Cloud-Resolving-Model Simulations of Nocturnal Precipitation over the Himalayan Slopes and Foothills

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ABSTRACT: A numerical experiment with a 2-km resolution was conducted using the Weather Research and Forecasting (WRF) Model to investigate physical processes driving nocturnal precipitation over the Himalayas during the mature monsoon seasons between 2003 and 2010. The WRF Model simulations of increases in precipitation twice a day, one in the afternoon and another around midnight, over the Himalayan slopes, and of the single nocturnal peak over the Himalayan foothills were reasonably accurate. To understand the synoptic-scale moisture transport and its local-scale convergence generating the nocturnal precipitation, composite analyses were conducted using the reanalysis dataset and model outputs. In the synoptic scale, moisture transport associated with the westward propagation of low pressure systems was found when nocturnal precipitation dominated over the Himalayan slopes. In contrast, moisture was directly provided from the synoptic-scale monsoon westerlies for nocturnal precipitation over the foothills. The model outputs suggested that precipitation occurred on the mountain ridges in the Himalayas during the afternoon and expanded horizontally toward lower-elevation areas through the night. During the nighttime, the downslope wind was caused by radiative cooling at the surface and was intensified by evaporative cooling by hydrometeors in the near-surface layer. As a result, convergence between the downslope wind and the synoptic-scale flow promoted nocturnal precipitation over the Himalayas and to the south, as well as the moisture convergence by orography and/or synoptic-scale circulation patterns. The nocturnal precipitation over the Himalayas was not simulated well when we used the coarse topographic resolution and the smaller number of vertical layers.

SIGNIFICANCE STATEMENT: Nocturnal precipitation is observed over the Himalayan slopes and foothills during summer. We investigated physical factors to cause the nocturnal precipitation there using a reanalysis dataset and a numerical simulation with 2-km horizontal resolution. Water vapor was intensively transported to the Himalayas between midnight and morning in the synoptic scale, and then generated precipitation under the influence of mountain topography. A downslope wind, which occurred in association with nocturnal cooling of land surface, enhanced a moisture convergence and caused precipitation over the Himalayan slopes and foothills. Evaporative cooling from the precipitation also contributed the cooling in the near-surface layer, which assisted an intensification of the nocturnal downslope wind.

KEYWORDS: Mountain meteorology; Complex terrain; Topographic effects; Monsoons; Precipitation; Water vapor; Regional models; Diurnal effects

1. Introduction

Summer precipitation systems over the Himalayan slopes and foothills make an essential contribution to the Asian monsoon circulation via the process of condensational heating in the middle and upper troposphere (e.g., Boos and Kuang

2010). The notable characteristics of Himalayan precipitation patterns during the monsoon season, as revealed by in situ observations, are the elevation-dependent diurnal variations. For example, a diurnal precipitation maximum is observed during the nighttime at Syangboche in the Nepal Himalayas (elevation 3833 m, 27.8°N, 86.7°E; Ueno et al. 2001). A single nocturnal peak in precipitation also occurs in Himalayan valleys situated at altitudes of 500–1500 m between 84.2° and 84.5°E, while two peaks in precipitation, one during the afternoon and another around midnight, is found over the mountain ridges between 2100 and 4400 m along the similar longitude (Barros et al. 2001, 2004). These differences in diurnal precipitation patterns that depend on altitude have been confirmed at other in situ observation sites (Ouyang et al. 2019), that is, the observations from the Yadong Valley (along ~89°E) indicate a single nocturnal peak below 3000 m, but

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two peaks (in the afternoon and around midnight) between 3000 and 4300 m. The satellite measurements also provide evidence of the diurnal variation in precipitation and its spatial distribution over the Himalayas and its surrounding regions. The satellite observations show that precipitation maxima occur between midnight and early morning along the Himalayas, and the nocturnal precipitation peak over the Himalayan foothills is consistent with the occurrence frequency of large precipitation systems ($>10\,000\text{ km}^2$) whereas the afternoon precipitation is associated with small- to medium-sized precipitation systems ($<10\,000\text{ km}^2$; Hirose and Nakamura 2005). Shrestha and Deshar (2014) recorded high precipitation amounts along the Lesser Himalayas at approximately 2000 m during the afternoon, which expands to the southern margin of the Himalayas around midnight and in the early morning. Furthermore, two peaks in daily precipitation are clearly captured between 500 and 1000 m, and at around 2000 m, especially in the western and eastern Nepal Himalayas, of which location was associated with the altitude of steep gradient in the Himalayas (Fujinami et al. 2021). These observational studies indicated that the diurnal variations in precipitation over the Himalayas are complex and depend on location; therefore, we should consider differences of them between slopes and foothills along the different longitude. In addition, nocturnal precipitation occurrence is one of the noteworthy features over the Himalayas.

Many researchers have investigated the synoptic-scale moisture supply to the Himalayas and its surrounding regions because it is an important factor in the precipitation occurrence over the high-altitude regions where semiarid conditions dominate. The westward propagation of low pressure systems generated over the Bay of Bengal (BoB) effectively transports water vapor to the Himalayan slopes, which affects the monsoon onset and extreme rainfall there (Barros and Lang 2003; Ueno et al. 2008; Vellore et al. 2016). Meanwhile, the monsoon flow reaches the Tibetan Plateau through the Himalayan valleys when synoptic-scale troughs pass over the plateau, which contributes to the precipitation that develops during the evening and night over the southern and southeastern plateau (Sugimoto et al. 2008). One hypothesis that may explain the mechanisms associated with the diurnal variations in precipitation over the Himalayas is that the upslope wind along the Himalayan slopes induces orographic precipitation during the daytime along the mountain ridges, whereas the synoptic-scale moisture flow converges with katabatic downslope winds observed on the Himalayan slopes around midnight and in the early morning (Egger et al. 2000; Barros and Lang 2003; Ueno et al. 2008; Sahany et al. 2010; Rthrich et al. 2013). However, a more careful investigation for the moisture transport and its convergence processes driving nocturnal precipitation is required over the Himalayas because it is known that the downslope wind is relatively light, or even absent, during the mature monsoon season (e.g., Ueno et al. 2008). In addition, we need to consider the diurnal changes in the synoptic-scale flows that may enhance nighttime precipitation. For example, Fujinami et al. (2017) suggested that the

development of a low-level jet in the stable surface layer enhances the synoptic-scale monsoon flow during the nighttime as well as the barrier jet effect west of the Arakan Mountains, which helps the development of nocturnal precipitation over the southern slopes of the Meghalaya Plateau. The nocturnal acceleration of the low-level monsoon flow has also been identified over the south of the Himalayas (Fujinami et al. 2021).

A numerical experiment with a high spatial resolution is a powerful tool that can be used to generate local-scale physical evidence for the mechanisms associated with diurnal variation in precipitation hypothesized from the limited observational dataset. However, a reasonable simulation of the precipitation over the complex mountainous regions of the Himalayas remains a challenging issue. A seasonal precipitation maximum is observed in July, which is produced in June or another month in most of the dynamical downscaling experiments run by the Coordinated Regional Climate Downscaling Experiment–South Asia (a horizontal resolution of approximately 50 km; Ghimire et al. 2018). Numerical experiments with a horizontal resolution of 6.7 km simulate the summertime nocturnal precipitation over the Himalayan slopes as well as the seasonal variations in precipitation, although the diurnal precipitation maximum occurs slightly earlier than is seen in the observations due to an overestimation of evening precipitation (Norris et al. 2017). Sugimoto and Takahashi (2016) also indicated that even a 5-km resolution experiment cannot capture the nocturnal maximum of precipitation frequency over the Himalayan foothills during the mature monsoon season.

Recently, several studies have conducted convection-permitting and cloud-resolving simulations with horizontal resolutions of less than 3 km over the Himalayas, and with calculation periods between 10 days and 1 year. Numerical simulations with a 1-km horizontal resolution improve the accuracy of the diurnal cycle in precipitation compared with coarser resolution simulations over the Himalayan slopes, particularly for nocturnal precipitation during summer, because they are able to simulate the local-scale physical processes such as the mountain–valley winds over the complex topography (Karki et al. 2017). The simulated precipitation over the Himalayas is also sensitive to microphysical processes in the cloud-resolving experiment, i.e., snow growth by riming, graupel formation, and an increase in rain and cloud droplets via the melting of solid particles (Orr et al. 2017). A momentum budget analysis using a 1-km sensitivity simulation revealed that the local upslope wind reaches the higher-elevation areas when the glacier is absent during the summer season, playing an important role in driving near-surface valley winds (Potter et al. 2018). At higher horizontal resolutions, the complex topography increases surface friction and decelerates the surface southerly moisture fluxes over the Himalayan slopes, which improves the simulated precipitation amounts over the Tibetan Plateau in a similar manner to the effect of a turbulent orographic form drag scheme (Lin et al. 2018; Wang et al. 2020). Although these studies have contributed greatly to our understanding of the physical and local-scale processes associated with the diurnal cycles in precipitation and the local-scale circulation over the Himalayas, the

effect of higher resolutions on the simulation of the precipitation climatology remains an issue. A larger domain experiment would be also required to investigate the interaction between the synoptic-scale wind and topographically induced wind with respect to the development of the nocturnal precipitation.

To improve our understanding of the physical processes that cause the diurnal variations in precipitation over the Himalayan slopes and foothills, and especially for the nocturnal precipitation occurrence, we conducted numerical experiments with a 2-km horizontal resolution during the mature monsoon season over a period of 8 years. The climatology of the diurnal variations in simulated precipitation was evaluated using satellite estimations. We identified days on which nocturnal precipitation dominated over the Himalayan slopes and foothills and analyzed the associated synoptic-scale moisture transport processes using a reanalysis dataset. Furthermore, we also examined land surface and microphysical processes in local scale, which are associated with the moisture convergence to cause nocturnal precipitation, using the simulation results. Finally, we discussed two issues: 1) the difference in simulated precipitation with reference to the representations of topography and vertical resolution using sensitivity experiments during the summer of a single year and 2) whether the cloud-resolving-model simulation is acceptable for the one-by-one comparison of precipitation between the in situ observation and the model experiments.

2. Numerical setup and data

To better understand the local-scale precipitation processes over the Himalayan slopes and southern foothills, we carried out a numerical simulation in the domain shown in Fig. 1a using the Weather Research and Forecasting (WRF) Model version 3.9.1.1 (Skamarock et al. 2008). A 2-km horizontal resolution with 50 vertical layers from surface to 25 hPa were used in the simulation, and outer boundaries were forced by a reanalysis data. Topography in this simulation also have 2-km horizontal resolution, which was made from the Global 30 Arc-Second Elevation (GTOPO30). Our calculation period was between 0000 UTC 1 June and 0000 UTC 1 September for the 8 years from 2003 to 2010, which covers the summers of 2003–09 and for which rain gauge data are available. We also included the summer of 2010 in the simulation to allow a balance of dryer and wetter years in Nepal; that is, the standardized anomaly of observed precipitation between 1987 and 2015 was negative in 2004, 2005, 2006, and 2009, but positive in 2003, 2007, 2008, and 2010 (Sharma et al. 2020). As the first month used as the spinup period for the land surface scheme, our analysis period was between 0000 UTC 1 July and 0000 UTC 1 September in each year. The time step for integration was 0.5 s, and the output interval was 30 min. We used the Dudhia scheme for shortwave radiation (Dudhia 1989), the Rapid Radiative Transfer Model for longwave radiation (Mlawer et al. 1997), the Mellor–Yamada–Nakanishi–Niino 2.5-level turbulence kinetic energy (TKE) scheme to calculate a planetary boundary layer condition (Nakanishi

and Niino 2004, 2006), the unified Noah land surface scheme (Chen and Dudhia 2001), and the modified Thompson scheme for microphysics (Thompson et al. 2008). This combination of physical schemes is similar to that used by Lin et al. (2018), who examined moisture transport to the Tibetan Plateau along the central Himalayan valleys using a 2-km-resolution experiment, although they used the Mellor–Yamada–Janić TKE scheme (Janić 2001) in their study.

The highest-resolution reanalysis from the European Centre for Medium-Range Weather Forecasts, i.e., ERA5 (Hersbach et al. 2020), was used for the initial and lateral boundary conditions of the 2-km WRF simulation. As downloaded, the ERA5 dataset had a Gaussian latitude grid, which we regridded to a 0.28125° mesh grid to match the longitudinal mesh size. The lateral boundary conditions were provided at 6-hourly intervals to reduce the computational load associated with data input/output, although the ERA5 has an hourly resolution. We also used ERA5 alongside the NOAA interpolated outgoing longwave radiation (OLR) data, which is at a resolution of $2.5^\circ \times 2.5^\circ$ (daily; Liebmann and Smith 1996), to analyze the synoptic-scale atmospheric conditions.

The precipitation patterns simulated by the WRF Model were assessed using the Integrated Multisatellite Retrieval for Global Precipitation Measurement (IMERG) V06B product for the period between 2003 and 2010 (Huffman et al. 2019; Tan et al. 2019). The IMERG provides hourly precipitation (mm h^{-1}) at half-hourly intervals and a $0.1^\circ \times 0.1^\circ$ resolution, and these relatively higher temporal and spatial resolutions are useful for the analysis of diurnal cycles in precipitation. The half-hourly precipitation data from IMERG [mm (30 min)^{-1}] were estimated as one-half of the hourly precipitation (mm h^{-1}) for comparison with the model output. In section 4b, we also discussed the one-by-one comparison of precipitation between the simulation and in situ rain gauge observations from four sites (Fig. 1b) between 2003 and 2009 contained in the Coordinated Enhanced Observing Period Asian Monsoon Project (CEOP_AP): Himalayas hourly surface meteorology and radiation dataset. For the gauge observations, we used a composite of the hourly precipitation with flag G (Good) to obtain the climatology of its diurnal cycle. To compare the precipitation between the gauges and the model experiments, simulated hourly precipitation was calculated from the numerical outputs at half-hourly intervals and was bilinearly interpolated from the closest four native simulation grids to the latitude and longitude of the observation site to avoid a chaotic error for the precipitation occurrence over the complex mountainous regions in the model. Hereafter, the local time (LT) indicates UTC + 6 h, the time along 90°E .

3. Results

a. Evaluation of precipitation climatology over the Himalayas

The spatial distributions of the simulated climatological precipitation during July and August over the 8 years from 2003 to 2010 were compared with the satellite-based estimates derived from the IMERG data. Figure 2 shows typical spatial distributions of precipitation during the day and night, using

FIG. 1. (a) Topography of the calculation domain. (b) The enlargement over 27.5° – 28.5° N and 86.3° – 87.2° E shows the in situ observation sites (black stars; from south, Lukla, Namche, Pheriche, and Pyramid) for the CEOP_AP: Himalayas hourly surface meteorology and radiation dataset. Light blue indicates a water body. The large rectangle in (a) is the analysis domain, and the W_{slope} , W_{foothill} , E_{slope} , and E_{foothill} areas are also shown in (a).

the half-hourly precipitation data for 1500 and 0300 LT, respectively. The half-hourly IMERG precipitation was slightly low over the Himalayan slopes relative to that over other regions in the analysis domain at 1500 LT (Fig. 2a), whereas the WRF Model simulated increased precipitation along the Himalayas (Fig. 2b), which caused a remarkable difference in precipitation between the IMERG estimation and the model over the Himalayan slopes (Fig. 2c). At 0300 LT, precipitation increased markedly over the mountain slopes between 83° and 85° E (W_{slope}) and the Himalayan foothills between 88° and 90° E (E_{foothill}), and extended widely to the south in both the IMERG data and the WRF Model (Figs. 2d,e). The difference in half-hourly precipitation between the two datasets was relatively small during the nighttime compared with that during the daytime, except for the overestimation in the E_{foothill} area (Fig. 2f). As a result, the spatial correlation for precipitation between the IMERG and WRF output was not high from 1300 to 1900 LT (correlation coefficient; CORR = 0.26–0.34), and the root-mean-square error (RMSE) was 0.17 – $0.24 \text{ mm (30 min)}^{-1}$ for the same period, and in particular, it exceeded 0.2 between 1330 and 1730 LT (Fig. S1 in the online supplemental material). Meanwhile, from 0000 to 1030 LT, the spatial correlation was higher (CORR = 0.70–0.80) and the RSME was lower [0.11 – $0.17 \text{ mm (30 min)}^{-1}$] than in the daytime. These results indicate that the simulated nocturnal precipitation was more similar with the IMERG precipitation than that in the daytime although its difference was slightly large over the eastern foothill region.

We focus here on the area with higher nocturnal precipitation shown in Fig. 2, and selected four subregions for further analysis; i.e., two where nocturnal precipitation dominates (W_{slope} and E_{foothill} ; Figs. 1a and 2d,e) and two neighboring regions for comparison (W_{foothill} and E_{slope} ; Fig. 1a). The

climatological diurnal variation in the area-averaged precipitation during July and August over our 8-yr study period are shown in Fig. 3. Over both foothill regions (Figs. 3a,b), the area-averaged precipitation was high between night and morning, and the peak occurred at 0400 LT (0400 LT) in the IMERG data and 0700 LT (0500 LT) in the WRF Model over the W_{foothill} (E_{foothill}) area. In the WRF Model, precipitation tended to last longer by a few hours. The simulated precipitation amount was comparable with the IMERG data over the W_{foothill} area, but was 1.4–2.1 times stronger than the IMERG data over the E_{foothill} area, which is consistent with the result shown in Fig. 2f. Meanwhile, the simulated precipitation over the W_{slope} area (Fig. 3c) increased between 1200 and 1500 LT, then remained almost the same amount until 2200 LT, and reached a maximum at 0100 LT on the following day. The IMERG precipitation was approximately half of the simulated precipitation over the W_{slope} area during the afternoon, and its increase lasted until 1700 LT. Although we found these slight differences of the diurnal variation in precipitation over the W_{slope} area between the two datasets, they were qualitatively consistent. Over the E_{slope} area (Fig. 3d), the precipitation maximum at 1500 LT was shown in the model due to the increase in the daytime precipitation as shown in Fig. 2b, which was not captured in the IMERG datasets. The simulated precipitation increased again during the night and reached the maximum at the midnight as well as the IMERG precipitation.

Over the Himalayas and its surrounding regions, the precipitation amounts simulated using the WRF Model tended to exceed those recorded in the IMERG dataset, particularly over the slopes during the daytime and over the E_{foothill} area in the nighttime. Although the WRF Model may have a high sensitivity for precipitation over the mountainous regions, we should discuss characteristics of the IMERG precipitation

FIG. 2. Spatial distribution of climatological half-hourly precipitation in (a),(d) the IMERG and (b),(e) the WRF simulation at 1500 and 0300 LT during the analysis period. (c),(f) The anomalies (WRF minus IMERG) are also shown. Thin black contours indicate elevations of 1000 and 3000 m. The W_{slope} , W_{foothill} , E_{slope} , and E_{foothill} areas also shown in each figure.

carefully. According to the previous studies, convection is shallow over the Himalayan slopes during the daytime (Kurosaki and Kimura 2002), and shallow convection embedded in stratiform clouds generates increased precipitation around midnight over the slopes and foothills (Lang and Barros 2002; Barros et al. 2004). The passive microwave sensor underestimates precipitation from shallow and warm convection because the retrieval algorithm assumes that intense

precipitation results from deep convection containing ice particles (Shige et al. 2013; O and Kirstetter 2018), and it is also discussed in more detail by Fujinami et al. (2021). Therefore, precipitation amounts in the IMERG data will be less than the ground truth. In particular, the afternoon precipitation maximum over the Himalayan ridges was observed by the rain gauge and analyzed in the TRMM PR dataset (Barros and Lang 2003; Shrestha and Deshar 2014; Ouyang et al.

FIG. 3. Diurnal variations in the climatological half-hourly precipitation averaged over the (a) W_{slope} , (b) E_{slope} , (c) W_{foothill} , and (d) E_{foothill} areas during July and August over the 8 years from 2003 to 2010. The half-hourly precipitation in the IMERG data (WRF Model) is shown by blue and gray lines (red and black lines), respectively.

2019; Fujinami et al. 2021), which was captured in the WRF Model but not in the IMERG datasets (Figs. 2 and 3). In addition, the $0.1^\circ \times 0.1^\circ$ resolution of the IMERG data better captures the wider-spreading precipitation caused by stratiform cloud with shallow convection during the nighttime than the scattered daytime precipitation over the Himalayan slopes, the distribution of which is strongly associated with the complex mountain ridge–valley orography (Bhatt and Nakamura 2005). From this evidence, we suggest that the WRF Model simulated the spatial distribution of precipitation and its area-averaged diurnal cycle over the Himalayas reasonably well. The output of the model experiments with a 2-km horizontal resolution were found to be appropriate to diagnose the physical processes associated with the nocturnal precipitation occurrence.

b. Diagnosis of the nocturnal precipitation occurrence over the Himalayan slopes and foothills

1) DETECTION OF NOCTURNAL PRECIPITATION EVENTS

In the WRF Model, a single precipitation peak was simulated during the early morning over the Himalayan foothills, and the increase in precipitation during both the daytime and nighttime occurred over the Himalayan slopes as described in section 3a. These characteristics of the simulated diurnal precipitation pattern matched those seen in the satellite estimations and the previous observational studies. In this section, we will examine how nocturnal precipitation over the Himalayan slopes and foothills is generated from the viewpoint of both synoptic-scale and local-scale processes. We selected the events to be analyzed using the original procedure described below. Note that, hereafter, we define a day as running from 1200 LT on the first day to 1200 LT on the following day.

First, to define the time that the nocturnal precipitation occurred, we normalized the area-averaged climatological diurnal variations in the simulated precipitation over the W_{slope} , W_{foothill} , E_{slope} , and E_{foothill} areas (as shown in Fig. 3) using the following equations:

$$X_{\text{mean}} = \frac{1}{n} \sum_{i=1}^n X_i,$$

$$\sigma_i = \sqrt{\frac{1}{n-1} \sum_{i=1}^n (X_i - X_{\text{mean}})^2},$$

$$Y_i = \frac{X_i - X_{\text{mean}}}{\sigma_i} \quad (i = 1, 2, 3, \dots, n).$$

Here Y_i is the normalized climatological half-hourly precipitation, X_i is the area-averaged climatological half-hourly precipitation, X_{mean} is the area-averaged daily mean precipitation, σ_i is the area-averaged standard deviation for the climatological half-hourly precipitation, and n represents the number of half-hour steps in a day (i.e., $n = 48$). The diurnal variations in the normalized precipitation (Y_i) showed that, between 2300 and 0900 LT, mean precipitation exceeded one standard deviation ($Y_i > 1$) in at least one of

the four focused areas; i.e., W_{slope} , E_{slope} , W_{foothill} , or E_{foothill} (Fig. 4). Hence, we defined nocturnal precipitation as precipitation occurring between 2300 and 0900 LT. We excluded the E_{slope} area from our analysis because daytime precipitation was more active than nighttime precipitation in the WRF Model. Next, for each half hour of the original precipitation time series, a *candidate* nocturnal precipitation event was identified during which the half-hourly precipitation exceeded 3.0 mm between 2300 and 0900 LT, and its horizontal extent was greater than or equal to 400 grids (i.e., 1600 km²). We determined the thresholds applied to the precipitation amounts and size using an iterative approach. The result was that if the candidate nocturnal precipitation event covered over 80% of an area (i.e., W_{slope} , W_{foothill} , or E_{foothill} area) at least once a day, then that date was selected for analysis. This procedure identified 119 (proportion of selected days relative to the total number of days in the studied period = 24%), 103 (21%), and 202 (41%) days as having nocturnal precipitation events (NP events) in the W_{slope} , W_{foothill} , and E_{foothill} areas, respectively. Because the number of NP events over the E_{foothill} area was too large, we slightly increased the half-hourly precipitation threshold to 5.0 mm. This left 142 (29%) NP events in the E_{foothill} area, and these were used in our analysis.

The spatial distributions of precipitation anomalies; i.e., the composite of the NP events minus climatology, are shown in Fig. 5 to demonstrate the effectiveness of our event detection method. In the composite of the NP events for the W_{slope} area, the simulated precipitation increased significantly over the mountain slopes and a basin-like area in 27.5° – 28.5° N and 83° – 85° E at 0000 LT, relative to that in the climatology (Fig. 5a). Meanwhile, a positive precipitation anomaly and its southward extension were significant over the western and eastern side of the Himalayan foothills at 0600 LT in the composite of the NP events for the W_{foothill} and E_{foothill} areas (Figs. 5b,c). The spatial distribution of precipitation anomalies was consistent with the IMERG data across all three

FIG. 4. Normalized diurnal variations in the climatology of the simulated half-hourly precipitation averaged over the W_{slope} , W_{foothill} , E_{slope} , and E_{foothill} areas (black solid, red solid, black dashed, and red dashed lines, respectively). The blue rectangle indicates the time of the nocturnal precipitation occurrence defined in section 3b.

FIG. 5. Spatial distributions of precipitation anomalies (composites for the NP events minus climatology) for the (a),(d) W_{slope} ; (b),(e) W_{foothill} ; and (c),(f) E_{foothill} areas from the WRF simulation and the IMERG data. To consider the peak of diurnal variations in precipitation, 0000 LT was selected for (a) and (d) and 0600 LT for (b), (c), (e), and (f). Hatching indicates statistical significance at the 90% confidence level. Thin black contours indicate elevations of 1000 and 3000 m. The W_{slope} , W_{foothill} , E_{slope} , and E_{foothill} areas also presented in each figure.

areas, although its absolute value was higher in the model than in the IMERG data (Figs. 5d–f). The simulated precipitation anomalies in other time between the night and the morning were also similar to those in the IMERG data (not shown). Note that the accumulated precipitation amounts from 2300 to 0900 LT for the NP events accounted for 25.8%, 39.1%, and 44.6% of accumulated daily precipitation during July and August over the 8 years for the W_{slope} , W_{foothill} , and E_{foothill} areas, respectively. These results suggest that the NP events are good representations of nocturnal precipitation occurrence for each area.

We will now briefly consider the intraseasonal-scale characteristics of the NP events. The daily mean precipitation averaged over each area (W_{slope} , W_{foothill} , and E_{foothill}) is shown in Fig. 6. High daily precipitation of approximately 20 mm day^{-1} occurred frequently in the W_{slope} area, but the NP events occurred randomly. Most of the NP events over the W_{slope} area happened intermittently (62 days out of 119 selected days were nonconsecutive), and we found only two episodes in which the NP events occurred on more than four consecutive days. Meanwhile, the pattern of the NP events over the W_{foothill} and E_{foothill} areas differed from that over the W_{slope} area, as in these two areas they tended to occur consecutively. Periods with more than four consecutive NP days occurred 10 times in both W_{foothill} and E_{foothill} areas, with individual NP

events being recorded on 41/103 days for W_{foothill} and 32/142 days for E_{foothill} . Figures 6b and 6c also suggest intraseasonal oscillations (ISOs) with submonthly-scale in the occurrence of NP events over both foothill regions, which will be discussed in the next section.

2) SYNOPTIC-SCALE MOISTURE FLOW AND ITS LOCAL-SCALE CONVERGENCE CAUSING NOCTURNAL PRECIPITATION

To find evidence of moisture transport at the synoptic scale in the case with the NP events occurrence, we conducted composite analysis using the ERA5 data. The composite of the synoptic-scale geopotential height and water vapor flux at 900, 700, and 500 hPa for the NP events at 0000 LT are shown for each area in Figs. 7 and 8. When nocturnal precipitation occurred over the W_{slope} area, a cyclonic circulation developed over the northwest of the BoB in the lower and midtroposphere, suggesting the presence of low pressure systems (Fig. 7). The center of low pressure system (marked by X in Fig. 7) tilts slightly southeastwards with height, and its location at 500 hPa correspond to the deep convection, as indicated by the low OLR (blue hatching in Fig. 7c). Water vapor was transported along the northeastern and northern side of this cyclonic circulation. Indeed, according to our subjective

FIG. 6. Simulated daily precipitation averaged over the (a) W_{slope} , (b) W_{foothill} , and (c) E_{foothill} areas during the analysis period. Red bars indicate NP events and blue bars other events.

estimation using the geopotential height and moisture flux at 700 hPa by eyes, a low pressure system existed between south and southeast of the W_{slope} area in 67% of the NP events and contributed the westward and northwestward transport of moisture there. Our results are consistent with several previous studies which indicated that the low pressure systems

generated over the BoB, and their westward movement, contribute to precipitation occurrence over the Himalayan slopes (Barros and Lang 2003; Ueno et al. 2008; Vellore et al. 2016).

The synoptic-scale circulation pattern was similar for the NP events in the W_{foothill} and E_{foothill} areas (Fig. 8). The area with lower geopotential height (below 940 gpm) at 900 hPa

FIG. 7. Composites of synoptic-scale geopotential height (shading) and water vapor fluxes (vectors) at (a) 900, (b) 700, and (c) 500 hPa for NP events over the W_{slope} area. The blue hatched area in (c) indicates OLR less than 200 W m^{-2} . Topography below 1000 and 3000 m is shaded gray in (a) and (b). The thin contours in (c) indicate an elevation of 3000 m. The X marks are approximate centers of low pressure systems. The black square indicates the W_{slope} area.

was limited to only over the northern Indian subcontinent, i.e., south of the Himalayas, in those two cases (Figs. 8a,d). As a result, both the W_{foothill} and E_{foothill} areas received moisture directly from the monsoon westerlies, which turn to southerlies near the Himalayas. The northwesterly flow along the western Himalayas also supplied water vapor to the W_{foothill} area at 700 and 500 hPa (Figs. 8b,c), while the southerlies and southeasterlies were dominant over the E_{foothill} area in the layer between 900 and 500 hPa (Figs. 8d–f). The daily OLR showed that there was deep convection over the W_{foothill} and E_{foothill} areas, which is consistent with the higher percentages of convective rain over the Himalayan foothills as shown in Romatschke and Houze (2011).

The differences in synoptic-scale circulation patterns between W_{slope} and W_{foothill} or E_{foothill} relates with active and break phases of the ISOs with submonthly scale exhibiting over northeast India, Bangladesh, Nepal and its surrounding regions (Fujinami et al. 2014, 2017). The low pressure systems over the northwest of the BoB are captured in the break phase of the ISOs in above regions, whereas we find the southwesterly flows in the lower troposphere and the increase in precipitation amount over the Himalayan foothills in the active phase of it. The large-scale precipitation patterns for the NP events over each area (Fig. S2) also correspond to the active and break phases of that kind of ISOs and seem to be opposite of the ISOs shown in the Indian monsoon activity

(e.g., Krishnamurthy and Shukla 2007). More careful investigation would be required for the analysis of the ISOs in the Himalayan precipitation in future work, but this is beyond the scope of the present study.

The WRF Model would provide better understanding of how synoptic-scale moisture flow converged with topography and/or local-scale flows to produce nocturnal precipitation and maintain it; therefore, we diagnosed the local-scale atmospheric conditions over W_{slope} , W_{foothill} , and E_{foothill} areas (corresponding to the square regions in Figs. 7 and 8). For NP events that occurred over the W_{slope} area, precipitation appeared in the afternoon along the mountain ridges in the Himalayas where the altitude was approximately 2000–2500 m (Fig. 9a). At this time, the wind at 10 m above the surface was relatively strong during the day (Fig. 9e), and its spatial distribution was similar to the vertically integrated moisture fluxes from surface to 100 hPa. That is, southerly and southeasterly winds transported the water vapor to the Himalayan slopes and the Tibetan Plateau across the lower-altitude mountains in front of the Himalayas. Observational studies in this region suggest that the combination of a synoptic-scale southerly flow and local-scale upslope winds associated with the mountain topography effectively transport water vapor toward the Himalayan mountains during the afternoon (e.g., Barros and Lang 2003). The WRF Model was able to simulate this type of large-scale moisture flow.

FIG. 8. As in Fig. 7, but for NP events over the (a)–(c) W_{foothill} and (d)–(f) E_{foothill} area.

FIG. 9. Composites of the half-hourly precipitation (shading) and vertically integrated moisture fluxes (vectors) for NP events over the W_{slope} area at (a) 1500, (b) 2100, (c) 0000, and (d) 0600 LT. (e)–(h) As in (a)–(d), but for the anomaly of precipitable water between the composite and climatology (shading) and wind at 10 m above the surface (vector). Thick black contours indicate elevations of 1000, 2000, and 3000 m, thin black contours are elevation of 1500 and 2500 m, and red lines are elevations of 250 and 500 m.

Precipitation over the Himalayan slopes increased gradually from the evening onward, and the precipitation area extended southward over the Himalayan slopes (Figs. 9b,c). Then, the precipitation over the lower-altitude mountains in front of the Himalayas, at approximately 1000 m and along 27.8°N, persisted until the early morning (Fig. 9d). A northerly and northeasterly wind appeared in the near-surface layer over the Himalayas and to the south during the night and morning (Figs. 9f–h), while the vertically integrated moisture flux indicated a northward nocturnal moisture supply toward the Himalayan mountains as well as the daytime (Figs. 9b–d), which was also captured at the synoptic scale over the northern and northeastern side of the low pressure system as shown in Fig. 7. According to the time series of the vertically integrated hydrometeors (i.e., the sum of cloud water, cloud ice, rain, snow, and graupel from the surface to 100 hPa) and surface wind speed with a northerly component along 84.0°E (Fig. 10a), the hydrometeors increased after 1600 LT between 27.8° and 28.5°N, and after 2000 LT to the south of 27.8°N, which is consistent with the expansion of precipitation over the Himalayan slopes and the lower-elevation mountains in front of the Himalayas. Cloud development corresponded to

the occurrence of surface winds with a northerly component. The northerly surface wind appeared over the mountain slopes with negative net radiation at the ground (Fig. 10b), suggesting that this northerly surface wind was downslope wind caused by radiative cooling at the surface of the slopes. The vertically integrated moisture flux intensified south of 28.2°N after 1900 LT (Fig. 10c), and this is similar to the development of a synoptic-scale nocturnal low-level jet described by Fujinami et al. (2017, 2021). The convergence of this enhanced synoptic-scale moisture flow with the local-scale downslope wind promoted an increase in hydrometeors (Fig. 10), precipitable water (the vertically integrated water vapor from surface to 100 hPa; Figs. 9f–h), and precipitation (Figs. 9b–d), as well as the development of orographic precipitation along the mountain ridges. Although radiative cooling was suppressed by the cloud cover over time, the downslope wind persisted until the morning. The evaporative cooling caused by hydrometeors simulated below 800 hPa between 28° and 28.4°N (Fig. S3a), and below 900 hPa between 27.2° and 27.8°N (Fig. S3b), which was particularly enhanced in association with the cloud development after 2100 LT, suggests an intensification of the gravity current over the mountain slopes.

FIG. 10. Composites of time series for (a) wind speed with a northerly component at 10 m above the surface (shading) and vertically integrated hydrometeors (contours with 0.5-mm intervals) along 84°E for NP events over the W_{slope} area. (b) As in (a), but for net radiation at the surface (the sum of sensible heat flux, latent heat flux, and ground heat flux) by contours. The contours intervals in (b) are 50 and 5 W m^{-2} for positive (upward) and negative (downward) values, respectively. (c) As in (a), but for magnitude of vertically integrated moisture flux by shading. Left panel shows topography (black) and precipitation at 0300 LT (blue) along 84°E. In (a) and (b), wind speed with a southerly component is shaded gray. The blue line in (b) shows 0 W m^{-2} . Dashed yellow lines in (a) indicate the latitude of 27.7° and 27.9°N.

Indeed, the downslope wind along 27.7°N was accelerated and decelerated in association with the increase and decrease in the hydrometeors between 2100 and 0900 LT, which was also found along 27.9°N (yellow lines in Fig. 10a). Furthermore, condensational heating (Fig. S3) would help to form continuous upward motion and convection for several hours in the nearly saturated troposphere as a result of the decrease in temperature and increase in specific humidity caused by moist southerlies during the nighttime (not shown).

Although the basic mechanisms associated with the occurrence of nocturnal precipitation over the W_{foothill} and E_{foothill} areas were similar to those over the W_{slope} area, there were several differences as described below. In the W_{foothill} area, the moisture flux was weak throughout the day compared with the W_{slope} area (Figs. 11a,b and S4), and this can be related to the differences in their synoptic-scale circulation patterns. Therefore, the absolute values of precipitation and precipitable water were small over the mountain slopes between 27.8° and 28.5°N during the daytime and nighttime relative to the NP events over the W_{slope} area (Figs. 11c,d). Meanwhile, the convergence of the synoptic-scale moisture flow was encouraged over the area bounded by 26.5°–27.5°N and 83.5°–85°E (Fig. 11b), which corresponds to the monsoon westerlies turning to southerlies near the Himalayas, as shown in Fig. 8. This effectively increased the precipitable water content and precipitation over W_{foothill} .

For NP events over the E_{foothill} area, precipitation developed along the mountain ridges during the afternoon, as it did in W_{slope} and W_{foothill} , and the nocturnal precipitation occurred zonally along 26.6°N (Figs. 11e,f). Although the synoptic-scale circulation pattern was similar for the NP events in the W_{foothill} and E_{foothill} areas as shown in Fig. 8, the southerly moisture flow did not turn to westerlies clearly in front of the Himalayas over the E_{foothill} area. Instead,

the wind with a northerly component, i.e., the downslope wind, was generated over the mountain slopes at approximately 1800 LT in association with the radiative cooling (Figs. 11g,h and 12a,b), which converged with the southerlies in the near-surface layer over the foothill regions (Fig. 11h), such as the NP events over the W_{slope} area. The southerly moisture fluxes were intensified during the nighttime (Fig. 12c), and we found the increase in hydrometeors (Figs. 12a,c) and precipitation (Fig. 11f) in the early morning. The condensational heating was stronger than in the W_{slope} area (not shown), and the associated lifting of moist air mass also affected the maintenance of the surface convergence zone, which allowed suitable conditions for continuous precipitation prevail.

Over the E_{foothill} area, other interesting features for the relationship between the atmospheric moisture and vertical wind were found in the midtroposphere (Fig. 13). Near-surface air mass was lifted up to the midtroposphere between 26.5° and 27°N in the early morning, which was not captured in the afternoon. In association with the nocturnal intensification of the vertical wind, atmospheric moisture increased in the lower and midtroposphere relative to that in the climatology, and moist area expanded to north particularly between 500 and 400 hPa. Since the latitude with the strong vertical wind corresponded to the precipitation occurrence (Fig. 11f) and the increase in hydrometeors (Figs. 12a,c), we implied that a convective system lifted near-surface moisture up to the midtroposphere and transported it to the Tibetan Plateau and the higher-elevation region of the Himalayas during the nighttime. This type of moisture transport process was indicated by Dong et al. (2016) over the region between the northern Indian subcontinent and the southwestern Tibetan Plateau when mesoscale convective systems developed there. Our simulation results suggest that the nocturnal precipitation system over the E_{foothill} area also affects the moisture transport process over the Himalayas and the Tibetan Plateau.

FIG. 11. As in Fig. 9, but for NP events over the (a)–(d) W_{foothill} and (e)–(h) E_{foothill} areas at 1500 and 0600 LT.

4. Discussion

a. Sensitivity to representation of topography and vertical resolution

The WRF Model, with its 2-km resolution and 50 vertical layers, was able to simulate the diurnal variations in precipitation over the Himalayan slopes and foothills. In this section, we discuss how such high-resolution settings affected the model's ability to simulate nocturnal precipitation over the Himalayas. We conducted two kinds of 1-yr sensitivity experiments, in which the topographic resolution or the number of vertical layers were changed. Other settings remained

unchanged from the experiment presented in section 2. To examine the benefits of using high-resolution topography in the original experiment, the topography with lower resolution (i.e., a 5-arc-min resolution) was regridded to a resolution of 30 s and then used in a simulation with a 2-km horizontal resolution (Fig. S5; LowTOPO experiment). For the other experiment, to understand the impact of the vertical resolution, we reduced the number of vertical layers from 50 to 40 but with a fixed top height of 25 hPa (the 40L experiment), in which the thickness of the vertical layers was coarser in the middle and upper troposphere. The summer of 2010 was selected for these sensitivity experiments

FIG. 12. As in Fig. 10, but along 89°E for NP events over the E_{foothill} area. The blue line in the left panel indicates precipitation at 0600 LT.

FIG. 13. Latitude–vertical cross sections of the anomaly of the water vapor mixing ratio between the composite and climatology (shading and contours) and the v - w wind (vectors) averaged over 88° – 90° E for NP events over the E_{foothill} area at (a) 1500 and (b) 0600 LT. Topography is shaded gray.

because nocturnal precipitation was active over the whole target area (Fig. 6). Hereafter, the original experiment (exp) for summer 2010 will be described as the CTL exp.

The difference in precipitation for the NP events between the CTL exp and the LowTOPO exp was larger over the complex mountain slopes, particularly the W_{slope} area. The precipitation expanded widely over the Himalayan slopes during the night in the CTL exp (at 0000 LT in Fig. 14a), which was consistent with the 8-yr composite shown in Fig. 9c. It was, however, limited over the area between 28.1° and 28.4° N in the LowTOPO exp for the NP events in the W_{slope} area (Fig. 14b). This difference in the spatial distribution of nocturnal precipitation was also found for the NP events in the W_{foothill} area (Fig. S6). Precipitable water tended to be

lower in the CTL exp than the LowTOPO exp because the topographic height was generally higher in the CTL exp (not shown); therefore, the higher precipitation in the CTL exp suggested that having more complex topography is beneficial for precipitation than having higher precipitable water. This is likely to be due to the effective triggering of convective initiation for nocturnal precipitation over the mountain slopes, as well as the moisture convergence between the downslope wind and synoptic-scale moisture flow [as described in section 3b(2)], and this is also presumably important for the spatial distribution of precipitation. In the W_{foothill} area, the difference in area-averaged precipitation over the foothills area 26.5° – 27.5° N and 83° – 85° E; i.e., the area dominated by nocturnal precipitation, was

FIG. 14. As in Fig. 9, but for precipitation and wind at 10 m above the surface at (a),(b) 0000 LT for NP events over the W_{slope} area and (c),(d) 0600 LT for NP events over W_{foothill} area for the CTL and the LowTOPO exps.

FIG. 15. As in Fig. 14, but at 0600 LT for the NP events over the E_{foothill} area for (a) CTL exp and (b) 40L exp.

small between the CTL exp and the LowTOPO exp, whereas the precipitation tended to occur discretely during the early morning in the LowTOPO exp (Figs. 14c,d). These sensitivity experiments should also be conducted for other years, so that a more comprehensive assessment of the altitudinal dependency of precipitation can be obtained and used to validate the robustness of our results.

On the other hand, the effect of the vertical resolution in the model on the nocturnal precipitation was found over the E_{foothill} area. Both the CTL exp and 40L exp simulated the convergence between the monsoon flow and the downslope wind from the Himalayas in the near-surface layer, although the convergence zone shifted slightly to the south in the 40L exp (Fig. 15). The nocturnal precipitation occurred along this convergence in the CTL exp, which is similar to the composite analysis of the 8-yr period as shown in Fig. 11f, whereas precipitation over the Himalayan slopes was dominant in the 40L exp. In other words, the precipitation associated with the surface convergence over the foothill region was not simulated well in the 40L exp. The vertical wind generated by the near-surface convergence was not intensified in the middle and upper troposphere if the thickness of vertical layer was coarse, which reduced the hydrometeors around 26.5°N (Fig. 16). As a result, the moisture flow excessively reached the Himalayan slopes and caused the precipitation there in the 40L exp, which was unrealistic because in reality (IMERG data; e.g., Figs. 2d and 5f) the location of the maximum nocturnal precipitation anomaly is over the foothills region.

b. One-by-one comparison of precipitation between the rain gauges and the simulations

The diurnal cycle in area-averaged precipitation and its difference between the Himalayan slopes and foothills are

simulated reasonably well in the model as described in section 3a. The 2-km horizontal resolution experiments of our study also allow a direct comparison with the hourly precipitation

FIG. 16. Latitude-vertical cross section along 89°E for the vertical wind (shading) and the hydrometeors (contour interval of 0.1 g kg^{-1}) at 0000 LT for the (a) CTL exp and (b) 40L exp. Topography is shaded gray.

data recorded by the rain gauges. We obtained the rain gauge observations over the Himalayan slopes between 86.3° and 87.3°E (Fig. 1b) for the years 2003 to 2009 from the CEOP_AP website. To understand what the 2-km-resolution model can and cannot simulate with respect to precipitation, the diurnal variations in the simulated precipitation were compared with the CEOP_AP rain gauge observations, although the gauges were located between $W_{\text{slope/foothill}}$ and $E_{\text{slope/foothill}}$ areas.

For the Pyramid and Pheriche observation sites, precipitation amounts were simulated relatively well between 1200 and 1800 LT (Figs. 17a,b). In both the gauge observation and the model experiment, two peaks in precipitation were found at the Pyramid site in the evening and around midnight (at 1800 and 0000 LT in the observation and at 1700 and 0200 LT in the model, respectively), although the simulated nocturnal precipitation was lower than the observation (Fig. 17a). At the Pheriche site, the afternoon maximum in precipitation was unclear in the observation, whereas the double peaks in precipitation were clearly simulated in the WRF Model (Fig. 17b). The characteristic of diurnal cycle in precipitation at Namche was similar with that in the Pheriche site both in the observation and the simulation (Fig. 17c), and the simulated precipitation was quite low at Lukula throughout day comparing with the gauge observation (Fig. 17d). Precipitation at the lower elevation sites was underestimated because the occasional, but extremely intense and short-lived precipitation during the night was not captured even though we used the 2-km-resolution model (not shown). This underestimation of

precipitation at lower elevations meant that the model also found it difficult to simulate the altitude-dependent decrease in nocturnal precipitation along the specific Himalayan valleys between 86.3° and 87.3°E (also see Fig. S7 with same range y axis for each site). The diurnal cycle in observed precipitation was also not captured in the IMERG datasets particularly in the higher-elevation sites (Fig. S7) since the $0.1^\circ \times 0.1^\circ$ resolution could not represent heterogeneity of the precipitation patterns affected by the complex mountain–valley topography. We suggest that additional improvements might be required to do the one-by-one evaluation of precipitation between the rain gauge observation and the model experiment.

5. Conclusions

In this study, we conducted experiments using a cloud-resolving model to improve our understanding of the occurrence of nocturnal precipitation over the Himalayan slopes and foothills during the mature monsoon season from 2003 to 2010. The model was able to simulate the diurnal variations in precipitation, and its spatial distribution, reasonably well. The area-averaged precipitation has a peak over the Himalayan foothills between the night and morning and increases twice a day over the Himalayan slopes in the afternoon and around midnight.

We selected NP events over the W_{slope} , W_{foothill} , and E_{foothill} areas to examine the synoptic-scale moisture flows and its local-scale convergence. For NP events over the

FIG. 17. Composite of hourly precipitation observed at the (a) Pyramid, (b) Pheriche, (c) Namche, and (d) Lukla sites during July and August from 2003 to 2009 (black bars). Blue bars are same as black bars but for the climatology of the simulated hourly precipitation from 2003 to 2010 interpolated to the latitude and longitude at the observation site.

W_{slope} area, a low pressure system was formed over the northwest BoB, which effectively transported water vapor by southeasterlies to the Himalayan slopes. Meanwhile, the large-scale monsoon westerlies turned to southerlies near the Himalayas and directly supplied moisture to the foothills when the NP events occurred over the W_{foothill} and E_{foothill} areas. The model outputs suggested that the convergence between the synoptic-scale moisture flow and topography or local-scale flows was the cause of the NP events. Over the W_{slope} area, the daytime precipitation appeared along the ridges at 2000–2500 m over the Himalayas, and the southward extension of the precipitation area was found around midnight and in the early morning. During the nighttime, a downslope wind was generated by radiative cooling at the surface and was enhanced by the evaporative cooling by the hydrometeors in the near-surface layer. In addition, the intensification of the moisture fluxes associated with the synoptic-scale circulation converged with the local-scale downslope wind around the Himalayas, which increased the clouds and precipitation through the night. The condensational heating in the cloud tended to assist the development and maintenance of convection. Although the physical mechanisms associated with nocturnal precipitation occurrence in W_{foothill} and E_{foothill} areas were similar to those in the W_{slope} area, the synoptic-scale convergence of the monsoon flow controlled the area of active nocturnal precipitation over W_{foothill} . Over the E_{foothill} area, the convergence zone between the monsoon flow and local-scale downslope wind was clearer than elsewhere, which enhanced nocturnal precipitation and convection. Furthermore, we implied that the nocturnal moisture transport to the Tibetan Plateau was active in the midtroposphere associated with the convection in E_{foothill} . The topographic resolution and the number of vertical layers used in the model are critical factors for the simulation of nocturnal precipitation over the Himalayas and its surrounding regions.

The lack of observational data with which to validate model results has been a common problem in previous studies of cloud-resolving simulations. In particular, in situ measurement of precipitation is quite difficult over mountainous regions because precipitation at near-freezing temperatures comprises both solid and liquid precipitation types whose collection rates are significantly different. The intensive observation collaborating with the satellite measurement and the cloud-resolving-model simulation would help to improve our understanding of these precipitation processes over the complex mountainous regions. Therefore, we expect high-resolution modeling to play an important role in the development of high-performance models and precipitation products including in situ measurement and satellite remote sensing techniques.

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Archival and Retrieval System (MARS). The NOAA Interpolated OLR data were downloaded from the NOAA Physical Sciences Laboratory (PSL), and the IMERG V06B product was obtained from Goddard Earth Sciences Data and Information Services Center (GES DISC). Please visit the web site for the CEOP_AP: Himalayas hourly surface meteorology and radiation dataset to obtain the rain gauge observations used in this study.

Data availability statement. The ERA5 dataset analyzed during the current study is available in the RDA (<https://rda.ucar.edu/datasets/ds630.0/#description>; the web access has been removed now) and MARS (<https://apps.ecmwf.int/data-catalogues/era5/?class=ea>). The NOAA Interpolated OLR data are available from the PSL (https://psl.noaa.gov/data/gridded/data.interp_OLR.html). The IMERG V06B product is available from the GES DISC (https://disc.gsfc.nasa.gov/datasets/GPM_3IMERGHH_06/summary). The contact information and the detail of dataset for the CEOP_AP: Himalayas hourly surface meteorology and radiation dataset to obtain the rain gauge observations are provided from the EOL data archive (<https://data.eol.ucar.edu/dataset/76.113>). The WRF Model outputs simulated in this study are too large to archive or to transfer. Instead, all simulation outputs and the information needed to replicate the simulations are available from the corresponding author, Shiori Sugimoto at Japan Agency for Marine-Earth Science and Technology. We used the WRF Model version 3.9.1.1, which is available from the WRF Users Page (https://www2.mmm.ucar.edu/wrf/users/download/get_source.html).

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