The Impacts of Taiwan Topography on the Predictability of Typhoon Morakot’s Record-Breaking Rainfall: A High-Resolution Ensemble Simulation

XINGQIN FANG
Department of Atmospheric Sciences, School of Environmental Sciences and Engineering, Sun Yat-Sen University, Guangzhou, China, and National Center for Atmospheric Research, Boulder, Colorado

YING-HWA KUO
National Center for Atmospheric Research, Boulder, Colorado

ANYU WANG
Department of Atmospheric Sciences, School of Environmental Sciences and Engineering, Sun Yat-Sen University, Guangzhou, China

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ABSTRACT

In this study, the impacts of Taiwan topography on the extreme rainfall of Typhoon Morakot and the predictability of this rainfall are examined with a high-resolution (4 km) ensemble simulation using the Advanced Research core of the Weather Research and Forecasting Model (WRF-ARW). Ensemble prediction with realistic topography reproduces salient features of orographic precipitation. The 24- and 96-h accumulated rainfall amount and distribution from the ensemble mean compare reasonably well with the observed precipitation. When the terrain of Taiwan is removed, the rainfall distribution is markedly changed, suggesting the importance of the orography in determining the rainfall structure. Moreover, the peak 96-h rainfall amount is reduced to less than 20%, and the total rainfall amount over southern Taiwan is reduced to less than 60% of the experiments with Taiwan topography. Further analysis indicates that Taiwan’s topography substantially increases the variability of rainfall prediction. Analysis uncertainties as reflected in the perturbed initial state of the ensemble are amplified due to orographic influences on the typhoon circulation. As a result, significant variability occurs in the storm track, timing, and location of landfall, and storm intensities, which in turn, increases the rainfall variability. These results suggest that accurate prediction of heavy precipitation at a specific location and at high temporal resolution for an event such as Typhoon Morakot over Taiwan is extremely challenging. The forecasting of such an event would benefit from probabilistic prediction provided by a high-resolution mesoscale ensemble forecast system.

1. Introduction

Taiwan is an island that possesses complicated meso-scale topography. Its Central Mountain Range (CMR) has dimensions of about 200 km × 100 km and includes many peaks that exceed 3000 m. Heavy rainfall resulting from the orographic effects on the typhoon circulation associated with the CMR is one of the most serious threats to Taiwan. From 6 to 10 August 2009, Typhoon Morakot brought extraordinary rainfall over Taiwan, breaking 50-yr precipitation records and causing the loss of more than 700 lives with an estimated property damage exceeding $3.3 billion (USD). Figures 1a and 1b show the analysis of the observed 96- and 24-h rainfall amounts, based on the hourly gauge data collected from about 450 automatic stations scattered throughout Taiwan. During the 4-day period, more than half of southern Taiwan received more than 800 mm of accumulated rainfall (Fig. 1a). Some of the mountainous areas recorded more than 2500 mm, with the maximum 96-h gauge value of 2874 mm recorded at Chiayi County ending at 0000 UTC 10 August 2009 (marked with the white star in Fig. 1a). From 0000 UTC 8 August to 0000 UTC 9 August, the most intensive
rainfall period, 1504 mm was recorded at the same Chiayi County (marked with the white star in Fig. 1b).

The westward-moving Typhoon Morakot landed in central eastern Taiwan at around 1530 UTC 7 August 2009. Figure 2 shows the infrared cloud imagery of Typhoon Morakot before, during, and after landfall. Although this was not a particularly intense storm (e.g., a category 2 storm), the size of its circulation was fairly large. In particular, the structure of its circulation possessed distinct asymmetry, with strong northwesterly, westerly, or southwesterly flow sustained in its southern quadrants before, during, and after landfall. Significant moisture flux and convergence resulting from the sustained westerly and southwesterly flows impinging upon the mountainous area of southern Taiwan produced continuous heavy rainfall over a very large area, causing severe flooding and landslides.

Typhoon Morakot’s rainfall distribution (Figs. 1a and 1b) exhibited significant small-scale orographic rainfall features as a result of time-varying typhoon
circulation impinging upon the fixed Taiwan mountains. To better understand the predictability of the extreme rainfall associated with Typhoon Morakot, it is important to examine the impacts of Taiwan’s topography. Recently, many high-resolution sensitivity simulations were performed to study orographic effects on typhoon circulations for different typhoon cases (e.g., Jian and Wu 2008; Yang et al. 2008). Most of these studies are deterministic modeling studies. Previous studies have suggested that there are many sensitive parameters associated with the orographic effects on typhoon circulations. As reviewed by Wu and Kuo (1999), the behavior of a typhoon approaching Taiwan varies as a function of the intensity and size of the typhoon, as well as the environmental flow. The control parameters influencing the continuity and deflection of
typhoon tracks are related to the typhoon strength and its structure; the scale, height, and dimension of the mountain; and the large-scale background features (Chang 1982; Yeh and Elsberry 1993a,b; Lin et al. 2002, 2005, 2006). As shown in Wu et al. (2002), even if the track is not sensitive to the existence of topography and is well simulated, the rainfall can be very sensitive to model resolution and topography. In addition, the effects of planetary boundary layer physics, precipitation parameterization, impinging angles, and landfall location can significantly influence the orographic effects on typhoon circulations. Practical predictability is limited by uncertainties in both the initial states and the forecast models. For these reasons, it is necessary to investigate the impacts of Taiwan’s topography on the typhoon mesoscale structures and precipitation forecasts from a stochastic perspective.

In this paper, we present results from high-resolution ensemble forecast experiments on Typhoon Morakot with and without Taiwan’s topography. Section 2 presents the experiment design. Section 3 discusses the role of Taiwan’s topography in Typhoon Morakot’s extreme rainfall. Section 4 examines the rainfall variability impacted by Taiwan topography. Section 5 investigates the variability of the storm features impacted by Taiwan topography and their relationship with the rainfall variability. Section 6 discusses the ensemble probability prediction and the performance of the high-resolution ensemble simulation. Finally, our conclusions are presented in section 7.

2. Methodology

a. Data and experiment design

Two sets of 96-h 32-member ensemble forecast experiments initiated at 0000 UTC 6 August 2009 are performed with the version 3.1.1 of the Advanced Research core of the Weather Research and Forecasting Model (WRF-ARW; Skamarock et al. 2008). Set 1 (hereafter referred to as EN0600) uses the real topography (at 30° resolution) based on Moderate Resolution Imaging Spectroradiometer (MODIS) land-use data, while set 2 (hereafter referred to as NTEN0600) reduces the terrain height of Taiwan to zero but still retains its land surface character. The terrain everywhere else remains unchanged. The ensemble initial and boundary conditions (ICBCs) are randomly perturbed from the high-resolution (0.225° × 0.225°) European Centre for Medium-Range Weather Forecasts (ECMWF) analysis by adding analysis uncertainties based on WRF three-dimensional variational data assimilation (3DVAR) background error covariance. Basically, we use a random number generator to generate 32 sets of Gaussian perturbations for all of the control variables. These perturbations are then transformed into analysis increments through the WRF 3DVAR system. Thirty-two analysis members are generated every 6 h during the entire 96-h simulation period, and then the boundary conditions are updated for each member. The background error statistics used is the CV5 option of WRF 3DVAR system, where the control variables operate in eigenvector space. This procedure ensures that the final perturbations are consistent with the errors of the background fields in terms of their scales and amplitudes and are compatible with the 36-km resolution of our outermost domain. Two corresponding deterministic simulations without any perturbations in the ICBCs (hereafter ec0600 and ecNT0600) were performed for comparison. Figure 3 shows the model domain configuration and the geophysical height distribution, with the illustration for the removal of Taiwan’s topography for the sensitivity experiment. The two-way interactive, triple-nested domains include 36- (280 × 172), 12- (430 × 301), and 4-km (364 × 322) meshes, respectively, extending vertically to the top at 20 hPa, with 36 η levels. The model physics include the WRF single-moment five-class (WSM5) microphysics package (Hong et al. 2004; Hong and Lin 2006), the Yonsei University (YSU) planetary boundary layer scheme (Hong et al. 2006; Hong 2007), the Noah land surface model (Chen and Dudhia 2001), the Rapid Radiative Transfer Model (RRTM) longwave radiation scheme (Mlawer et al. 1997), the Goddard shortwave radiation scheme (Chou and Suarez 1994), and the Betts–Miller–Janjic (BMJ) cumulus convective parameterization (Betts 1986; Betts and Miller 1986; Janjic 1994, 2000). These physics schemes are used for all three meshes.

b. Predictability measure

The standard deviation (SD) between the ensemble members is often used to quantify the variability (i.e., the ensemble spread) of a particular variable. The SD is calculated as

\[
SD(i, j, t) = \sqrt{\frac{1}{M} \sum_{m=1}^{M} [R_m(i, j, t) - \bar{R}(i, j, t)]^2}; \quad (1)
\]

and the associated root-mean-square error (RMSE) and mean error (ME) are defined as

\[
RMSE(i, j, t) = \sqrt{\frac{1}{M} \sum_{m=1}^{M} [R_m(i, j, t) - O(i, j, t)]^2}; \quad (2)
\]
\[ ME(i, j, t) = \frac{1}{M} \sum_{m=1}^{M} [R_m(i, j, t) - O(i, j, t)] \]  

where \( m \) denotes the ensemble member index, \( M \) the total number of ensemble members, \( R \) the verification variable, \( \bar{R} \) the ensemble mean, \( O \) the observation, \( i \) and \( j \) the gridpoint two-dimensional spatial coordinate indices, and \( t \) the time coordinate index.

Some authors (e.g., Mitchell et al. 2002; Zhang et al. 2006) use the root mean of the difference total energy (RM-DTE) calculated from the model variables as a measure of the predictability in ensembles. In our study, we focus on the rainfall predictability within a targeted verification area, the worst-hit area (hereafter referred to as HA) in southern Taiwan, which is defined by the box in black in Fig. 1a. The target area’s terrain is shown in Fig. 3. As pointed out by Jolliffe and Stephenson (2003), rainfall is a highly discontinuous variable with highly skewed distributions; it is very challenging to verify rainfall and measure rainfall variability using the instantaneous rainfall prediction on the model grid. Therefore, in order to reduce rainfall discontinuity, we use 3-h rainfall on the model grid as the basic verification variable; this is equivalent to performing time averaging before the calculation of the rainfall SD. The rainfall SD between the members of the ensemble simulation is used as a measure of rainfall predictability. Moreover, the SDs of some other storm features, such as storm position, storm intensity, etc., are used as measures of storm variability.

3. The role of Taiwan topography in Typhoon Morakot’s extreme rainfall

Figure 4 shows the spatial distribution of the simulated 96-h accumulated rainfall ending at 0000 UTC 10 August by the 32 ensemble members. Almost all members of EN0600 with Taiwan topography reproduce the key observed orographic rainfall features, with most of the precipitation concentrated along the windward side of the mountain, which is very different from the members of NTEN0600 without Taiwan’s topography. Almost all of the members of EN0600 predict 96-h accumulated rainfall over 2500 mm; while only a few members of NTEN0600 produce 96-h accumulated rainfall exceeding 800 mm.

Figures 1c and 1d show the spatial distributions of the simulated 96-h accumulated rainfall ending at 0000 UTC 10 August from the ensemble means of EN0600 and NTEN0600, respectively. The EN0600 mean predicts three heavy rainfall areas with amounts exceeding 800 mm (Fig. 1c) and several spots with amounts over 2500 mm. The rainfall extremes are distributed along the general orientation of the southern mountain range and along the windward slope, which closely resemble the observed orographic rainfall features shown in Fig. 1a. On the contrary, when the terrain of Taiwan is removed, the simulated rainfall is evenly distributed and exhibits no
FIG. 4. The spatial distribution of the simulated 96-h accumulated rainfall ending at 0000 UTC 10 Aug over Taiwan and surrounding islands (mm) from the 32 ensemble members of EN0600 (labeled Tm1–32) and NTEN0600 (labeled NTm1–32).
obvious local extremes (see Fig. 1d). The peak rainfall amount in the NTEN0600 mean is only 616 mm, which is less than 20% of that (3128 mm) in the EN0600 mean. Table 1 shows some areal-average statistics of 24- and 96-h rainfall in the HA. The areal-average 96-h accumulated rainfall in the HA is 917 mm for the EN0600 ensemble mean and 535 mm for NTEN0600. If we view the difference in the rainfall between EN0600 and NTEN0600 as being “orographically additive rainfall” (or OAR), the areal-average OAR in terms of 96-h total rainfall is 382 mm, which means that the total rainfall amount in the HA is increased by 70% by Taiwan’s topography.

Figure 5 shows the time series of some areal-average statistics of the 3-h rainfall in the HA. The areal-average OAR in terms of 96-h total rainfall is 382 mm, which means that the total rainfall amount in the HA is increased by 70% by Taiwan’s topography.

4. The rainfall variability in the HA impacted by Taiwan topography

a. Areal-average 24- and 96-h rainfall SD in the HA

If we view the difference in the SD between EN0600 and NTEN0600 as being “orographically additive rainfall variability (or OARV),” as shown in Table 1, the areal-average 24-h rainfall OARV in the HA is positive for each of the four simulation days, and the areal-average 96-h accumulated rainfall SD is significantly increased with the existence of Taiwan's topography. The maximum OARV occurred on the third day, when the observed rainfall is also the greatest.

b. Spatial distribution of 24- and 96-h rainfall SD in the HA

As shown in Fig. 6, significant 24- and 96-h rainfall SDs in the HA are induced by Taiwan’s topography, especially on the second and third simulation days. For instance, in EN0600, there exist two areas with SDs exceeding 250 mm on the second day and three areas with SDs exceeding 350 mm on the third day, while the maximum SD in NTEN0600 on these 2 days is less than 150 mm.

c. Time series of areal-average 3-h rainfall SD in the HA

The areal-average 3-h rainfall OARV in the HA is positive during the entire simulation period except for a short 9-h period from 1500 UTC 7 August to 0000 UTC 8 August (see the dotted curve in Fig. 5b). The OARV is about 5–8 mm on average before this 9-h period. After that, the OARV becomes even larger, with amounts exceeding 15 mm.

<table>
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<tr>
<th>Rainfall variables</th>
<th>24-h rainfall (0000 UTC 6 Aug–0000 UTC 7 Aug)</th>
<th>24-h rainfall (0000 UTC 7 Aug–0000 UTC 8 Aug)</th>
<th>24-h rainfall (0000 UTC 8 Aug–0000 UTC 9 Aug)</th>
<th>24-h rainfall (0000 UTC 9 Aug–0000 UTC 10 Aug)</th>
<th>96-h rainfall (0000 UTC 6 Aug–0000 UTC 10 Aug)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Analyzed observation (OBS, mm)</td>
<td>84</td>
<td>321</td>
<td>625</td>
<td>179</td>
<td>1209</td>
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<td>EN0600_mean (mm)</td>
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<td>356</td>
<td>48</td>
<td>917</td>
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<td>NTEN0600_mean (mm)</td>
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<td>535</td>
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<td>OAR* (mm)</td>
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<td>107</td>
<td>165</td>
<td>36</td>
<td>382</td>
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<tr>
<td>EN0600_SD (mm)</td>
<td>39</td>
<td>115</td>
<td>183</td>
<td>32</td>
<td>216</td>
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<tr>
<td>NTEN0600_SD (mm)</td>
<td>27</td>
<td>91</td>
<td>114</td>
<td>11</td>
<td>142</td>
</tr>
<tr>
<td>OARV** (mm)</td>
<td>12</td>
<td>24</td>
<td>69</td>
<td>21</td>
<td>74</td>
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<td>Correlation coefficient of SD and ensemble mean in EN0600</td>
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<td>0.89</td>
<td>0.62</td>
<td>0.22</td>
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<tr>
<td>Correlation coefficient of SD and ensemble mean in NTEN0600</td>
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<td>0.48</td>
<td>0.89</td>
<td>0.62</td>
<td>0.22</td>
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<td>EN0600_ME (mm)</td>
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<td>−131</td>
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<td>EN0600_RMSE/EN0600_SD (%)</td>
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<td>1.96</td>
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<td>Correlation coefficient of EN0600_SD and EN0600_RMSE</td>
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<td>0.69</td>
<td>0.58</td>
<td>0.41</td>
<td>0.63</td>
</tr>
</tbody>
</table>

* OAR denotes orographically additive rainfall, the rainfall difference between EN0600 and NTEN0600.

** OARV denotes orographically additive rainfall variability, the rainfall variability difference between EN0600 and NTEN0600.
The comparison of the time series of the SD (the dotted curves) and ensemble mean (the solid curves) of NTEN0600 in Fig. 5a shows that they are generally in phase with each other: the larger the ensemble mean, the larger the SD. However, it is not always the case in EN0600; they are out of phase during the 18-h period from 1500 UTC 7 August to 0900 UTC 8 August, when the ensemble mean is near a peak while the SD is exhibiting a local minimum. This is unique in EN0600 and may be a result of the orographic effects on typhoon circulations, which will be discussed later.

According to the time evolution of the simulated areal-average 3-h rainfall SD in the HA and its relationship with the ensemble mean shown in Fig. 5a, the total 96-h simulation period of EN0600 with Taiwan topography can be separated into five stages: stage 1 (30-h period from 0000 UTC 6 August to 0600 UTC 7 August), when the SD and ensemble mean are well in phase with each other, both reaching a local maximum from 2100 UTC 6 August to 0000 UTC 7 August; stage 2 (9-h period from 0600 UTC 7 August to 1500 UTC 7 August), when the SD and ensemble mean have similar positive trends; stage 3 (18-h period from 1500 UTC 7 August to 0900 UTC 8 August), when the SD and ensemble mean are out of phase with each other and when the ensemble mean reaches its global maximum from 2100 UTC 7 August to 0000 UTC 8 August, while the SD persist, although the ensemble mean decreases rapidly; and stage 5 (33-h period from 1500 UTC 8 August to 0000 UTC 10 August), when the SD and ensemble mean both have similar decreasing trends.

Obviously, the most significant OARV is at stage 4 and the first half of stage 5 (see the dotted curve in Fig. 5b). Note that this is also the period when the model significantly underpredicts the observed rainfall (see the dotted–dashed curve in Fig. 5b). As will be shown later in section 5, the storm (both in the observations and the simulation) has left Taiwan during this period.

d. Spatial distribution of 3-h rainfall SD in the HA

Figure 7 shows the spatial distribution of the simulated 3-h rainfall SD in the HA at 3-h intervals from 0000 UTC 7 August to 0000 UTC 9 August. This 2-day period is selected for detailed examination because it includes all of the five stages mentioned above and as shown in Fig. 5a, and it is also the period when the simulated rainfall and rainfall variability are high.

Consistent with the results of SD and OARV presented in Fig. 5, EN0600 has much larger rainfall variability than NTEN0600, in terms of the maximum SD values and the area coverage with SD at large thresholds. For example, for a threshold of 40 mm, EN0600 has a much larger areal coverage than does NTEN0600 after 0900 UTC 8 August, except for the short period from 1500 UTC 7 August to 0300 UTC 8 August. For NTEN0600, there is only a brief period, from 1500 UTC to 2100 UTC 7 August, when the simulated typhoon rainbands propagate over the HA and produce significant rainfall and rainfall variability. The maximum 3-h rainfall SD in EN0600 is over 90 mm at the southern tip of the CMR from 1200 UTC to 1500 UTC 7 August. For NTEN0600, there is also a region of large SD near the southern boundary of the HA, but the amount of SD is 10 mm less and it occurs 6 h later. The areas with SDs exceeding 50 mm only exist for a very short period (12 h) in NTEN0600 in the southern part of the HA. On the contrary, in EN0600, regions with SD exceeding 50 mm persist over a much longer period (36 h) and cover larger areas, with the distribution pattern changing with time. From 0600 UTC to 1800 UTC 7 August, one large SD area is located at the northern part of the HA and the other one is near the southern tip of the CMR; while from 0300 UTC 8 August to 0000 UTC 9 August, a string of small areas with SDs exceeding 50 mm is located along the mountain range. The distribution of large SD
in EN0600 tends to orient along the CMR, showing the distinct orographic influence.

e. Spatial correlation between the rainfall SD and ensemble mean in the HA

As shown in Figs. 6 and 7, the rainfall SD generally matches well with its corresponding ensemble mean in terms of spatial distribution, for both EN0600 and NTEN0600. The orographically enhanced small-scale, large-value SD structures are consistent with the orographically enhanced small-scale structures of ensemble mean rainfall. This relationship suggests that the spatial distribution of the rainfall ensemble spread is not independent of the ensemble mean; the larger the mean, the larger the spread. Such a relationship between ensemble mean precipitation and spread was discussed by Hamill and Colucci (1998). In our experiments, significant (at the 99% confidence level with the Student’s t test) spatial correlation between the rainfall SD and the ensemble mean in the HA exists in both EN0600 and NTEN0600. However, a higher correlation exists in EN0600 with Taiwan topography (see Table 1).

Figure 8 is a scatterplot of the 3-h rainfall SD and ensemble mean in the HA at 3-h intervals from 0000 UTC 7 August to 0000 UTC 9 August. The correlation coefficient between the 3-h rainfall SD and the ensemble mean is, in general, larger in EN0600 than in NTEN0600. The smaller regression coefficient of the 3-h rainfall SD against the ensemble mean in EN0600 than in NTEN0600 may be related to the much larger rainfall amount in EN0600 with Taiwan’s topography. The correlation coefficient is relatively low (between 0.66 and 0.76) from 1200 UTC to 2100 UTC 7 August, when the storm is very close to or on top of Taiwan in EN0600. Also, there is an obvious cluster separation in SD between large and small ensemble rainfall thresholds, with a very large range of rainfall SDs occurring at smaller values of ensemble mean rainfall. The minimum regression coefficient of 0.18 in EN0600 occurs from 2100 UTC 7 August to 0000 UTC 8 August, which is consistent with the results shown in Fig. 5, with relatively small SDs tied to the large ensemble mean rainfall.

We note that the amplitude ratio between the SD and ensemble mean (or the SD normalized by the ensemble mean) may vary significantly in space. For instance, as
shown in T00/6–00/10 of Fig. 6, of the several areas with large SD values, one area with SD exceeding 400 mm, located at the northern part of the HA (hereafter referred to as area A), is tied to the ensemble mean precipitation of about 1500 mm. On the other hand, another area with large SD values located at the southern tip of the CMR (hereafter referred to as area B) is associated with ensemble mean precipitation of over 2500 mm. These two areas have similar values of rainfall SD but with very different ensemble mean precipitation amounts. The loci of areas A and B are shown in Fig. 9.

f. Rainfall predictability at specific locations

Ensemble spread has been used as a measure to gauge flow-dependent errors through the ensemble spread–skill relationship (see, e.g., Houtekamer 1993; Wobus and Kalnay 1995; Whitaker and Loughe 1998). However, this may not be applicable to a discontinuous variable with a skewed distribution such as rainfall.

Figure 9 shows the spatial distribution of the mean error (ME) of the 24- and 96-h accumulated rainfall in the HA in EN0600. The relationship between the rainfall SD and the magnitude and distribution of rainfall errors (as reflected in ME–RMSE) shown in Table 1 and Figs. 5, 6, and 9 suggests that the rainfall SD can serve as an indicator for ensemble rainfall errors in some sense, at least for integrated quantities such as 24- or 96-h total rainfall. However, for rainfall forecast verification at a specific location and time, it may not be so simple.

Two rain gauge stations within areas A and B are selected to examine the 3-h rainfall SD and model rainfall prediction skills, respectively: station A (23.51°N, 120.80°E) at Chiayi County within area A, which received the maximum 96-h gauge value shown in Fig. 1a, and station

![Figure 7. The spatial distribution of the 3-h rainfall SD (mm, color shaded in levels as in the color bar) and ensemble mean (mm, contoured in levels of 10, 30, 50, 70, 100, 150, and 200 mm) for EN0600 and NTEN0600 (labeled T and NT, respectively, and time stamped) in 3-h intervals from 0000 UTC 7 to 0000 UTC 9 Aug in the HA.](image)
B (22.59°N, 120.60°E) in area B, which is near the maximum of model 96-h rainfall in Fig. 1c. Figure 10 shows the time series of selected 3-h rainfall statistics at these two stations. Significant orographically enhanced rainfall and rainfall variabilities (i.e., OAR and OARV) are introduced at both stations. The OARV values of these two stations are comparable, with relatively small values just after the storm makes landfall, and the maximum value after the storm leaves Taiwan, consistent with the OARV curve in Fig. 5b. However, their OARs are quite different. The very large OAR at station B is due to its orographically amplified erroneous heavy rainfall prediction.

As shown in Fig. 10, the EN0600 ensemble mean significantly underpredicts the observed rainfall at station A after 0900 UTC 8 August but significantly overpredicts the rainfall at station B, especially before 2100 UTC 7 August. Although the ensemble spread has been used to estimate or predict large-scale flow-dependent errors for traditional meteorological variables, this may not be applicable to rainfall. It is evident from Fig. 10 that there is no clear relationship between model rainfall errors and rainfall SD at a fixed geographical location at 3-h intervals. The model rainfall RMSE at station A varies significantly with time and reaches extreme values after 0900 UTC 8 August. However, there is no apparent time

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**Fig. 8.** The scatterplot of the 3-h rainfall SD (Y axis, mm) and ensemble mean (X axis, mm) for EN0600 and NTEN0600 (labeled T and NT, respectively, and time stamped) in 3-h intervals from 0000 UTC 7 Aug (00/7) to 0000 UTC 9 Aug (00/9) in the HA.

**Fig. 9.** The spatial distribution of the 24- and 96-h accumulated rainfall ME (mm) in the HA in EN0600: (a) 00/6–00/7; (b) 00/7–00/8; (c) 00/8–00/9; (d) 00/9–00/10; (e) 00/6–00/10. The blue (red) box illustrates area A (B).
variation in SD around that time. Likewise, no clear relationship exists between the ensemble rainfall errors and the rainfall SD for station B. It is quite evident from Fig. 10 that when the model has large systematic errors, the SDs among members will not adequately reflect these errors.

5. The relationship between the variability of storm features and rainfall variability

Figure 11 shows the simulated tracks of Typhoon Morakot of the EN0600 and NTEN0600 ensemble members. For EN0600, the model storm tracks generally follow the observed track well during the first 36-h simulation, until right before landfall. At about 50 km before landfall, the observed storm started to turn northwestward and moved through northern Taiwan after making landfall. The model storms make landfall south of the observed landfall location and show only a very slight northwestward turning after landfall. Consequently, the model storms move across central Taiwan, and the storm tracks deviate farther away from the observed storm with time. The model storm tracks do not have a significant amount of spread in the ensemble before the storms leave Taiwan. The systematic southward and westward storm track biases may account for, at least partially, the large model rainfall error at the Chiayi station on the third day (0000 UTC 8–9 August).

The model storms in NTEN0600 behave similarly to those of EN0600 before landfall for the first 30 h. However, at about 250 km before landfall, the model storms in NTEN0600 veer to the north of the observed track, as well as that of the model storms in EN0600. The model storms then move across Taiwan with nearly a straight westward track. For NTEN0600, the ensemble storm tracks start to diverge earlier than those in EN0600, at around landfall time. It is important to note that the tracks shown in Fig. 11 do not completely represent differences in storm positions with time. Some storms whose overall tracks look similar to each other may move slower or faster than the others.

Figure 12 shows the time series of the storm position spread in EN0600 and NTEN0600. Note that the dispersion radius of storm position, the latitude spread, and the longitude spread are calculated separately, because they are each important parameters when referring to the location of the storm relative to the fixed Taiwan topography. And in this way, the storm position dispersion can be better presented than the track dispersion shown in Fig. 11.

For EN0600, the storm position spread increases sharply during 1200–1800 UTC 7 August at around landfall time. By comparison with the storm position

\[ \text{FIG. 10. The time series of some 3-h rainfall statistical measures (mm) at stations (a) A and (b) B. See text for station descriptions and locations.} \]

\[ \text{FIG. 11. The simulated track of ensemble members (green lines), the single deterministic simulation (blue line), and the average track of the ensembles (red line) of (top) EN0600 and (bottom) NTEN060. The JMA best track (modified by analysis from the Taiwan Central Weather Bureau) is superimposed as the thick black line labeled OBS. The track is plotted from 00/6 to 06/9 August, and the storm positions are shown by the red circles at 00/6, 00/7 and 00/8, blue circles at 12/6 and 12/7, black circles at 12/8 and cyan circles at 00/9.} \]
spread in NTEN0600, the impacts of topography on the storm position spread become quite obvious. The difference between the two will be called orographically additive storm position variability (OASPV), which can be attributed to increased variability in the latitudinal and longitudinal directions. Although the storm ensemble mean propagation speed in the east–west direction is not significantly impacted by Taiwan’s topography, the storm position longitude variability substantially increases near the landfall time, which is probably due to variability in the timing of landfall among ensemble members in EN0600. Figure 12b shows that between 0000 and 1800 UTC 7 August, the NTEN0600 ensemble mean shows a distinct northward track bias when compared with the observed track. Such bias is largely eliminated in EN0600 with the inclusion of topography. After 1800 UTC 7 August both EN0600 and NTEN0600 show southward and westward track biases, consistent with Fig. 11. There are slight differences in storm position spread between EN0600 and NTEN0600 after 1800 UTC 7 August (after the model storms leave Taiwan). The differences in position variability between these two sets of ensemble experiments in the latitudinal and longitudinal directions appear to cancel each other.

The storm intensity is weakened around landfall time in both EN0600 and NTEN0600, but it is weakened 3–5 hPa more with the impacts of Taiwan’s topography, and the orographically additive storm intensity variability (OASIV) around landfall time is about 5–8 hPa (Fig. 12d). After landfall, the OASIV is negative when the storms are moving toward the mainland, related perhaps to the significant impacts of the mainland topography on the more intense storms in NTEN0600 at that time.

To better understand the relationship between the variability of the storm features and the rainfall variability over southern Taiwan impacted by Taiwan’s topography, three 3-h periods, 1200–1500 UTC 7 August, 2100 UTC 7 August–0000 UTC 8 August, and 1200–1500 UTC 8 August in stages 2, 3, and 4, respectively, are selected for further examination.

a. 1200–1500 UTC 7 August (model simulation time 36–39 h)

This period is around the time when the model storms make landfall, with very large rainfall SD in the HA in both EN0600 and NTEN0600 (see Figs. 5a and 7). Figure 13 shows the spatial distribution of the model 3-h rainfall and the associated sea level pressure (SLP) at 1500 UTC 7 August. Both the inner and outer circulations of the typhoon are moving over Taiwan during this period. The simulated SLP field and rainfall distribution involved with the typhoon rainbands in EN0600 with Taiwan’s topography are totally different from those in NTEN0600 without Taiwan’s topography.

As noted by Lin et al. (1999), the low pressure and circulation centers tend to split when a tropical cyclone passes over the CMR, which makes it difficult to trace the path of the storm center. Based on the analysis of many historical typhoons influencing Taiwan, Wang...
(1980) showed that when a typhoon passes over Taiwan, its track may be continuous or discontinuous. When the center of a typhoon with a discontinuous track is about to make landfall, a secondary center (or secondary low) often forms on the lee side of the CMR. As the primary typhoon center above the mountain moves over and becomes aligned with the secondary center, the secondary low develops and replaces the original surface
center. The formation of the secondary center often influences the spatial distribution and intensity of local rainfall and makes the rainfall forecast even more difficult. Some typhoons may experience a “quasi-continuous track” phenomenon (Lee et al. 2008), in which the convective bands and the circulation associated with the secondary low interact with the topography to produce local heavy rainfall.

In our simulation with Taiwan’s topography, as shown in Tm1–32 in Fig. 13, the typhoon circulations are significantly modulated by topography, and the formation of a secondary low is clearly indicated by the minimum SLP. Most of the ensemble members form a secondary low on the lee side over northwest Taiwan that is presumably associated with orographic effects on the typhoon’s inner circulations. Before the storm completely passes over the mountains, a low pressure trough extends southward on the lee side over the southeast Taiwan seashore and a transient low pressure center shows up in some members. Together with the original low, a total of three lows show up on the SLP map. This transient low pressure center can be slightly deeper than that of the secondary low over northwest Taiwan or even the main typhoon low center, as found in members 1, 10, 12, 17, 19, 20, and 28.

Due to the variability in the timing of landfall and the orographic effects on the typhoon circulations, the relative positions between the storms and the CMR vary among the ensemble members, which in turn results in large rainfall variability. For 3-h rainfall exceeding 100 mm, the two areas with the largest variability among Tm1–32 of Fig. 13 are the two regions, areas A and B, with large SDs in T00/7–00/8 and T00/6–00/10 of Fig. 6 and T12/7–15/7 of Fig. 7. Obviously, the large rainfall SD within area A is related to the different positions of the inner typhoon circulation relative to the CMR; while the even larger rainfall SD within area B is related to the larger diversity of positions of the outer typhoon circulation patterns relative to the topography over southern Taiwan.

In sharp contrast, the storm position spread for ensemble members without Taiwan topography is much smaller during the period of 1200–1500 UTC 7 August (see Fig. 12), and no secondary low is found in any member of NTm1–32 in Fig. 13. It is obvious that without Taiwan’s topography, the typhoon circulation structure is not disturbed, and both the inner and outer rainbands are intact. Neither the rainfall distribution nor its variability exhibits any topographical influence.

b. 2100 UTC 7 August–0000 UTC 8 August (model simulation time 45–48 h)

During this period, the simulated storms in EN0600 with Taiwan’s topography have a relatively small rainfall SD concurrent with the greatest ensemble mean rainfall in the HA. Figure 14 shows the distribution of the simulated 3-h rainfall during this period and the associated SLP at 0000 UTC 8 August in EN0600. The simulated storms are about to leave or have just left the island of Taiwan. On average, the storm is located on the west coast of Taiwan with sizeable position spread. However, regardless of the storm position spread, the typhoon circulations for almost all the members are positioned very favorably for topographically enhanced precipitation, with broad and strong moist southwesterly flows impinging upon the CMR at a very favorable angle. Under these circumstances, the orographic effects produce almost identical heavy rainfall that is closely tied to Taiwan’s topography. As a result, the 3-h rainfall in the HA of almost each member is very large and the rainfall is distributed along the mountain range in a very similar pattern (see Tm1–32 in Fig. 14); consequently, there is relatively little rainfall variability. This presents a unique situation in which a relatively small rainfall SD is associated with maximum ensemble mean rainfall when the storm circulations are positioned at a very favorable position relative to Taiwan’s topography.

As an illustration, Figs. 15a and 15b show the west–east and south—north cross sections for member 31 of EN0600 along the lines AB and CD, as shown in Fig. 3. Intensive convection occurs over a very large area along the west–east cross section (see Fig. 15a). There are several low-level westerly jets with speeds up to 36 m s\(^{-1}\) (see Fig. 15b), providing northward and upward moisture transport along the windward slope of CMR (see Fig. 15a). Consequently, deep convection associated with the inner rainband takes place at the northern end of the low-level westerly jet (see Fig. 15b). These are very favorable environmental conditions for orographically enhanced convection and precipitation.

c. 1200–1500 UTC 8 August (model simulation time 60–63 h)

This period has the largest areal-average rainfall SD in the HA (see Fig. 5) after the simulated storm moves farther away from Taiwan. Figure 16 shows the distribution of the simulated 3-h rainfall and the associated SLP at 1500 UTC 8 August. At this stage, the storm’s inner core has moved far enough away from Taiwan that only the outer circulation can impact the precipitation forecast in the HA. Due to the large spread of the storm positions in EN0600 (especially the longitude spread; see Fig. 12), some members are now beginning to be affected by the topography of mainland China, and the rainfall SD in the HA consequently reaches a maximum. On one hand, some members of EN0600, with the outer circulation at the back of the typhoon impinging upon
the CMR, produce large amounts of precipitation. On the other hand, some members have almost no precipitation over the HA; for these members the storm’s circulation is no longer interacting with the CMR. The large storm spread in the east–west direction, amplified by the orographic effects, results in very large rainfall variability. Although the rainfall variability in NTEN0600 over the HA also reaches its maximum at this period due to very large track spread, without the orographic effects, the rainfall discrepancy among the dispersed members is not amplified and thus the rainfall variability in NTEN0600 is much smaller than that in EN0600.

6. Probabilistic rainfall prediction

a. A rainfall probability forecast

An important advantage of ensemble prediction is its ability to provide probabilistic forecasts and estimates of

![Image](https://example.com/image1.png)

**Fig. 14.** As in Fig. 13, but for EN0600 at 21/7–00/8.

![Image](https://example.com/image2.png)

**Fig. 15.** (a) West–east cross section of reflectivity along the line AB in Fig. 3 (color shaded, dBZ), wind bar ($u$, $w$) (m s$^{-1}$, 0.05 m s$^{-1}$) with one full bar of 10 units and meridional wind $v$ (blue line, m s$^{-1}$) for member 31 of EN0600 and (b) south–north cross section of reflectivity along the line CD in Fig. 3 (color shaded, dBZ), wind bar ($v$, $w$) (m s$^{-1}$, 0.05 m s$^{-1}$) with one full bar of 10 units and zonal wind $u$ (blue line, m s$^{-1}$) for member 31 of EN0600. The X axis is the model grid number with grid length of 4 km.
forecast uncertainties. Palmer (2002) gave an example of an extratropical cyclone that devastated parts of Europe on 26 December 1999. Although the single deterministic prediction from ECMWF did not forecast the storm, about a third of the simulations from the ECMWF’s 50-member Ensemble Prediction System produced intense cyclogenesis.

Based on the forecasts from our high-resolution ensemble system, the rainfall probability distribution for different thresholds at different forecast periods is shown.

**FIG. 16.** As in Fig. 13, but for 12/8–15/8.
in Fig. 17. Obviously, for an extreme rainfall event such as Typhoon Morakot, probabilistic forecasts for high thresholds are very important, even if the probability is small. For example, as shown in Fig. 17d, the southern part of the observed main rainfall area with 24-h rainfall ending at 0000 UTC 9 August having over 1000 mm in southern Taiwan is predicted by 30%–60% of the members, and 10% of the members indicate that such extraordinary rainfall could take place over the northern part of HA (where the maximum observed rainfall was recorded). This is important probability forecast information that a single deterministic prediction cannot provide. Initialized with a global ensemble Kalman filter data assimilation scheme based on the National Centers for Environmental Prediction’s (NCEP) Global Forecast System (GFS), Zhang et al. (2010) also showed that cloud-scale ensemble prediction could provide useful probability forecasts of precipitation for Typhoon Morakot.

b. The performance of the high-resolution ensemble simulation

As indicated by Roebber et al. (2004), the accuracy of ensemble probabilistic forecasts can be compromised by model biases. How well the ensemble (probability forecast) performs has a strong dependence on how well the “ensemble mean forecast” performs.

As shown in Figs. 1c and 1e, the two forecasts of the 96-h accumulated rainfall ending at 0000 UTC 10 August,
7. Conclusions and discussions

This paper investigates the impacts of Taiwan’s topography on the record-breaking rainfall of Typhoon Morakot and its predictability through a high-resolution ensemble simulation. Analysis of results from ensemble experiments with and without Taiwan’s topography leads to the following conclusions.

1) Taiwan’s topography plays a key role in making Typhoon Morakot a record-breaking rainfall event. With realistic topography, the high-resolution (4 km) ensemble with the WRF-ARW model reproduced salient features of orographic precipitation. The 96-h total rainfall amount and distribution from the ensemble mean compares reasonably well with the observed precipitation. When the terrain of Taiwan is removed, the orographic rainfall distribution features completely disappear and the maximum 96-h accumulated rainfall amount is reduced to less than 20% and the total rainfall amount in the HA is reduced to less than 60%. The orographically enhanced rainfall in the HA reaches 25 and 382 mm for the ensemble mean 3- and 96-h accumulated rainfall, respectively.

2) Rainfall variability is substantially enhanced by Taiwan’s topography. The standard deviation of the areal-averaged rainfall in the HA among members of the ensemble system is increased by up to 15 and 75 mm for 3- and 96-h accumulated rainfall, respectively, as a result of topography. For a single-gridpoint rainfall prediction, the variability can be more than doubled, with a value less than 150 mm for no terrain experiment and more than 400 mm with terrain. The spatial distribution of rainfall variability is strongly modulated by the underlying topography.

3) The relationship between the ensemble mean precipitation and rainfall variability is complex and time varying. There is no simple linear relationship between the amount of the ensemble mean precipitation and the rainfall variability. There exists a unique period when the ensemble mean precipitation reaches its maximum, while the rainfall variability has a local minimum. This is the time when the storm is located over the west coast of Taiwan, and shortly after it left Taiwan, a period when the typhoon circulation patterns are positioned most favorably relative to Taiwan’s topography for producing topographic precipitation.

4) It is not possible to use the rainfall forecast standard deviation to estimate model rainfall forecast error at high temporal and spatial resolutions. Although ensemble spread has been considered as a useful variable when measuring flow-dependent model forecast uncertainties for large-scale variables, this does not apply for rainfall prediction. The high temporal and spatial variabilities of rainfall, coupled with the enhanced variability due to topography, prevents the use of rainfall ensemble spread as a useful measurement of model rainfall forecast errors. This is especially true when the model exhibits a significant systematic bias in certain specific regions. We do note that for integrated quantities, such as 96-h accumulated rainfall, this relationship becomes more robust. For example, the correlation coefficient for rainfall standard deviation and rainfall RMSE is 0.63.

5) The orographic effects on typhoon circulations amplify uncertainties in the initial model states and substantially increase variability in storm track, positions, storm propagation speed, landfall location and timing, and the detailed pressure structures (e.g., secondary lows) among the ensemble members. The variability in these mesoscale storm features, in return, results in significant variability in rainfall. The topographically enhanced variability is particularly robust near the time of landfall. This makes it very challenging to prepare an accurate quantitative precipitation forecast at a specific location for a landfalling typhoon over Taiwan.

6) Probabilistic rainfall prediction derived from the high-resolution ensemble system does show some skill in estimating where significant rainfall may take place. For example, the 90% probabilities of accumulated rainfall exceeding 1000 mm between 0000 UTC 7 August and 0000 UTC 9 August match well with the distribution of the observed 1000 mm. In comparison with one single deterministic forecast, the ensemble mean is shown to possess superior skill in forecasting precipitation.
Careful evaluation of the ensemble mean prediction with observations also points out some notable deficiencies. These include a westward and southward track bias after landfall, significant underprediction of the rainfall over Chiayi County (represented by station A) and significant overprediction of rainfall at the southern tip of the CMR (represented by station B). The systematic track bias is most likely associated with the use of the BMJ cumulus parameterization scheme in the simulations. We have performed additional sets of ensemble experiments with limited ensemble numbers (eight) to assess the impacts of cumulus parameterization schemes in the three nested domains (36, 12, and 4 km). We found that different cumulus parameterization schemes together with the explicit microphysics would produce storms with different sizes, structures, and intensities. The topographic effects would also vary, and they would subsequently produce different storm tracks. The underprediction of rainfall at station A is most likely related to the systematic track bias after landfall, which may be related to the cumulus parameterization used in the model simulation. With the use of BMJ convective parameterization in the three domains the storms tend to move faster and have a westward and southward bias. The systematic storm track bias would alter the impinging angles of the flow with the mountain. The faster movement also reduces the accumulation of rainfall. However, the reason for the overprediction at station B is still not completely clear. We found that ensemble forecasts with different combinations of moist physics would all have similar overpredictions at station B. Thus, it might be related to the representation of a sharp terrain variation in the model. In particular, the actual terrain is high and narrow, and may not be well represented in the model. To reduce the computational cost for the 32-member ensemble nested down to the 4-km grid, we use the WSM5 microphysics scheme in this study. The deficiency of not including the graupel particles in the WSM5 microphysics scheme might also contribute to the overestimated peak rainfall. The impacts of detailed microphysical processes and their interaction with cumulus parameterization on precipitation forecasts in a typhoon environment with sharp terrain variations are an important topic for future study. Some would also argue that there are not sufficient rain gauge stations over this part of the mountain, and, consequently, the rainfall observations may be “underestimated.” Further investigation is needed to better understand this rainfall overprediction problem. The analysis of these additional results will be presented in a subsequent paper.

As we know, the best method for generating initial perturbations and setting model physics packages for ensemble forecasts is still open to debate. Toth and Vannitsem (2002) indicated that, unless model related uncertainty is properly simulated, ensembles will not be able to capture variations in forecast uncertainty arising due to model imperfections. Similarly, such ensembles will be deficient in the sense that in cases where model error is present they will not always be able to correctly identify the variations. This is reflected, for example, by a difference between the time evolution of the error in the ensemble mean forecast and the spread of ensemble members around this mean. In an ideal ensemble, these two quantities should statistically be identical. In this study, the reason for the poor performance of rainfall spread at some specific locations and times is complicated. It may be caused by three factors: 1) the skewed nature of rainfall distribution, 2) the ensemble generation method, and 3) model errors. These issues need to be examined carefully in future studies.

Ensemble prediction at cloud-resolving resolution is in its early stage of development. There are many challenging issues related to ensemble generation, model uncertainties, statistical postprocessing and calibration, and ensemble verification. The work presented in this paper is very preliminary in nature. The main purpose of this paper is to illustrate the challenges of predicting an extreme heavy rainfall event associated with Typhoon Morakot. There are considerable deficiencies in several aspects of this study, including the ensemble generation, the lateral boundary conditions, and our handling of the model errors. Despite these deficiencies, this paper represents the first study on the impacts of topography in a cloud-scale ensemble for a heavy rainfall event associated with the interaction of a typhoon and a mesoscale mountain. The results are useful for the design of future cloud-ensemble prediction systems for torrential rainfall associated with the landfall of typhoons. The salient impacts of topography are captured, and the main conclusions should be valid. A considerable amount of additional work is needed to design and develop an operational ensemble forecast system at cloud-resolving resolution. This paper represents a first step in this direction.

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REFERENCES


