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Identifying mechanisms of tropical cyclone generated orographic precipitation with Doppler radar and rain gauge observations

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The tropical cyclone (TC) generated orographic precipitation frequently causes severe floods and landslides over coastal and land areas, but its underlying processes remain largely unresolved. This study explored this issue using a high-density rain gauge network and Doppler radar observations to investigate an intense orographic precipitation event over Da-Tun Mountain (DT) in northern Taiwan associated with Typhoon Meari (2011). Detailed examination of observations and the quantification of precipitation enhancement showed that the seeder–feeder mechanism, rather than the widely known upslope lifting mechanism, was a primary contributor to heavy precipitation. Smaller-scale, landfalling convective elements embedded within TC background precipitation. These rapidly evolving scenarios represent a secondary contributor to the modulation of precipitation intensities. The results from the study provide important insights into the relative importance of the different processes of orographically enhanced precipitation for TCs.

Major natural disasters, such as severe floods and landslides, are frequently caused by the approach or landfall of tropical cyclones (TCs). For many geographical locations influenced by TCs, interactions of TC circulations with topography have generally been recognized as one of the most important processes contributing to continuous, torrential rains¹⁻⁷. Understanding these underlying processes is essential for improving TC rainfall forecasts over coastal and inland regions. Nevertheless, in contrast to a very historical and comprehensive study of orographic precipitation occurring in midlatitude environments of synoptic cyclones and fronts⁸⁻¹⁰, our knowledge of the mechanisms responsible for orographic enhancement of precipitation in TC environments remains limited due not only to a general lack of detailed observations over mountains but also to complicated features of preexisting TC precipitation and realistic topography¹¹⁻¹⁵.

TCs are approximately circular, Rankine-like vortices and consist of intense winds and heavy precipitation over a broad area. Given the typically strong and moist low-level ambient winds associated with TCs, a relatively large Froude number ($F_r = U/NH$, where U is the upstream wind speed, N is the Brunt-Väisälä frequency, and H is the mountain height) flow regime can easily occur; thus, the oncoming winds would most likely climb over mountains instead of flowing around the topography. Consistent with this speculation, upslope lifting has been documented as one of the primary

forcing mechanisms contributing to precipitation enhancement over the windward side of mountains under the influence of TCs^{2,16-19}. This "upslope enhancement" scenario is widely known and is helpful for interpreting large-scale rainfall patterns over mountainous regions with TC track-terrain characteristics (i.e., the relative location of the TC circulation center to the topography)²⁰⁻²². However, owing to the complicated terrain features and highly variable precipitation inherently associated with TCs, it seems possible that the intensity and detailed distribution of TC-produced orographic precipitation would be not simply governed by the mechanism of upslope lifting.

A very limited number of recent TC investigations have provided some clues of other processes beyond the scientific scope of the upslope enhancement of precipitation described above. With detailed analyses of radar observations collected over northern Taiwan for Typhoon Xangsane (2000), Yu and Cheng¹¹ first showed that the simple concept of upslope enhancement was inadequate to explain the distribution and intensity of precipitation over mountains in TC environments. Instead, the precipitation inherently associated with TC circulations (i.e., the TC background precipitation) was also found to be an important contributor to the actual characteristics of the observed orographic precipitation during that typhoon event. In support of findings from Yu and Cheng¹¹, Smith et al.²³ reported

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the importance of the TC background precipitation on the precipitation enhancement observed over Dominica Island during the passage of Hurricane Dean (2007). With the aid of a linear, idealized model of orographic precipitation, Smith et al.²³ hypothesized that raindrops from higher-level background precipitation can gather low-level orographically generated cloud water as they fall to the ground and thus increase the precipitation intensity over mountainous regions in a manner similar to the seeder-feeder mechanism²⁴. Yu and Cheng¹³ investigated the detailed distribution of intense orographic precipitation associated with Typhoon Morakot (2009) over different topographical regions of southern Taiwan. Their observations revealed that the rainfall maximum evident over a low and narrow mountain barrier tended to shift to regions near and downstream of the mountain crest, and these observed characteristics of enhanced precipitation are consistent with the theoretical predictions of the seeder-feeder processes. Although these previous studies of TCs have mentioned the potential role of the seeder-feeder effect in precipitation enhancement, our understanding of this topic is still rather incomplete because of not only the diverse characteristics of TC background precipitation (i.e., seeder clouds)^{15,25-28} but also the lack of observationally quantitative confirmation.

The spatial and temporal variability of upstream oncoming winds associated with TCs and the structural modulation of landfalling rainbands of TCs by topography have also been documented to be additional factors influencing the intensity of TC-generated orographic precipitation^{6,11,29}. These aspects are complicated in nature and are more challenging to inspect because they involve both the rapid evolution of convective motions of individual landfalling rainbands and the dynamic interactions between topographically generated disturbances and rainband-scale circulations and precipitation. It is clear that there are still considerable gaps in our knowledge of the relative importance of various possible mechanisms of orographically enhanced precipitation that are active in the different TC environments and topographic features.

The primary objective of this study is to investigate the mechanisms responsible for the intense orographic precipitation occurring over Da-Tun Mountain (hereafter DT) as Typhoon Meari (2011) affected Taiwan on 24-25 June 2011. DT is a three-dimensional, isolated mountain barrier located adjacent to the northern coast of Taiwan and has terrain peaks at approximately 1 km (MSL, mean sea level). Climatologically, DT is one of the most concentrated areas of heavy precipitation in Taiwan, as it is influenced by the approach or landfall of TCs^{6,11}. A similar scenario was also evident for Meari, which caused the highest rainfall accumulation over DT compared to other regions of Taiwan, as the typhoon's outer circulation brought strong north-northwesterly/ northwesterly flow impinging on the windward slopes of the mountain barrier. Moreover, a high-density, automatic rain gauge network [the DT rain gauge network (DTRGN)³⁰, see Methods for details] was established over DT by the end of 2010. The DTRGN has provided very good coverage of rainfall measurements across this mountain barrier since 2011. This unique rain gauge dataset, together with the availability of high-resolution Doppler radar observations over northern Taiwan, makes it possible to document the detailed aspects of precipitation intensity and distribution over DT for Typhoon Meari and to explore the underlying processes responsible for the observed orographic enhancement of precipitation in the TC environment.

Results

Case overview and environmental conditions

Typhoon Meari formed over the western Pacific Ocean east of the Philippines on 22 June 2011. Meari initially headed northwest on the first day and then continued its north-northwestward journey at a speed of ~8–10 m s⁻¹ over the next three days along the western boundary of the western Pacific subtropical high. Meari was classified as a weak typhoon (with maximum winds less than 32.6 m s⁻¹) and maintained a similar intensity during its lifetime. Meari did not make landfall on Taiwan; instead, it passed over the oceanic area east of Taiwan during 24-25 June (Fig. 1).



Fig. 1 | **Best track of Typhoon Meari (2011) provided by the CWA.** The position of the typhoon center is indicated by solid gray circles every 6 h. The primary analysis in this study is focused on the duration of occurrence of heavy precipitation over DT from 1700 UTC on 24 June to 0300 UTC on 25 June 2011, and the typhoon track during this particular period is highlighted by the red line.

Figure 2a shows the spatial distribution of the lowest available radar reflectivity from the Wu-Fen-San (WFS) radar at 1857 UTC on 24 June to illustrate the major precipitation patterns associated with Meari. During this time, Meari's center was located ~250 km east-northeast of northern Taiwan, and the primary precipitation associated with the typhoon circulation was confined to the outer, western portion of the typhoon center. Satellite images (not shown) at a larger scale also support the highly asymmetric distribution of typhoon precipitation seen from radar observations. Heavy rain started to occur over northern Taiwan as it was influenced by the arrival of the outer precipitation of Meari. The outer typhoon precipitation was somewhat organized into multiple elongated, banded zones of moderate to strong reflectivity (~25-40 dBZ) and typically moved onshore and made landfall on the northern and/or northwestern coast of Taiwan. There were also some smaller-scale, irregular patterns of locally enhanced reflectivities (>40 dBZ) embedded within the banded features of precipitation, such as those observed ~10-50 km offshore to the northwest of the northern coast (highlighted by a hollow arrow in Fig. 2a). The landfall of these convective elements appeared to contribute to a rapid increase in rainfall over land. Meari's outer precipitation maintained its intensity with time but gradually became more stratiform and continuous in space (>35 dBZ) during later periods of time, as seen from radar observations valid at 2331 UTC on 24 June (Fig. 2b). Smaller-scale regions of locally enhanced reflectivities (> 40 dBZ) were also evident within the region of widespread stratiform precipitation. Some of these convective elements moved with the typhoon circulation and made landfall on DT. Prominent influences of the landfalling typhoon precipitation on northern Taiwan persisted in the following hours until Meari moved further northward from Taiwan after 0300 UTC on 25 June (Fig. 1).

As shown by the whole-island rain gauge observations, the occurrence of heavy rainfall was confined to northern Taiwan (Fig. 3a), as Meari's outer circulations brought considerable precipitation and strong north-northwesterly and/or northwesterly flow impinging on the northern coast of Taiwan. In particular, maximum rainfall accumulations were present locally over the DT region. The detailed distributions of radar-derived rainfall further indicate a rainfall amount of more than 260 mm during the 10 h study period of primary rainfall from 1700 UTC on 24 June to 0300 UTC on 25 June 2011 (see Methods), as shown in Fig. 3b. The region with the heaviest precipitation tended to collocate with the windward



Fig. 2 | **Larger-scale precipitation distribution of Typhoon Meari (2011).** Lowest available radar reflectivity (dBZ, color shading) from the WFS radar at **a** 1857 UTC on 24 June and **b** 2331 UTC on 24 June 2011. The terrain height is depicted by contours at 500, 1500, and 2500 m MSL. The typhoon center is marked with a



typhoon symbol. The hollow arrows in **a** and **b** denote convective elements with enhanced reflectivities (>40 dBZ) embedded within the banded features of precipitation (i.e., typhoon rainband) or the region of widespread stratiform precipitation (see details in the text).

(i.e., northwestern) mountain ridges of DT. Much less rainfall (60–100 mm) was received well downstream over the low terrain regions to the southeast of DT.

The evolving aspect of orographic precipitation and the upstream oncoming flow associated with Meari's circulation can be best illustrated by a sequence of mean reflectivities at 1 km MSL obtained from the WFS radar and mean dual-Doppler-derived winds valid at 1.5 km MSL (see Methods), averaged within the elongated box and the shaded area, respectively (locations in Fig. 4a). The box is oriented approximately parallel to the oncoming northwesterly flow and encompasses DT and its upstream region. The results are shown in Fig. 4b. During the early and middle portions of the study, roughly before 2300 UTC on 24 June, DT was influenced frequently by the arrival of landfalling typhoon rainbands with embedded convective elements of intense reflectivity (i.e., roughly at 1730, 1830, 1930, 2100, and 2230 UTC). The low-level upstream oncoming flow (15–20 m s⁻¹) was primarily north-northwesterly during this stage. During the later portion of the study (i.e., after 2300 UTC on 24 June), the outer typhoon precipitation became more stratiform and continuous in space. Nevertheless, small-scale regions of enhanced reflectivity, not shown in Fig. 4b due to the averaging procedure, were occasionally embedded within the typhoon precipitation during this stage (highlighted by a hollow arrow in Fig. 2b). These landfalling precipitation elements also influenced the precipitation intensity over DT, as will be described in the Discussion section. The oncoming flow became more northwesterly and generally stronger $(20-25 \text{ m s}^{-1})$, with persistent, heavy precipitation (>40 dBZ) in the vicinity of the mountain ridges of DT. The orographic precipitation started to weaken after 0200 UTC on 25 June in association with changes in low-level prevailing flow (i.e., more westerly winds) and the lack of significant background precipitation associated with Meari as the typhoon left northern Taiwan during these later times.

The modified sounding from Banciao (see Methods for more details), which is valid at 0000 UTC on 25 June, is chosen to demonstrate the major characteristics of the upstream kinematics and thermodynamics. Nearly saturated conditions and neutral convective instability were present in the low to mid-troposphere (Fig. 5a, b). The convective available potential energy (CAPE) was calculated to be near zero. Very moist conditions, neutral instability, and a lack of CAPE have typically been observed in the presence of considerable precipitation associated with the outer circulations of TCs^{11,13}. The corresponding vertical profile of environmental winds indicates that a deep layer of the troposphere from the near-surface to the 250 mb level was dominated by prevailing north-northwesterly or northwesterly flow over northern Taiwan, as expected from the influence of Meari's cyclonic circulations.

With the sounding information above, the flow regime characterizing this case can be evaluated. The representative height of DT is approximately 1 km MSL. Given the saturated conditions in the lower troposphere, the saturated Brunt-Väisälä frequency $(N_m)^{31}$ is more appropriate for approximating static stability in this case. The N_m calculated below the mountain height is equal to 4.6×10^{-3} s⁻¹. The dual-Doppler-derived winds at 1 km MSL during the study period are approximately 20-25 m s⁻¹. These values yield a Froude number of 4-5. In such a relatively large Fr flow regime, oncoming winds are expected to climb over mountains instead of flowing around them⁸. Moreover, given a saturated, convectively neutral environment (Fig. 5), the buoyant force acting on air parcels displaced by orographically forced lifting would be rather small, implying the insignificance of mountain-wave motions9. This thermodynamic characteristic suggests that the spatial pattern and intensity of vertical motions over DT for the present case would be primarily governed by upslope lifting as oncoming winds impinged on the windward slopes of DT rather than by mountain waves and buoyant convection due to the release of convective instability or CAPE.

Quantification of the orographic enhancement of precipitation

To quantitatively understand the degree of precipitation enhancement over mountains in TC environments, it is essential to define the so-called typhoon background precipitation that is inherently associated with typhoon circulations and embedded within the region of orographically enhanced precipitation. However, separating typhoon background



Fig. 3 | **Spatial distribution of TC generated orographic rainfall derived from rain gauge and radar observations during the study period. a** Horizontal distribution of rainfall accumulation (mm, color shading) measured by rain gauges available in Taiwan during the study period from 1700 UTC on 24 June to 0300 UTC on 25 June 2011. The terrain height is depicted by contours of 500, 1500, and 2500 m MSL. **b** Horizontal distribution of rainfall accumulation over the DT region based on radar-derived rainfall. The terrain height is indicated by contours with 200-m intervals from 100 to 900 m MSL. The locations of the CWA surface stations and DTRGN rain gauges are denoted by solid circles and asterisks, respectively. The inset box (8 × 4 km²) highlights the primary region of heavy precipitation over DT.

precipitation from orographically enhanced precipitation is usually not straightforward because the latter is a result of complex dynamic and microphysical interactions between typhoon-induced and topographically generated circulations/precipitation. A common way adopted by previous studies of TCs to measure the strength of typhoon background precipitation is to average the rainfall intensity over a certain area upstream of the considered topography^{11–13}. Once the typhoon background precipitation information is obtained, one can then quantify the orographic enhancement of precipitation by calculating the difference in the intensity between the typhoon background precipitation and the precipitation observed over mountains. Although this kind of quantification is simple and logically reasonable, it is not practical to evaluate specific processes contributing to the estimated amount of precipitation enhancement due to the lack of physically based calculations.

In an effort to provide a more robust estimate and evaluation for the orographic enhancement of precipitation in the present case, the precipitation trajectories are analyzed using the three-dimensional wind fields derived from dual-Doppler observations and the terminal velocity information of precipitation computed empirically from radar reflectivities. For a given time period, the precipitation trajectories are calculated by tracking backward in time from the 1-km MSL level over the upstream region (shading in Fig. 4a) and the primary region of heavy precipitation observed over DT (i.e., the inset box in Fig. 3b). These calculations allow the construction of vertical profiles of precipitation intensities along each precipitation trajectory. The vertical profiles generated from the DT trajectories are supposed to be affected by orographic effects and thus can physically reflect the growth/decay of orographically influenced precipitation (P_{DT}) as it falls from upper to lower levels. On the other hand, the evolving nature of typhoon background precipitation (P_{bg}) , presumably without orographic influences, may be represented by the vertical profiles generated from the upstream trajectory.

An example illustrating the mean vertical profiles of P_{DT} and P_{bg} at 0000 UTC on 25 June is shown in Fig. 6a. The DT and upstream trajectories during this period are also shown in Fig. 6b for reference. Along the upstream trajectory, a rather minor change with height was observed for the intensities of the typhoon background precipitation $(P_{bg} \approx 16 \text{ mm h}^{-1})$, indicating a lack of precipitation growth in the absence of orographic effects. The intensities of the orographically influenced precipitation (i.e., P_{DT}) are initially close to those of its corresponding background precipitation (i.e., P_{bg}) at upper levels (above 2.5-3 km), but they intensify persistently with decreasing height (Fig. 6a). P_{DT} valid at ~3 km MSL may be representative of the initial intensity of background precipitation before it starts to be intensified by orographic effects. P_{DT} reaches a maximum of approximately 33 mm h⁻¹ at the lowest level of trajectory analysis (1 km MSL), with the precipitation originating predominantly from the coastal, upper-level region northwest of DT (Fig. 6b), as expected from the low-level prevailing northwesterly flow associated with Meari's cyclonic circulations. Figure 6a suggests that the orographic enhancement of precipitation was exclusively related to the significant growth in upper-level typhoon background precipitation as it fell below ~3 km MSL into the mountainous region of DT. Moreover, although the mean vertical profiles of P_{DT} and P_{bg} obtained from different analysis periods vary during the 10 h study period (not shown), the characteristics of orographic enhancement of precipitation evident in Fig. 6a are found to be typical. A specific, quantitative measure of the orographic enhancement of precipitation over DT (ΔP_{DT}) in the present study can thus be quantified by the difference in the precipitation rate between 1 and 3 km MSL (P_{DT} at 1 km MSL minus P_{DT} at 3 km MSL) along the DT trajectory. In such a Lagrangian framework of precipitation analysis, it is possible to evaluate the relative importance of forcing mechanisms responsible for the calculated $\triangle P_{DT}$, as will be elaborated in the next section.

Upslope lifting vs. seeder-feeder mechanism

As discussed in the earlier section, given a flow regime with a large Froude number, one of the possible forcing mechanisms contributing to the observed precipitation enhancement in the present case is upslope lifting that occurred as Meari's outer circulations encountered the windward slopes of DT. Meari brought considerable typhoon precipitation to the vicinity of northern Taiwan (Figs. 2 and 4b); therefore,





Fig. 4 | Evolution of mean radar reflectivities in the area encompassing DT and its upstream region during the study period. a Topographic features of DT (m MSL, color shading). The inset box ($40 \times 15 \text{ km}^2$) oriented approximately parallel to the oncoming northwesterly flow and encompassing DT and its upstream region indicates the area used to calculate the mean reflectivities at 1 km MSL from the WFS radar shown in **b**. The smaller inset box with gray shading ($15 \times 12 \text{ km}^2$), located

upstream of DT, indicates the area used to calculate the mean dual-Doppler-derived winds at 1.5 km MSL shown in **b**. **b** A sequence of mean radar reflectivities (dBZ, color shading) as a function of time and the long axis of the inset box during the 10 h analysis period. The calculated mean dual-Doppler-derived winds are indicated by wind barbs. Full wind barbs correspond to 5 m s^{-1} ; half barbs correspond to 2.5 m s^{-1} . The right panel shows the corresponding mean terrain height along the inset box.



Fig. 5 | Environmental conditions from the upstream sounding. a Skew T-logp plot for the modified Banciao sounding valid at 0000 UTC on 25 June 2011 and (b) the corresponding vertical profiles of potential temperature and equivalent potential temperature. In (a), full wind barbs correspond to 5 m s⁻¹; half barbs correspond to 2.5 m s⁻¹.

microphysical interactions between ambient typhoon precipitation and topographically generated precipitation may also likely play a role in facilitating the occurrence of prominently enhanced precipitation observed over DT. If the orographic precipitation exclusively generated by upslope lifting and ambient typhoon precipitation are regarded as lower-level feeder and higher-level seeder clouds, respectively, the low-level precipitation enhancement involved in such microphysical processes can be physically considered the so-called seeder–feeder mechanism^{24,32–34}. In this section, the relative importance of the upslope lifting and seeder–feeder mechanism in contributing to the orographic enhancement of precipitation (i.e., ΔP_{DT}) is evaluated.

Three-dimensional precipitation information obtained from the simulation of the upslope model (see Methods for more details), is used to provide a theoretically quantitative estimation of precipitation production due to upslope lifting, denoted as P_{UL} . Given that ambient typhoon precipitation is not present in the upslope model, the precipitation enhancement caused by upslope lifting is exactly the modeled precipitation rate (P_{UL}) valid along each precipitation trajectory. It should be noted that the P_{UL} obtained from the upslope model may represent an upper limit of precipitation production through upslope lifting because in reality cloud water generated by upslope lifting may have been partially washed out by TC background precipitation due to the seeder-feeder effect. For the theoretical



Fig. 6 | Vertical variations of precipitation rates along DT and upstream precipitation trajectories, respectively. a The mean vertical profiles of P_{DT} (black line, mm h⁻¹) and P_{bg} (blue line, mm h⁻¹) valid at 0000 UTC on 25 June 2011. P_{DT} is obtained from precipitation trajectories over DT, while P_{bg} is obtained from precipitation trajectories over the upstream region. **b** Horizontal distribution of radar reflectivities (dBZ, color shading) at 1 km MSL from the WFS radar at 0000

prediction of the orographic enhancement of precipitation due to the seeder–feeder process along DT trajectories, an attempt is made to derive an equation describing the vertical variation in ambient typhoon precipitation (i.e., the seeder cloud) as it interacts microphysically with low-level feeder clouds. All variables/symbols used in the formula derivation below are summarized in Table 1 for reference. We start with a simplified growth equation of a seeder cloud droplet by assuming perfect collection and neglecting the size and terminal velocity of feeder clouds³⁵ as follows:

$$\frac{dD}{dt} = \frac{M_f v_t}{2\rho_w} \tag{1}$$

where D, M_f , v_t , and ρ_w are the diameter of the seeder cloud droplet, liquid water content, terminal velocity of the seeder cloud droplet, and density of the liquid water, respectively. Moreover, following Yu and Cheng¹³, the precipitation rate (P_r) can be expressed in terms of D, ρ_w , v_t , and the number concentration (N) with the monodisperse assumption as follows:

$$P_r = \frac{\pi}{6} D^3 \rho_w v_t N \tag{2}$$

By differentiating (2) with respect to time and with the aid of (1), we have

$$\frac{1}{P_r}\frac{dP_r}{dt} = \frac{3M_f v_t}{2D\rho_w} \tag{3}$$

The empirical relationship between the median volume diameter (D_0) of precipitation particles and the precipitation rate can be approximated by a power law expression³⁶:

$$D_0 = A P_r^{\ B} \tag{4}$$

where *A* and *B* are constant values of 9.6578947 × 10^{-2} and 0.14, respectively. Given that the seeder clouds are assumed to be monodisperse in this study, D_0 is actually identical to *D*. With the approximation of (4), (3) may be



UTC on 25 June 2011. The terrain height is indicated by contours with 200 m intervals from 100 to 900 m MSL. The inset boxes indicate the areas used to calculate the DT and upstream precipitation trajectories. The positions of the backward precipitation trajectories at 4 km MSL for each dual-Doppler-synthesized grid are marked by white dots, with black lines representing their corresponding trajectories.

rewritten as:

$$\frac{1}{P_r}\frac{dP_r}{dt} = \frac{3M_f v_t}{2A P_r^{\ B} \rho_w}$$
(5)

Mathematically, the change in precipitation rate with height can be expressed by the following equation:

$$\frac{dP_r}{dz} = \frac{dP_r}{dt} \times \frac{dt}{dz} = \frac{dP_r}{dt} \times \frac{1}{-\nu_t}$$
(6)

Combining (5) and (6) yields

$$P_r^{B-1}dP_r = -\frac{3M_f}{2A\rho_w}dz \tag{7}$$

Moreover, (7) can be integrated vertically as:

$$\int_{P_{r1}}^{P_{r2}} P_r^{B-1} dP_r = \int_{Z_1}^{Z_2} -\frac{3M_f}{2A\rho_w} dz$$
(8)

where Z_1 and Z_2 represent different heights, and P_{r1} and P_{r2} represent the precipitation rates at Z_1 and Z_2 , respectively. (8) can be written in numerical form as:

$$P_{r1} = \left[P_{r2}{}^{B} + \frac{3B\overline{M_{f}}}{2A\rho_{w}}(Z_{2} - Z_{1})\right]^{1/B}$$
(9)

where $\overline{M_f}$ is the average liquid water content calculated between Z_1 and Z_2 and can be obtained from the upslope model. Based on (9), the precipitation rates at different heights can be calculated along the DT trajectories from the initial precipitation rate and height. Following the quantitative measure for the observed orographic enhancement of precipitation described in the previous section, the precipitation enhancement due to the seeder–feeder mechanism, denoted as ΔP_{SF} , can then be estimated by calculating the difference between the integrated precipitation rate at 1 km MSL and its

 Table 1 | Variables/symbols used in the derivation of the formula for estimating precipitation enhancement through the seeder-feeder mechanism

Variable/Symbols	Description
D	Diameter of seeder cloud droplet
M _f	Liquid water content of feeder cloud
V _t	Terminal velocity of seeder cloud droplet
ρ _w	Density of liquid water
P _r	Precipitation rate
Ν	Number concentration of seeder cloud
Do	Median volume diameter of seeder cloud
Α	Empirical constant: 9.6578947×10 ⁻²
В	Empirical constant: 0.14
Z ₁	Height at level 1
Z ₂	Height at level 2
P _{r1}	Precipitation rate at level 1
P _{r2}	Precipitation rate at level 2

initial precipitation rate observed at 3 km MSL (i.e., P_{DT} valid at 3 km MSL). Comparisons of the theoretical ΔP_{SF} and observed ΔP_{DT} can provide a strong hint on the relevance of the seeder–feeder mechanism to the observed precipitation enhancement over DT.

The relative importance of upslope lifting and the seeder-feeder mechanism for the present case can be illustrated by the time series of P_{UL} , ΔP_{SF} , and ΔP_{DT} during the study period, as shown in Fig. 7. Rapid temporal variations in the observed precipitation enhancement (i.e., ΔP_{DT}) are evident, and most of these fluctuations are greater than 5-10 mm h⁻¹, with a maximum of ~22 mm h⁻¹. These observed magnitudes of precipitation enhancement are roughly comparable to those reported over the mountainous region of southwestern Taiwan by Yu and Chengli (~10-30 mm h⁻¹) for Typhoon Morakot (2009), although the methods used to measure the orographic enhancement of precipitation are somewhat different between the two studies. It is clear that the precipitation enhancement observed over DT is generally well captured by those of ΔP_{SF} (dashed curve in Fig. 7) in terms of trends and intensities. However, the precipitation rates due to upslope lifting (P_{UL} , red dashed curve in Fig. 7) exhibit much less pronounced temporal variations with a narrow range of precipitation enhancement between 5 and 10 mm h⁻¹. The slow increasing trends of P_{UL} during the early and middle portions of the analysis are consistent with some intensification of upstream oncoming flow associated with TC circulations during the study period (Fig. 4b). Most of the time, significant underestimations of the PUL compared to the observations are present. One of the most interesting aspects shown in Fig. 7 is that the seeder-feeder mechanism appears to be potentially more efficient than upslope lifting in intensifying precipitation over mountains in the studied TC environment. Seeder-feeder enhancement (ΔP_{SF}) greater than and/or even twice the upslope-lifting enhancement (P_{UL}) is common. These quantitative analyses support that the seeder-feeder mechanism should be much more important than upslope lifting for explaining the temporal variations and intensities of precipitation enhancement observed over DT in the present study.

A statistical visualization to complement what we see in Fig. 7 is provided by a scatterplot of ΔP_{DT} , ΔP_{SF} , and P_{UL} as shown in Fig. 8. There is good agreement between ΔP_{DT} and ΔP_{SF} , and both are generally distributed along the diagonal line. This characteristic is in distinct contrast to the data points of ΔP_{DT} and P_{UL} , which mostly lie along the horizontal line. The correlation coefficients for ΔP_{DT} versus ΔP_{SF} and ΔP_{DT} versus P_{UL} are calculated to be 0.76 and 0.55, respectively, with RMSEs equal to 4.3 and 6.7 mm h⁻¹, respectively. Note that the RMSE of ΔP_{SF} is close to that of the radar-estimated precipitation intensities (i.e., 4.6 mm h⁻¹, see Methods). It is possible that ΔP_{SF} may be able to provide a more accurate prediction of





Fig. 7 | Time series of the observed and theoretically estimated precipitation enhancement during the study period. The observed precipitation enhancement (ΔP_{DT} mm h⁻¹), the precipitation enhancement due to the seeder–feeder process (ΔP_{SF} , mm h⁻¹), and the precipitation enhancement due to upslope lifting (P_{UL} , mm h⁻¹) are indicated by solid black lines, dashed black lines, and dashed red lines, respectively.

precipitation enhancement if the uncertainty of radar precipitation estimates could be considered well and reasonably mitigated.

The mean horizontal distribution for ΔP_{DT} , ΔP_{SF} and P_{UL} over the mountainous area used to calculate the DT precipitation trajectories (Fig. 6a) during the study period is shown in Fig. 9. ΔP_{DT} exhibits two local maxima of precipitation enhancement just downstream of the mountain ridge, and both ΔP_{SF} and P_{UL} are characterized by a spatial pattern of precipitation enhancement similar to that of ΔP_{DT} . However, the intensities of ΔP_{DT} and ΔP_{SF} are generally comparable, but P_{UL} provides obvious underestimates of precipitation enhancement over the mountainous region with ~4–6 mm h⁻¹ weaker than observations. These analyses also suggest a better prediction of the seeder-feeder process for the observed precipitation enhancement, further supporting what we see in Fig. 7.

Considering airflow-terrain interactions, one of the important factors influencing the degree to which upslope lifting enhances windward precipitation is the spatial scale of the mountain's lifting section⁹. For the present case, the windward slope of DT is narrow and only approximately 10 km wide (Fig. 3b) and the upstream, low-level oncoming flow associated with typhoon circulations retrieved from the dual-Doppler synthesis is 20-25 m s⁻¹. With these characteristics, the time scale for an air parcel to experience orographic lifting over the windward slope (i.e., width/wind speed) is 7-8 min. This time scale is insufficient for cloud droplets to grow into raindrop-size hydrometeors, as the typical time required for rain formation is around 0.5-1 h³⁵. Therefore, theoretically, a direct, significant contribution of upslope-lifting condensation to surface rainfall should not occur easily in the present case, consistent with our results shown in Fig. 7. Hence, if one observes a rainfall enhancement over a narrower barrier, hydrometeors from preexisting precipitation associated with TCs (i.e., TC background precipitation) would probably work as collector drops to wash out orographically generated clouds at low levels. In this seeder-feeder context, when active rainbands/precipitation prevail in the TC environment and approach the mountainous region, like the present case, they can provide abundance of background precipitation to serve as seeder clouds. The seeder clouds can then continually collect the feeder cloud generated by upslope lifting, enhancing orographic precipitation efficiently over a sustained period, as supported by Fig. 7. On the other hand, the relative importance of the upslope lifting and seeder-feeder effect on contributing to precipitation enhancement in the TC environment would be probably altered for a wider mountain barrier (i.e., longer lifting section), as elaborated in Yu and Cheng¹³.

Discussion

The results elaborated in the previous section support the general importance of the seeder-feeder mechanism, rather than the widely known upslope lifting mechanism, on the orographic enhancement of precipitation observed in the present case. Although this finding is consistent with several previous modeling and observational studies of TCs suggesting a potential impact of the seeder-feeder effect on contributing to heavy orographic precipitation^{6,13,23}, the present study provides more quantitative and robust evidence to confirm the relative contribution of seeder-feeder and upslope lifting mechanisms in the TC environment. Nonetheless, obvious discrepancies between the theoretical prediction of the seeder-feeder



Fig. 8 | A statistical comparison of observed and theoretical precipitation enhancement. Scatterplot of the observed precipitation enhancement (ΔP_{DT}) vs. the precipitation enhancement due to upslope lifting (P_{UL} , red solid circles) and the precipitation enhancement due to the seeder–feeder process (ΔP_{SF} , black solid circles) during the study period. The calculated correlation coefficients (CORRs) and root-mean-square errors (RMSEs) are also indicated at the top left.

mechanism and the observed precipitation enhancement do occur in some particular time zones, as shown in Fig. 7. This implies not only the presence of inherent errors due to assumptions required to derive the quantitatively predicted equation of the seeder–feeder mechanism but also the possible existence of processes other than the seeder–feeder mechanism contributing to the modulation of precipitation enhancement observed over DT.

To explore the possibility above, the specific differences between the seeder-feeder precipitation enhancement and observed orographic enhancement of precipitation (i.e., $\Delta P_{SF} - \Delta P_{DT}$) at different times were calculated during the study period, as shown in Fig. 10. Considering the uncertainty of radar-estimated precipitation intensities with an RMSE of 4.6 mm h⁻¹ (see Methods), it is reasonable to interpret the seeder-feeder mechanism as a suitable, key contributor to the observed orographic enhancement of precipitation if the differences between ΔP_{SF} and ΔP_{DT} are within $\pm 5 \text{ mm h}^{-1}$. As indicated in Fig. 10, this condition occurred most of the time during the study period, which is consistent with the aforementioned conclusion. However, significant differences well beyond ±5 mm h⁻¹ also occurred at several particular times, as highlighted by the thick arrows in Fig. 10. With the aid of Fig. 4b and the detailed inspection of individual radar images, we found that these large errors typically occurred as smaller-scale convective elements of enhanced reflectivities embedded within propagating rainbands or background stratiform precipitation moved into the DT region and modulated the precipitation intensities over the topography. Two examples of radar observations shown below are used to illustrate these scenarios.

The first example is chosen from the time period of 1717–1735 UTC on 24 June, encompassing a local maximum underestimate for the seeder–feeder mechanism (denoted by a thick arrow in Fig. 10). As evident from a sequence of horizontal distributions of radar reflectivity at 1 km MSL, an approximately east-west-oriented typhoon rainband propagated southward, approached the coast of northern Taiwan, and then made landfall over the DT region (Fig. 11a, c, e, g). The precipitation over DT was observed to intensify significantly during the arrival of landfalling convective elements (>44 dBZ) embedded within the rainband. The vertical cross-sections passing through one of the convective elements indicate the highly convective nature of precipitation (Fig. 11b, d, f, h). This precipitation

Fig. 9 | Horizontal distribution of the observed and theoretically estimated precipitation enhancement during the study period. The observed precipitation enhancement (ΔP_{DT} , mm h⁻¹), the precipitation enhancement due to the seeder–feeder process (ΔP_{SF} , mm h⁻¹), and the precipitation enhancement due to upslope lifting (P_{UL} , mm h⁻¹) are shown with color shading. The terrain height is indicated by contours with 200 m intervals from 300 to 900 m MSL. The analysis domain (8 × 4 km²) is corresponding to the area used to calculate the DT precipitation trajectories as highlighted in Fig. 6b.





Fig. 10 | Differential values between the theoretical and observed precipitation enhancement. The time series of differences (mm h⁻¹) between the seeder–feeder precipitation enhancement (ΔP_{SF}) and the observed orographic enhancement of precipitation (ΔP_{DT}) during the study period. The gray shading indicates the range of the differences within ± 5 mm h⁻¹. The thick arrows highlight the times with the most prominent overestimation and underestimation in the theoretical calculation.

intensification cannot be well captured by the seeder-feeder theory on the basis of precipitation trajectory analyses because the propagation of the rainband and its associated convective elements did not actually follow ambient northwesterly winds that resulted in the typical northwest-southeast-oriented precipitation trajectory (Fig. 6b). Note that rainbands and convective elements usually propagate in a manner more complicated than precipitation trajectories because of continuous and discrete propagations due to the generation of new convection^{37,38}. This highlights the limitation in the precipitation trajectory analyses and contributes to the prominent discrepancy in the orographic enhancement of precipitation estimated by the seeder-feeder mechanism during this period. These rapidly evolving aspects suggest that, in addition to the seeder-feeder mechanism, the interactions of typhoon rainbands and their embedded convective elements with topography²⁹ are also important processes for modulating precipitation intensities over DT.

The other example is selected from the period of 2325-2343 UTC on 24 June, which includes a local maximum overestimate for the seeder-feeder mechanism (denoted by a thick arrow in Fig. 10). As shown in Fig. 12a, c, e, g, multiple smaller-scale (< ~4 km) convective elements (>42 dBZ) embedded within background stratiform precipitation propagated southeastward, approached the coast of northern Taiwan, and subsequently made landfall over the DT region. The convective elements initially weakened slightly during landfall (Fig. 12b, d, f). However, an obvious intensification of the convective elements was observed as they just reached the mountain crest of DT (Fig. 12g, h), further modulating the precipitation intensities over DT. Similar to the first example, such a complicated scenario cannot be well resolved by the theoretically idealized seeder-feeder mechanism. Owing not only to the insufficient temporal and spatial resolution of dual-Doppler synthesis but also to the lack of microphysical measurements in this study, it is difficult to clarify the dynamic and microphysical processes involved in these small-scale interactions.

The two examples shown above suggest that the modification of convective elements embedded within typhoon rainbands/circulations by topography also influences the intensity and distribution of orographic precipitation. Compared to the overall importance of the seeder–feeder mechanism, these rapid processes should not be a primary but a secondary contributor to the precipitation enhancement over DT because they occurred only in particular, relatively short periods of time with the landfall of convective elements (i.e., highlighted by arrows in Fig. 10). However, one important implication is that the nature of the background precipitation inherently associated with TC circulations is one of the key factors determining the relative importance of the different processes of orographically enhanced precipitation discussed in this study. Future research should pay more attention to these aspects to better understand and predict heavy precipitation over mountainous regions in TC environments.

In addition, the details of microphysical processes involved in the seeder-feeder mechanism cannot be addressed in the present case due to the lack of microphysical measurements over DT. Considering the potential importance of the seeder-feeder mechanism in the TC environment, making microphysical measurements available over the mountainous region would be thus crucial for an in-depth understanding of TC generated orographic precipitation. These observational efforts, together with detailed kinematic and thermodynamic datasets collected over topography from future specially designed field experiments, will be required to provide a more complete picture of orographic enhancement of precipitation associated with TCs. The sensitivity experiments of high-resolution numerical modeling by removing terrain features would be also helpful to clarify the relative roles of TC background precipitation and orographic forcings in contributing to the seeder-feeder-like precipitation enhancement.

Methods

Data

The datasets used in this study include surface rain gauge measurements, Doppler radar observations, and sounding data. The locations of rain gauge stations from the Central Weather Administration (CWA) and the DTRGN are indicated in Fig. 13. The DTRGN composed of 22 tipping-bucket gauge stations (locations indicated by asterisks) and largely complements the sparse distributions of CWA rain gauges over the mountainous region of DT and its surrounding area³⁰. In this study, 1 min temporal resolution rainfall measurements recorded by the DTRGN are accumulated hourly to coordinate with the CWA rain gauge data.

Doppler radar observations are provided by the CWA operational S-band (10-cm) Doppler radar (WSR-88D) on WFS and the Civil Aeronautics Administration (CAA) operational C-band (5 cm) Doppler radar located at Taoyuan International Airport (Fig. 13). Both radars provide volumetric distributions of reflectivity and radial velocity with temporal intervals of ~6 min (for the WFS radar) and ~10-20 min (for the CAA radar) between each volume. Details on the characteristics of both the WFS and CAA radars are described in Yu et al.28. The WFS radar is located ~25 km southeast of DT, and the lower boundary of the radar beam from the lowest plan position indicator (PPI) scan (0.5° elevation) over the DT region has an altitude of ~800 m (MSL), which is mostly above or close to the slope surface of DT. Compared to the CAA radar, the WFS radar has a better observational geometry and longer wavelength to provide important lowlevel precipitation information over the mountainous region of DT. However, some radar beams from the lowest PPI scans of the WFS radar may be partially blocked by topography near the mountain peaks of DT. Following Cheng and Yu³⁰, contamination due to mountain clutter and blockage is identified by comparing the altitudes of the lower boundary for each range gate of a given radar beam with the terrain heights. Any remaining gates beyond a gate with a calculated altitude lower than the underlying terrain height are considered to be mountain clutter and are thus removed.

Z-R empirical relationship

For better documentation and quantification of the orographic enhancement of precipitation observed in this study, the WFS radar measurements with clutter removal, as described above, are used to estimate precipitation intensities in the vicinity of DT. The lowest available radar reflectivities from the WFS radar are converted to hourly rainfall rates by applying the well-known power law $Z = aR^b$, where Z is the reflectivity factor (mm⁶ m⁻³), R is the rainfall rate (mm h⁻¹), and a and b are experimentally determined coefficients. For the present case, a least-square regression fitting is performed between the available rain gauge data and the lowest available radar reflectivities over the DT area during the 10-h period of primary rainfall (to be

Fig. 11 | Evolving aspects of precipitation for a selected landfalling rainband and its associated convective elements. a Horizontal distributions of radar reflectivity (dBZ, color shading) at 1 km MSL from the WFS radar at 1717 UTC on 24 June 2011. The terrain height is indicated by contours with 200 m intervals from 100 to 900 m MSL. Line segment A-A' marks the location of the vertical crosssection shown in **b**. The thick arrow indicates the position of the target convective element. b The A-A' vertical cross-section of radar reflectivity (dBZ, color shading) and the underlying topography (km MSL, brown shading) along the section. The thick arrow indicates the position of the target convective element. c-h As in a and b but showing the results at 1723, 1729, and 1735 UTC on 24 June 2011, respectively.



described later). This statistical analysis yields the best values of *a* and *b*, which are equal to 35.2 and 1.77, respectively. A scatterplot of the radar-based rainfall estimates versus the rain gauge measurements is shown in Fig. 14, and the correlation coefficient and root-mean-square error (RMSE) are calculated to be 0.93 and 4.6 mm h⁻¹,

respectively. Note that the empirical Z-R relationship ($Z = 32.5R^{1.65}$) commonly adopted for convective storms and typhoons^{13,19,39} produces larger errors in rainfall estimates (RMSE ~ 7.8 mm h⁻¹). The slightly larger values of coefficients *a* and *b* used herein give rise to relatively smaller errors in the radar-estimated rainfall intensity.

Fig. 12 | Evolving aspects of precipitation for selected landfalling convective elements. As in Fig. 11 except for the horizontal distributions and vertical cross-sections of the radar reflectivity (dBZ, color shading) along line segment B-B' at 2325 UTC (**a**, **b**), 2331 UTC (**c**, **d**), 2337 UTC (**e**, **f**), and 2343 UTC (**g**, **h**) on 24 June 2011. The two target convective elements are marked by thick arrows.



Upstream soundings

The sounding from Banciao (BC), which is located in northern Taiwan (Fig.13), can be used to understand the environmental conditions related to the occurrence of intense orographic precipitation over DT. Considering the Meari-induced north-northwesterly/northwesterly flow over northern

Taiwan, the Banciao sounding would probably be influenced by the lee-side effect of DT, although previous studies have supported a relatively minor modification of thermodynamics by this smaller-scale mountain barrier in typhoon environments¹¹. To mitigate potential uncertainty, we followed the procedure of modifying the Banciao sounding proposed by Cheng and Yu³⁰,

Fig. 13 | Topographic features of northern Taiwan and the study region. The terrain height (m MSL) is indicated by color shading. The locations of the Wu-Fen-San (WFS) Doppler radar and the Civil Aeronautics Administration (CAA) Doppler radar at Taoyuan International Airport are denoted by triangles. The locations of surface stations from the Central Weather Administration (CWA) of Taiwan are denoted by solid and hollow circles, while rain gauges of the Da-Tun rain gauge network (DTRGN) are indicated by asterisks. The surface station at Tamsui (TS) is also marked. The rain gauges used to define the duration of intense orographic precipitation in this study are provided by the DTRGN (asterisks) and some of the CWA surface stations (solid circles) located closer to DT. The location of the Banciao (BC) sounding is marked by a square. The larger inset box denotes the dual-Doppler synthesis domain and the spatial coverage of the upslope model simulation $(70 \times 60 \text{ km}^2)$. The smaller inset box with gray shading $(15 \times 12 \text{ km}^2)$ indicates the area where vertical profiles of mean dual-Doppler-derived horizontal winds are calculated, which are used as the input of upstream kinematics when calculating topographically forced vertical motions in the upslope model.



namely, the sounding data below 1 km (MSL) were replaced by a linear interpolation between the surface data measured at the Tamsui (TS) station located at the coastline of northwestern Taiwan (i.e., upstream of DT) and the sounding data at the height of DT (i.e., 1 km MSL). As required for the thermodynamic input to the simulation of the upslope model, the Banciao soundings with 12 h resolution are first interpolated temporally to obtain hourly sounding profiles, and then a sequence of hourly modified, upstream soundings through the adjusted procedure above are generated.

Dual-doppler synthesis

Dual-Doppler radar synthesis of multiple-view reflectivity and radial velocity data from the WFS and CAA radars is performed to provide a unique depiction of the three-dimensional kinematic field over DT and its upstream areas. In this study, the multiple-Doppler synthesis and continuity adjustment technique (MUSCAT) software⁴⁰ is used to retrieve the wind field. The inset box in Fig. 13 marks the synthesis domain $(70 \times 60 \text{ km}^2)$ extending from the DT region to the oceanic region ~40-60 km off the northern coast of Taiwan. The cross-beam angles of the two radars within the synthesis domain are mostly between 30° and 150°, thus producing relatively smaller uncertainties and errors in the dual-Doppler-derived winds due to the inherent limitations of synthesized geometries⁴¹. By using the Cressman weighting function⁴², reflectivities and radial velocities from raw PPI scans are interpolated into Cartesian coordinates with a horizontal grid spacing of 1 km and a vertical grid spacing of 0.5 km, and the lowest and highest analysis levels are located at 0.5 and 15 km MSL, respectively. The three wind components (u, v, w) are obtained through least squares adjustment of the continuity equation in mass flux form, which allows one to determine the wind field over both flat and complex terrain regions⁴³. In this study, a total of 41 sets of dual-Doppler-synthesized winds are derived with a temporal interval of 10-20 min during the 10-h study period. The synthesis time window (10 h) chosen herein is consistent with the duration of intense orographic precipitation, which is defined as more than 50% of the rain gauges (the rain gauges used are marked with solid circles and asterisks in Fig. 13) over the DT region with rainfall intensities greater than 5 mm h⁻¹. Detailed analyses for the study are also focused on this particular period of primary orographic precipitation.

Upslope model

In this study, an orographic precipitation diagnostic model (i.e., upslope model³⁰) is also used to evaluate the specific contribution of upslope lifting to the orographic enhancement of precipitation observed over DT. The methodology of the upslope model is similar to that of Sinclair⁴⁴ but has been modified by Cheng and Yu³⁰ to make it more capable of resolving spatial variabilities in precipitation due to upslope lifting over smaller-scale mountain barriers, such as DT. The upslope model adopted herein is relatively simple, and its precipitation physics involve only two primary processes: condensation due to topographically forced upward motions and the advection of condensated liquid water by ambient horizontal winds. Note that the upslope model is most suitable for environmental conditions with a relatively large Froude number close to or greater than unity because these flow regimes favor the occurrence of upslope lifting⁸. In addition, a convectively neutral or stable ambient condition is more relevant because the model does not take the contribution of vertical motions by the release of convective instability into account. As elaborated in the earlier section, the thermodynamic characteristics of Meari's outer circulations meet these environmental scenarios.

The spatial coverage of the upslope model simulation corresponds to the dual-Doppler synthesis domain shown in Fig. 13, with a 1 km (0.1 km) horizontal (vertical) grid spacing. The upstream thermodynamical information required for the upslope model to simulate precipitation is assumed



Fig. 14 | **Statistical comparison of radar-based rainfall estimates and rain gauge measurements in the vicinity of DT.** Scatterplot of radar-based rainfall estimates versus rain gauge measurements (mm h⁻¹) in the vicinity of DT. The calculated correlation coefficient, linear regression equation, *Z-R* relationship, and root mean square error (RMSE) are all indicated.

to be horizontally uniform and is provided from the hourly modified sounding profiles, as described earlier. The vertical profiles of mean dual-Doppler-derived horizontal winds averaged over the oceanic domain off the northwestern coast of Taiwan (gray shading in Fig. 13) are used as the input for upstream kinematics when calculating topographically forced vertical motions in the upslope model. Moreover, considering the advection effect of the condensated water by ambient flow, the dual-Doppler-derived horizontal winds, instead of upstream environmental winds, should be more representative of mountainous flow over DT and are thus used in this model simulation. In this case, the upslope model is run continuously from 1600 UTC 24 June (i.e., one hour prior to the beginning of our study period) to 0300 UTC 25 June 2011 with an integration time step of 10 s. The three-dimensional distributions of modeled precipitation rates (mm h⁻¹) and liquid water content (g m⁻³) are output every minute for relevant analyses.

Data availability

The observational data (radar, surface, and sounding) used in this study are accessible via https://asrad.pccu.edu.tw/dbar/. The TC best track dataset can be accessed at https://rdc28.cwa.gov.tw/.

Code availability

The codes to reproduce the analyses presented in this study are available upon request from the corresponding author.

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Author contributions

L.-W. Cheng performed the data analysis and drafted the manuscript, C.-K. Yu led the study, interpreted the results, and wrote and revised the manuscript, and S.-P. Chen helped with data collection and performed preliminary analyses.

Competing interests

The authors declare no competing interests.

Additional information

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