

Mechanism of Daytime Strong Winds on the Northern Slopes of Himalayas, near Mount Everest: Observation and Simulation

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ABSTRACT

The seasonal variability of strong afternoon winds in a northern Himalayan valley and their relationship with the synoptic circulation were examined using in situ meteorological data from March 2006 to February 2007 and numerical simulations. Meteorological observations were focused on the lower Rongbuk valley, on the north side of the Himalayas (4270 m MSL), where a wind profile radar was available. In the monsoon season (21 May–4 October), the strong afternoon wind was southeasterly, whereas it was southwesterly in the nonmonsoon season. Numerical simulations were performed using the Weather Research and Forecasting Model to investigate the mechanism causing these afternoon strong winds. The study found that during the nonmonsoon season the strong winds are produced by downward momentum transport from the westerly winds aloft, whereas those during the monsoon season are driven by the inflow into the Arun Valley east of Mount Everest. The air in the Arun Valley was found to be colder than that of the surroundings during the daytime, and there was a horizontal pressure gradient from the Arun Valley to Qomolangma Station (QOMS), China Academy of Sciences, at the 5200-m level. This explains the formation of the strong afternoon southeasterly wind over QOMS in the monsoon season. In the nonmonsoon season, the colder air from Arun Valley is confined below the ridge by westerly winds associated with the subtropical jet.

1. Introduction

The climate at high altitudes in the Himalayas is believed to be strongly influenced by the subtropical westerly jet (STJ), and the seasonal meridional translation of the jet stream signals the transition of the natural synoptic season (Schiemann et al. 2009; Bonasoni et al. 2010; Li et al. 2011). The onset of the South Asian summer monsoon (SASM) coincides with the shifting of the STJ from the south side to the north side of the Tibetan Plateau (Yanai and Wu 2006; Wu and Zhang 1998). Observations on a glacier on Mount Shishapangma in the Himalayas (6900 m MSL) revealed that the weather in this high-

altitude area was strongly influenced by the STJ outside the monsoon season (Li et al. 2011). The northward shift of the STJ is sudden and abrupt, owing to significant inhibition that is primarily caused by the height of the Himalayas, which tend to block and split the westerlies (Allaby 2010). The response of the weather—specifically, the response of local circulations to this process—is not fully understood.

Diurnal wind circulation patterns, which are thermally driven and aligned with the axis of the valley, are common in mountain valleys. Up-valley winds generally occur during the daytime and down-valley winds during nighttime (Whiteman 2000). Well-developed valley winds have been observed on the southern slopes of the Himalayas. Using continuous data over a full year, Ohata et al. (1981)

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found a very persistent valley circulation system in the Khumbu Valley (on the south side of Mount Everest). Here, the measured daytime up-valley wind speed reached 8.5 m s^{-1} in May and June, while the down-valley wind speeds were only 3.5 m s^{-1} , on average. In the same area, Bonasoni et al. (2010) identified a clear mountain wind regime during nonmonsoon months, with a strong daytime southerly up-valley breeze and a reversed nighttime flow; during the monsoon season, the southerly wind was strengthened in the daytime, and even observed at night.

Besides the typical diurnal valley winds, other types of strong winds have been described both on the north and south sides of the Himalayas. An example is provided by Egger et al. (2000) in Kali Gandaki, a very deep valley located in the Nepal Himalayas, where the southerly wind speed can reach 14 m s^{-1} . Zängl et al. (2001) found that the strength of these up-valley winds is enhanced by a heat low over the Tibetan Plateau (TP), and they speculated that this heat low is sufficiently strong to draw these winds over to the north side of the Himalayas. Strong daytime winds have also been observed on the north side, at QOMS (28.36°N , 86.95°E ; 4270 m MSL), in particular during the SASM in the afternoon (Sun et al. 2007). This persistent afternoon wind was characterized by a stable speed of about 9 m s^{-1} on average, a southeasterly direction, and a depth of 500–700 m (Sun et al. 2007). The southeasterly wind at QOMS mostly occurred during the monsoon season and had a strong statistical relationship with the weakening of the STJ westerly winds at high levels (Sun et al. 2017). Similar daytime strong winds were reported at the north base camp of Mount Everest in the upper Rongbuk valley (mean wind speeds of 10 m s^{-1} and vertical extent of over 1000 m) (Gao 1985; Song et al. 2007; Cai et al. 2007; Zou et al. 2008; Zhou et al. 2008; Ma et al. 2013).

The strong afternoon winds at QOMS and the north base camp cannot be attributed to the typical thermally driven diurnal valley circulation (mountain breeze), since these down-valley winds prevailed when up-valley winds would normally be expected (e.g., as observed on the southern slopes of Mount Everest). This reveals that the local circulations on the northern slopes of Mount Everest are considerably different from those on the southern slopes. Investigation of the underlying causes may therefore provide valuable insights into the response of local weather to the interaction of topography and synoptic circulation. The study takes on further importance now that biomass burning tracers from the south of Himalayas have been observed at QOMS (Cong et al. 2015).

The Arun Valley is oriented north–south and lies to the east of Mount Everest. Up-valley winds would be

expected in this valley because its characteristics are very similar to those of the Kali Gandaki valley (Egger et al. 2000). Zängl et al. (2001) speculated that, under favorable conditions, strong up-valley winds on the southern slopes of the Himalayas may migrate onto the TP, thereby influencing the local wind on the northern slopes. Ueno et al. (2008) reported that the weakening of upper westerlies may create favorable conditions for the enhancement of meridional flows crossing the Himalayas.

This paper investigates the relationship between the strong afternoon wind observed at QOMS (before and during the monsoon season) and strong up-valley winds in the Arun Valley, and analyzes the underlying mechanism and its relationship to the monsoon.

In situ observations at QOMS are used in the study. The data and observations are presented in section 2, and the data analysis is presented in section 3, focusing on the intra-annual variation of the surface wind on the northern slopes. Since wind profiler observations at a single site are insufficient to prove the hypothesized relationship, numerical simulations using the Weather Research and Forecasting (WRF) Model (Skamarock et al. 2008) are used in this study. WRF is capable of reproducing the overall temporal and vertical distribution of the wind over mountainous topography (Lu et al. 2012), and was reported to have better reproducibility of wind direction at high wind speeds over complex terrain than at low wind speeds (Jiménez and Dudhia 2012, 2013). The model configuration and the results of the numerical simulation are presented in sections 2 and 4, respectively.

2. Data and methodology

a. Observations and study site environment

On the north side of the Himalayas, QOMS is situated on the bottom of the lower Rongbuk valley to the north of Mount Everest, at an altitude of 4270 m (Fig. 1a). Around QOMS, the valley has a north–south orientation, with a flat bottom of about 1.5 km width. Ridges on both sides of the valley are of 600–900 m in height above the valley, sloping from 25° to 30° . The valley turns eastward in the north and then turns to the south, joining the Arun Valley. To the south of QOMS, three valleys converge. The southwest valley extends to the upstream Rongbuk valley; the middle valley drains a catchment to the south; and the third (from the southeast) is called Repu, originating at a mountain ridge near the Arun Valley. Arun is a large valley, crossing the Himalayan ranges to the east of Mount Everest. The mountain ridges between the Repu Valley and the Arun Valley vary in height from 800 to 1600 m relative to QOMS.

QOMS is instrumented with an automatic weather station (AWS) that measures air temperature, relative

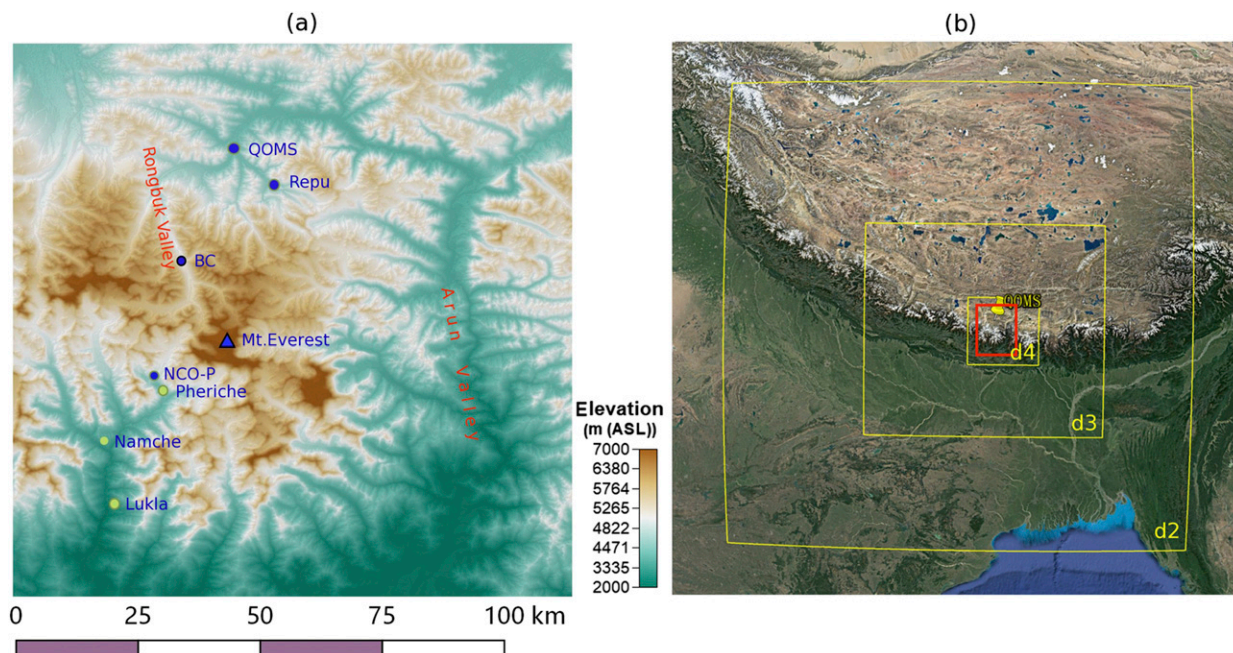


FIG. 1. (a) Topography of the Mount Everest region with locations and sites discussed in the text: Arun Valley, Rongbuk valley, QOMS, Repu Valley, north base camp (BC), NCO-P, Pheriche, Namche, and Lukla. (b) Nested WRF domains [the first domain (d1) is not shown] outlined in yellow, and the Mount Everest region outlined in red. The grid spacing for the first, second (d2), third (d3), and fourth (d4) domains is 27, 9, 3, and 1 km, respectively.

humidity, air pressure, wind speed, and direction. The AWS was set up with a PBL tower, and the measurements were taken at five levels (1.5, 3, 5, 10, and 20 m). A Vaisala, Inc., model LAP3000 1290-MHz wind profile radar (WPR) was deployed at QOMS to measure wind speed and direction in the boundary layer. The WPR was operated in two modes: low mode and high mode. In low mode, the maximum observable height is approximately 2200 m with a vertical resolution of 100 m; in high mode, it is 4000 m with a vertical resolution of 205 m.

For reference, data from the south-side observatory, the Nepal Climate Observatory–Pyramid (NCO-P), were also used. NCO-P (27.95°N, 86.82°E) is located on the south slopes of Mount Everest, at an altitude of 5079 m (Bonasoni et al. 2008; Salerno et al. 2015). AWS observations at QOMS started in September 2005, while AWS records at NCO-P are available starting in February 2006. The AWS data from both sites from March 2006 to February 2007 were used for the analysis of meteorological conditions. The measuring instruments of NCO-P are located at 5-m height, so 5-m-height data at both sites were used. The WPR at QOMS operated from May 24 to 30 June 2006 and from March 29 to 6 April 2007. WPR data for the two periods were used to describe the vertical structure and seasonal variation of low-level atmospheric flow. Additionally, AWS data of three Everest–K2–National Research

Council (Ev-K2-CNR) sites on the south side of Mount Everest—Pheriche, Namche, and Lukla (Fig. 1a)—were used to evaluate the simulation results.

b. Model configuration

The WRF Model, developed by the National Center for Atmospheric Research (NCAR), is a nonhydrostatic mesoscale meteorological model that is suitable for a broad spectrum of applications with options for physical parameterizations (Skamarock et al. 2008). For this study, the WRF Model, version 3.8.1, was used. It was configured with four two-way nested grids, at horizontal grid spacings of 27, 9, 3, and 1 km. A vertical resolution of 49 vertically stretched sigma levels was used (15 levels in the lowest 2 km; model top at 20 km). The innermost 1-km grid covered the Mount Everest region, while the outermost 27-km domain covered the whole TP. The time step of the innermost domain was set to 6 s to maintain the model's stability. The grid information and parameterizations used for this study are summarized in Table 1. Final analysis (FNL) data from the operational analysis performed every 6 h at 1° horizontal resolution at the National Centers for Environmental Prediction (NCEP 2000) were used as the initial and boundary conditions.

WRF numerical simulations were used to investigate processes responsible for the development of the

TABLE 1. WRF physics schemes used in the present study.
YSU = Yonsei University.

	Domain setting/physics scheme (option number in WRF)
Grid spacing	d1: 27 km, d2: 9 km, d3: 3 km, d4: 1 km
No. of grid points	134×128 , 190×190 , 286×256 , 253×238
Microphysics	Thompson scheme (8) (Thompson et al. 2008)
Cumulus	Kain–Fritsch (1) (none for d4) (Kain and Fritsch 1993)
PBL	YSU (1) (Hong et al. 2006)
Surface layer	MM5 similarity (1) (Fairall et al. 2003)
Land surface	Noah (2) (Chen and Dudhia 2001)
Radiation	RRTM for longwave radiation (4) (Mlawer et al. 1997) Dudhia for shortwave radiation (1) (Dudhia 1989)

afternoon southeasterly wind at QOMS in the monsoon season. For comparison afternoon winds in the nonmonsoon season were also simulated. Therefore, two simulations were performed: one for the period from 0000 LST 2 April to 2400 LST 4 April 2006 (S1) and another for the period from 0000 LST May 27 to 2400 LST 29 May 2006 (S2), representing nonmonsoon and monsoon seasons, respectively. The delimitation of the monsoon and nonmonsoon in 2006 is presented in section 3. To exclude spinup effects, the analysis of results was limited to the second simulation day, that is, from 3 April 0000 to 0000 LST the following day for S1 and from 28 May 0000 to 0000 LST the following day for S2. Figure 1b depicts the domains used in the simulation. Note that local solar time (LST, coordinated universal time + 6 h) is used throughout this work.

To give a visual indication of how the strong wind in the Arun Valley influences the surface wind at QOMS, passive tracers were emitted every time step to visualize the movement of air in the Arun Valley. The source region of the tracers was an area of $8 \text{ km} \times 2 \text{ km}$ in the Arun Valley, 61 km southeast of QOMS. For each time step, 100 tracer elements were introduced into the lowest three model levels of each grid box in the source region and then are advected within the domain by the wind field.

3. Observation results

a. Synoptic and local meteorology

Based on the SASM index (Wang and Fan 1999), the duration of the monsoon in 2006 was determined to be from 21 May to 4 October.

The large-scale SASM circulation has a strong impact on the local mountain wind systems in high-altitude

areas of the southern Himalayas. SASM brings southerly winds, strengthening the daytime upslope winds and weakening the nighttime downslope winds (Ueno et al. 2008; Bonasoni et al. 2010). This pattern has been used to characterize the seasonal variation of local weather conditions (Barros and Lang 2003; Ueno et al. 2008). The onset of the monsoon season on the southern slopes was defined as the period characterized by high humidity and the presence of southerly winds that continued to prevail at night at the measurement sites (Ueno et al. 2008). According to this definition, the duration of the monsoon season in 2006 on the high southern slope was determined to be from 21 May to 26 September (Bonasoni et al. 2010). This is close to the duration of SASM.

Since the distance between QOMS and NCO-P is 47.8 km, we examined whether the duration of the monsoon season was similar at the two stations. Comparable patterns were observed in both the daily average temperature and humidity (Fig. 2). The daily average air temperature markedly increased both at QOMS and NCO-P after the onset of SASM (Fig. 2a), as did the water vapor mixing ratio (Fig. 2c). The relative humidity (RH) at NCO-P abruptly reached values over 90% at the monsoon onset and maintained similarly high levels throughout the monsoon season. In contrast, the RH at QOMS increased slightly after the onset of the monsoon (Fig. 2b). The delay in the RH increase at QOMS was partially attributable to the air temperature at QOMS being higher than that at NCO-P. Another reason is that the monsoon rain had not reached QOMS by the beginning of the monsoon season; instead, only occasional rain occurred at QOMS until mid-July 2006. The increase in the water mixing vapor ratio at QOMS in mid-July was a response to the onset of the rainy season.

Considering the significant water vapor mixing ratio change at QOMS and NCO-P after the SASM onset, it can be concluded that in 2006, the monsoon weather began after the SASM onset on both the southern and northern slopes of the high-altitude Himalayas in the Mount Everest region. Therefore, SASM is used to define the onset of the monsoon season in the subsequent analysis.

b. Strong afternoon winds in the monsoon and nonmonsoon seasons at QOMS

The strong afternoon wind was the most notable phenomenon at QOMS throughout the year, and it showed clear seasonal variability. Figure 3a shows the diurnal pattern of 29-day mean wind speed and wind direction, generated with 10-min data, from 24 May to 21 June 2006. This period was in the early monsoon season with little precipitation and was chosen

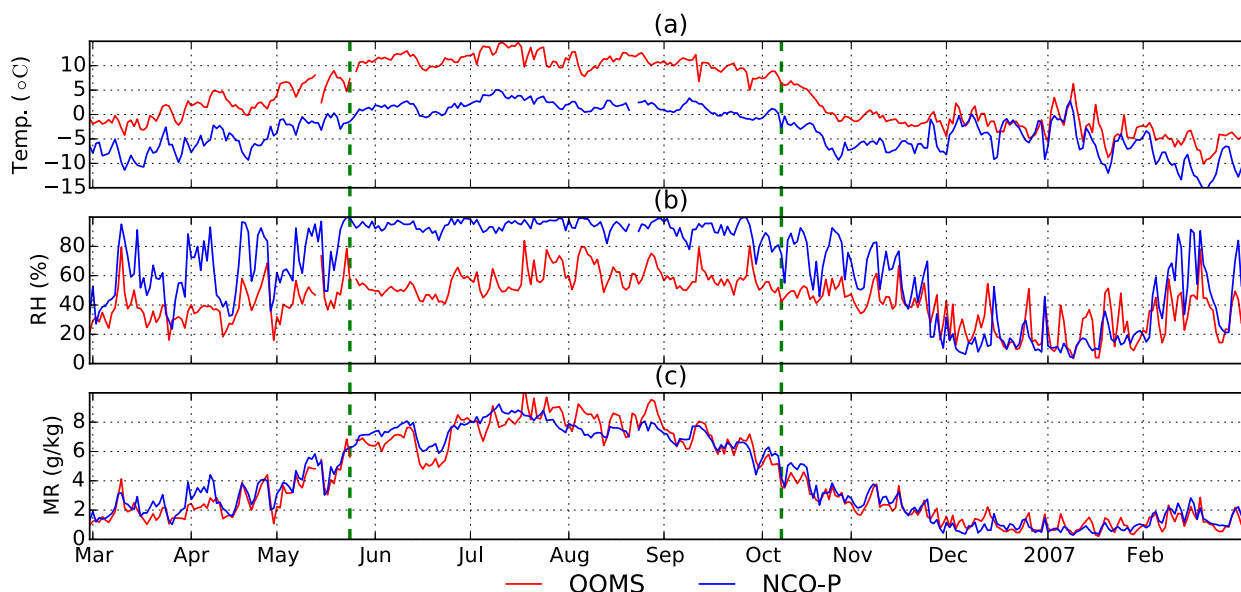


FIG. 2. Temporal development of (a) air temperature, (b) relative humidity, and (c) water vapor mixing ratios at QOMS and NCO-P, from 1 Mar 2006 to 28 Feb 2007. Monsoon onset and end dates are represented by the dashed green vertical lines.

based on the availability of WPR observations. A strong southeasterly wind developed between 1300 and 2200 LST, with a persistent direction of about 150° . It was strongest from 1400 to 1800 LST, with a mean speed of about 8.9 m s^{-1} . Observations showed little day-to-

day variability. The daily maximum speed varied between 13.4 and 20.0 m s^{-1} , with a mean value of 16.7 m s^{-1} .

A thermally driven wind was also observed. In Fig. 3a, wind variations from 0300 to 1200 LST show a typical

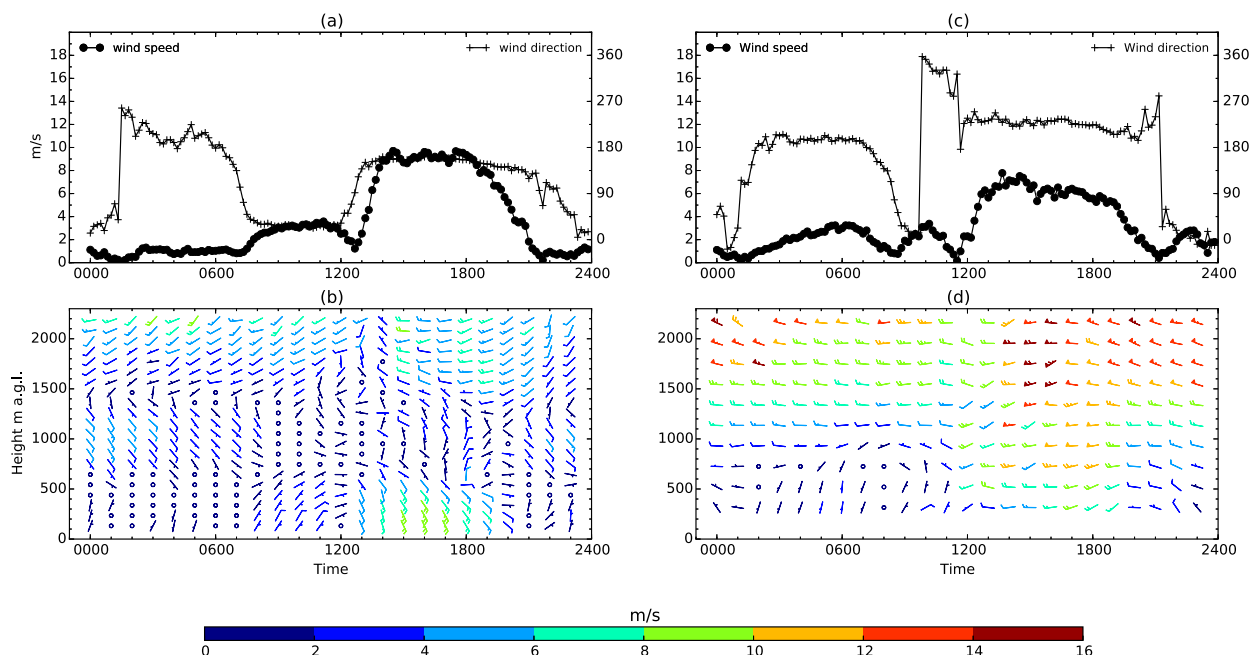


FIG. 3. Near-surface winds observed at QOMS in monsoon and nonmonsoon seasons, using both AWS and WPR data. (a) Diurnal variation of near-surface wind at QOMS, averaged from 24 May to 21 Jun 2006. (b) Wind barbs of the average wind velocity obtained by WPR for the same period as (a). A half barb is 2.5 m s^{-1} , and a full barb is 5 m s^{-1} ; a circle represents $0\text{--}2.5 \text{ m s}^{-1}$. (c) As in (a), but from 29 Mar to 6 Apr 2007. (d) As in (b), but for the same period as (c) and for a different range of heights.

valley wind system, with a nighttime southwesterly down-valley wind and a northeasterly up-valley wind after 0730 LST. The mean wind speed in the morning was about 3.8 m s^{-1} , higher than that in the nighttime (of about 1 m s^{-1}). The difference in direction between the down-valley wind and up-valley wind was about 180° .

The strong southeasterly wind occurring at QOMS from 1300 LST was apparently a different feature from that of the thermal up-valley wind and interrupted the latter. The diurnal variation of mean wind profiles (calculated using WPR data) for the same period as that in Fig. 3a is shown in Fig. 3b, presenting a southeasterly flow in the afternoon with a depth of about 600 m. Above this flow, the wind became weaker and showed a significant change in wind direction. Westerly wind was evident at heights greater than 1600 m AGL. The low-level southeasterly wind was distinct from the wind aloft as a result of the significant difference in wind direction and therefore was not caused by the downward momentum transfer from the upper strong westerly.

For comparison, the strong afternoon wind in the nonmonsoon season is also analyzed here. Observations from early April 2007 were chosen because there was no corresponding WPR operation during that period in 2006. Figure 3c shows the diurnal variation in mean wind over 11 days, generated with 10-min data, from 27 March to 6 April 2007. A strong westerly wind prevailed throughout the afternoon, peaking at 1300–1900 LST, with a direction of about 230° and mean speed of 5.2 m s^{-1} . Although the mean wind speed was not high, the daily maximum speed varied between 12.8 and 22.8 m s^{-1} . The mean speed was significantly lower than that of the southeasterly wind in the earlier summer monsoon, while the daily maximum values were close to the latter. In the nonmonsoon season, the strong wind feature had a more variable wind speed than the southeasterly wind earlier in the summer monsoon. Mean wind profiles in the same period are shown in Fig. 3d. There was a persistent westerly wind above 1000-m height throughout the day, reflecting the significant influence of the STJ on the wind in the free atmosphere. The strong wind below 1000 m during the afternoon had the same direction as the air aloft, indicating that the strong wind was caused by downward momentum transfer, in contrast to the case of the southeasterly wind in the monsoon season (Fig. 3b).

c. Annual variation of surface winds at QOMS

The dominant afternoon winds at QOMS showed significant seasonal variation, according to the AWS wind data from March 2006 to February 2007. As shown in Fig. 4a, from 1 March to mid-April 2006, the dominant afternoon wind was southwesterly, with an average

speed of 7.9 m s^{-1} . Subsequently, there was a transition period, when the surface wind direction was highly variable, with frequent southeasterly winds. After 23 May, a persistent southeasterly wind prevailed in the afternoon, strengthening before the main rainfall period and achieving a mean wind speed of 8.9 m s^{-1} . From 6 July to 10 September the wind was generally south-easterly at QOMS, but it was variable and also had periods of northerlies and easterlies. In November, the wind changed to a southwesterly direction again. The annual variation of the northward (v) and eastward (u) wind components of the surface wind are contoured in Figs. 4c and 4d, which clearly show that most of the afternoon strong winds had a southerly wind component through the year, with an easterly wind component during the monsoon and a westerly component during the remainder of the year.

The influence of the STJ on the high-altitude Himalayas has been confirmed by observations through an AWS installed on a glacier on Mount Shishapangma (6900 m MSL) in 2005 and 2006 (Li et al. 2011). The records reveal that this region was strongly influenced by the westerly jet from 10 October 2005 to 21 April 2006 and by the Indian monsoon from May to September. The STJ caused high wind speeds at the AWS site; therefore, the movement of the STJ should be considered when analyzing the seasonal transition and seasonal variability of strong afternoon winds at QOMS. The annual variation in the diurnal average u wind component (zonal wind) at 200 hPa (NCEP–NCAR reanalysis) over the Mount Everest region is plotted in Fig. 4b. Positive values represent a westerly wind.

The relationship between u wind at 200 hPa and the surface winds at QOMS was investigated. Figures 4b–d indicate good consistency between the southwesterly surface winds in the afternoon and high values (exceeding $\sim 30 \text{ m s}^{-1}$) of mean u winds at 200 hPa, suggesting that the STJ has a strong influence on the local weather at QOMS. Correspondingly, consistency was also found between the southeasterly winds in the afternoon and low values (below $\sim 30 \text{ m s}^{-1}$) of mean u winds at 200 hPa, indicating a possible relationship between the occurrence of the southeasterly surface winds and the weakening of westerly winds at high levels.

Since the variability in the dominant afternoon strong wind at QOMS is linked to the variability in the westerly wind of the STJ, such changes are an indicator of seasonal progression. The principal annual characteristics of the dominant afternoon surface wind were a southeasterly wind prevailing from mid-April to mid-November and a southwesterly wind during the remainder (from March to mid-April 2006 and from mid-November to 28 February 2007) (Fig. 4). This

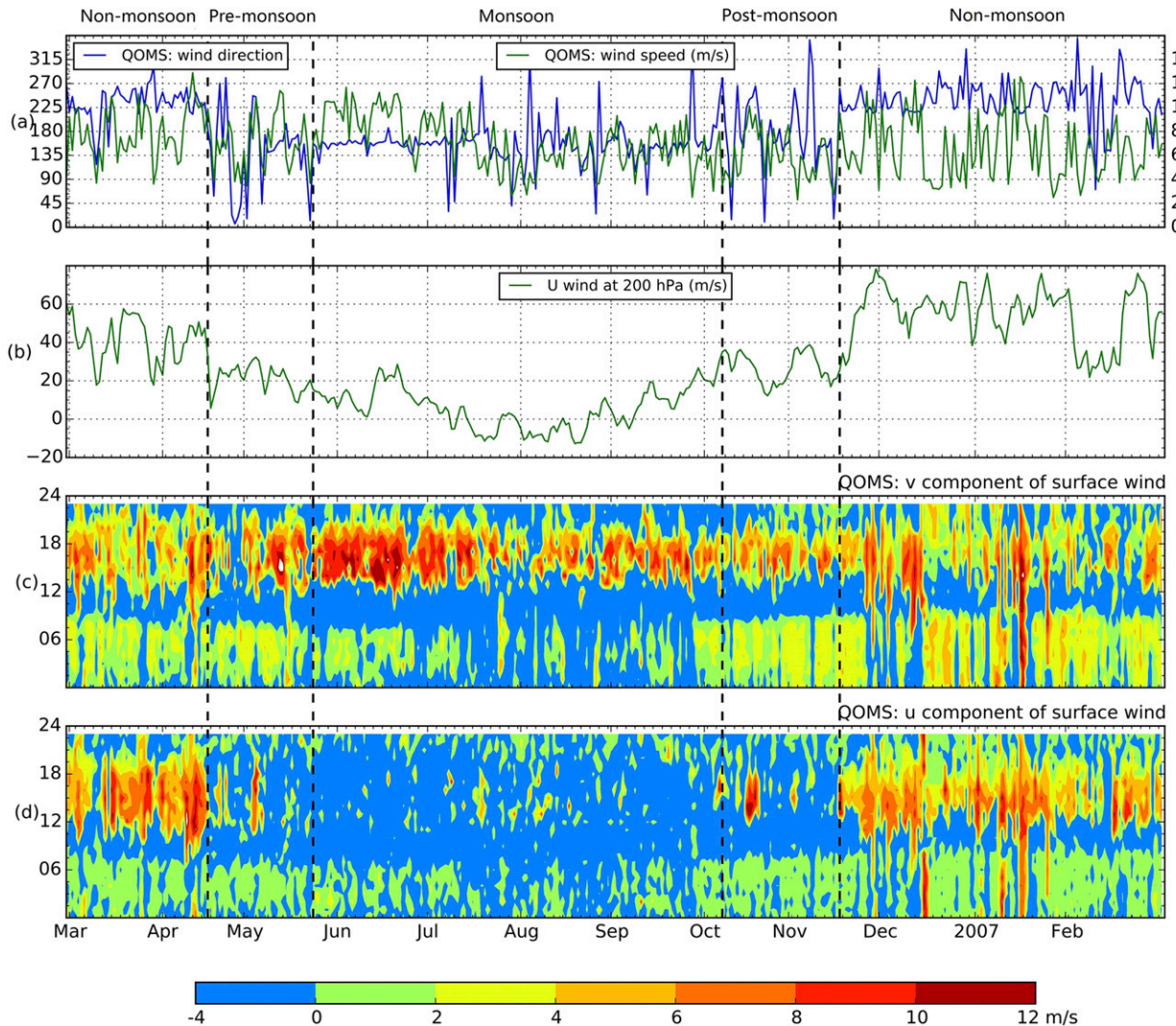


FIG. 4. (a) Annual march of mean afternoon wind averaged from 1400 to 1800 LST at QOMS. (b) Annual variations of the u component (daily mean value) at the 200-hPa level over the Mount Everest region (27.5°N , 87.5°E). (c) Annual v components of surface winds at QOMS. (d) As in (c), but for the u components. In (c) and (d), the northerly and easterly winds at QOMS were usually weak, with the absolute values mostly under 4 m s^{-1} ; thus, all negative values are given blue colors, and the positive values are given green through red colors, representing southerly winds in (c) and westerly winds in (d).

southwesterly wind coexisted with the dry and cold weather as shown in Fig. 2. Accordingly, we define this period as the nonmonsoon season at QOMS. The afternoon southeasterly wind was most persistent during the monsoon season (Figs. 4a, 4c, and 4d), and it was relatively weak and variable in direction before and after the monsoon season. These two transition periods before and after the monsoon were defined as the premonsoon and postmonsoon seasons, respectively. Table 2 shows the durations of the four seasons in the observation period.

The relationship between the seasonal transition and STJ movement was investigated. As shown in Fig. 4b, the u wind at 200 hPa was high in the nonmonsoon

season, ranging from 30 to 80 m s^{-1} with an average speed of 48 m s^{-1} , and fell to lower values in the monsoon season, with an average speed of 5 m s^{-1} . The average values in the premonsoon and postmonsoon seasons were 22 and 26 m s^{-1} , respectively. These results are consistent with the annual variations of the STJ position, as shown in Fig. 5. In the nonmonsoon season, between 80° and 90°E , the axis of the STJ was located near 27°N , directly over the southern edge of the TP (Fig. 5a); hence, a strong westerly wind belt was situated over the Himalayas, resulting in a near-surface strong westerly wind at high altitudes over the northern slopes of Mount Everest in the afternoon.

TABLE 2. Seasons in the observation period.

Seasons	Duration
Premonsoon	17 Apr–20 May 2006
Monsoon	21 May–4 Oct 2006
Postmonsoon	5 Oct–16 Nov 2006
Nonmonsoon	Until 16 Apr 2006; 17 Nov 2006 onward

In the premonsoon season, the STJ over the TP weakened notably and moved northward. The mean u wind over the Mount Everest region during this time was about 25 m s^{-1} , as compared with the speed of about 45 m s^{-1} in Fig. 5a. Correspondingly, the strong southwesterly afternoon wind at QOMS developed only occasionally and gradually came under the influence of the southeasterly wind. During the monsoon season, the STJ was located over the northern TP (Fig. 5c). The westerly u winds weakened further over the Himalayan region at 200 hPa and even easterly winds were able to develop (Fig. 4b). Under this synoptic background shown in Fig. 5c, a strong southeasterly wind prevailed in the afternoon throughout the monsoon season.

The mechanism causing the persistent southeasterly wind at QOMS during the monsoon season remains unclear. However, its consistency with the weakening of u winds at 200 hPa indicates that the STJ's shifting trajectory could be an important factor. Furthermore, considering the topography, Repu Valley is located to the southeast of QOMS, and the lowest part of the ridge between Repu Valley and the Arun Valley is only 800 m higher than QOMS. Thus, if a strong southerly daytime up-valley wind developed in the Arun Valley under a similar mechanism to that generating strong up-valley winds in other south-to-north-oriented valleys such as Kali Gandaki (Egger et al. 2000; Zängl et al. 2001), then it is reasonable to relate the wind at QOMS to the up-valley wind in the Arun Valley, east of QOMS (Fig. 1a). Once the up-valley wind has become sufficiently strong in the Arun Valley, it is possible that this flow pattern will extend as far north as QOMS, where the topography will deflect its direction to southeasterly. The northward movement of the STJ is assumed to yield suitable conditions for the development of the deep up-valley wind in the Arun Valley. However, there are only limited observations in this region; thus, numerical simulations are necessary to understand the underlying mechanism.

The southeasterly wind was strongest from the monsoon onset (21 May) to mid-July, for example, the rain-free period after monsoon onset. A 3-day simulation was performed for this period to reveal the mechanism

of the southeasterly wind, as introduced in section 2. The southeasterly winds in the early monsoon may not suitably represent those in the monsoon season as a whole; however, southeasterly winds also prevailed during days with rain but were weaker (Figs. 4c,d). Thus, the results may also be relevant to the rest of the monsoon season. Similarly, from mid-March to mid-April, southwesterly afternoon winds were typical to this region. Hence, the results of another 3-day simulation may be used to understand the mechanism of this southwesterly wind.

4. Model simulation results

a. Examination of simulation results

Simulated near-surface winds and air temperatures are compared with observations at QOMS and at three Ev-K2-CNR sites (Pheriche, Namche, and Lukla; Fig. 1a), located on the southern slopes of the Himalayas. Generally, as shown in Fig. 6, the WRF Model reproduced wind direction reasonably well at these four sites, especially during the afternoon. At QOMS, the model reproduced the strong southwesterly wind in the simulation S1 and the occurrence of the afternoon southeasterly wind in the simulation S2, as shown in Fig. 6. In S1, a westerly wind dominated from 0900 to 2100 LST, of which both the duration and direction were close to the observed wind at QOMS, but the speed was lower than observed. In S2, the simulated southeasterly wind during the afternoon was close to the observed wind in terms of both wind direction and speed, but it developed approximately 1 h later than observed. For the nocturnal winds, the simulated wind direction was significantly different from observations, especially in S1, showing a poor performance of the WRF in simulating weak flow over complex terrain, as has been reported by Jiménez and Dudhia (2012, 2013).

Near-surface temperature is important, since it governs forecasts of thermally driven flow. However, the simulated near-surface temperature was considerably lower than the observations at QOMS (Fig. 6a), partly because of the model's inability to reproduce accurate near-surface atmospheric conditions in complex terrain. Also, problems arise as a result of mismatches between the DEM and the actual terrain (Hanna and Yang 2001; Zhang et al. 2013). The underestimate of near-surface temperature at QOMS probably resulted in the underestimate of near-surface wind in the morning, but the forecast of the afternoon wind speed was not affected significantly (Fig. 6a). Thus, the model results can be used to assess the processes underlying the strong afternoon winds.

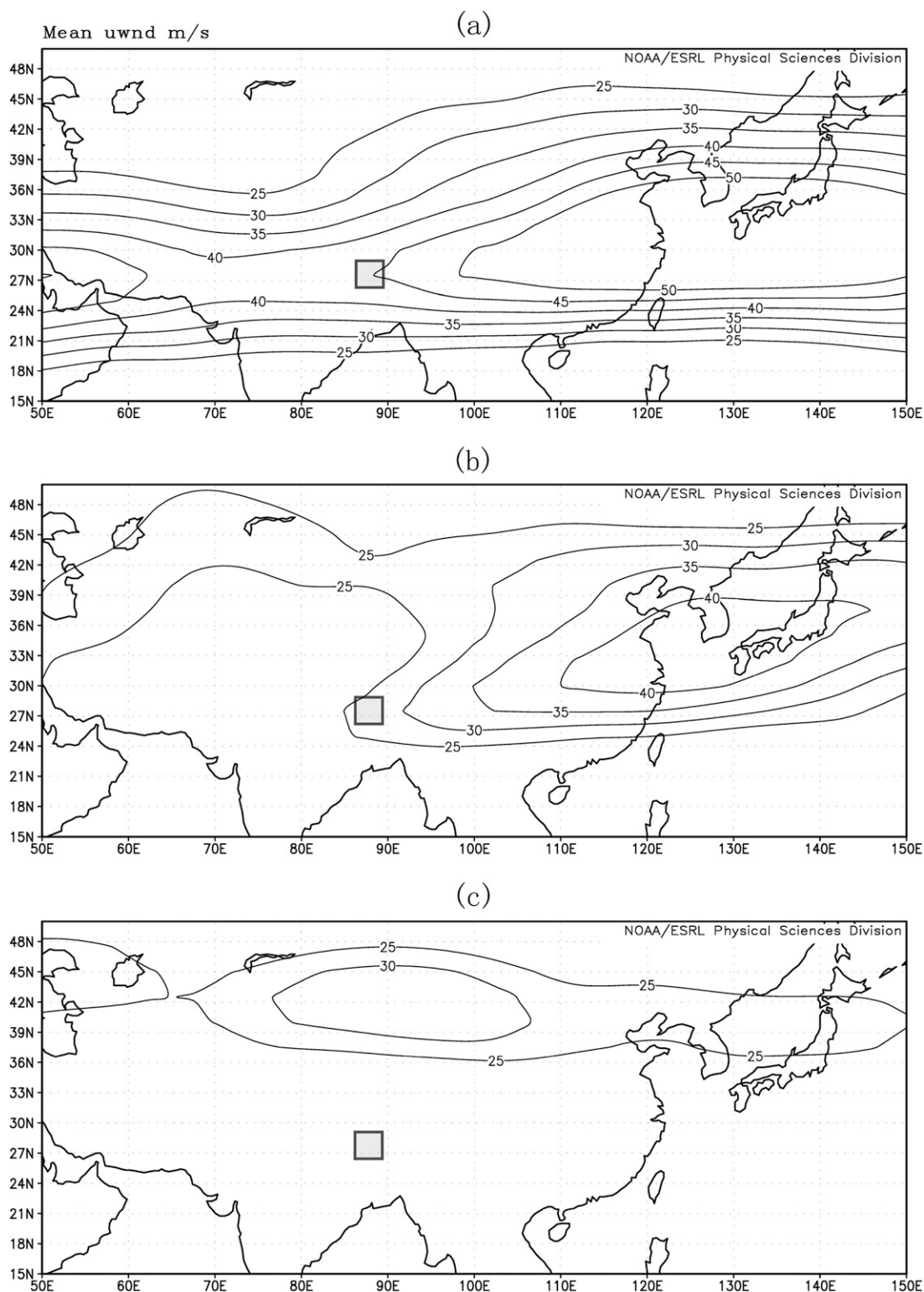


FIG. 5. Subtropical jet stream positions during (a) nonmonsoon, (b) premonsoon, and (c) monsoon seasons, indicated by contour maps of mean u winds (m s^{-1}) at the 200-hPa level (NCEP–NCAR reanalysis). The Mount Everest region is marked with the shaded rectangle.

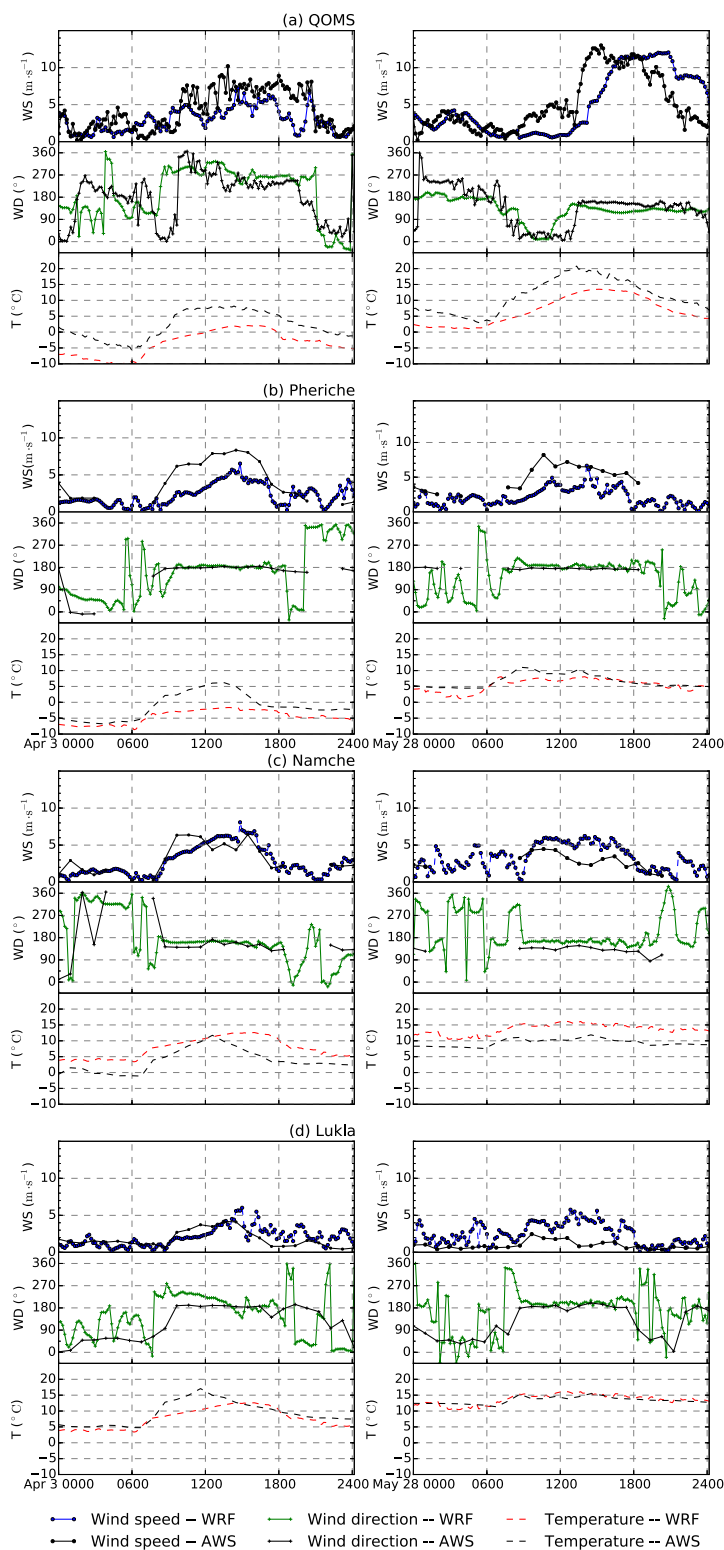


FIG. 6. Simulated and observed near-surface wind speed (m s^{-1}), wind direction ($^{\circ}$), and air temperature ($^{\circ}\text{C}$) at (a) QOMS, (b) Pheriche, (c) Namche, and (d) Lukla on (left) 3 Apr and (right) 28 May.

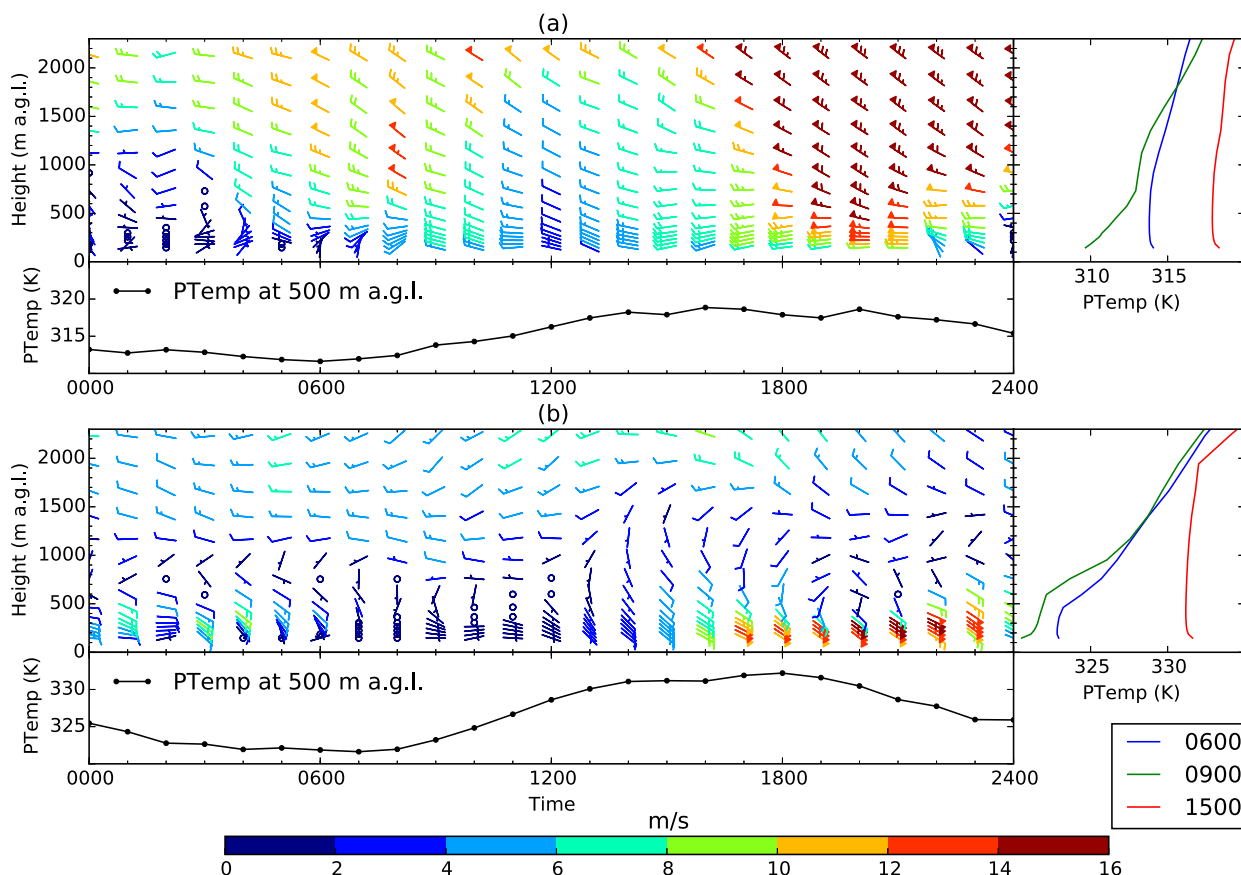


FIG. 7. Simulated diurnal variation of wind profiles, diurnal variation of potential temperature at 500 m AGL, and profiles of potential temperature at 0600, 0900, and 1500 LST at QOMS on (a) 3 Apr and (b) 28 May.

b. Analysis of simulation results

The discussion of the simulation results focuses on the strong afternoon wind at QOMS, as well as its possible relationships with the STJ and the up-valley wind in the Arun Valley.

The simulations yielded a plausible mechanism for the vertical developments of the strong afternoon winds. On 3 April, a persistent westerly wind prevailed at all levels from 0900 to 1900 LST, and the wind speed increased with height (Fig. 7a). Although there were no WPR observations for that period, this result is consistent with observations in April 2007 (Fig. 3d). At 0900 LST, there was a notable increase in the diurnal variation of potential temperature at 500 m AGL, indicating warming of the lower atmosphere as a result of ground heating after sunrise. Meanwhile, the convective boundary layer reached approximately 1000-m height at 0900 LST as determined from potential temperature profiles (Fig. 7a). At and above this level, the westerly wind prevailed. The development of the convective boundary layer then resulted in the downward transport of the

eastward momentum at high levels. On 28 May, the simulated southeasterly flow occurred at 1300 LST and lasted to 2200 LST (Fig. 7b), generally in accordance with observations (Fig. 3b).

Simulated wind fields over the Mount Everest region on 3 April and 28 May are shown at 0300 and 1500 LST in Fig. 8 (domain 4, 1-km resolution). The left and right panels in Fig. 8 show wind fields at 5200 m MSL and on the surface, respectively. In both simulations, the wind patterns at 5200 m MSL over the Arun Valley are characterized by a daytime southerly flow and a nighttime northerly flow. The winds changed directions at 0100 and 1100 LST on 3 April, and at around 0600 and 1100 LST on 28 May (not shown). At 0300 LST 3 April, strong northwesterly winds prevailed at the 5200-m level over QOMS and to its north, while the surface winds around QOMS were weak. At the same time on 28 May, northwesterly winds occurred to the north of QOMS at the 5200-m level and southeasterly winds, which correspond to the overnight southeasterly jet present at about 1000 m AGL shown in Fig. 3d, developed to the south (Fig. 8c).

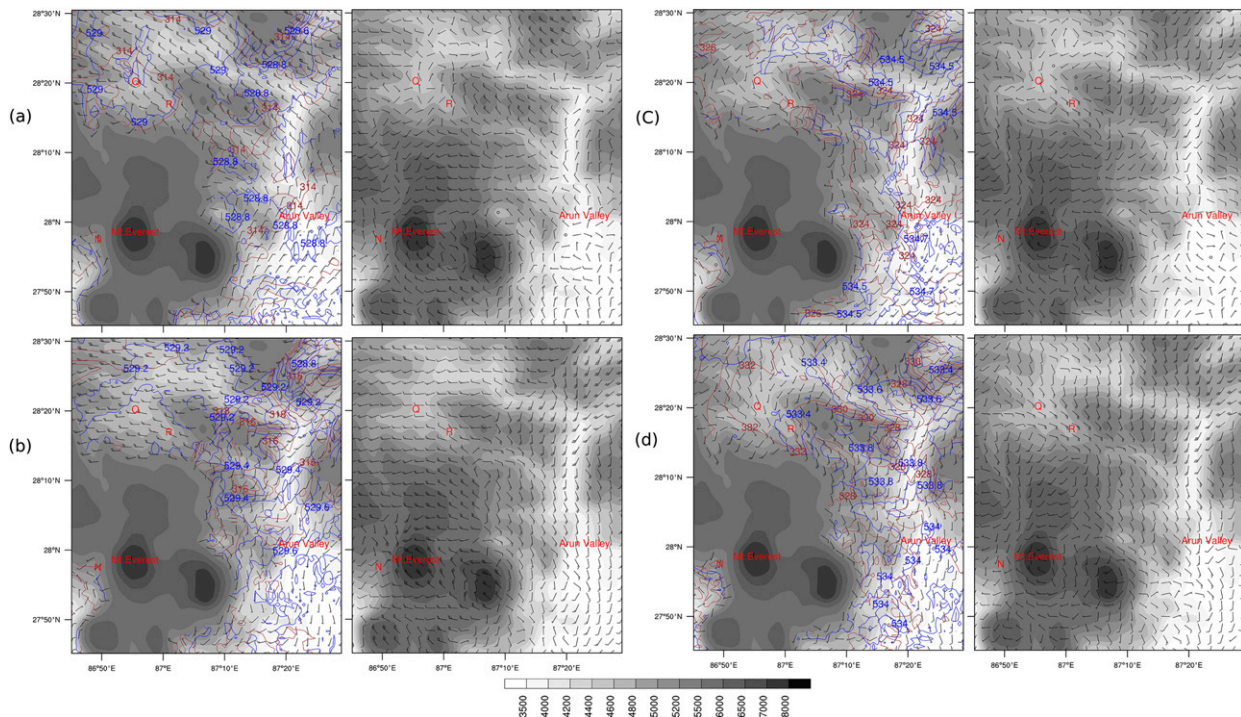


FIG. 8. Simulated horizontal wind, air pressure (blue contours at intervals of 0.1 hPa), and potential temperature (brown contours at intervals of 1 K) at 5200 m (left plots) and surface wind (right plots) at (top) 0300 and (bottom) 1500 LST on (a),(b) 3 Apr and (c),(d) 28 May (full barb = 5 m s^{-1}).

The most notable phenomenon in the 5200-m wind field is a strong daytime southerly wind over the Arun Valley on both 3 April and 28 May, which is considerably colder than the air at higher altitudes outside the valley, as illustrated by the contours of potential temperature (Figs. 8b and 8d). The resulting southerly inflow to the Arun Valley during the daytime is similar to the up-valley flow in the Kali Gandaki valley reported by Egger et al. (2000) and Zängl et al. (2001). On 3 April, there were very strong northwesterly winds to the north of the Mount Everest, including the QOMS area (Fig. 8b). These results suggest that the strong westerly winds in the nonmonsoon season are capable of preventing up-valley winds in the Arun Valley from extending farther into its western branch toward QOMS. The southerly winds appeared only to the east of Mount Everest, probably because the mountains shielded lee areas from the westerly flow. Correspondingly on 28 May, there was no strong westerly wind at 5200 m MSL, in contrast to conditions on 3 April, when the southerly flow in the Arun Valley extended to a wider area, including QOMS. It is noteworthy that a horizontal pressure gradient from the Arun Valley to Repu Valley developed over the ridge at 1500 LST 28 May, which favored the westward turn of the flow near the ridge (Fig. 8d).

At the near surface, winds showed a close resemblance to those at the 5200-m level. At 1500 LST 3 April, strong westerly winds prevailed over most of the region at the northern foot of Mount Everest, including QOMS; in the Arun Valley region, a strong up-valley wind had developed with a southerly direction and velocity of $8\text{--}13 \text{ m s}^{-1}$ (Fig. 8b). In the Arun Valley, near the Repu Valley, southeasterly winds parallel to the valley direction occurred in the branch valleys and met the westerly winds from QOMS on the ridge. Meanwhile at 1500 LST 28 May (Fig. 8d), a southerly wind covered most of the high-altitude area over 4000 m and the southerly wind in the Arun Valley extended to cover a larger area. The surface wind was southeasterly at QOMS, consistent with the wind at the 5200-m level. The origin of this southeasterly wind was likely to be an extension of the wind in the Arun Valley, which reached QOMS along the Repu Valley after crossing the ridge. Thus, there is a possible relationship between the monsoon season afternoon southeasterly wind at QOMS and the wind circulation in the Arun Valley.

To confirm this relationship, the maps of tracer counts on 3 April and 28 May are shown at 1300, 1500, and 1700 LST in Fig. 9 and give a visual indication of how the local air at QOMS is influenced by the wind from Arun

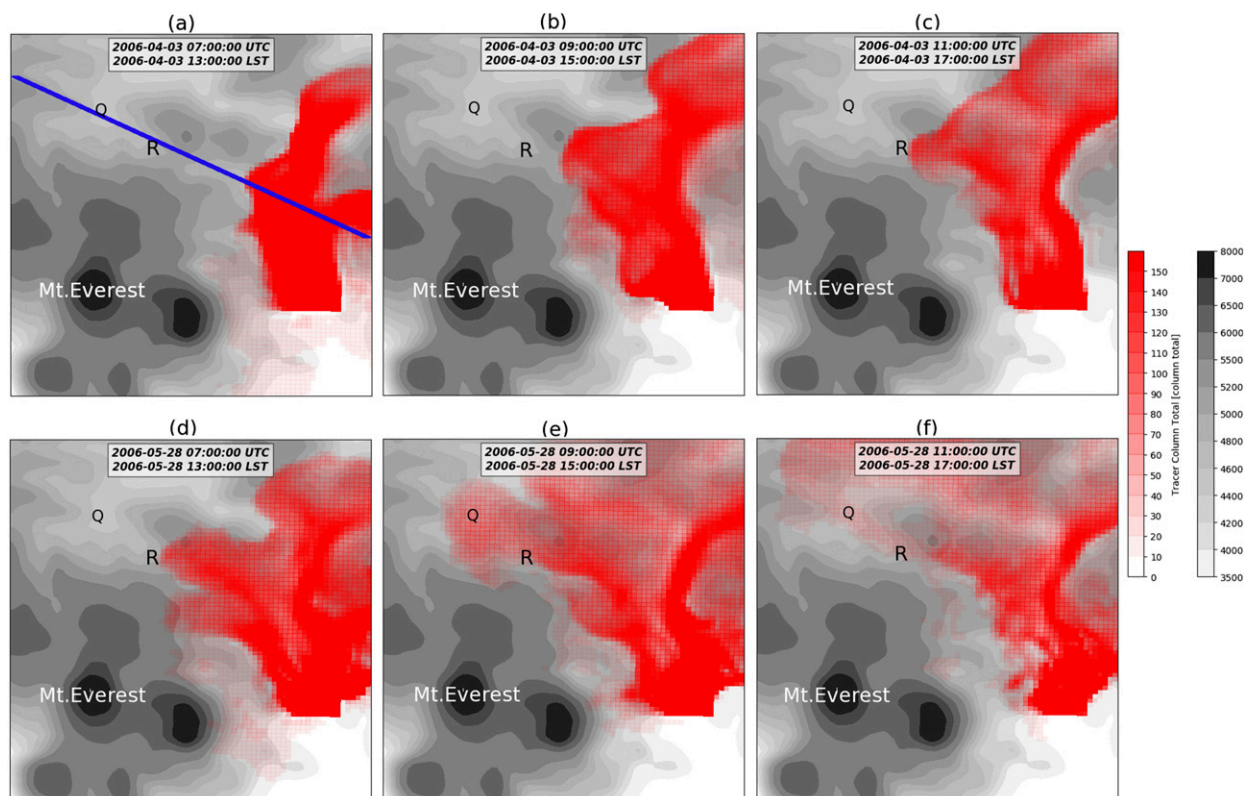


FIG. 9. The tracer counts at (a) 1300, (b) 1500, and (c) 1700 LST 3 Apr and (d) 1300, (e) 1500, and (f) 1700 LST 28 May. The straight blue line in (a) shows the position of the cross section Everest (EV). Letters Q and R represent the positions of QOMS and Repu Valley, respectively.

Valley. Tracers were transported toward the Repu Valley during the afternoon in both simulations. However, on 3 April tracers stopped at the ridge between Arun Valley and QOMS valleys (Figs. 9a–c), while they reached QOMS on the other side of the ridge on 28 May (Figs. 9d–f).

Simulated horizontal wind and potential temperature cross sections along the line passing approximately through QOMS, Repu Valley, and Arun Valley on 3 April and 28 May are shown at 1300, 1500, and 1700 LST in Fig. 10. On both days, the southerly flow in Arun Valley reached a depth of approximately 1500–2000 m. A strong horizontal gradient of potential temperature developed from the Arun Valley to Repu Valley at the ridge in S1 (Figs. 10a–c), and even reached QOMS in S2 (Figs. 10d–f). The extent of the gradient of potential temperature in the direction from Arun Valley to QOMS coincided with that of the southeasterly winds at low levels and the distributions of the red tracers. Simulated tracer counts, potential temperature fields, and wind direction clearly indicate a difference in air mass between the Arun Valley and QOMS. Similarly, the

temperature difference between air in the Arun Valley and Repu Valley (the location of QOMS) may also contribute to the formation of the afternoon southeasterly wind at QOMS in S2.

5. Discussion and conclusions

The overall features of the WRF numerical results and the wind observations at QOMS are consistent with the development of an up-valley wind in the Arun Valley, which is similar to an equivalent feature found in the Kali Gandaki valley (Zängl et al. 2001). According to the analysis of seasonal variations in the strong afternoon winds at QOMS, simulations S1 and S2 were able to reproduce the observed local surface wind patterns in the nonmonsoon and monsoon seasons, respectively.

We consider two kinds of afternoon strong winds as analyzed above: 1) southwesterly winds prevailing during the nonmonsoon season and 2) southeasterly winds during the monsoon season. These two persistent strong winds cannot be satisfactorily explained by mesoscale

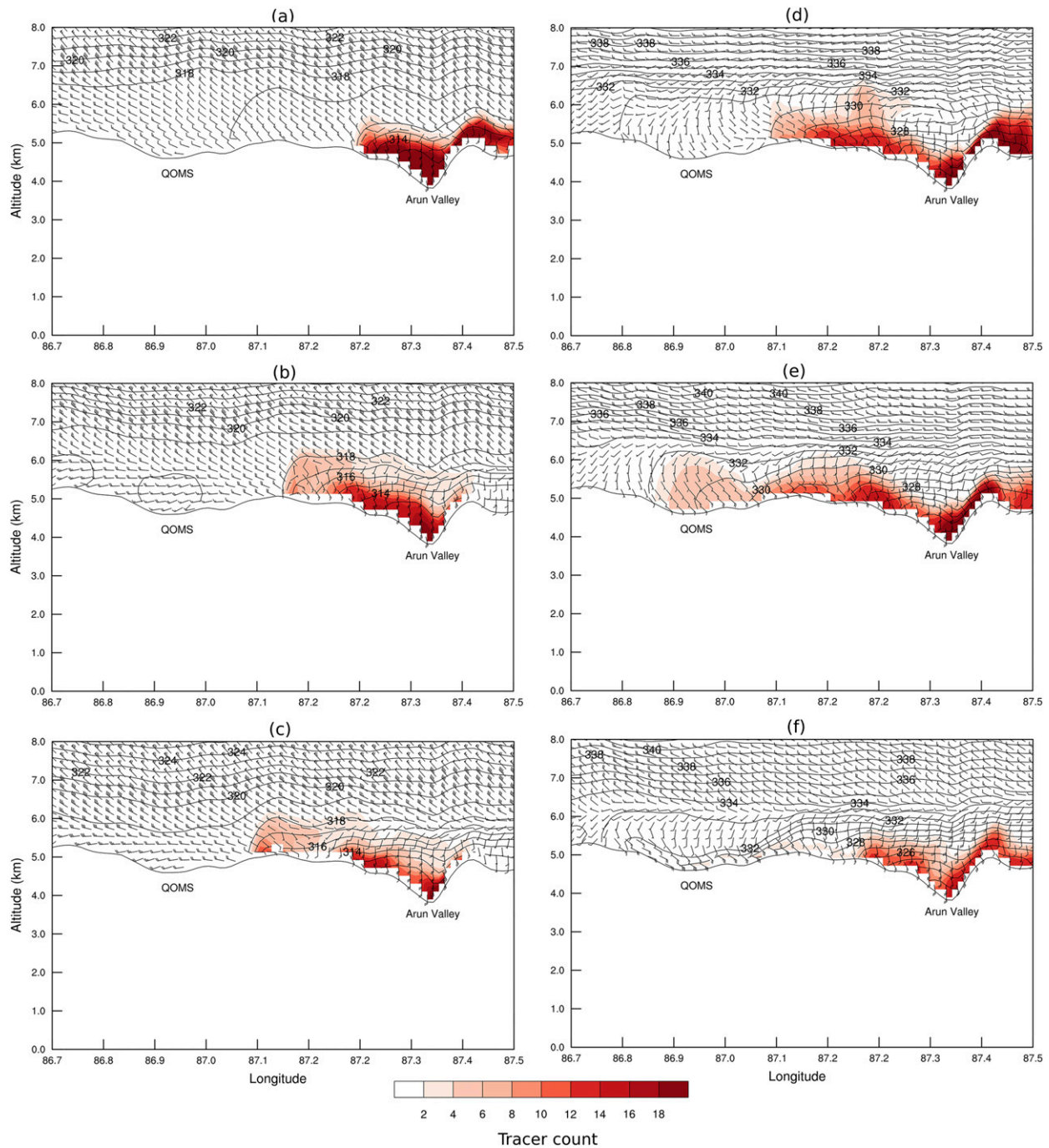


FIG. 10. Horizontal wind (barbs, with full barb = 5 m s^{-1}), potential temperature (black contours at intervals of 1 K), and tracer concentration at (a) 1300, (b) 1500, and (c) 1700 LST 3 Apr and (d) 1300, (e) 1500, and (f) 1700 LST 28 May, along the cross section EV (Fig. 9a).

weather systems, such as cold frontal activity. Meanwhile, the strong southwesterly wind in the nonmonsoon season seems to be driven primarily by downward momentum transport from strong winds aloft.

The annual variability of local winds coincides with the variations in the westerly winds associated with the

meridional shift of the STJ (Fig. 4). Development of the convective boundary layer also provides evidence that the lower-level southwesterly flow is driven by the strong westerly wind at high levels. This result agrees with observations of the STJ's influence on weather at a glacier on Mount Shishapangma at 6900 m MSL (Li et al. 2011).

Meanwhile, the southeasterly wind at QOMS during the monsoon season originates from the Arun Valley, according to the wind field and passive tracer transport (Figs. 8d and 9d–f), which shows that horizontal pressure gradients and thermal gradients between the air over the Arun Valley and QOMS may play a key role in governing the flow. There is good consistency in the southeasterly wind and the pressure gradients from the Arun valley to QOMS, suggesting that the pressure gradient drives the southeasterly wind. The thermal difference between the air within and outside of the Arun Valley should also intensify the pressure gradient. Moreover, once the colder air crosses the ridge, a drainage flow is expected to occur as a result of the thermal difference, which can accelerate the wind speed. However, the extent to which cold air drainage may affect the wind speed in the Repu Valley should be evaluated through a higher-resolution simulation.

The simulation results show that the southerly inflows in the Arun Valley are colder than the air outside the valley, both in a nonmonsoon and monsoon day (Fig. 10). Meanwhile, there is no apparent pressure gradient from the Arun Valley to QOMS over the ridge in the nonmonsoon season, probably because the strong westerly winds of the STJ dominate in this area and balance the inflow into the Arun Valley. As a consequence, the STJ and the inflow to the Arun Valley are the two key factors controlling the afternoon strong winds at QOMS.

This study has analyzed the mechanism through which inflow into the Arun Valley affects the local circulation on the northern slopes of Mount Everest: The results show that when the STJ is located over the southern TP, strong westerly winds prevail over the Himalayan region. As a result, the southerly winds in the Arun Valley are confined within the valley walls. After the monsoon onset, the STJ moves northward to the northern part of the TP and then the westerly winds over the Himalayas weaken significantly (Figs. 4b and 5). This synoptic-scale change over the northern slopes of Mount Everest triggers two local events: 1) The up-valley wind in the Arun Valley develops further in terms of its range and height—for example, the southerly winds occur over the northern Arun Valley up to 6000-m height MSL (Fig. 10f), which is greater than the height of hills or ridges bordering the valley; and 2) the strong afternoon southeasterly wind developing at QOMS is caused by the pressure gradients from the cold inflow in the Arun Valley to QOMS. Therefore, the occurrence of the afternoon southeasterly wind can be an indicator of the STJ's transition away from this region.

The deep and strong southerly inflow in the Arun Valley found in our simulations resembles the strong

up-valley winds reported in the Nepal Himalayan valley of Kali Gandaki (Egger et al. 2000; Zängl et al. 2001). Zängl et al. (2001) found that the heat low over the TP enhances the strength of the up-valley wind in the southern Himalayas and speculated that the inflow might migrate into the TP under favorable conditions. In this regard, Ueno et al. (2008) considered the STJ as an important factor. In our simulations, domain 3 (3-km resolution) covers the Kali Gandaki region, and the strong up-valley winds in this region were also reproduced by the model (Fig. 11). In both the nonmonsoon and monsoon seasons, significant daytime southerly inflows through Kali Gandaki valley were found. The southerly wind over the valley covered a larger area and developed a greater thickness during the monsoon season than during the nonmonsoon season (Figs. 11a and 11c). This result is similar to that in Arun Valley. However, the daytime surface winds in the Kali Gandaki valley during the nonmonsoon season are not significantly different than those during the monsoon season. This is because the valley is deep and long and the high west sidewall of the valley blocks the strong winds from the west (Fig. 11b).

Our study provides insights into the local circulation in a northern Himalayan valley. The afternoon surface winds at QOMS exhibited significant seasonality, owing to the variable influence of the STJ as its position shifts north and south. We considered whether the shift between these two seasonal wind patterns indicates the onset of the South Asian monsoon at this site. According to the jet stream theory, the onset of the monsoon is driven by the sudden shift of the STJ, after a significant period of inhibition caused by the high-elevation Himalayas (Allaby 2010). Although the seasonality of the afternoon strong winds at QOMS coincided with the changes in westerly winds at high levels, a further study using several years of data should be performed in the future.

The results of S1 and S2 should not be generalized to the respective nonmonsoon and monsoon seasons as a whole; nevertheless, they can serve as a case study to highlight the impact of the characteristic STJ location on the near-surface flow field near QOMS. Moreover, our model simulation at 1-km grid spacing was capable of reproducing the salient features of the strong afternoon winds despite the considerable challenges associated with the modeling of finescale complex terrain. However, a higher resolution (e.g., under 100 m) could be used to more accurately capture complex terrain and slopes in this region, yet the standard terrain-following coordinates used by WRF are unable to handle very steep terrain because of the grid distortion and related numerical errors (Thompson et al. 1985; Satomura 1989). In that case, an alternative gridding technique,

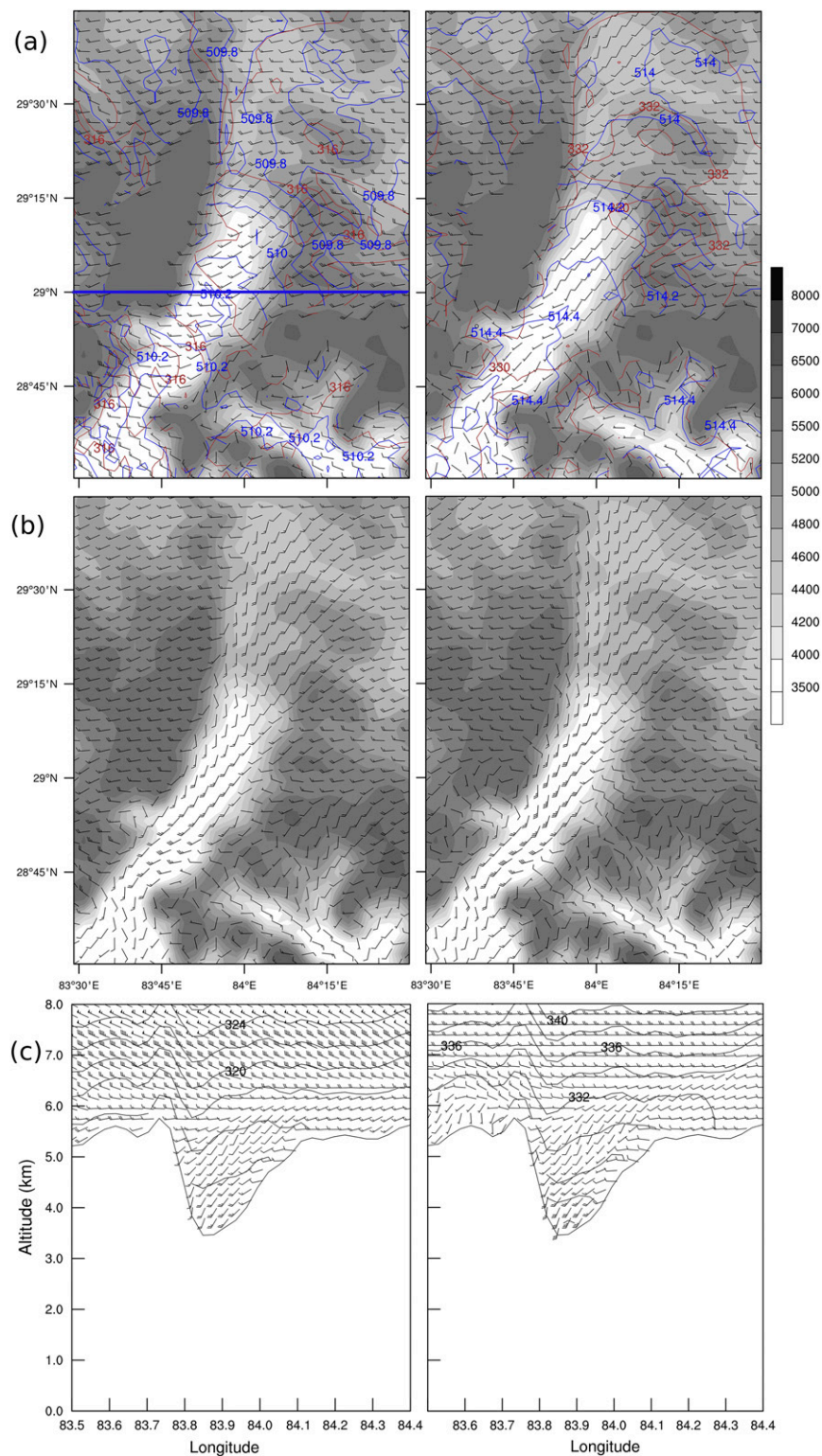


FIG. 11. (a) Simulated horizontal wind fields, air pressure (blue contours at intervals of 0.1 hPa), and potential temperature (brown contours at intervals of 1 K) at 5500 m over the Kali Gandaki region at 1500 LST (left) 3 Apr and (right) 28 May. (b) As in (a), but for surface wind. (c) Horizontal wind and potential temperature (black contours at intervals of 2 K) at 1500 LST (left) 3 Apr and (left) 28 May, along the cross section marked by the blue line in the left panel of (a) (full barb = 5 m s^{-1}).

such as the immersed boundary method that has been implemented into the WRF Model and eliminates conforming grids and errors associated with terrain-following coordinates (Lundquist et al. 2010, 2012), should be applied in the future.

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