledge, and T. J. Lang, 2011a: Investigation of microphysical processes occurring in isolated convection during NAME. *Mon. Wea. Rev.*, **139**, 424–443.
- Rowe, A. K., S. A. Rutledge, and T. J. Lang, 2011b: Investigation of microphysical processes occurring in organized convection during NAME. Submitted to *Mon. Wea. Rev.*
- Tao, S., Ed., 1980: *Severe rainstorms in China*. Science Press, 225 pp.
- Tessendorf, S. A., L. J. Miller, K. C. Wiens, and S. A. Rutledge, 2005: The 29 June 2000 supercell observed during STEPS. Part I: Kinematics and microphysics. *J. Atmos. Sci.*, **62**, 4127–4150.

Tong, C.-C., 2009: Precipitation characteristics and associated environment regimes during SoWMEX/TiMREX. Master's Thesis, Atmospheric Science, National Taiwan University, Taipei, Taiwan, 90 pp.

Ulbrich, C. W., and D. Atlas, 2007: Microphysics of raindrop size spectra: Tropical continental and maritime storms. *J. Appl. Meteor. Climatol.*, **46**, 1777–1791.

Wang, T. C.-C., Y. J. Lin, H. Shen, and R. W. Pasken, 1990: Characteristics of a subtropical squall line determined from TAMEX dual-Doppler data. Part I: Kinematic structure. *J. Atmos. Sci.*, **47**, 2357–2381.

Williams, C. R., W. L. Ecklund, and K. S. Gage, 1995: Classification of precipitating clouds in the Tropics using 915-MHz wind profilers. *J. Atmos. Oceanic Technol.*, **12**, 996–1012.

Xie, S. P., H. Xu, N. H. Saji, Y. Wang, and W. T. Liu, 2006: Role of narrow mountains in large-scale organization of Asian monsoon convection. *J. Climate*, **19**, 3420–3429.

Yeh, H. C., and Y. L. Chen, 1998: Characteristics of rainfall distributions over Taiwan during the Taiwan Area Mesoscale Experiment (TAMEX). *J. Appl. Meteor.*, **37**, 1457–1469.

Yuter, S. E., and R. A. Houze, Jr., 1995: Three-dimensional kinematic and microphysical evolution of Florida cumulonimbus. Part III: Vertical mass transport, mass divergence, and synthesis. *Mon. Wea. Rev.*, **123**, 1964–1983.

Yuter, S. E., and R. A. Houze, Jr., 2003: Microphysical modes of precipitation growth determined by S-band vertically pointing radar in orographic precipitation during MAP. *Q. J. R. Meteorol. Soc.*, **129**, 455–476.

Zipser, E. J., 1994: Deep cumulonimbus cloud systems in the Tropics with and without lightning. *Mon. Wea. Rev.*, **122**, 1837–1851.

Zipser, E. J., D. J. Cecil, C. Liu, S. W. Nesbitt, and D. P. Yorty, 2006: Where are the most intense thunderstorms on Earth? *Bull. Amer. Meteor. Soc.*, **87**, 1057–1071.

PART IV: OVERALL SUMMARY AND CONCLUSIONS

The North American and Asian monsoons are characterized by a reversal of the mean low-level winds, ushering moist flow onshore from the adjacent warm ocean water. In northwestern Mexico and southwestern Taiwan, both influenced by this monsoon environment, potentially unstable air is lifted along the Sierra Madre Occidental (SMO) and Central Mountain Range (CMR), respectively, resulting in intense convection over the terrain and adjacent coastal plain. Convective cells are also triggered within preexisting systems due to interaction with topography, as has been shown in radar observations in other monsoon regions (Medina and Houze 2003; Rotunno and Houze 2007). In particular, mesoscale convective systems (MCSs) embedded within the Mei-Yu front in Taiwan move onshore within southwesterly environmental flow, producing localized heavy rainfall along the western slopes of the CMR (e.g., Chen et al. 1999), suggesting an enhancement of convection within these features. Local mountain circulations also influence precipitation in this region, with more than 50% of rainfall associated with diurnally driven orographic convection (Yeh and Chen 1998). Diurnally forced land-sea and mountain-valley flows also affect the timing and distribution of precipitation in the core North American Monsoon (NAM) region of northwestern Mexico (Douglas et al. 1993; Dai et al. 1999). Furthermore, upscale growth of convection along the SMO, followed by propagation toward the coast, leads to diurnal trends in rainfall, with maximum rain totals located along the coastal plain (e.g., Lang et al. 2007, Rowe et al. 2008). The predictability of warm-season rainfall in these regions therefore requires an understanding of the nature of the precipitating systems, including the diurnal variability and microphysical processes as a function of the local topography. The objective of this dissertation was to investigate these potential elevation-dependent trends

and compare convective features in these two regions to each other, as well as to other regions influenced by warm-season, orographic rainfall.

Using polarimetric data from the S-Pol radar, deployed during the NAME and TiMREX field campaigns, which focused on northwestern Mexico and southwestern Taiwan, respectively, the evolution of hydrometeor characteristics and microphysical processes leading to intense warm-season rainfall in these mountainous regions were analyzed. A cell identification and tracking algorithm was applied to the radar data set to locate and track individual convective elements, contained within both isolated and organized features, as a function of terrain. Examples of isolated cells from a range of topography during NAME revealed characteristics comparable to other studies of tropical (e.g., Carey and Rutledge 2000; Cifelli et al. 2002) and midlatitude convection (e.g., Bringi et al. 1996; Zeng et al. 2001; Tessendorf et al. 2005), including the presence of Z_{DR} columns. This attribute indicated the lofting of drops above the melting level, where subsequent freezing and growth by riming led to the production of graupel along the western slopes of the SMO and adjacent coastal plain, suggesting a combined role of warm-rain and ice-based processes (Rowe et al. 2011a). Melting of large ice hydrometeors was also noted in convection over the high peaks of the SMO, leading to short-lived, intense rainfall despite the truncated warm-cloud depth, similar to convection observed over the high terrain of the Tibetan Plateau (Houze et al. 2007).

Case studies of organized convection during NAME (Rowe et al. 2011b) also revealed an additional dependence on ice-based processes via drop freezing and subsequent riming growth, leading to maximum rain intensities along the western slopes. The melting of ice not only enhanced precipitation, but also led to the production of

mesoscale outflow boundaries, which provided an additional focus mechanism for convective initiation over the lower elevations (especially when colliding with upslope flow in the opposing direction) resulting in propagation of the systems toward the coast. This occurrence of organized systems upstream of and over higher terrain has also been observed in the Himalayas, where convergence between outflow boundaries and the larger-scale monsoonal flow allowed for movement of these systems off the terrain (Medina et al. 2010), as well as in other mountainous regions including the western Ghats (Grossman and Durran 1984), the European Alps (Houze et al. 2001), and the Pyrenees (Romero et al. 2001).

In the Himalayan region, during active phases of the monsoon, MCSs also form within synoptically favorable regions over the Bay of Bengal, and are advected over land by the prevailing southwesterlies, resulting in organized systems characterized by maritime DSDs as they encounter the topography (e.g., Houze and Churchill 1987). Medina et al. (2010), using numerical simulations of this scenario, noted a strengthening of the MCSs over the mountain ranges. Similar trends were observed in organized systems during TiMREX, where an enhancement of rain rates occurred as MCSs moved onshore within the southwesterly flow and intercepted the steep terrain of the CMR. Elevated maxima in polarimetric variables, similar to observations in cells during NAME, indicated a contribution from melting ice to rainfall at these elevations.

In addition to elevation-dependent trends, differences as a function of organization also existed. Cells embedded within organized systems during TiMREX were deeper than isolated cells; a trend also observed in convection during NAME (Rowe et al. 2011b), as well as in TOGA COARE (Rickenbach and Rutledge 1998). However,

shallow isolated cells, characterized by broader DSDs compared to organized cells (as evidenced by wide variance in the Z_{DR} values) produced slightly greater rain rates, suggesting efficient warm-rain processes in convection during TiMREX. Convection during NAME, on the other hand, exhibited a greater dependence on ice-based processes, especially over the higher elevations. This was emphasized through comparisons between vertical profiles of mean ice mass for both regions, which revealed greater amounts throughout the entire vertical depth of convection during NAME for both cell types. Broader distributions in maximum Z_{DR} during TiMREX further emphasized the differences in microphysical processes between these two regions.

Regardless of these differences, the greatest contribution of hourly accumulated rainfall from both experiments was associated with deep organized systems, characterized by extended periods of intense rainfall over the western slopes due to enhanced rain rates and upstream regeneration of convection. This scenario poses threats along the steep topography due to flash flooding and subsequent landslides, emphasizing the need for accurate prediction and understanding of the processes that lead to intense rain in these vulnerable regions. Analyses of hydrometeor characteristics was possible due to the deployment of a polarimetric radar during both experiments, providing valuable information for increasing understanding of the nature of precipitating systems in these regions, which help to validate and improve modeling of the microphysical processes and their affect on the broader monsoon circulation. In addition, characteristics of the environment in which these features form were studied during both field experiments using a variety of instrumentation. Further studies need to relate these conditions to convection in order to advance the ability to forecast excessive rainfall in

these regions. Tong (2009) described environmental conditions associated with land versus oceanic initiation of convection during TiMREX, and studies by Lang et al. (2007) and Pereira (2008) related environmental CAPE and shear to characteristics of precipitating features during NAME. Additional observational and modeling studies relating the environmental conditions to the longevity of these systems in both regions (e.g., through analysis of interactions between cold pools and the environmental flow), as well as calculations of vertical motions and corresponding mass fluxes within convection, will further improve the understanding and predictability of these extreme events and contribute to the overall knowledge of warm-season, orographic rainfall.

REFERENCES

- Bringi, V. N., L. Liu, P. C. Kennedy, V. Chandrasekar, and S. A. Rutledge, 1996: Dual multiparameter radar observations of intense convective storms: The 24 June 1992 case study. *Meteor. Atmos. Phys.*, **59**, 3–31.
- Carey, L. D., and S. A. Rutledge, 2000: The relationship between precipitation and lightning in tropical island convection: A C-band polarimetric radar study. *Mon. Wea. Rev.*, **128**, 2687–2710.
- Chen, T.-C., M.-C. Yen, J.-C. Hsieh, and R. W. Arritt, 1999: Diurnal and seasonal variations in the rainfall measured by the automatic rainfall and meteorological telemetry system in Taiwan. *Bull. Amer. Meteor. Soc.*, **80**, 2299–2312.
- Cifelli, R., W. A. Petersen, L. D. Carey, S. A. Rutledge, and M. A. F. Silva-Dias, 2002: Radar observations of the kinematic, microphysical, and precipitation characteristics of two MCSs in TRMM LBA. *J. Geophys. Res.*, **107**, 8077, doi:10.1029/2000JD000264.
- Dai, A. G., F. Giorgi, and K. E. Trenberth, 1999: Observed and model-simulated diurnal cycles of precipitation over the contiguous United States. *J. Geophys. Res.*, **104**, 6377–6402.
- Douglas, M., R. A. Maddox, K. Howard, and S. Reyes, 1993: The Mexican monsoon. *J. Climate*, **6**, 1665–1677.
- Grossman, R. L., and D. R. Durran, 1984: Interaction of low-level flow with the Western Ghat mountains and offshore convection in the summer monsoon. *Mon. Wea. Rev.*, **112**, 652–672.
- Houze, R. A., Jr., and D. D. Churchill, 1987: Mesoscale organization and cloud microphysics in a Bay of Bengal depression. *J. Atmos. Sci.*, **44**, 1845–1867.
- Houze, R. A., Jr., C. N. James, and S. Medina, 2001: Radar observations of precipitation and airflow on the Mediterranean side of the Alps: Autumn 1998 and 1999. *Q. J. R. Meteorol. Soc.*, **127**, 2537–2558.
- Houze, R. A., Jr., D. C. Wilton, B. F. Smull, 2007: Monsoon convection in the Himalayan region as seen by the TRMM precipitation radar. *Quart. J. Roy. Meteor. Soc.*, **133**, 1389–1411.

- Lang, T. J., D. A. Ahijevych, S. W. Nesbitt, R. E. Carbone, S. A. Rutledge, and R. Cifelli, 2007: Radar-observed characteristics of precipitating systems during NAME 2004. *J. Climate*, **20**, 1713–1733.
- Medina, S., and R. A. Houze, Jr., 2003: Air motions and precipitation growth in alpine storms. *Q. J. R. Meteorol. Soc.*, **129**, 345–371.
- Medina, S., R. A. Houze, Jr., A. Kumar, and D. Niyogi, 2010: Summer monsoon convection in the Himalayan region: Terrain and land cover effects. *Q. J. R. Meteorol. Soc.*, **136**, 593–616.
- Pereira, L. G., 2008: Characteristics and organization of precipitating features during NAME 2004 and their relationship to environmental conditions. Ph.D. Dissertation, Atmospheric Science, Colorado State University, Fort Collins, Colorado, 229 pp.
- Romero, R., C. A. Doswell III, and R. Riosalido, 2001: Observations and fine-grid simulations of a convective outbreak in northeastern Spain: Importance of diurnal forcing and convective cold pools. *Mon. Wea. Rev.*, **129**, 2157–2182.
- Rotunno, R., and R. A. Houze, Jr., 2007: Lessons on orographic precipitation from the Mesoscale Alpine Programme. *Q. J. Meteor. Soc.*, **133**, 811–830.
- Rowe, A. K., S. A. Rutledge, T. J. Lang, P. E. Ciesielski, and S. M. Saleeby, 2008: Elevation-dependent trends in precipitation observed during NAME. *Mon. Wea. Rev.*, **136**, 4962–4979.
- Rowe, A. K., S. A. Rutledge, and T. J. Lang, 2011a: Investigation of microphysical processes occurring in isolated convection during NAME. *Mon. Wea. Rev.*, **139**, 424–443.
- Rowe, A. K., S. A. Rutledge, and T. J. Lang, 2011b: Investigation of microphysical processes occurring in organized convection during NAME. Submitted to *Mon. Wea. Rev.*
- Tessendorf, S. A., L. J. Miller, K. C. Wiens, and S. A. Rutledge, 2005: The 29 June 2000 supercell observed during STEPS. Part I: Kinematics and microphysics. *J. Atmos. Sci.*, **62**, 4127–4150.
- Tong, C.-C., 2009: Precipitation characteristics and associated environment regimes during SoWMEX/TiMREX. Master's Thesis, Atmospheric Science, National Taiwan University, Taipei, Taiwan, 90 pp.
- Yeh, H. C., and Y. L. Chen, 1998: Characteristics of rainfall distributions over Taiwan during the Taiwan Area Mesoscale Experiment (TAMEX). *J. Appl. Meteor.*, **37**, 1457–1469.

Zeng, Z., S. E. Yuter, R. A. Houze, Jr., and D. E. Kingsmill, 2001: Microphysics of the rapid development of heavy convective precipitation. *Mon. Wea. Rev.*, **129**, 1882–1904.