The Precipitation Hotspots of Afternoon Thunderstorms over the Taipei Basin: Idealized Numerical Simulations

Kuan-Ting KUO and Chien-Ming WU

Department of Atmospheric Sciences, National Taiwan University, Taiwan

(Manuscript received 21 August 2018, in final form 16 January 2019)

Abstract

In this study, the mechanism for precipitation hotspots (PHs) of locally developed afternoon thunderstorms in the Taipei Basin is investigated using a three-dimensional Vector Vorticity equation cloud-resolving Model (VVM) with an idealized topography and surface properties. A 500 m horizontal grid resolution is used in all experiments. The results show that the local circulation is a key for PHs at the south of the Taipei Basin. The two valleys guide background southwesterly (SW) flow along with the sea breezes to penetrate into the basin. The urban heat island (UHI) effect enhances the sea breeze convergence at the south of the basin and produces strong convection there. The interactions between cold pools generated by the convection and the sea breezes produce northward propagating new convective cells. Besides, the background wind direction is important in determining the location of sea breeze convergence. If the background wind direction changes from westerly (W) to west-northwesterly (WNW), there might be no precipitation at all in the basin. This study suggests that the idealized experiments also provide a useful framework for studying the impacts of future climate changes on the PHs in the Taipei Basin by applying the pseudo–global warming approach.

Keywords idealized simulations; deep convection; thunderstorms; sea breezes


1. Introduction

The Taipei Basin is located at northern Taiwan and is surrounded by the Tatun Volcano Group to the north, the Linkou Tableland to the west, and the Snow Mountain Range to the southeast (Fig. 1). Two river valleys that serve as channels connecting the Taipei Basin and its surrounding ocean are located at the northwest and northeast of the Taipei Basin, called Tamsui River and Keelung River valleys. Also, the Taipei Basin is covered mostly by the metropolitan area providing a strong urban heat island (UHI) effect there. With these geographical features, Taipei city, among other metropolitans, is a unique location to study the city scale precipitation hotspots (PHs) caused by the afternoon thunderstorms (TAS) in the summertime through the interactions of background wind fields, topographic effects, and sea breezes. In this study, PHs are defined as regions with heavy rainfall on an hourly time scale, especially for the urban area, which is basically the same as the basin area in Taipei.

Observational studies (Chen et al. 2007, 2014; Lin et al. 2011, 2012; Yen 2010) suggested that TAS are the major source of summertime precipitation over the Taipei Basin. Chen et al. (2007) and Lin et al. (2011) suggested that the cool and moist air from the ocean can be transported into the Taipei Basin by sea breezes through the valleys, and the convergence of the sea
breezes further moves southeastward and produces heavy precipitation near the mountains. They also showed that the increase of summer mean rainfall in the past four decades is attributed to the increase of precipitation caused by TS. This increasing trend implies that Taipei city could be more vulnerable to flash flood events due to lack of in-time warning.

Lin et al. (2011, 2012) studied the preferred environmental conditions for TS in warm seasons during 2005–2008 under weak synoptic conditions. They used the morning soundings (08:00 local standard time (LST); 00:00 UTC) at Banqiao weather station (triangle mark in Fig. 1) and surface data from weather stations around northern Taiwan to analyze the differences between the TS days and the non-TS days. Their results showed that the sounding profiles during the TS days tend to be warmer and wetter in the morning, and the background wind has the highest probability of a weak southwesterly (SW) flow. On the other hand, in the non-TS days, the northeasterly sea breeze from the Keelung River valley dominates the wind field in the Taipei Basin, providing no sea breeze convergence in the basin with no dominant background wind directions. The temporal/spatial evolution of TS was further analyzed by Chen et al. (2009) and Chen et al. (2014) using radar echoes and weather station data. Their results suggested that the life cycles of TS and land–sea breezes are tightly coupled. In the initial stage, the sea breezes from Tamsui and Keelung River valleys crossing to the southern Taipei Basin trigger convection at the southern slope. Later on, a cold pool front forms at the north of a heavy precipitation area caused by TS. When the convection is active, the radar shows that the convective systems align along a northeast–southwest axis in the Taipei Basin. In this study, the importance of the topography, in particular, the wind channeling through the valleys, the UHI effect, and the background wind directions on the evolution of TS is examined. Therefore, simulations with a cloud-resolving model (VVM) are performed. Idealized topography and surface fluxes are used to analyze the impact on TS evolution as well as different background wind directions. Section 2 presents the model description and the idealized experiment setup. The analysis of the temporal/spatial evolution of the PHs is presented in Section 3. Summary and discussion are presented in Section 4.

2. Model and experiment setup

2.1 The model

The Vector Vorticity equation cloud-resolving Model (VVM) used in this study was developed by Jung and Arakawa (2008) based on the three-dimensional anelastic vorticity equation. A unique aspect of the model is that the model predicts the horizontal components of vorticity and diagnoses the vertical velocity using a three-dimensional elliptic equation. Therefore, the pressure gradient force is eliminated in this framework, and the horizontal vorticities can respond directly to the buoyancy force to better capture the local-scale circulations associated with the strong heating difference, such as the sea breeze circulation and the cold pool fronts. Although predicting vorticity is mathematically equivalent to predicting momentum in solving fluid equations, the latter could be hindered by numerical problems such as solving Poisson’s equation for pressure or filtering out sound waves. The implementation of topography is performed through the immersed boundary method with the topography forcing added to the vorticity fields (Wu and Arakawa 2011; Chien and Wu 2016). A bulk three-phase cloud microphysics parameterization including cloud droplets, ice crystals, rain, snow, and graupel is used in this model (Krueger et al. 1995). This model has also been used to study the unified parameterization of deep convection (Arakawa and Wu 2013; Wu and Arakawa 2014) and the impacts of heterogeneous land surface fluxes on the diurnal cycle precipitation (Wu et al. 2015), as well as the environment of aggregated convection (Tsai and Wu 2017).

2.2 Experiment design

a. The general settings

To resolve this topography appropriately, especially the valleys at the northwestern and northeastern side of the basin, a 500 m grid resolution in the x- and y-directions was used in all experiments. A large 256 × 256 km horizontal domain is used to avoid the effects from the boundary due to the use of doubly periodic boundary conditions, and a 512 × 512 km domain is used for a real orography experiment. The vertical grids are fixed to 100 m under 1 km and stretched to 1.3 km near the model top (17 km). Random temperature perturbation is added for each simulation to every single grid point in the whole domain below 1 km during the first 120 time steps (4 minutes) with a random value between −0.5 K and +0.5 K to break up the model symmetry, and the first two hours are treated as model spin-up time. Coriolis force is not considered in the simulations for simplicity owing to the use of a small domain and short time integration. The detailed model settings are listed in Table 1.
b. The idealized topography

In order to identify the essential mechanisms responsible for PHs caused by TSAs, a simple topographic framework is presented in Fig. 2. For a comparison with the real topography of the Taipei Basin, all figures on an x-y plane in this paper are rotated 45 degrees so that the north will be pointing up in these figures. In this framework, we use a simple square area surrounded by mountains to represent the basin; this area will be set to an urban area for the experiments with urban-type heat fluxes (red area in Fig. 2). The gray colors from dark to light represent the heights of mountain areas from low to high. The mountain shape is idealized with a Gaussian function:

\[
\text{Height} = H_{\text{max}} e^{-\frac{D^2}{2\sigma^2}},
\]

where the maximum mountain height \(H_{\text{max}}\) is 1000 m and \(D\) is the distance between a grid point and the central line of mountains. \(\sigma\) is about 1316 m, such that the half width of the mountain bottom is 3 km in the model. The length of each side of the mountains is 30 km (from -15 to 15 km in Fig. 2). The valleys can be idealized by gaps of the mountains at the corresponding locations in the northeast and northwest. Despite the low elevation at the southwest side of the

![Fig. 1. The topography of Taiwan (left panel; satellite photograph from Google Earth) where the red box shows the location of the Taipei Basin (right panel). The color shading represents the orographic height over the Taipei Basin.](image-url)
Taipei Basin (the right panel in Fig. 1), the idealized mountain on the lower left side exists to avoid the SW wind from blowing into the basin area directly. This idealization is to mimic the effect of real topography such that the weak SW wind tends to flow around the Taiwanese topography. We will focus on the precipitation patterns in the basin throughout the study. Sensitivity tests of topography shapes are provided in Appendix A. The results suggest that the current topography setting best captures the Taipei Basin’s PHs compared with observation.

c. The sounding

To find out the essential factors affecting the development of TS$_A$ in the Taipei Basin, we use summertime composite sounding in all experiments in this study (Fig. 3). This sounding is composed of TS$_A$ events in 1999—2006 summer at 00:00 UTC (08:00 LST) at Banqiao weather station (Fig. 1) by Chen et al. (2009) representing the morning environment for TS$_A$. They firstly chose weak synoptic days, which are without influences of typhoons, fronts, and other low-pressure systems. Then, the TS$_A$ grid was identified when the area of high radar reflectivity (>15 dBZ) was larger than one-fifth in a 0.1-degree grid over northern Taiwan, and the maximum reflectivity reached 30 dBZ within that grid. When there are at least seven TS$_A$ grids during 11:00—20:00 LST, the day is defined as a TS$_A$ day. There are 110 TS$_A$ days in the composite sounding. The initial wind profile is adapted from Lin et al. (2011, 2012). They showed that the occurrence of TS$_A$ in northern Taiwan is related to the wind speed and the wind direction between 0—3 km and 3—6 km. Here, we follow their composite wind profiles and make idealization by setting the wind fields to 3 m s$^{-1}$ SW under 6 km and zero above in all experiments. Although this background wind is not the same as that described by Lin et al. (2011, 2012), this simplified setting can be closer to near-surface condition and focuses more on the effects of topography and wind direction on TS$_A$ over northern Taiwan.
d. The surface heat fluxes

The primary focus of this study is to understand the mechanisms of determining the PHs caused by the interactions among sea breezes through channels, convective systems, and the UHI effect. The radiative processes are idealized through surface fluxes to avoid complications from the land initialization problems. Therefore, all the surface heating processes can be easily represented by surface heat fluxes. This method is widely used by studies of local circulation simulations (see, e.g., Crosman and Horel 2010). Following Kirshbaum (2011), the land surface properties in this study are decided only by modifying the Bowen ratio of the prescribed surface heat fluxes, where the Bowen ratio is the sensible heat flux divided by the latent heat flux. The whole areas are divided into rural type, urban type, and ocean. Note that the urban type has a lower latent heat flux than that of the rural type, especially at noon. Moreover, to give a sense of the diurnal cycle of the surface heat flux in the real world, the experiment begins at 06:00 LST. The above idealizations might not represent the real urbanized effects, but they roughly capture the land–sea temperature contrast from observation (e.g., the mean highest temperature in the model is 32°C, whereas that at Taipei station was 34°C in the summer of 2018; the ocean’s temperature was around 28°C in both the model and the observation). The detailed settings are presented in Appendix B.

We present three sets of experiments to investigate the effects of valleys and the UHI and six sets for the effects of wind directions. Each experiment has six ensemble members to account for the convective variability in producing PHs. A random temperature perturbation is added for each ensemble member with different random seeds. We use the first three sets of experiments to investigate the impact of topography shapes and UHI effects on the PHs. The control experiment (CTL) represents the topography with two valleys and urban-type surface heat flux in the basin area (Fig. 2). In experiment with no UHI effects (noUH), the surface heat flux is of a rural type in the basin area.

Moreover, gaps are replaced with Gaussian-shaped mountains in experiment with no valleys (noVL) to examine the effects of the valleys. To investigate the influence of large-scale wind direction on thunderstorm initiation, we performed five additional experiments. The topography with two valleys and urban-type surface heat fluxes was used, and large-scale wind direction was varied by 22.5° from the south-southwesterly (SSW) (202.5°) to the northwesterly (NW) (315°) direction. The experiment names corresponding to the settings are shown in Table 2.

3. Results of idealized experiments

3.1 Realistic topography simulation

To evaluate the performance of simulating $T_{S_A}$ precipitation in the Taipei Basin, we perform a VVM simulation covering the whole Taiwanese area. This includes the use of homogeneous morning soundings responsible for $T_{S_A}$ development (Fig. 3) and uniform weak SW winds. The surface fluxes are as described in Section 2.2.4d. The setup of this experiment is identical to those introduced in the previous section except for using the real topography of Taiwan and a larger domain size ($512 \times 512$ km). Snapshots of near-surface wind at noon and 12-hour accumulated precipitation taken from the simulation are presented in Fig. 4. It is clear that the sea breezes converge through the valleys and heavy precipitation at the south of the Taipei Basin. The results correspond well with the high $T_{S_A}$ frequency area analyzed by Lin et al. (2012). This area is marked as a white circle in the right panel of Fig. 4, demonstrating the model’s capability in capturing the primary processes for $T_{S_A}$. To qualitatively

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**Table 2. Experiments settings.**

<table>
<thead>
<tr>
<th>Exp. Name</th>
<th>Land-use type in the basin area</th>
<th>Gaps at the NW and NE mountain areas</th>
<th>Background wind speed / direction</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTL</td>
<td>Urban</td>
<td>Yes</td>
<td>3 m s$^{-1}$/SW</td>
</tr>
<tr>
<td>noUH</td>
<td>Rural</td>
<td>Yes</td>
<td>3 m s$^{-1}$/SW</td>
</tr>
<tr>
<td>noVL</td>
<td>Urban</td>
<td>No</td>
<td>3 m s$^{-1}$/SW</td>
</tr>
<tr>
<td>SSW</td>
<td>Urban</td>
<td>Yes</td>
<td>3 m s$^{-1}$/SSW</td>
</tr>
<tr>
<td>SW</td>
<td>Urban</td>
<td>Yes</td>
<td>3 m s$^{-1}$/SW</td>
</tr>
<tr>
<td>WSW</td>
<td>Urban</td>
<td>Yes</td>
<td>3 m s$^{-1}$/WSW</td>
</tr>
<tr>
<td>W</td>
<td>Urban</td>
<td>Yes</td>
<td>3 m s$^{-1}$/W</td>
</tr>
<tr>
<td>WNW</td>
<td>Urban</td>
<td>Yes</td>
<td>3 m s$^{-1}$/WNW</td>
</tr>
<tr>
<td>NW</td>
<td>Urban</td>
<td>Yes</td>
<td>3 m s$^{-1}$/NW</td>
</tr>
</tbody>
</table>
understand the evolution of TS$_A$ and the impact of the complex topography and the UHI effect on the PHs, results from the idealized experiments are presented in the next subsection.

**3.2 Temporal evolution**

The impacts of the UHI effect and topography can be first examined through analyzing the time evolution of precipitation patterns in the individual experiment. The time evolution of low-level winds and precipitation for a selected member in CTL is presented in Fig. 5. In CTL, the experiment with valleys and the UHI effect, sea breezes build up because of the strong heating difference near the coast and blow onshore into the basin through the two valleys at 11:00. The northern mountain surrounding the basin splits the SW flow producing the air flow through the northwest valley. In the northeast valley, on the other hand, due to the blocking of the mountains surrounding the basin, the SW creates a weak flow around the northern side of the valley. Combined with the sea breezes from the two valleys, a convergence area develops at the center of the basin and moves southward to the southern corner of the mountain areas. At 12:00, the southward confluent flow gathers at the southern corner of the basin and triggers deep convection there. The southern branch of the mountain serves as a barrier to block the SW from directly interacting with the sea breezes in the basin. Furthermore, the convection produces strong precipitation at the southern corner, and the cold pool associated with downdraft produces a strong outflow in the low levels at 12:00 in Fig. 5. The feature can be better visualized by the vertical cross section of the convection in the north–south direction (Fig. 6). The cold pool region is represented by the negative horizontal anomaly of the virtual potential temperature near the surface. This can be seen from −12 to −5 km. Strong cold pool outflow converges with the sea breezes confluent flow at −4 km inducing a strong updraft at the north side of the convection, causing subsequent convective systems moving northward. This propagation can be seen from 12:00 to 13:00 at the center of the basin (Fig. 5). At 14:00, the convective system moves to the northeastern side of the mountains as a result of SW background wind at the upper levels. During the rest of the simulation, precipitation over the basin is generally caused by the movement of the convective systems over the mountain area.

The development of convective systems in experiment noUH is presented in Fig. 7. The results show that, without the urban heating, there is no convection being triggered at the southern corner of the basin at 12:00 compared to the result of CTL. The wind speed of sea breezes averaged over the two valleys decreases from 6.55 m s$^{-1}$ to 5.23 m s$^{-1}$ at 11:00. The weaker sea breeze convergence from the valleys delays the development of convection in the center of the basin.
Fig. 5. Snapshots of CTL_1 (the first ensemble member of CTL). Time is labeled at the upper-right corner. The color shading is the precipitation rate with intervals labeled in the color bar, and the areas with transparent white represent the thicknesses of clouds. The white arrows are averaged low-level winds from the surface to 500 m. The white contour indicates the mountain areas above sea level in Fig. 2.
until 14:00. In general, the surface precipitation in the basin is weaker compared to CTL.

To understand the effects of the valleys on basin precipitation, we present the temporal evolution of experiment noVL in Fig. 8. The results show that, without the valleys, there is no convective system developed in the basin area. On the other hand, most precipitation is located at mountain areas, especially at the northeastern side. The convection is triggered by the convergence of background flow in the lee of the mountains (northeast) and upslope winds driven by the heating mountain areas. Previous studies (Baik et al. 2007; Huff and Changnon 1973; Shem and Shepherd 2009; Shepherd 2005) have discussed the role of UHI effect in increasing convective instability over the downwind side of the background wind, which is the northeastern area in this experiment. In the basin area, the results of noVL demonstrate that, without the convergence of the sea breezes, it is hard to produce convective systems inside the basin.

### 3.3 Ensemble results

To examine the statistical characteristics of PHs, we perform six ensemble members for each experiment with different random seeds. The ensemble-averaged accumulated precipitation is presented in Fig. 9. Here, we define the areas with ensemble-averaged accumulated precipitation larger than 15 mm as PHs. The results showed that the PHs in the basin area only exist in CTL along the north–south axis in the center. In noUH, on the other hand, the precipitation is a lot weaker, and the PHs in the south of the basin are missing. In noVL, there is only weak precipitation in the basin, suggesting the importance of the valleys in controlling the PHs through guiding sea breezes into the basin and creating a convergence zone at the center of the basin area. The movement of the convective systems over the mountain is generally random, hence no PHs over the basin (Fig. 9c). The results emphasize the importance of enhanced sea breeze convergences from the two valleys due to the UHI effect in producing PHs at the south of the Taipei Basin.

The causes of the precipitation pattern difference between CTL and noUH can be evaluated in Fig. 10. The sea breezes in CTL penetrate deeper into the basin than those in noUH at 11:00 owing to the stronger inflow speeds. The ensemble-averaged wind speed of near-surface inflow within the northwest valley is

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Fig. 6. Cross section of CTL_1 at 12:00. The vertical slice is taken from the red dashed line shown in the upper-right figure. The color shading represents the horizontal anomaly of the virtual potential temperature, \( \theta_v = \theta(1 + 0.61q) \), where \( \theta \) is the potential temperature and \( q \) is the specific humidity. The gray dots (hatchings) represent cloud (rain) water higher than 0.01 (0.1) g kg\(^{-1}\). The purple contour represents a region where the updraft velocity is higher than 3 m s\(^{-1}\). The arrows only show wind speeds higher than 3 m s\(^{-1}\), and the bottom-right arrow scale presents the arrow length of 20 m s\(^{-1}\). The gray shading is the mountain area.
Fig. 7. As in Fig. 5 but for snapshots of noUH_1.
Fig. 8. As in Fig. 5 but for snapshots of noVL_1.
Fig. 9. Ensemble averages of accumulated surface precipitation in (a) CTL, (b) noUH, and (c) noVL from 6:00 to 18:00. The shaded area represents precipitation with intervals labeled in the color bar. The unit is millimeters. The arrows represent near-surface wind in the morning (10:30). Regions of mountain areas are circled with black lines and filled with transparent gray.

Fig. 10. Cross sections of ensemble-averaged horizontal wind speed and near-surface temperature of (a) CTL and (b) noUH at 11:00. The vertical slices are taken from the red line shown in the upper-right figures. The shaded area represents the wind speed with intervals labeled in the color bar. The lower figures represent the near-surface air temperature. The blue dashed lines show the position of the coastline.
7.20 m s^{-1} in CTL and 5.78 m s^{-1} in noUH. The stronger inflow in CTL is caused by the larger temperature gradient along the sea breeze front. The ensemble-averaged surface temperature over the basin area in CTL is 0.84 K higher than that in noUH at 11:00, whereas it is less than 0.1 K over the ocean area. The stronger sea breezes cause a stronger southward confluent flow that geographically locks the PHs at the south of the basin.

To further understand the PHs at the north–south axis of the basin, the temporal evolution of the ensemble-averaged surface precipitation rates in the basin area (red area in Fig. 2) is presented in Fig. 11. It is clear that CTL begins to rain earlier in the basin and has the most significant precipitation rate among these three experiments. Comparing CTL with noVL, the result implies that the early deep convection in CTL is mechanically lifted by the sea breeze fronts rather than the boundary layer turbulence in the favorable environment for TSa.

Besides, the diurnal evolution of the precipitation rate in CTL presents apparent double peaks at 13:30 and 15:30 compared with noUH and noVL in which a single peak occurs around 14:00. To understand how the two peaks vary in the ensemble members, we mark the precipitation peaks in each ensemble member in CTL in Fig. 11. The results show that the first peak in each ensemble member occurs roughly within 30 min, and the difference of the precipitation rate among ensemble members is within 1 mm h^{-1}. On the other hand, the second precipitation peak shows larger spread with a magnitude of over 2.5 mm h^{-1}. This result suggests that the precipitation events that occur in the first precipitation peak are dominated by the sea breeze convergence, as shown in Fig. 5, during 11:00 to 13:00. The random perturbation of temperature has a little effect on the timing and strength of the PHs. On the other hand, the second peak could be contributed by a collection of random events of moving convective systems, which may be triggered over mountains or by cold pools in the basin. In noUH, the formation process of the only one peak is similar to that of the first peak in CTL (Figs. 5, 7), but the initial time is later, so the random convections are developed at the same time. Therefore, the peak in noUH is more like a weaker combined result of the two peaks in CTL.

The spatial pattern of the temporal evolution of PHs is presented in Fig. 12. This figure shows the temporal evolution of the ensemble-averaged precipitation rate of the CTL runs higher than 15 mm h^{-1}. Colored areas represent the same spatial and temporal features of precipitation in all six members. The time evolution of PHs can be visualized by the changes of warm to cold colors. The ensemble-averaged PHs move...
northward from 11:00 to 14:30. This result further implies that the PHs are directly on the north–south axis at the center of the basin. This moving feature of PHs corresponds well to an observation case study by Chen et al. (2014). The results emphasize that sea breezes are key for the position of PHs at the south of the Taipei Basin. The two valleys guide background SW flow along with the sea breezes to penetrate into the basin. The sea breeze convergence enhanced by the UHI effect produces PHs at the south of the basin.

3.4 Impacts of wind directions

Figure 13 presents the sensitivity experiments on wind directions to evaluate the robustness of the PHs. The results show that when the wind direction changes gradually from SSW to westerly (W) clockwise, the PHs move rapidly but still locate along the north–south axis. When the wind directions change to westnorthwesterly (WNW) and NW, the PHs disappear in the basin.

One possible mechanism for the disappearing PHs is that when the low-level background wind directly blows into the basin through one of the valleys, the convergence will not be at the center of the basin. Therefore, there will be no PHs in the basin area. This mechanism can be examined by the low-level wind fields before convection is triggered in the basin area. Figure 14 shows the wind rose diagrams of near-surface wind fields including all grid points in the basin area from 10:00 to 11:00. This period is selected after the onset of the sea breeze, but before the precipitation initiation in the basin area. The wind fields in the basin are between W and east-northeasterly owing to the existence of valleys, but with different frequency ratios, which is defined as the balance of sea breezes from the two valleys. If the wind frequen-

![Fig. 13. Ensemble-averaged accumulated precipitation patterns where the background wind directions are (a) SSW, (b) SW, (c) WSW, (d) W, (e) WNW, and (f) NW. The shaded area represents precipitation with intervals labeled in the color bar, which is the same as in Fig. 9. Regions of mountain areas are filled with transparent gray.](image-url)
cy from the two valleys is the same, the ratio will be close to 1. If the wind blows more frequently from one valley than from the other, the ratio will be larger than 1 (with the dominant frequency in the numerator). The frequency ratios of near-NW winds (281.25°–348.75°) are higher than those of near-NE winds (11.25°–78.75°) in experiment W (5.8 times), experiment WNW (4.3 times), and experiment NW (3.7 times). These three experiments in which near-NW winds dominate wind fields also have weaker precipitation in the basin, which demonstrates the importance of the sea breezes from both valleys in producing PHs in the basin. If the sea breezes cannot converge at the center of the basin, the PHs will tend to locate at the downwind side of mountain regions.

Another possible mechanism is that even if cold pools by convective systems at mountain areas triggered new convection at the margin of the basin, the core of the convection will be advected away from the basin by the background wind. For example, the frequency ratios of near-NW winds and near-NE winds are similar between experiment SSW (2.1) and experiment WSW (2.4), but experiment SSW has 1.6-fold ensemble-averaged precipitation compared with experiment WSW. Their precipitation in the basin before 14:30 is similar (Table 3), suggesting that both experiments can trigger convection in the basin first, but the movement of convective systems in the later period is different. The sensitivity experiments correspond well with the observation (Lin et al. 2012) in that the background wind direction highly links to TSA in the basin.

Fig. 14. Wind rose diagrams including all ensemble members in each experiment of near-surface wind in the whole basin area from 10:00 to 11:00. The background wind directions are (a) SSW, (b) SW, (c) WSW, (d) W, (e) WNW, and (f) NW. The colors of fan-shaped areas present the frequency of different ranges of wind speeds in each wind direction with intervals labeled in the color bar. The percentages of frequency at the circles are from 6% to 30% with the interval of 6%.
4. Summary and discussion

In this study, a three-dimensional VVM with a 500 m horizontal resolution was used to investigate the PHs of afternoon thunderstorms (TS_A) in the Taipei Basin. Interactions among sea breezes, complex topography, and convective systems make it difficult to understand the fundamental mechanisms of controlling PHs. In this study, an idealized topography and land use settings are used to clarify its essential mechanisms.

The results demonstrate the importance of enhanced sea breeze convergences from the Keelung River valley and the Tamsui River valley on geographically locking TS_A in the south of the Taipei Basin. The results also suggest that if the two valleys do not exist, there will be no PHs in the basin. Thus, without the sea breeze convergence through the valleys, precipitation in the basin is mainly produced by the random convective systems from the mountains moving into the basin. The UHI effect causes enhanced sea breezes through the enhanced thermal contrasts between Taipei city and its surrounding ocean. The PHs then move northward from 11:00 to 14:30 to the north of the basin due to new convective systems initiated by cold pool fronts produced by heavy precipitation. The essential mechanisms for the PHs can be illustrated with a schematic diagram in Fig. 15, emphasizing the importance of topography, sea breezes, confluent flow, background wind direction, the evolution of cold pool fronts, and associated location of the PHs. Sea breezes and their confluent flow play essential roles in the initiation process, whereas the cold pool fronts provide another mechanism for the propagation of PHs.

Sensitivity experiments on the wind directions show that the PHs are generally located in the north–south axis when the wind direction changes clockwise from SSW to W. On the other hand, if the background wind blows into the basin through one valley directly, it will act as a blocking effect on the sea breeze from the other valley. Therefore, the sea breeze convergence in the basin dictates the surface PHs in the basin.

The idealized experiments proposed in this study can be used as a framework for studying the variabilities of TS_A on the PHs under future climate changes.

Table 3. Ensemble-averaged accumulated precipitation over the basin area with different background wind directions.

<table>
<thead>
<tr>
<th>Wind Direction</th>
<th>SSW</th>
<th>SW</th>
<th>WSW</th>
<th>W</th>
<th>WNW</th>
<th>NW</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basin-Averaged Accumulated Precipitation (mm)</td>
<td>7.45</td>
<td>8.99</td>
<td>4.58</td>
<td>3.00</td>
<td>0.49</td>
<td>1.04</td>
</tr>
<tr>
<td>Basin-Averaged Accumulated Precipitation (mm) Before 14:30</td>
<td>3.78</td>
<td>5.24</td>
<td>3.04</td>
<td>1.98</td>
<td>0.40</td>
<td>0.51</td>
</tr>
</tbody>
</table>

Fig. 15. Three-dimensional schematic diagram of the (a) initiation process and (b) propagating process for PHs in CTL. The base maps are (a) the ensemble-averaged near-surface temperature with intervals labeled in the color bar and wind fields at 11:00 and (b) ensemble time evolution of PHs, which is the same as in Fig. 12. The purple arrows represent the background wind. (a) The blue arrows represent sea breezes, and the yellow arrow represents confluent flows caused by sea breezes from the valleys. The cloud indicates the location triggering convection. (b) The blue and light-blue fronts represent cold pool fronts caused by the PHs at 12:00 and 13:00. The positions of the fronts are judged by the convergence of wind fields under 500 m in the basin area.

Instead of the traditional regional downscaling method, a pseudo–global warming approach, in which the thermodynamics changes estimated from the future projections using the general circulation models, can be translated into atmospheric stability and moisture
changes. Applying these changes to our idealized setup, the variability of PHs caused by global warming can be estimated.

Finally, it should be pointed out that the interpretation of our results is based on an idealized topography and simplified land surface properties, and the model uses an idealized microphysics parameterization. Also, the thermodynamic environment for triggering deep convection might be another important issue. As discussed by Wu et al. (2009), the initiation of the diurnal cycle deep convection is controlled by the development of shallow convection. The pattern of the surface heterogeneity within the city can also generate aggregated convective systems produced by the convergence of inland breezes (Wu et al. 2015) and can modify the location of PHs. We believe that this study demonstrates the essential mechanisms for PHs in the Taipei Basin and provides useful guidance for extreme precipitation warning when the conditions are met.

Appendix A:
Sensitivity tests of the idealized topography

To better understand the characteristics of the idealized topography used in this study, we separate mountain areas to an upper half and a lower half, conducting additional experiments with the same settings mentioned in Section 2. Ensemble-averaged precipitation patterns and wind fields at 10:30 for two additional topography shape experiments are provided in this appendix. The results show that, without the top panel of the topography, as shown in Fig. A1, the SW cannot penetrate directly into the basin. It mainly converges with the sea breeze at the east side of the original basin area because the background wind flows around the southwestern mountain and speeds up the W sea breeze. This result suggests that the top panel of the topography is essential for splitting the background flow, creating equally strong sea breezes and making their confluent flow penetrate into the basin. The results also show that, without the bottom panel of the topography, as shown in Fig. A2, the background wind field can blow directly into the basin, causing convergence in the northeast corner of the basin and heavy precipitation there. Both scenarios confirm the importance of the topography surrounding the basin in producing PHs close to observation.

Appendix B:
Detailed settings of the surface heat fluxes

The sensible and the latent heat fluxes are determined from the Bowen ratio of the total surface heat fluxes as follows:

\[ \text{SH} = F_{\text{tot}} \times \frac{\text{BR}}{1 + \text{BR}}, \quad \text{LE} = F_{\text{tot}} \times \frac{1}{1 + \text{BR}} \]

\[ F_{\text{tot}} \text{ [W m}^{-2}\text{]} \]

\[ = \begin{cases} 
500 \sin(\Omega r), & \text{for the rural and urban areas} \\
318, & \text{for the ocean} 
\end{cases} \]

and
where $F_{\text{tot}}$ is the total surface heat fluxes, $BR$ is the Bowen ratio, $SH$ is the sensible heat fluxes, $LE$ is the latent heat fluxes, $\Omega$ is the angular speed of the earth, and $t$ is the time from 6:00 to 18:00. This period may not indicate an exact LST, but it gives a sense of the diurnal cycle of the surface heat fluxes on land. The difference between the rural and the urban areas is that the Bowen ratio is much higher over the urban one. Over the ocean, on the other hand, the fluxes remain the same over the entire simulation. The total heat flux of the ocean is $318 \text{ W m}^{-2}$, which is the temporal average of $F_{\text{tot}}$ over the land area providing the same energy input over land and ocean.

Acknowledgments

The model results of VVM used in this study can be made available by contacting the corresponding author. We thank Taida Institute for Mathematical Sciences (TIMS) for providing computation resources. The authors are supported by Taiwan’s MoST through grant 107-2111-M-002 -010 -MY4 to National Taiwan University.

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