A Diabatically Driven Mesoscale Vortex in the Lee of the Tibetan Plateau

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ABSTRACT

An analysis of a diabatically driven and long-lived midtropospheric vortex in the lee of the Tibetan Plateau during 24–27 June 1987 is presented. The large-scale conditions were characterized by the westward expansion of the 500-mb western Pacific subtropical high and the amplification of a trough in the lee of the plateau. Embedded within the lee trough, three mesoscale convective systems (MCSs) developed. A vortex emerged following the dissipation of one MCS, with its strongest circulation located in the 400–500-mb layer. Low-level warm advection, and surface sensible and latent heating contributed to the convective initiation. Weak wind and weak ambient vorticity conditions inside the lee trough provided a favorable environment for these MCSs and the vortex to develop and evolve. The organized vortex circulation featured a coherent core of cyclonic vorticity extending from near the surface to 300 mb, with virtually no vertical tilt. The air in the vicinity of the vortex was very moist, and the temperature profile was nearly moist adiabatic, with moderate convective available potential energy. The wind near the vortex center was weak, with little vertical shear. These characteristics are similar to those of mesoscale convectively generated vortices found in the United States. The vortex circulation persisted in the same area for 3 days. The steadiness of large-scale circulation in the region, that is, the presence of the stationary lee trough and a geopotential ridge that developed to the east of the trough, likely contributed to the persistence of the vortex over the same area.

Potential vorticity (PV) diagnosis suggests that the significant increase in the relative vorticity associated with the vortex development was largely a result of diabatic heating associated with the MCS. An elevated PV anomaly was found near 400 mb in situ after the dissipation of the MCS. The PV anomaly was distinctly separated from those associated with baroclinic disturbances located to the north of the Tibetan Plateau, and the region of the PV anomaly was nearly saturated (with relative humidity exceeding 80%). Further support for this hypothesis was provided by the estimated heating profile and the rate of PV generation due to diabatic heating. The heating peaked at 300 mb, while the diabatic generation of PV reached its maximum at 500 mb. The preexisting ambient vorticity contributed about 20% to the total PV generation near the mature stage of the MCS.

The vortex was also associated with heavy precipitation over the western Sichuan Basin of China. The persistent, heavy rainfall took place in the southeasterly flow associated with the vortex circulation, about 300 km north of the vortex center.

1. Introduction

The low-level shear lines and vortices that form in the vicinity of the Tibetan Plateau are among the most important rain-producing systems in China during the warm season (Tao et al. 1980; Tao and Ding 1981). One such circulation system that forms at the eastern and southeastern flanks of the Tibetan Plateau is the southwest (SW) vortex; this term is used by Chinese meteorologists because the vortices are located in southwestern China (Tao and Ding 1981). Southwest vortices are cyclonic circulations that are frequently observed at low levels (primarily at 700 mb, and 850 mb over lower elevations). The horizontal scale of the SW vortices is typically several hundred kilometers, and the vertical scale is relatively shallow in their for-
mative stage. The cyclonic circulation is often observed to be the strongest at low levels. For cases of more intense development, the circulation may extend to 500 mb and above. The vortices normally form in regions of relatively warm and horizontally uniform temperature, away from major baroclinic zones. Some remain in such an environment throughout their life cycle. Others are affected in the later stages of their development by cold fronts arriving from the north along the eastern edge of the Tibetan Plateau.

The interest in these vortices arises from the observations that 1) they frequently occur in certain geographical areas, 2) they are often associated with severe weather over the Sichuan Basin (located immediately to the east of the Tibetan Plateau; see Fig. 1), and 3) they have a great influence on the precipitation downstream when they move away from their place of origin (Tao et al. 1980; Tao and Ding 1981; Huang 1986). The flash flood case of 12–14 July 1981 (Kuo et al. 1986) is one of many examples of such events. In that case, an SW vortex centered over the western Sichuan Basin intensified at low levels, as a plateau disturbance migrated eastward and became superimposed upon it. With daily rainfall exceeding 200 mm for more than two consecutive days, it resulted in a severe flood in the basin. In a study of a heavy-rain event along the Yangtze River valley during the period of 23–25 June 1983, Ma and Bosart (1987) reported that a 24-h rainfall of 200 mm over the basin was associated with an SW vortex. Huang (1986) composed six heavy-rainfall events over the Sichuan Basin from May to September of 1983, including the case presented by Ma and Bosart, and attributed the cause of the heavy rainfall to the development of the SW vortices.

Despite the recognition of the importance of SW vortices and their association with heavy rainfall, many issues remain unresolved. For example, it is not clear from available climatological surveys (e.g., Sichuan Weather Bureau and Meteorological Program, Department of Physics, Yun-Nan University 1975; Ma and Wang 1992, unpublished manuscript) whether these vortices are forced by the same mechanisms. The climatological surveys and case studies suggest that there may exist a spectrum of SW vortices ranging from those where terrain effects are essential to those where large-scale forcing and/or diabatic effects are dominant. This possibility is supported by the fact that these vortices form and develop under a variety of large-scale environmental conditions, and dry vortices are observed. More case studies are needed to document the kinematic and thermodynamic structures, and the physical processes, contributing to the formation and development of these vortices under different environmental conditions. Although the importance of orographic forcing has been recognized, the exact physical mechanisms by which the terrain interacts with the airflow in the genesis of these vortices remain elusive. On the large scale, the interaction of the synoptic flow with the Tibetan Plateau, and the relationship between the vortex and large-scale flow pattern also warrant more research. Additionally, the relationship between the vortex and diabatic heating requires quantitative assessment to better understand the impact of the diabatic processes, since many of the developing vortices

![Diagram](image.png)

**Fig. 1.** Geographic locations and smoothed terrain of East Asia. Dots indicate places referred to in the text. Terrain contours are every 1000 m, with 0-m contour omitted.
are accompanied by clouds and heavy precipitation at various stages of the vortex life cycle.

In this paper, we examine an SW vortex that formed over the Yun-Gui Plateau (YGP, the southeast extension of the Tibetan Plateau; see Fig. 1) through diabatic heating associated with a mesoscale convective system. This case is interesting because after its development the vortex circulation was linked to an extremely heavy rainfall event over the western Sichuan Basin (Huang and Xiao 1989). Figure 2 shows the 3-day accumulated precipitation over the western Sichuan Basin beginning at 0000 UTC 24 June 1987. The maximum rainfall, 492 mm in 3 days, was recorded at Anxian (see Fig. 1 for location), located in the northwestern basin. A major portion of the rainfall (100-mm isohyet) occurred in a 100-km × 500-km area along the sloping terrain at the eastern edge of the Tibetan Plateau. The objectives of this study are 1) to document the evolution and the structure of this vortex, with a focus on the initial stage of the development, and 2) to show that this vortex was strongly diabatically driven, with its characteristics similar to the mesoscale convectively generated vortices (MCVs, Bartels and Maddox 1991) associated with mesoscale convective complexes (MCCs, Maddox 1980) documented over the United States (e.g., Johnston 1981; Menard and Fritsch 1989; Bartels and Maddox 1991). The term SW vortex will be used here, since its physical appearance at low levels does not differ from other SW vortices reported in the literature.

Section 2 will briefly describe the data and analysis method used in this study. A synoptic overview is provided in section 3. In section 4, a comprehensive examination of the synoptic- and regional-scale environment in which the SW vortex evolved is presented. The kinematic and thermodynamic structure of the vortex is described in section 5. Section 6 provides a discussion on the vortex development, the role of terrain, and the relationship between the vortex and the heavy rainfall to the north of the vortex. A summary of our findings is given in the final section.

2. Data and method of analysis

Analysis of the June 1987 SW-vortex case was performed using conventional data. Radiosonde, pibal, and surface reports were obtained from the archives of the National Center for Atmospheric Research (NCAR). The radiosonde network over most of China is fairly dense, with a station separation of about 250 km. However, the soundings available from the NCAR archives contained mainly the mandatory-level data for most stations, with very limited significant-level reports. To improve the quality of the analysis, soundings data with higher vertical resolution from 11 stations in the area of interest were obtained from the Beijing Meteorological Center (BMC), China. The locations of the stations are shown in Fig. 3. The additional data were mainly temperature and dewpoint temperature observations, while the supplementary wind data were more limited. The additional winds were at 300, 600, and 900 m above ground, and at 2, 4, 5, and 6 km above mean sea level. These winds were then converted to the pressure levels at which the other observations were available, and objective analyses were performed.

Gridded data for diagnostic calculations were obtained, using the existing analysis package of the Pennsylvania State University–NCAR mesoscale modeling system (Manning and Haagenson 1992) by enhancing the 2.5° × 2.5° global analyses from the European Centre for Medium-Range Weather Forecasts with ra-
diosonde data on a 90-km grid mesh using a Cressman-type (Cressman 1959) objective analysis scheme. The radius of influence is 520 km for the first scan, decreasing to 178 km on the last (fourth) scan. There is an additional fifth scan for the relative humidity (RH) field. After the objective analysis, the temperature and RH fields are smoothed but the wind field is not.

Satellite data available for this case study include 3-h infrared (IR) images from 0000 UTC 22 June to 0000 UTC 27 June (missing 0300 UTC 26) and visible images taken three times a day at 0100, 0400, and 0700 UTC on 23, 24, and 25 June. Six-hourly enhanced IR imagery from 0000 UTC 22 June through 12 UTC 26 June (missing 0000 UTC 26) is also available. These satellite data were obtained from secondary data utilization stations in China and Taiwan, and have a typical resolution of approximately 20 km. The exceptions are the IR images shown in Fig. 9, which were obtained from the archives of the Japan Meteorological Agency, and have a 5-km resolution.

Precipitation analyses were performed based on the BMC's 6-h rainfall data, which were supplemented by a denser network of raingages over the heavy rainfall area in the western Sichuan Basin (50 stations with a spacing of approximately 25 km).

3. Synoptic overview

The synoptic-scale conditions in the upper troposphere during the study period are typical of those over East Asia in late June. A large anticyclone, known as the Tibetan high, was established over East Asia at 200 mb and above. Major baroclinic systems in the westernlies had shifted to the north of the Tibetan Plateau, as is typical for this time of the year. Figure 4 shows geopotential height and isolach analyses at 200 mb for 0000 UTC 23 June and 0000 UTC 25 June 1987. At 0000 UTC 23 June, the Tibetan high occupied much of the Tibetan Plateau and southern China, with its center located over the eastern plateau. An intensifying upper-level jet (Fig. 4) was located on the northern edge of the Tibetan Plateau, and remained more than 1000 km away from the YGP and the Sichuan Basin.
over the entire study period. The Tibetan high was quasi-stationary, but its intensity increased after 0000 UTC 24 June (cf. Figs. 4a, b). The SW vortex and the heavy rainfall occurred under this quasi-stationary upper-level anticyclone. A deep tropospheric disturbance in the westerlies moved across the region to the north of the Tibetan Plateau during the study period. Although it did not move sufficiently far south to directly influence the development of the SW vortex, a lower-level front accompanying the disturbance was a very important factor in producing the later stage of the heavy rainfall over the Sichuan Basin.

Geopotential height and temperature analyses at 500 mb for 0000 UTC 20 June and 0000 UTC 23 June (Fig. 5) show that much of the eastern Tibetan Plateau and southern China were situated under both a geopotential and a ridge, with a very weak temperature gradient south of 35°N. The lack of baroclinicity at 500 mb over the area of interest persisted throughout the study period, and is consistent with the lack of transient upper-level disturbances in the region. However, the circulation over the southern part of the analysis domain underwent significant changes between the two time periods shown in Fig. 5. A large anticyclone, centered over the western Pacific (known as the western Pacific subtropical high, or WPSH), was retreating southward over southeastern China prior to 0000 UTC 23 June. Meanwhile, the WPSH began to build westward to western Indochina and the Bay of Bengal as a low pressure system (known as the monsoon depression), earlier located to the north of the Bay of Bengal (Fig. 5a), was moving westward and dissipating. The movement of the WPSH is clearly reflected by the changes of the 588-dam contour. The change in the large-scale circulation was evident at all levels from the surface up to 300 mb.

4. Development and evolution of the SW vortex

a. Amplification of a lee trough

As the WPSh was building over western Indochina, the wind over the western YGP switched from southwesterly to northwesterly, and a trough began to appear downstream of the southeastern Tibetan Plateau and the YGP. By 0000 UTC 23 June, the trough and its associated northwesterly flow had become well established at the southeastern corner of the Tibetan Plateau and the western YGP (Fig. 6). The ridge over the southeastern Tibetan plateau and upstream of the YGP is also clearly illustrated at both 500 and 700 mb. The formation of the ridge–trough couplet upstream and downstream of the YGP was possibly a consequence of the interaction between the southerly to southwestern flow along the western branch of the WPSh, the topography of the Tibetan Plateau, and the YGP. The hypothesis that the trough was topographically induced is supported by the time series of sea level pressures (Fig. 7) at station 42410 (upstream of the YGP at 26.1°N, 91.5°E; marked by “X” in Fig. 6b) and station
59211 (downstream of the YGP; marked by “Y” in Fig. 6b). Starting from 0000 UTC 20 June, the sea level pressure at 42410 (open circles) steadily increased, while that at 59211 (crosses) gradually decreased. As a result, the pressure difference between the upstream and downstream stations varied from −7.5 mb at 1200 UTC 20 June to +9.5 mb at 0600 UTC 23 June (dots). Above the surface, the geopotential height over the area immediately downstream of the YGP typically fell about 30 m at 850 mb, and 20–25 m at 700 mb between 0000 UTC 22 June and 0000 UTC 23 June, as indicated by the formation of a trough (24°N, 106°E, Fig. 6) immediately downstream of the strong northwesterly flow over the western YGP. The observation that the strongest height fall occurred at lower levels is consistent with the general description of a lee trough (Palmén and Newton 1969, 345–347). In a comparison with the typical circulation patterns during SW-vortex events (e.g., Lu 1986), which exhibit southwesterly flow at 700 mb over the YGP, the ridge–trough pattern across the YGP in this case was atypical. At 500 mb, a weak trough was also present over the YGP, with another short-wave trough appearing farther to the east (27°N, 107°E, Fig. 6).

Associated with the established ridge upstream and trough downstream of the YGP, changes in the kinematic and thermodynamic conditions over the southern, southeastern Tibetan Plateau and the YGP were observed. From 22 to 27 June, convection occurred every day after 0600 UTC (1400 LST) and lasted until early the next morning (2100 UTC, or 0500 LST) along the geopotential ridge, which spanned the southern and southeastern Tibetan Plateau and the YGP (Fig. 6). In the vicinity of Xichang (station 56571, 27.8°N, 102.2°E; see Fig. 3 for location) and the northeastern YGP, which were located within the broad trough, the wind became very light below 500 mb (generally less than 6 m s⁻¹; see Fig. 6) and the relative humidity of the air increased.

The low pressure system that was located at 700 mb over the Sichuan Basin at 0000 UTC 23 June (Fig. 6b) was a separate, shallow vortex, which gradually moved eastward and did not develop into a deep circulation system. This system was not the focus of this study.

b. Convection over the Yun-Gui Plateau

Figure 8 shows the sounding of 0000 UTC 23 June (0800 LST) at Xichang, about 5–6 h prior to the initial convective outbreak over the northeastern YGP during the day. At this time, the sounding had relatively little convective available potential energy (CAPE) compared to a midlatitude severe convective storm environment, and more closely resembled that of the tropical atmosphere; that is, the temperature profile was close to moist adiabatic and the air was fairly moist throughout the troposphere. Warm advection was present near Xichang and was indicated by the veering of the wind below 400 mb at the station (Fig. 8). By 0600 UTC (1400 LST) 23 June, the surface temperature had risen from 22.2°C to 29.9°C, and the dewpoint temperature had reached 18.5°C from 16.3°C 6 h earlier. This low-level warming and moistening destabilized the sounding considerably and increased the CAPE to an estimated value of 1300 J kg⁻¹, as indicated by the stippled area in Fig. 8. Beginning at about 0600 UTC (1400 LST) 23 June, convection started over the southeastern Tibetan Plateau, the YGP, and the area immediately to its east. Apparently, the warm advection, and the surface sensible and latent heating, contributed to the initiation of the convective activity over the YGP.

Figure 9 depicts the GMS-3 IR imagery that shows the convective activity between 0600 UTC 23 June and 0000 UTC 24 June at 3-h intervals. The cloud shields expanded considerably between 0600 and 1200 UTC (cf. Figs. 9a, c). The near-circular and continuous cold cloud shields suggest the organization of mesoscale circulations similar to those documented by Maddox (1980). Three distinct mesoscale convective systems (MCSs) emerged (A, B, and C in Fig. 9). The sizes of these cloud shields—with temperatures less than or equal to −33°C, as estimated from the enhanced IR imagery at 1200 UTC 23 June (not shown)—were approximately 88 000 km² for A, 145 000 km² for B, and 180 000 km² for C, with eccentricities of 0.74, 0.63, and 0.67, respectively. (Note that systems A and B
may not have reached their maximum extents at 1200 UTC 23 June, but an enhanced IR imagery was not available to assess their sizes at a later time.) In terms of their appearances from satellite imagery, these systems are very similar to the MCCs as described by Maddox (1980) and Velasco and Fritsch (1987).

The wind and geopotential height fields at 500 and 700 mb for 1200 UTC 23 June (Fig. 10) continued to exhibit the ridge–trough pattern across the YGP with strong northwesterly flow over the western YGP, and weak wind over the eastern YGP and inside the trough. The weak trough was still present over the YGP, and there was no indication of a closed circulation at this time. The 500-mb short-wave trough to the east of the YGP (27°N, 107°E) was intensified, perhaps as a result of a possible interaction between the wave and the ongoing convective activity of MCS C (Fig. 9c). The shallow cyclonic vortex that was observed over the Sichuan Basin at 0000 UTC 23 June (Fig. 6b) had moved eastward by this time.

By 0000 UTC (0800 LST) 24 June, the nearly circular cloud shields associated with MCSs A and B had largely dissipated. Meanwhile, a small convective system appeared over the western Sichuan Basin at 1800 UTC 23 June (30°N, 104°E; Fig. 9e). It gradually grew with time and produced locally heavy rainfall in the vicinity of Chengdu (station 56294, 30.4°N, 104°E; see Fig. 3 for location) in the 12-h period ending 0600 UTC 24 June. This system (denoted by the letter D in Figs. 9f, g), which was located to the north of the dissipating MCS A, was the first of the three heavy precipitation episodes that occurred over the western Sichuan Basin during the study period.

Figure 11 shows the 24-h accumulated precipitation ending at 0000 UTC 24 June. Due to the lack of high-resolution precipitation data over the area of the MCSs, and the unavailability of radar data, the rainfall associated with these MCSs may be underestimated. However, the analysis does indicate that the MCSs produced widespread precipitation with moderate intensity over the area of 24°–29°N, 101°–112°E, and the three centers of maximum precipitation, located at (27.8°N, 102.2°E), (25.2°N, 106°E), and (27°N, 108.5°E), corresponded well with MCSs A, B, and C, respectively.
Vertical profiles of divergence and kinematically computed vertical \( \rho \) velocity (\( \omega \)) averaged over an area of 130,000 km\(^2\) (box B in Fig. 3), which partially encompasses MCSs A and B, before (0000 UTC 23 June) and during (1200 UTC 23 June) the MCS episode, are shown in Fig. 12. Both the \( \omega \) and divergence profiles were adjusted using the O'Brien (1970) technique, in which the integrated divergence was removed in a column between the surface and 100 mb. Prior to the convective activity, weak convergence (<0.5 \( \times 10^{-5} \) s\(^{-1}\)) was present in the column, resulting in weak upward motion (<1.5 \( \mu \)b s\(^{-1}\)). At the time of the nearly mature MCSs, the area was characterized by increased convergence in the lower and middle troposphere and by strong upward motion, which peaked at 300 mb (~4 \( \mu \)b s\(^{-1}\)). Strong divergence also existed above 200 mb, with a maximum value of approximately 3.7 \( \times 10^{-5} \) s\(^{-1}\), indicating a region of convectively generated outflow at upper levels. The general characteristics of these profiles compare well with those of composite MCCs over the United States at the comparable stages (Maddox 1983; Cotton et al. 1989). At 0000 UTC 24 June, descending motion was observed to the west and southwest of the vortex center, and ascending motion was still present to the east and northeast of the vortex center, which was apparently consistent with the observed lingering convection in the area (see Fig. 9g).

c. The vortex

By 0000 UTC 24 June, a mesoscale cyclonic circulation (or the SW vortex), which was embedded within the relatively large-scale lee trough, had intensified over the northeastern YGP (cf. Figs. 6 and 13). The cyclonic vorticity extended from near the surface up to 400 mb (the average surface pressure was about 770 mb). The center of the cyclonic circulation was vertically aligned between 700 and 400 mb, and located slightly to the south of Xichang (27°N, 103°E). The position of the vortex center coincided with the area that was previously occupied by MCS A (Fig. 9). By 0300 UTC 24 June, the convection over the upper YGP had largely ceased (and so had the convection over the southeastern Tibetan Plateau).

Figure 14 shows the analyses of the wind and geopotential height at 700 and 500 mb for 0000 UTC 25 June. The vortex is clearly evident over the northeastern YGP at both 700 and 500 mb. There was very little vertical tilt of the circulations between the two levels.
FIG. 10. The same as Fig. 6 except for 1200 UTC 23 June 1987.

FIG. 11. Twenty-four-hour accumulated precipitation (mm), beginning at 0000 UTC 23 June 1987. Isohyets are 1, 10, 25, and 50 mm. Dashed lines indicate terrain contours of 1500 and 3000 m.

FIG. 12. Divergence ($10^{-5}$ s$^{-1}$) and kinematically computed vertical motion ($10^{-3}$ mb s$^{-1}$) averaged over the area of box B (see Fig. 3 for location) at 0000 and 1200 UTC 23 June 1987.

FIG. 13. The same as Fig. 6 except for 0000 UTC 24 June 1987. Line $A'A'$ in (a) indicates position of cross section in Figs. 20, 22, and 26.
Fig. 14. The same as Fig. 6 except for 0000 UTC 25 June 1987.

To the east of the SW vortex, a strong ridge developed, which was the result of the reintensifying WPsh circulation over southern China. Convective activity near the vortex center was mostly absent at 0000 UTC 25 June. However, convection was very intense over the western Sichuan Basin at this time (Fig. 15), about 300 km north of the vortex center. This was the second heavy precipitation episode over the western basin during the study period.

Fig. 15. GMS-3 enhanced infrared satellite imagery for 0000 UTC 25 June 1987. The black indicates the temperature range of −58.2° to −62.8°C, and the inner white denotes temperatures less than −62.8°C.

Fig. 16. The same as Fig. 15 except for 1200 UTC 25 June 1987.

Fig. 17. The same as Fig. 6 except for 0000 UTC 26 June 1987.
The vortex maintained its identity (especially at 500 mb) during the day, with a moderate increase in intensity at 1200 UTC 25 June (not shown). After a pausing period of 12–15 h at night and in the morning hours, convection redeveloped in the vicinity of the vortex sometime between 0600 and 0900 UTC (1400 and 1700 LST) 25 June, as shown in the enhanced IR image at 1200 UTC 25 June (Fig. 16). By 0000 UTC 26 June, the cyclonic circulation of the SW vortex had noticeably strengthened at both 500 and 700 mb, accompanying the increase of the southerly and south-easterly flow (Fig. 17). The 500-mb vorticity associated with the vortex reached its maximum intensity at this time. Rain showers were reported at Xichang throughout the day of 26 June, while heavy precipitation had been occurring over the western Sichuan Basin in the past 18 h, and continued during most of the day (not shown), marking the third heavy precipitation episode over the basin. A low-level front, as indicated by a wind shift at 700 mb, was located at the northern edge of the basin (32.5°N, Fig. 17b) and began to move southward after 1200 UTC 26 June.

At 0000 UTC 27 June, three days after its initiation, the cyclonic circulation remained intact over the YGP (Fig. 18). At 500 mb, a closed vortex was replaced by a trough, as the westerly flow became considerably stronger. The low-level front that was located at the northern edge of the basin 24 h earlier had moved to the Sichuan Basin, and the wind across the shear line had increased greatly by this time. As the low-level front moved into the basin, the area of precipitation began to shift to the east along with the warm and moist air mass (not shown), and the heavy rainfall over the western basin finally terminated.

5. Structure of the southwest vortex

In the last two sections, we described the synoptic and regional-scale environment in which the SW vortex developed and evolved. The analysis suggests that the development of the SW vortex was followed by the establishment of a very weak flow regime over the YGP and the decay of an MCS. In this section, we will describe the kinematic and thermodynamic structures of the vortex to gain further insight into its formation and development processes.

a. The kinematic structure

Figure 19 presents a time–height section of relative vorticity averaged over an area of 130 000 km² (25 grid points), centered on the 600-mb vorticity maximum over YGP for the period of 22–27 June, with the shading indicating relative humidity greater than 80%. The approximate location of the area is indicated in Fig. 3 (box A; the position of the vorticity center did not change much during the period). A very pronounced feature in this figure is the significant increase in cyclonic vorticity in the 300–600-mb layer between 0000 UTC 23 June (prior to the convection over the YGP) and 0000 UTC 24 June (the time when the SW vortex was first observed). For example, the average vorticity at 500 mb increased from less than 0.5 × 10⁻³ s⁻¹ before 0000 UTC 23 June to about 2.5 × 10⁻³ s⁻¹ at 0000 UTC 24 June. Simultaneously, anticyclonic vorticity also increased in the upper troposphere, suggesting the presence of the convectively generated outflow. The low-level vorticity was present well before the development of the midlevel vortex, and also intensified after 0000 UTC 24 June. The existence of the low-level vorticity reflects the interaction of the low-level flow with the terrain. The intensity of the 500-mb vortex reached its maximum at 0000 UTC 26 June (also see Fig. 17a).

The vorticity structure of the SW vortex at 0000 UTC 24 June is illustrated in the cross section of Fig. 20 (position of the cross section is indicated in Fig. 13a). A coherent core of cyclonic vorticity was located over the YGP with very little vertical tilt. The horizontal scale of this vortex, as measured by the 1 × 10⁻³ s⁻¹ relative vorticity isopleth, was about 400
downward motion to the west of the vortex center and over the western YGP in the northwesterly flow, which was associated with the lee trough.

The wind near the center of the vortex was fairly weak throughout the troposphere. Wind speeds generally were less than 8 m s\(^{-1}\) between the surface and 300 mb, and no more than 15 m s\(^{-1}\) at upper levels. The wind shear was also very weak, with a typical value of 1 m s\(^{-1}\) km\(^{-1}\) in the 300–700-mb layer. Bartels and Maddox (1991) showed that the formation and longevity of the mesoscale convectively generated vortices are often associated with environments of light wind and weak vertical shear. The wind averaged over the area of the vortex (not shown) indicates weak westerly flow below 300 mb and moderate easterly flow (10–15 m s\(^{-1}\)) above.

b. The thermodynamic and moisture structure

The thermodynamic structure of the vortex is illustrated in Fig. 21, which shows the southwest–northeast cross section of \(\theta_v\) (solid) and RH (dashed) at the same location as in Fig. 20. The relative vorticity core is outlined in the figure by the 5 \times 10^{-5} \text{ s}^{-1} isopleth. Relatively low-\(\theta_v\) air was present in the midtroposphere over a layer several hundred millibars deep in the general environment of the vortex (the region outside the YGP). This indicates the conditional instability of the large-scale monsoon air in which the vortex developed. To the northeast of the vortex core, \(\theta_v\) was relatively high and the air was very moist. This column of high-\(\theta_v\) air extended to the southeast of the vortex center below 500 mb (not shown), and was a reflection of the lingering convective activity associated with the decaying MCSs. In comparison, the vorticity core co-
incided with the area occupied by relatively lower-\(\theta_e\) air, and \(\theta_e\) decreased toward the southwest of the vortex center. The contrast was mainly due to the moisture difference across the vortex, between the more moist and higher-\(\theta_e\) air that was affected by the earlier convection and the drier and lower-\(\theta_e\) environmental air to the southwest of the vortex center. The displacement of the vorticity core and higher-\(\theta_e\) air suggests that the development of the vorticity core occurred to the west of the most convective region. In contrast, the high-\(\theta_e\) air was located within the vortex center in the July 1981 case studied by Kuo et al. (1986). This apparent structural difference suggests that different organizational processes may be responsible for the formation of these two vortices. To the southwest of the vortex center (to the left of the vortex core in Fig. 21), the air was relatively dry and warm near the surface. This region was in the northwesterly descending flow. Although the cyclonic vorticity was positive within the descending flow, it was relatively weak (<4 \times 10^{-5} \, \text{s}^{-1}). The maximum vorticity associated with the vortex (\(\sim 8 \times 10^{-5} \, \text{s}^{-1}\)) was located at 500 mb, an elevation much higher than this shallow layer of dry descending flow. Additionally, the region of high vorticity had a relative humidity of about 80%. The nearly saturated condition in the elevated vortex center contrasted sharply with the shallow descending northwesterly flow, excluding the possibility that the vortex was generated and maintained by vortex stretching in a descending flow.

Strong warming and moistening at Xichang and Weining (station 56691; see Fig. 3 for location) were observed after the MCS activity. Figure 22 shows the virtual potential temperature difference between the soundings at 1200 UTC and 0000 UTC 23 June at Xichang (dashed), with a maximum warming of 6.7°C at 400 mb, and another maximum of equal magnitude at 200 mb. The warming was also detected 12 h later at Xichang in the layer between 470 and 220 mb, with cooling occurring above 220 mb and in the lower troposphere (Fig. 22, solid). Since the atmosphere was

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**Fig. 22.** Virtual potential temperature difference \(\Delta \theta_v\) at Xichang (station 56571) between soundings (a) at 1200 and 0000 UTC 23 June (dashed; missing observation above 200 mb), and (b) at 0000 UTC 24 June and 0000 UTC 23 June (solid) 1987.

**Fig. 23.** Soundings at Xichang (station 56571) for (a) 0000 UTC 24 June and (b) 1200 UTC 25 June 1987. Winds are in meters per seconds (full barb: 5 m s\(^{-1}\)). The shading indicates positive CAPE for a near-surface air parcel.
already fairly moist prior to the convection at Xichang, the actual warming, as measured by the change in potential temperature, was found to be very close to that presented in Fig. 22 in the 220-470-mb layer. Hence, the vortex may have a warm core at middle to upper levels. A similar structure has been documented for MCSs over the United States (Maddox 1983; Cotton et al. 1989; Bartels and Maddox 1991).

Figure 23 shows two soundings at Xichang for 0000 UTC 24 June and 1200 UTC 25 June. These soundings depict very moist conditions throughout the troposphere with a nearly moist-adiabatic temperature profile above the lowest 150 mb. The CAPE was small immediately following the convective outbreak over the area at 0000 UTC 24 June (about 190 J kg\(^{-1}\) for a near-surface air parcel; Fig. 23a). It increased after 1200 UTC 24 June. Figure 23b shows an example of such a sounding at 1200 UTC 25 June (CAPE of 1000 J kg\(^{-1}\)). The CAPE, computed for a near-surface air parcel, averaged to 560 J kg\(^{-1}\) for the morning soundings (0000 UTC) and 910 J kg\(^{-1}\) for the late afternoon soundings (1200 UTC) at Xichang during the 3-day period, with very little inhibition (negative area) beneath the level of free convection. These estimates are consistent with the observed convective activity in the vicinity of the vortex, which exhibited a strong diurnal variation. Weak low-level warm advection and sensible heating at the surface could have contributed to the generation of this moderate instability. Away from the surface, the moist-adiabatic core was maintained throughout the period. These conditions are similar to those found for MCVs over the United States (Bartels and Maddox 1991).

6. Discussion

In the previous sections, we have presented an analysis of a mesoscale vortex that occurred over the YGP, and persisted for 3 days in the same area where it first developed. The following are the immediate questions that arise from the analysis. 1) How did the vortex develop? Was it diabatically generated? 2) Did the terrain play a role in the genesis process? 3) Why was the vortex long-lived? 4) What was the relation between the vortex and the heavy precipitation that occurred over the western Sichuan Basin, since the vortex center was not located over the basin? In this section, we address these questions.

a. Generation of vorticity at midlevels

As the analysis from the previous sections shows, although there was some preexisting low-level vorticity, the most pronounced feature in the 500- and 700-mb wind fields (Figs. 6, 10, 13, 14, 17, and 18) is the development and persistence of the midlevel mesoscale vortex, which appeared immediately after the dissipation of MCS A at 0000 UTC 24 June. The development was manifested by a significant increase in cyclonic vorticity in the 400-600-mb layer. Thus, a very important question is, What is the source of the midlevel vorticity? Was the vorticity adveected into the region, or was it generated locally? Was the vortex development related to an upper-level disturbance? To answer these questions, we present, in Fig. 24, the potential vorticity (PV) analysis, where PV is defined in isentropic coordinates as \((\zeta + f)(\partial p/\partial \theta)^{-1}\) (where \(\zeta\) is relative vorticity in isentropic coordinates, \(f\) is the Coriolis parameter, \(p\) is pressure, and \(\theta\) is potential temperature) computed from the gridded data on the 340-K isentropic surface for 0000 UTC 23 June and 0000 UTC 24 June. The winds at 400 mb over the area of interest are also shown in Fig. 24, since the 340-K surface roughly corresponded to 400 mb over the YGP at the time. The distribution and evolution of PV fields may help identify the source of a disturbance.
At 0000 UTC 23 June (Fig. 24a), there were two zones of relatively high PV in the vicinity of the Tibetan Plateau. The major zone of PV was located to the north of the plateau, with values greater than 2 PVU (potential vorticity unit defined as $10^{-4}$ m$^2$ s$^{-1}$ K kg$^{-1}$). This zone of PV extended from the northwestern tail of the Tibetan Plateau to northeastern China, and was clearly associated with the baroclinic disturbances in the westerlies. The other zone of PV was of lower value (~0.6–0.8 PVU; hatching) and extended from the Bay of Bengal and over the southeastern Tibetan Plateau to the YGP. From the YGP, it linked with an east–west zone of relatively higher PV (~0.8–1 PVU), which was collocated with a quasi-stationary frontal zone (also known as the Mei-Yu front). This southern zone of relatively high PV coincided with a belt of convection (see, for example, Fig. 9a) and was located within the broad anticyclonic circulation regime. It was not connected with the higher PV farther to the north, as a belt of low PV (with values less than 0.4 PVU; light shading) separated this zone from the higher values to the north. This observation serves as further evidence to suggest that the formation of the SW vortex was not linked to the upper-level disturbances to the north of the Tibetan Plateau. Also, winds were generally weak within the anticyclonic circulation regime (wind speed less than 10 m s$^{-1}$ at 400 mb). Therefore, it is unlikely that any significant advection could occur in the region.

By 0000 UTC 24 June, the PV over the YGP had greatly amplified, increasing from about 0.8 to 1.29 PVU (Fig. 24b). Since the background PV value was approximately 0.5 PVU, the maximum anomaly value actually increased from approximately 0.3 to 0.8 PVU, or by nearly a factor of 3. Additionally, the area enclosed by the 0.8-PVU isopleth was much larger than that covered 24 h earlier. The region of high PV coincided with the area of the midlevel vortex (Fig. 24b), and with the region previously occupied by MCS A (Fig. 9). The vertical structure of the PV at 0000 UTC 24 June is shown in Fig. 25. In contrast to the vorticity, which extended from the surface to upper levels, the PV anomaly existed only aloft. The difference resulted partially from the lower stability near the surface within the northwesterly flow. This observation again points out that the descending flow was a dynamically separate feature and was not related to the midlevel vortex.

Figure 26 shows the time–height section of PV from 0000 UTC 22 June to 0000 UTC 27 June, averaged over the same area as that of Fig. 19. The region with RH greater than 80% is shaded, and the dashed line near the top of the figure denotes 60% RH value. This figure indicates a significant increase in PV between 0000 UTC 23 June and 0000 UTC 24 June in the 300–600-mb layer, with a distinct PV maximum located at 400 mb. In addition, the region of high PV was also very moist beginning at 0000 UTC 24 June, with the RH exceeding 80%. The region with RH below 60% was located above 200 mb, where PV was markedly decreased after 1200 UTC 23 June. This analysis indicates that the local increase in PV could not be due to an intrusion of stratospheric air via subsidence. On the other hand, the PV structure shown in Fig. 26 is qualitatively similar to that of a long-lived MCS, as discussed by Raymond and Jiang (1990) and Hertenstein and Schubert (1991). Raymond and Jiang (1990) discussed how such a PV distribution could result from an MCS.

The observations that the vortex appeared after the dissipation of the MCS and that the region of high PV was nearly saturated suggest that the PV anomaly over the YGP at 0000 UTC 24 (Fig. 24b) was generated locally by diabatic heating associated with the MCS. Following Hoskins et al. (1985), the vertical component of the source term due to diabatic heating for
small Rossby number–large Richardson number flow can be written in pressure coordinates as

\[ \frac{D(PV)}{Dt} \approx -g(f + \gamma) \frac{\partial \theta}{\partial p}, \]

(1)

where \( \dot{\theta} \) represents the rate of diabatic heating and \( D/Dt \) is the Lagrangian derivative. This implies that PV can be generated below the level of the heating maximum where there is positive absolute vorticity. Because relative vorticity is positive in the area of interest, the absolute vorticity is therefore positive below 300 mb over the YGP. To estimate the diabatic generation of PV, a heating profile is required.

To estimate a diabatic heating profile, the approach of calculating a large-scale apparent heat source (Yanai et al. 1973) is employed. As in the work of Kuo and Anthes (1984), the heat budget equation in the terrain-following \( \sigma \) coordinates is used for the computation. The estimated heating profile over an area of 130 000 km\(^2\) (box A in Fig. 3) for 1200 UTC 23 June is shown in Fig. 27, which indicates that the heating maximum occurred at 300 mb with a value of 25 K day\(^{-1}\). Note that the area of the box partially encompasses MCSs A and B, and both systems were close to their mature stage at this time. While the magnitude of the estimated heating in this case is comparable to that calculated for the July 1981 case (Kuo et al. 1986), a distinct difference exists between the two heating profiles. The maximum heating in the July 1981 case (Kuo et al. 1986) was found at 550 mb at the time when the strongest upward motion was observed. More significantly, the heating profile in the current case exhibits a strong vertical gradient in the 400–700-mb layer, which implies that there would be PV generation in that layer if the absolute vorticity were positive. The estimated PV generation at 400, 500, and 600 mb is 0.16, 0.26, and 0.20 PVU (12 h\(^{-1}\)), respectively. Therefore, the region of significant PV generation is in the 500–600-mb layer during the mature stage of the MCSs. With an air parcel ascending at a rate of approximately 110 mb in 12 h, the PV acquired by the parcel is comparable to the PV anomaly seen in Fig. 24. However, the 12-h sounding data available for the study allowed the estimate to be calculated only once during the life cycle of the MCSs, and the time corresponds closely to the mature stage of the MCSs. Judging from the scale of the PV anomaly, the satellite images of clouds, and the steady-rain reports in the vicinity, the decay stage of MCS A was likely characterized by a mesoscale precipitating regime.

The appearance of MCVs after the dissipation of their parent MCSs has been documented in the United States (e.g., Johnston 1981; Menard and Fritsch 1989; Bartels and Maddox 1991). Many of these MCVs have a horizontal scale ranging from 200 to 400 km, as estimated from satellite imagery after the cloud anvils of MCSs have dissipated or been advected away (Bartels and Maddox 1991). However, the scale over which the environmental winds may be disturbed can be larger: for example, approximately 500 km in the case of Menard and Fritsch (1989). The seemingly larger vortex scale in this case was due in part to the nonzero ambient vorticity in the environment of the vortex (i.e., the vortex developed within the lee trough). Nevertheless, the characteristics of these MCVs and their environmental conditions (Bartels and Maddox 1991) show remarkable similarities to those of the case presented here. These conditions include very light tropospheric wind, weak vertical wind shear, a moderate amount of CAPE, a very moist atmosphere, and a nearly moist-adiabatic temperature profile through a deep layer. Table 1 compares some of the physical characteristics of the MCVs observed in the United States with those of the June 1987 case. Even though it is widely accepted that these MCVs are diabatically driven, only rarely is an analysis performed to quantitatively assess the role of latent heating. The diagnosis of PV performed in this study enabled us to show that the midlevel mesoscale vortex was likely a consequence of the bulk effect of diabatic heating associated with the MCS activity. In particular, the elevated heating maximum and large vertical heating gradient in the 400–700-mb layer were crucial for the generation of the midlevel vortex. However, it should be emphasized that the environment in which the vortex developed and evolved may critically depend on the large-scale conditions. The establishment of the weak wind regime over the northeastern YGP within the subsynoptic-scale lee trough provided favorable conditions for the mesoscale vortex to develop and persist. The presence
of the weak trough and associated weak vorticity over the YGP prior to the MCS development (see section 4a) may have played a role in organizing the convection and, subsequently, the formation of the mesoscale vortex associated with the MCS. One way of looking at this interaction is to examine the contribution of the preexisting ambient vorticity over the YGP to the total PV generation. The contribution due to the preexisting vorticity alone was estimated to be approximately 20% of the total generation at 1200 UTC 23 June, averaged over the area of the vortex. Although it is not a trivial contribution, it does support our findings that the diabatic heating was the primary forcing for the development of the midlevel vortex. However, the detailed dynamics cannot be fully explored using the observational data alone, and high-resolution numerical simulations are needed to clarify these issues.

While the interaction of a midlevel short wave and an MCS in the generation of an MCV warrants more investigation, some recent research emphasizes the effects of diabatic processes in organizing the mesoscale circulation within an MCS. Raymond (1992) shows that integrating a semibalanced, nonlinear system of equations with a prescribed heating distribution (maximum 1.8 K h\(^{-1}\) at 5 km) can generate a pair of closed vortices in a sheared, nonrotating environment. The maximum perturbation wind reaches about 6 m s\(^{-1}\), and the maximum absolute PV anomaly is about 0.6 PVU at the level of the maximum heating by the eleventh hour of the model integration. Skamarock (1992, personal communication) has also simulated a closed mesoscale cyclonic vortex at 3 km in a weakly sheared environment using a modified version of the Klemp and Wilhelmson cloud model (Klemp and Klump 1993) without a preexisting large-scale disturbance. These results seem to suggest that diabatic effects alone are sufficient to generate circulation of appreciable strength in a weak flow environment.

### b. The role of terrain

To quantitatively assess the role of terrain in the evolution of the vortex is a formidable job, given the limited observational data available for this study. However, a qualitative assessment is possible. The analysis suggests that the role of terrain in the vortex evolution was twofold. First, the amplification of the lee trough was closely related to the terrain, and there was weak ambient vorticity associated with the trough in the vicinity where the vortex developed later. The ambient vorticity contributed to the PV generation as discussed previously. Because the lee trough was established prior to the vortex development, the northwesterly wind to the west of the vortex center was therefore part of the ambient flow. This explains why the winds near the vortex center exhibited such an asymmetry, with the wind to the west being much stronger than that elsewhere.

Second, the establishment of the lee trough left the area near Xichang with very weak wind by 0000 UTC 23 June. The weak wind conditions appear to be conducive to the formation and longevity of MCVs (Bartels and Maddox 1991). The reason for this observation may be explained with the assistance of the quasigeostrophic adjustment theory. In a region of light wind, any significant height perturbation could result in an imbalance between the mass and wind fields. The imbalance would then cause the air to converge to the area of the height fall in a geostrophic adjustment process, creating a cyclonic circulation (Mesinger and Pierrehumbert 1986). In the case of an MCS, the height depression may result from a decrease in the thickness field due to lower-tropospheric diabatic cooling, such as that shown in Fig. 22, and in other case studies (e.g., Bartels and Maddox 1991). If the Rossby number is not too large (as a result of the weak wind conditions), the geostrophic adjustment theory may explain the formation of an MCV. (In this case, the Rossby number is about 0.18, for \( U = 5 \text{ m s}^{-1} \), \( f = 6.8 \times 10^{-5} \text{ s}^{-1} \), and the scale of the vortex is approximately 400 km.)

It is interesting to note that a weak midlevel cyclonic circulation was also present in the area of the decaying convective systems C and C' in Fig. 9 (\( \sim 28^\circ\text{N}, 108^\circ\text{E} \) in Figs. 13b and 14a). However, it had dissipated by 1200 UTC 25 June (not shown). There was virtually no convective activity observed near this vorticity center after its formation. In addition, the strengthening of the WPSH over the area (Figs. 14a and 17a) may have contributed to the local destruction of the midlevel cyclonic circulation that resulted from the MCSs. In contrast, the vortex over the YGP was situated within a stationary lee trough, with geopotential ridges building on each side of the trough. Winds were very light inside the trough, which prevented the PV anomaly from being advected away, or absorbed into the large-
scale flow. Convection was also observed daily in the vicinity of the vortex. Menard and Fritsch (1989) argued that the new convective activity is necessary to maintain the moist-adiabatic core of the midlevel vortex. These observations seem to suggest that the location of the SW vortex (being embedded within the stationary trough), and the terrain-modulated diurnal convection, may have both contributed to the longevity and stationary nature of the SW vortex. The midlevel vortex over the YGP may have also helped to intensify the low-level cyclonic circulation by maintaining the mesoscale updraft aloft and, hence, low-level convergence; that is, the low-level vorticity increased after the development of the midlevel vortex (see Fig. 19).

c. The relationship between the vortex and the heavy rainfall over the western Sichuan Basin

We now return to the heavy rainfall that first led us to the investigation of this case. Unlike the July 1981 case studied by Kuo et al. (1986), where the vortex and the heavy precipitation were collocated over the Sichuan Basin during most of the period, the vortex in this case was separated by about 300 km from the localized area of extremely heavy rainfall over the western Sichuan Basin. Therefore, the relationship between the vortex and the heavy rain is not immediately clear. The precipitation analysis indicates that the heavy and persistent rain occurred in a low-level southeasterly flow in the basin. This southeasterly flow developed and became rather steady after the vortex formed over the YGP (see Figs. 13b, 14b, and 17b). The atmosphere over the basin was very moist and conditionally unstable (not shown). The southeasterly flow impinging upon the sloping terrain of the western basin may have helped to lift the moist and conditionally unstable air mass, and hence, caused the heavy precipitation. Figure 28 illustrates this point by showing the streamline analysis of the 700-mb flow at 1200 UTC 25 June, overlaid by the 24-h precipitation ending at 0000 UTC 26 June. It clearly depicts the relative position of the vortex (maximum 700-mb vorticity marked by “Ω”), the southeasterly flow associated with the vortex circulation over the basin, and the heavy rain to the north of the vortex and on the eastern slope of the Tibetan Plateau. Thus, the analysis suggests that the heavy rainfall over the western Sichuan Basin was likely a consequence of the SW vortex development, since the established southerly and southeasterly flow was capable of transporting warm and moist air from the south to the basin. The southerly flow was reinforced later, perhaps by the development of a low-level jet on the western side of the reintensifying WPSH (see Fig. 17b).

7. Summary

In this paper, we have presented an analysis of a mesoscale, long-lived midtropospheric vortex that developed within the subsynoptic-scale lee trough in the lee of the southeastern Tibetan Plateau. We have described the structure and evolution of the vortex, and shown that the significant increase in the midlevel vorticity likely resulted from the diabatic heating effects associated with the MCS activity over the YGP. The vortex first acquired its identity at 0000 UTC 24 June. It reached its maximum intensity by 0000 UTC 26 June, and persisted until at least 0000 UTC 27 June. The weak environmental wind condition, and the relatively dense observational network in China, allowed the vortex to be observed for a long period of time. The unique aspects of this vortex include its unusually long life cycle, its stationary nature, and its association with heavy precipitation over the western Sichuan Basin to the north of the vortex. The general characteristics of this vortex agree with the existing definition of the SW vortex in the literature. Specifically, there was a cyclonic circulation observed at 700 mb, and the vortex was located in the area where SW vortices frequently appear. The vortex developed and evolved in a fairly barotropic environment and a broad anticyclonic circulation regime at 500 mb. The major findings can be summarized as follows:

1) The SW vortex developed following the westward expansion of the WPSH over the southeastern Tibetan Plateau and upstream of the YGP, and following the amplification of a lee trough to the east of the plateau. Embedded within the lee trough, three MCSs developed, and each lasted for 12–15 h. The low-level warm advection, and the surface sensible and latent heating contributed to the convective initiation. The weak wind and weak ambient vorticity conditions provided a fa-
vorables environment for the MCSs and the vortex to develop and evolve.

2) The vortex developed after the dissipation of an MCS over the southeastern corner of the plateau and the YGP. Extensive and organized convection near the vortex center was absent during the first two days after the vortex development. Nevertheless, some scattered convective activity, with a strong diurnal variation, was observed during these two days in the vicinity of the vortex. Convective activity increased during the last day.

3) The kinematic structure of the SW vortex featured a coherent vertical core of cyclonic vorticity extending from the surface up to 300 mb, with light winds and weak vertical wind shear. The thermodynamic conditions near the vortex center were characterized by very moist air and a nearly moist-adiabatic temperature profile in the troposphere. These characteristics compare well with those of MCVs observed over the United States (Menard and Fritsch 1989; Bartels and Maddox 1991).

4) Potential vorticity analysis indicates that the development of the mesoscale vortex was also accompanied by a significant increase in PV at midlevels. The conservative property of the PV in an adiabatic and frictionless flow allowed the source of the elevated PV to be attributed to the diabatic heating associated with the MCS activity. The estimated diabatic heating distribution and PV generation strongly support the hypothesis that the vortex was generated by diabatic heating. The preexisting ambient vorticity contributed about 20% to the total PV generation near the mature stage of the MCS.

5) The steadiness of the large-scale circulation in the region likely contributed to the persistence of the vortex over the same area. In particular, the presence of the stationary trough to the east of the Tibetan Plateau, and the geopotential ridges on both sides of the trough, established an almost stagnant flow regime over the Yun-Gui Plateau. Very weak steering flow prevented the vortex from being advected away.

6) The continued heavy precipitation over the western Sichuan Basin occurred after the steady low-level southerly to southeasterly flow was in place. This flow structure was established after the formation of the vortex over the YGP.

7) The presence of the low-level vorticity, even before the formation of the midlevel vortex, was likely a result of the flow interaction with the terrain.

While PV diagnosis has been applied to the baroclinic cyclones (e.g., Hoskins et al. 1985; Davis and Emanuel 1991), using PV analysis for MCVs has been rare in the literature. In this study, PV diagnosis was used to quantitatively assess the role of diabatic heating in generating an MCV. We have shown, through such an exercise, that the SW vortex observed here was strongly diabatically driven. An SW vortex of this type has not been documented in detail in the literature. However, in view of the general meteorological conditions (conditionally unstable during most of the warm season, and a weak wind regime in the lee of the Tibetan Plateau), and the frequent occurrences of MCVs (Miller and Fritsch 1991) in the region of frequent SW vortex appearance, the mechanism, as it was seen to operate in this case, may also occur in many other instances. In fact, Ma and Bosart (1987) briefly mentioned the observation of a mesoscale, midlevel comma cloud pattern (Fig. 9 of their paper), and suggested that it may have originated from the MCV that occurred 12-h earlier over the Sichuan Basin. However, no detailed analysis on the vortex was performed in their study.

A great challenge that remains for future investigation is the role of terrain in the generation of the SW vortex in this case. Even though we made an attempt to address it, the effect of orography on the development and evolution of the vortex remains elusive. Quantitative assessment is desirable to explain the role of orography in the convective initiation, the relationship between the low-level vorticity and the midlevel development, and the stationary nature of the vortex. A high-resolution dataset, such as from a field program or a successful numerical model simulation, is very important for clarifying many of the foregoing issues.

Finally, it is worth noting that the observation made in this paper and in other SW-vortex studies, that the lee vortex develops in an environment that is removed from the major baroclinic zone to the north of the Tibetan Plateau, strongly suggests that the mechanism by which many of these vortices form is different from that proposed for the cyclogenesis occurring in the lee of the European Alps and to the east of the Rocky Mountains (e.g., Palmén and Newton 1969; Buzzi and Tibaldi 1978; Buzzi et al. 1987). In the latter situations, the argument usually requires the presence of a “parent” cyclone upstream of the mountain ridge, or an incoming baroclinic wave. The disturbance then moves across the mountain ridge to produce further intensification of the lee cyclone. Although a parent cyclone may appear to the north of the Tibetan Plateau in the case of a developing SW vortex (see Fig. 4), it cannot move over the massive plateau to reach the lee side, because of the blocking effect of the plateau (Staff Members, Academia Sinica 1958). This seems to put SW vortices into a separate category of lee cyclogenesis, and for this reason, they deserve special attention and investigation. More research is needed to study various problems associated with SW vortices. Detailed documentation of many cases and a systematic climatological survey over a long period of time are important steps toward a better understanding of the SW vortices and an improvement in their prediction.

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REFERENCES


Sichuan Weather Bureau and Meteorological Program, Department of Physics, Yun-Nan University, 1975: Formation of southwest vortices and their origin. Meteorology, 4. [In Chinese.]


