Final Report

Radar Rainfall Estimation for Flash Flood Forecasting in Small Urban Watersheds

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Abstract

The utility of WSR-88D rainfall estimates for flash flood forecasting is examined based on observations from the 14.3 km^2 Dead Run watershed in Baltimore County, Maryland. Radar rainfall estimates are computed from volume scan reflectivity observations for 25 storms during the period June - November 2003. Storm to storm variability of bias in the storm total rainfall estimates is assessed based on observations from 19 double-gage stations in the Dead Run watershed. For the convective Z-R relationship, bias ranges from 0.5 to 2.1. Analyses of ancillary forecast variables, including ETA model forecast fields, provide limited predictive capability for specifying the multiplicative bias. Gage - radar analyses show large spatial variability of storm total rainfall over the 14.3 km^2 basin for flash flood producing storms. The ability to represent the detailed spatial variation in storm total rainfall by WSR-88D rainfall estimates varies markedly from event to event. Paired analyses of WSR88D rainfall estimates and discharge observations in Dead Run at basin scales ranging from 1.2 to 14.3 km^2 are used to assess the accuracy of radar-based rainfall estimates for flash flood forecasting. Drop-size measurements from a Joss-Waldvogel disdrometer during July and August of 2004 are used to examine the variability of Z-R relationships for flash flood producing storms. They are also used to examine the potential of polarimetric rainfall estimation algorithms (based on reflectivity and differential reflectivity measurements) for estimating rainfall rate at small spatial scales for flash flood forecasting. Rainfall estimation problems for flash flood forecasting are illus-

trated through analyses of the 7 July 2004 storm, which produced the flood of record in Dead Run. The potential utility of Terminal Doppler Weather Radar (TDWR) observations for flash flood forecasting is also examined through analyses of the 7 July 2004 storm.

1 Introduction

In this report, we examine radar rainfall estimation for flash flood forecasting in small, urban catchments. The spatial scale of interest ranges from approximately 1 to 100 km^2. The time for flood response in small, urban catchments is short, with interest focusing on time scales ranging from 1 to 60 minutes.

Dead Run is a 14.3 km^2 tributary of the Gwynns Falls watershed (Figs. 1 and 2), which is the principal study watershed of the Baltimore Ecosystem Study (BES; see Smith et al. [2004]). The 2003 study period contained an unusually large number of heavy rainfall events, which contributed to the record annual rainfall accumulation in Baltimore of 1574 mm, breaking the record set in 1889.

Accurate, high-resolution (1 km horizontal resolution, 5 minutes time scale) rainfall estimates from the WSR-88D (Weather Surveillance Radar - 1988 Doppler) play a central role in flash flood forecasting. In this study, rainfall analyses are based principally on volume scan reflectivity observations from the Sterling, Virginia WSR-88D radar and storm total rainfall accumulations from a network of 19 double-gage platforms in the Dead Run watershed (Fig. 2). We also include analyses based on observations from the BWI Terminal Doppler Weather Radar (TDWR), a Joss-Waldvogel disdrometer, and a network of 5 stream gages in the Dead Run watershed. ETA model analysis fields are also used to examine the potential for operational specification of the multiplicative bias in radar rainfall estimates.

The rainfall distribution for flood-producing storms in urban environments can exhibit large variability in time and space (Zhang et al. [2003] and Smith et al. [2004]). A special focus of this study is examining the capability to resolve the spatial and temporal distribution of rainfall from WSR-88D reflectivity observations for flash flood forecasting.

The flood of record in Dead Run occurred on 7 July 2004. Observations from the 7 July storm are used to examine potential enhancements to procedures used for quantitative precipitation estimation. In particular we use disdrometer observations as the basis for examining polarimetric rainfall observations. Reflectivity observations from the BWI TDWR for the 7 July 2004 storm are used to examine the utility of higher spatial and temporal resolution reflectivity measurements for flash flood forecasting.

2   Data

WSR-88D reflectivity observations are used to derive rainfall fields for storm events in Dead Run during the 6-month period beginning 1 June 2003. Volume scan reflectivity data from the Sterling WSR-88D (time resolution of 5-6 minutes and spatial resolution of 1 km in range by 1 degree in azimuth) were obtained for 25 storm events, each with basin-averaged storm total rainfall accumulations exceeding 5 mm.

The Dead Run rain gage network consists of 19 double-gage platforms (Figs. 2

and 3). Storm total rainfall was collected for 35 storm events during the 6-month observing period. Analyses focus on the 25 events with mean accumulations exceeding 5 mm.

Drop size distribution measurements from a Joss-Waldvogel Disdrometer were made during the period July - August 2004 at the University of Maryland Baltimore County campus, which is 4 km south of the Dead Run watershed. Drop size distribution measurements are used to examine variability of Z-R relationships and the potential for improving rainfall rate estimates at small space and time scales using polarimetric rainfall algorithms. For the latter objective, the backscatter cross-section data of Goddard and Cherry [1984] are used to compute differential reflectivity, Z_{DR} from drop size measurements.

Five stream gages within the Dead Run watershed (Fig. 2), at basin scales ranging from 1.2 to 14.3 km^2, are used as tools for examining radar rainfall estimates. The U. S. Geological Survey (USGS) stream gage at Franklintown (drainage area of 14.3 km^2) has been in operations since the late 1950s. Gages DR1A and DR2A were installed in June 2003. Gages DR3A and DR4A were installed during the summer of 2004. Stage-discharge rating curves were developed for each of the stations based on direct discharge measurements made during the summers of 2003 and 2004.

The BWI TDWR is located south of the BWI airport and is 19 km from the Dead Run watershed. Reflectivity observations from the TDWR, which are available at 1 degree azimuth resolution and 150 m range resolution, provide far better spatial

resolution over Dead Run than do WSR-88D reflectivity observations. The time resolution of 0.5 degree elevation scans is 1 minute in hazardous weather mode.

ETA model analysis fields were obtained for 11 events during the June - November 2003 period (11-12 June, 12-13 June, 17 June, 18 June, 19 June, 6-7 July, 9 August, 11 August, 18-19 September, 22-23 September and 19 November). Derived fields that were examined include lifted index (and other stability indices), precipitable water (and other humidity variables), and 0 degree C elevation (and other temperature variables).

3 WSR-88D Rainfall Estimation

In this section, we examine WSR-88D rainfall estimates for 25 storms during the period June - November 2003. The observing period covers a time of record rainfall accumulations in Baltimore and numerous flash floods in Dead Run. The 19 November 2003 flash flood resulted in 3 fatalities in Dead Run.

For the ith storm, $R_i(t, x)$ denotes the rainfall rate at time t, relative to the time origin U_i of the storm, and location x within the drainage basin. The domain of the basin is denoted A. The duration of the ith storm event is denoted T_i. The rainfall field for the ith storm is $\{R_i(t, x) \; ; \; t \; \in \; (0, T_i]; \; x \; \in \; A \}$.

Rainfall rate fields $\{R_i(t, x)\}$ are estimated by applying a Z-R relationship ($R = aZ^b$; Z is the radar reflectivity factor in $mm^6 \; m^{-3}$) to reflectivity observations at

volume scan times $t_1, t_2, ...$ and for spatial locations $x_j; j = 1, ..., M$, where x_j denotes the center of the jth 1 degree by 1 km radar bin in the domain A and M is the number of radar bins. Analyses presented in this paper utilize the "convective" Z-R relationship for which a = 0.0174 and b = 0.71 and the "tropical" Z-R relationship for which a = 0.010 and b = 0.83 (Fulton et al. [1998] and Baeck and Smith [1998]). The time-interpolated rainfall rate estimates at radar bin locations are given by linear weighting between volume scan times, i.e.,

$$\bar{R}_i(t, x_k) = \frac{t - t_j}{t_{j+1} - t_j} a[Z_i(t_{j+1}, x_k)]^b + \frac{t_{j+1} - t}{t_{j+1} - t_j} a[Z_i(t_j, x_k)]^b \tag{1}$$

for $t_j < t \le t_{j+1}$. This computation provides estimates at an arbitrary time t, but for locations fixed to the center of radar sample bins x_k. The radar rainfall estimate for time t and location x, $\tilde{R}_i(t, x)$, is obtained by an inverse distance-squared weighting of the observations from radar bins:

$$\tilde{R}_i(t, x) = \sum_{k=1}^{M} w_k \bar{R}_i(t, x_k) \tag{2}$$

for

$$w_k = \frac{\frac{1}{||x - x_k||^2}}{\sum_{j=1}^{M} \frac{1}{||x - x_j||^2}}. \tag{3}$$

The sample bias, B_i, for the ith storm is the ratio of the mean storm total rainfall at rain gage stations to the mean storm total rainfall from radar at rain gage locations, i.e.

$$B_i \; = \; \frac{\sum_{j=1}^{m} G_{ij}}{\sum_{j=1}^{m} \int_0^{T_i} \tilde{R}_i(t, x_j) \, dt} \tag{4}$$

where m is the number of rain gage stations (typically 19 for the Dead Run rain gage analyses), x_j is the spatial location of the jth rain gage and G_{ij} denote storm total accumulation for the ith storm from the jth gage. Bias-corrected rainfall rate estimates are given by:

$$\hat{R}_i(t, x) \; = \; B_i \, \tilde{R}_i(t, x) \tag{5}$$

Storm event bias values for the convective Z-R relationship ranges from 0.52 to 2.14; for the tropical Z-R relationship the range is from 0.22 to 1.70 (Table 1). Of the 11 events that occurred between June 20 and September 1, 9 have bias values less than 1. None of the 14 events outside this time window had bias values less than 1. Bias values for the ten largest flood peaks were all greater than 1.

The multiplicative bias in WSR-88D rainfall estimates varies systematically with convective potential (Fig. 4), as reflected in ETA model fields of lifted index. The analyses in Fig. 4 are based on ETA model fields of lifted index averaged over 9 MDR bins with the center bin containing the Dead Run watershed (see Fig. 5 for similar analyses based on precipitable water). The analyses are all taken from the 21:15 UTC model analysis preceding the storm event. Analyses suggest that unstable conditions over the 120 km by 120 km domain (average lifted index values less

than 0) are associated with significant overestimation using the convective Z-R relationship (bias values less than 1). Highly stable environments can result in severe underestimation by the WSR-88D rainfall algorithms. Lifted index values between 2 and 5 are associated with bias values in the 1.1 - 1.6 range. ETA model fields of lifted index exhibited large temporal variability during the period 11 - 13 June 2003 (Fig. 6). Further analysis is needed to develop operationally useful representations of time-varying bias in terms of model-derived fields of lifted index (or related variables).

Flash-flood producing rainfall can exhibit large spatial variation over the 14.3 km^2 Dead Run basin. For the 12 June 2003 storm, which produced a flood peak of 1.31 m^3 s^{-1} km^{-2}, storm total rainfall derived from WSR-88D varied from less than 10 mm to more than 30 mm (Fig. 7a. Bias-corrected rainfall estimates for this storm capture the spatial variability quite well (Fig. 7b). The capability to represent spatial variation of rainfall at scales smaller than 14.3 km^2 varies significantly from event to event. WSR-88D rainfall estimates for flash flood producing storms on 18-19 September 2003 (Hurricane Isabel) exhibit spatial variation (Fig. 8a) that is not reflected in the rain gage observations (Fig. 8b). For the 22-23 September storm, which produced the annual flood peak in Dead Run, WSR-88D rainfall estimates resolve the markedly larger accumulations in the southern portion of the basin from the smaller accumulations in the central and northern portion of the basin (Fig. 9). Comparable analyses are shown for storms on 14 October (Fig. 10), 27 October (Fig. 11) and 5 November (Fig. 12).

Errors in the spatial representation of rainfall over the Dead Run basin will generally not present serious problems for flash flood forecasting. The accuracy of the time-varying rainfall estimates for the 18-19 September, 22-23 September, and 19 November storms will vary over the 14.3 km^2, but for all sub-basins they will point to a significant potential for flash flooding (which occurred in each case). In some settings, errors in the spatial distribution of rainfall can result in false alarms for flash flooding.

Bias-corrected rainfall estimates provide excellent monitoring of flash flooding at 14.3 km^2 scale of Dead Run. The critical time scale for rainfall estimates varies with basin scale for Dead Run, scale is approximately 30 - 60 minutes (Figs. 13 and 14). The analyses of Figs. 13 and 14 show maximum 60-minute and maximum 15-minute basin-averaged rainfall rate versus peak discharge (expressed as a unit discharge, i.e. $m^3 \ s^{-1} \ km^{-2}$.

4   Analyses of July 7 Storm

A storm on 7 July 2004 produced the flood of record throughout much of the Dead Run watershed, based both on USGS stream gaging observations and interviews with long time residents and merchants. The flood resulted from a 2 hour period of heavy rainfall (Fig. 15). A series of convective elements tracked over the region from 1830 UTC - 2000 UTC (Figs. 16 a-d). Convection organized into a squall line which

produced a period of heavy rainfall over the region after 2000 UTC (Figs. 16 e-g).

Storm total rainfall from WSR-88D shows a swath of heavy rainfall passing through Dead Run (Fig. 17). The rainfall distribution was tightly concentrated around Dead Run, reflecting the most extreme flooding observed in the Baltimore region. On the 14.3 km^2 scale, the WSR-88D rainfall estimates provided an excellent representation of flash flood producing rainfall.

Storm total rainfall estimates (Fig. 18) show large gradients over Dead Run. Unfortunately, these are not reflected in rain gage observations (Fig. 19). Like a number of the 2003 storms, the spatial pattern of heavy rainfall was not accurately reproduce by the WSR-88D rainfall estimates.

Disdrometer observations show significant variability in microphysical properties over the course of the storm. Rainfall rate estimates derived from the convective Z-R relationship (using a bias correction of 0.87 based on the disdrometer rainfall totals) show a period of serious overestimation followed by a period of underestimation during the critical 30 minute period of the storm beginning at 2030 UTC (Fig. 20). Overestimation is linked to a period in which the differential reflectivity takes anomalously large values peaking at more than 3 dB (Fig. 21). The temporal pattern of overestimation and underestimation can be linked to spatial patterns of overestimation and underestimation associated with spatial variations in storm microphysical properties.

TDWR provides very high time and space resolution representation of rainfall.

A comparison of TDWR reflectivity at 1 minute time scale with disdrometer reflectivity (Fig. 22) shows an excellent match, until the end of the storm when TDWR reflectivity values drop to the 40 - 44 dBZ and disdrometer reflectivity values exceed 50 dBZ. This is the first period for which there is significant rainfall in the radial between the disdrometer and the TDWR (Fig. 16g). Attenuation limits the utility of TDWR reflectivity measurements and a high priority for operational use of TDWR reflectivity values is incorporation of attenuation correction algorithms.

Comparison of reflectivity time series between TDWR and WSR-88D (Fig. 23) at the location of the USGS stream gage (and one of the Dead Run rain gages) shows good general agreement, but much larger variation than between disdrometer and TDWR. The spatial averaging of WSR-88D reflectivity observations and 5-6 minute sampling result in a coarse representation of rainfall, both in space and time, relative to that provided by the TDWR.

Storm total rainfall from the TDWR is similar in pattern to the WSR-88D (Fig. 18), but with an even more pronounced spatial gradient. The amplified gradient results from attenuation, which is most severe in the northwestern portion of the Dead Run basin. Time series of reflectivity from TDWR and WSR-88D for the north westernmost gage location shows pronounced attenuation of TDWR.

5 Disdrometer Analyses of Polarimetric Algorithms

Developing accurate radar rainfall estimates from reflectivity is inherently difficult.
Adding a second variable in the form of rain gage observations for bias correction
has also proven challenging, especially from an operational perspective. Bias varies
in space and time in a manner that has proven difficult to address through rain
gage networks. We examine in this section the utility of polarimetric measurements.
Interest focuses on small spatial scales, so we examine (Z, Z_{DR}) rainfall estimates.
K_{DP} measurements are also potentially of great utility, but the range resolution for
these measurements diminishes their utility for flash flood applications at the smallest
spatial scales.

The procedure for estimating rainfall rate from Z and Z_{DR} is based on the gamma
drop size distribution and follows ideas and methods introduced by Ulbrich and Atlas
[1984], Smith [1993] and Brandes et al. [2004]). The gamma drop size distribution
can be represented as:

$$N(x) = N_0 \ x^\mu \ exp(-\Lambda x) \tag{6}$$

Estimation of the parameters μ, and N_0 from disdrometer measurements of drop size
distributions follows the moment-based procedures used in Smith [1993]. Estimators
of the three gamma parameters for the July 7 storm show large variability in time (Fig.
24), highlighting the difficulty in estimating rainfall rate from a single measurement.

If we can capture this variability, however, rainfall rate can be estimated quite accurately. Rainfall rate can be computed from the gamma parameters as follows

$$R = \frac{6\pi \ v_0}{10^4} \ \Gamma(4 + \mu + p) \ N_0 \ \Lambda^{4+\mu+p} \tag{7}$$

where p and v_0 are parameters of the terminal velocity relationship

$$v(x) = v_0 x^p \tag{8}$$

(parameter values $v_0 = 3.866$ and p $= 0.67$ will be used).

Using equation 7 with the estimated gamma parameters (as illustrated in Fig. 24) results in near-perfect reconstruction of rainfall rate (Fig. 25). Thus the information contained in the drop size measurements for a given minute can be condensed to the three gamma parameters, maintaining an almost perfect representation of rainfall rate. The problem of rainfall rate estimation simplifies to estimating the parameters of the gamma distribution.

The crux of the estimation problem is reconstructing the three gamma parameters from the two observations, Z and Z_{DR}. We derive below a procedure for solving this estimation problem. The volume-weighted mean diameter of the drop size distribution

$$D_m = \frac{\int_0^\infty x^4 \ N(x) \ dx}{\int_0^\infty x^3 \ N(x) \ dx} \tag{9}$$

can be represented in terms of the gamma parameters as follows:

$$D_m = \frac{4 + \mu}{\Lambda} \qquad (10)$$

An estimator of volume-weighted mean diameter is obtained from differential reflectivity measurement:

$$\hat{D}_m = c_1 + c_2 \cdot Z_{DR} \qquad (11)$$

Brandeis et al. [2004] argue that there is a strong relationship between the location parameter, Λ, and the shape parameter, μ. Furthermore this relationship can be used to solve the estimation problem. For the Baltimore data, there is a strong relationship (Fig. 26) and the relationship varies with reflectivity. We use the following relationship:

$$\frac{\Lambda}{\mu + 1} = c_3 + c4 \cdot dBZ \qquad (12)$$

Equations 10 - 13 provide the estimation equations for obtaining μ and Λ from Z and Z_{DR}.

The parameter N_0 is estimated from Z and the estimators of μ and Λ using the following expression for reflectivity in terms of gamma parameters:

$$Z = \Gamma(7 + \mu) \, N_0 \, \Lambda^{7+\mu} \qquad (13)$$

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The procedures provides rainfall estimates for the July - August time period (Fig. 27) that are far more accurate than those obtained from Z-R relationships (Fig. 28; convective Z-R).

6 Summary and Conclusions

Principal conclusions of this study are as follows.

1. Bias of storm total rainfall estimates was computed for 25 storms in Dead Run during the 2003 warm season observing periods. Analyses considered convective and tropical Z-R relationships. Bias ranged from 0.51 to 2.14 for the convective Z-R relationship. Bias exhibits systematic temporal variation that may be exploited to improve WSR-88D rainfall estimates. Variables relating to convective potential derived from ETA model analysis fields hold some promise for specification of the multiplicative bias.

2. Basin-averaged rainfall rate at 30-60 minute time intervals are closely linked with peak discharge for Dead Run at 14.3 km^2 scale. Averaging over longer time periods reduces information content concerning peak discharge. The critical time scale for quantitative precipitation estimation decreases with decreasing spatial scale.

3. The bias-corrected, basin-averaged rainfall rate estimates at 14.3 km^2 are very accurate for the purposes of flash flood forecasting. The capability for representing spatial distribution of flash flood producing rainfall diminishes significantly as the

spatial scale decreases to 1 km^2.

4. The high temporal and spatial resolution of TDWR reflectivity observations is potentially of great utility for flash flood forecasting. For the 7 July 2004 storm, TDWR reflectivity observations match the "point" reflectivity observations from a Joss-Waldvogel disdrometer quite well, except when there is heavy rainfall in radial between the disdrometer and TDWR. Attenuation correction algorithms are needed for quantitative precipitation estimation from the TDWR.

5. A polarimetric rainfall estimation algorithm based on reflectivity and differential reflectivity measurements was developed and tested using the Joss-Waldvogel disdrometer data. The algorithm can provide very accurate rainfall estimates, based solely on the two measurements.

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Date	Mean Rain (mm)	Max Rain (mm)	CV	Bias 1	Bias 2	Max 15 Rain $mm\ h^{-1}$	Max 60 Rain $mm\ h^{-1}$	Peak Discharge $m^3\ s^{-1}\ km^{-2}$
June 3	17.1	17.9	0.11	1.43	1.13	4.0	3.1	0.26
June 7	38.1	41.9	0.07	1.97	1.39	15.5	10.6	1.71
June 12a	17.2	34.3	0.49	1.13	0.55	29.1	12.7	1.31
June 12b	11.3	25.5	0.54	0.59	0.31	14.9	4.7	0.47
June 17	13.7	15.2	0.06	2.14	1.70	9.7	6.3	0.49
June 19a	10.6	12.8	0.10	1.02	0.62	15.1	7.3	0.81
June 19b	8.6	9.9	0.08	0.81	0.51	10.5	4.2	0.49
July 2	34.9	41.7	0.07	1.57	1.10	17.5	8.9	0.63
July 7	5.2	15.6	0.82	0.79	0.48	9.2	3.5	0.26
July 23	23.6	27.7	0.10	1.42	0.88	17.0	13.1	0.93
July 28	7.1	8.8	0.14	0.77	0.54	6.2	3.3	0.11

Table 1. Storm event summaries for the 25 events with storm total rainfall accumulation exceeding 5 mm, including: storm date, storm total rainfall (mm), maximum point rainfall accumulation over the basin (mm), coefficient of variation of rainfall accumulation, bias (convective Z-R), bias (tropical Z-R), maximum 15 minute basin-averaged rainfall rate, maximum 60-minute basin-averaged rainfall rate, and peak discharge of Dead Run at the Franklintown stream gage (expressed as a unit discharge, i.e. discharge divided by drainage area).

Date	Mean Rain (mm)	Max Rain (mm)	CV	Bias 1	Bias 2	Max 15 Rain $mm\ h^{-1}$	Max 60 Rain $mm\ h^{-1}$	Peak Discharge $m^3\ s^{-1}\ km^{-2}$
August 3	8.7	13.2	0.28	0.63	0.32	9.9	3.7	0.24
August 9	9.9	15.8	0.26	0.95	0.45	30.9	10.5	0.70
August 11	5.0	13.0	0.74	0.95	0.41	13.6	5.0	0.31
August 16	15.7	21.5	0.23	0.70	0.38	19.2	7.6	0.36
August 22	10.9	21.6	0.38	0.52	0.22	39.1	11.7	0.59
Septem. 1	5.2	16.3	0.24	0.72	0.37	16.1	5.1	0.17
Septem. 12	26.7	29.0	0.08	1.91	1.56	9.6	5.1	0.37
Septem. 18	64.9	73.7	0.11	1.51	0.83	51.3	25.4	2.77
Septem. 22	73.3	85.7	0.11	1.47	0.85	36.5	26.5	3.31
October 14	41.3	48.0	0.11	1.83	1.10	48.7	16.9	1.78
October 27	53.9	56.6	0.05	1.36	0.90	22.0	10.6	1.45
October 28	27.9	29.2	0.02	1.11	0.80	9.4	7.2	0.85
Novemb. 5	22.5	28.0	0.19	1.15	0.59	54.6	17.1	1.68
Novemb. 19	43.5	48.0	0.07	1.00	0.70	30.4	13.5	2.00

Table 1. (continued)

Figure Legends

Figure 1. Overview of the Baltimore study region. Background image is topography derived from USGS 30 meter DEM. The basin boundary for the 14.3 km^2 Dead Run watershed is outlined. Also shown are the basin boundaries for Dead Run, Gwynns Falls and Moores run. The Baltimore City boundary is outlined in purple.

Figure 2. Overview image of the Dead Run study region. Background image is land-use land cover representation, with gray shading denoting impervious regions. Rain gages are denoted by red stars. Basin boundaries for DR1A, DR2A and DR3A are shown.

Figure 3. Rain gage station Security Mall Plaza.

Figure 4. Scatterplot of ETA model lifted index versus WSR-88D storm total rainfall bias for 11 events. All ETA model analyses are for 20:15 UTC prior to or during storm.

Figure 5. Same as in 4, but for ETA model precipitable water.

Figure 6. Time series of ETA model lifted index for 11 - 13 June 2003.

Figure 7. a) Storm total accumulation for 12 June 2003 storm (Dead Run boundary is shown) and b) gage radar plot of storm total accumulation (left panel - without bias correction; right panel - with bias correction).

Figure 8. a) Storm total accumulation for 18 September 2003 storm (Dead Run boundary is shown) and b) gage radar plot of storm total accumulation (bias

corrected).

Figure 9. Same as Fig. 8, but for 22-23 September 2003.

Figure 10. Same as Fig. 8, but for 14 October 2003.

Figure 11. Same as Fig. 8, but for 27 October 2003.

Figure 12. Same as Fig. 8, but for 5 Nobember 2003.

Figure 13. Scatterplot of maximum 60 minute rainfall rate vs. peak discharge (expressed as a unit discharge,i.e. $m^3 \ s^{-1} \ km^{-2}$) for Dead Run.

Figure 14. Same as Figure 13, but for maximum 15-minute rainfall rate.

Figure 15. Time series of basin-averaged rainfall rate and discharge for Dead Run at the Franklintown gage (drainage area of 14.3 km^2) for the 7 July 2004 event.

Figure 16. Reflectivity images from the BWI TDWR radar for the 7 July 2004 storm.

Figure 17. Storm total rainfall map from WSR-88D rainfall analyses for the 7 July 2004 storm (no bias correction).

Figure 18. Detail of the WSR-88D storm total rainfall map for the 7 July 2004 storm (left) and storm total rainfall map derived from BWI TDWR reflectivity observations.

Figure 19. Gage - radar scatterplots for WSR-88D storm total rainfall estimates, without bias correction (top) and with bias correction (bottom).

Figure 20. Time series of rainfall rate derived from disdrometer reflectivity using standard Z-R relationship (denoted by "x") and rainfall rate computed directly from disdrometer observations for the most intense period of the 7 July 2004 storm.

Figure 21. Time series of disdrometer reflectivity (dBZ; top) and differential reflectivity (dB; bottom) for the 7 July 2004 storm.

Figure 22. Time series of TDWR and disdrometer reflectivity for the 7 July 2004 storm. TDWR values are for the 150 m by 1 degree bin containing the disdrometer.

Figure 23. Time series of TDWR and WSR-88D reflectivity for the Franklintown stream gage locations for the 7 July 2004 storm.

Figure 24. Time series of gamma drop size distribution parameters for the 7 July 2004 storm (μ top; Λ middle and N_0 bottom).

Figure 25. Scatterplot of rainfall rate computed directly from disdrometer observations and computed from gamma model parameters.

Figure 26. Scatterplot of reflectivity vs. the ratio of Λ to $\mu + 1$ for disdrometer observations.

Figure 27. Scatterplot of rainfall rate computed directly from disdrometer observations vs. rainfall rate computed from disdromter reflectivity using standard Z-R relationship.

Figure 28. Scatterplot of rainfall rate computed directly from disdrometer observations vs. rainfall rate computed from (Z, Z_{DR}) estimation procedure (as described

in the text).