A Case Study of the Cutoff Process and Latent Heating Effect in an Upper-Level Cold-Core Low during the Mei-Yu Season in East Asia

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ABSTRACT

Using observational data and model sensitivity test, the present study diagnosed the evolution of an upper-level trough over 2–7 May 1998 during the mei-yu season in East Asia, with a focus on the role of latent heating downstream from the trough on the cutoff process and the formation of the cold-core low (CCL). Compared with conceptual models based on dry dynamics, the formation process of this cutoff low (COL) was inconsistent with the model of migratory jet streak–trough interaction, but the evolution of the upper-level trough (including its narrowing and rotation of axis) and the scale of the resultant COL generally agreed with the theory of interaction between trough and mean westerly flow with anticyclonic background shear. East/southeast of the trough, active and persistent convection appeared and induced 1) divergent ageostrophic flow near the tropopause, 2) an increase in mid- to upper-level thickness due to effective heating, and 3) amplification of an upper-level ridge downstream. Through a model sensitivity test, the diabatic heating is suggested to play a key role in the formation of the current CCL as the trough moved and extended southeastward to approach the amplifying ridge. Because of the east/southeast ageostrophic flow, the trough encountered a critical layer and became distorted, and its southern segment eventually detached to form a COL. When latent heating is turned off, the model produced a trough evolution anticipated by the dry theory but without cutoff. Such a role played by the moist process downstream of the trough during cutoff in the mei-yu season has not been previously noted in the literature.

1. Introduction

a. Climatology and impacts of CCL on weather

Upper-level cold-core lows (CCLs) are low pressure systems with closed cyclonic circulation and center temperature lower than their surroundings in the upper troposphere (e.g., Palmén 1949; Palmer 1953; Bluestein 1993, 135–136; Carlson 1998, p. 153). Some previous investigators have classified CCLs into two types based on their location and movement: Palmer and Palmén types (e.g., Chen et al. 1988, 1989; Tsai 1998; Kou 1998). The Palmer-type CCLs (Palmer 1953) are those that develop near an easterly trough or saddle point south of the subtropical high. Forming at lower latitudes, they usually move westward under the prevailing easterly flow in the tropics. On the other hand, the Palmén-type CCLs (Palmén 1949) commonly form through cutoff from a midlatitude trough and thus mostly move eastward and are also cutoff lows (COLs; e.g., Bell and Bosart 1993; Davis and Bosart 2002). Kelley and Mock (1982) analyzed the composite structure of Palmer-type CCLs over the western North Pacific (WNP) in four summers. They found that these CCLs have horizontal scales of about 2000 km and exist through deep layers of 700–100 hPa, with cold-core maximum at 300 hPa, warm-core maximum at 125 hPa, and strongest circulation at 200 hPa near the tropopause. Also associated
with this type of CCLs are significant rising motion and cloudy sky in the southeastern quadrant and mean sinking with little cloud cover in the northwestern quadrant. Besides these characteristics, Chen and Chou (1994) also obtained a mean speed of 4.1 m s⁻¹ toward the west and averaged lifespan of 6.3 days for Palmer-type CCLs, and their scale tends to be larger and lifespan longer if they exhibit upper-level jet streak (typically at northwestern and/or southern quadrants).

The Palmer-type CCLs are also referred to as tropical upper-level trough (TUTT) cells by some researchers (e.g., Colton 1973; Ferreira and Schubert 1999) and can affect both the formation and movement of tropical cyclones (TCs; Montgomery and Farrell 1993; Hodanish and Gray 1993). In addition, the cold anomaly aloft produces a steeper temperature lapse rate and weaker stability and provides ascent as implied by the quasigeostrophic (OG) omega (ω) equation (through both differential vorticity advection term and thermal advection Laplacian term) at synoptic scale (e.g., Holton 2004, section 6.4.1). Hence, local convection can be enhanced if its triggering mechanism at low levels also exists (Erickson 1971). Because of their presence, the CCLs tend to lower the level of 0°C isotherm and increase vertical wind shear, and thus this affects the intensity of local afternoon convection (e.g., Wu 1976; Chen et al. 1990; Chen and Chi 1990) and even provides favorable conditions for severe weather (e.g., hail; e.g., Chen and Tang 2004).

On the other hand, the COL is often regarded as resulted from the Rossby wave breaking (RWB) phenomenon (e.g., McIntyre and Palmer 1983; Postel and Hitchman 2001). After cutoff from midlatitude trough, the Pálmén-type CCLs can provide cyclonic relative vorticity $\zeta$ in the middle to upper levels under suitable synoptic setup and thus are favorable for the formation of TCs (e.g., Bosart and Bartlo 1991; Davis and Bosart 2001, 2002). Similar to Palmer-type ones, the Pálmén-type CCLs can also reduce the low-to-mid-level stability and affect the formation and/or intensification of the low-level jet (LLJ) and the subsequent moisture advection. Eventually, they can lead to local heavy rainfall downstream from the LLJ on the warm side (e.g., Garreaud and Fuenzalida 2007; Singleton and Reason 2007).

Price and Vaughan (1992) studied 200-hPa COLs in the Northern Hemisphere in five years and found that they often appear in summer and over western Europe and the North Pacific. They classified COLs into three types, polar, subtropical, and polar vortex, and considered all of them the result of interaction of jet streaks at scales comparable to COLs with upper-level trough. Bell and Bosart (1989, 1993, 1994) analyzed the climatology of formation regions of 500-hPa COLs in the Northern Hemisphere and found that the synoptic environment prior to cutoff is characterized by clear meridional flow near the trough. Before cutoff, the upper-level jet streak is located upstream from (west of) the trough and transports air with larger $\zeta$ or higher potential vorticity (PV) from the north toward the trough base via advection, thus causing the trough to deepen and eventually the cutoff (Bell and Bosart 1993, 1994). Here, PV in pressure ($p$) coordinate ($q$) is defined as (Hoskins et al. 1985)

$$q = -g(fk + \nabla_p \times \mathbf{V}) \cdot \nabla_p \theta,$$  

(1)

where $g$ is gravitational acceleration, $f$ is the Coriolis parameter, $k$ is the unit vector in the vertical, $\nabla_p$ is the three-dimensional ($x$, $y$, $p$) gradient operator, $\mathbf{V}$ is the wind vector, and $\theta$ is the potential temperature. Bell and Bosart (1994) also found that COLs in southwestern and eastern United States and southern Alps in Europe tend to form to the lee of major topography but consider the upper-level trough–jet streak interaction to be the primary mechanism for the formation of COLs, whereas the process of leeside cyclogenesis associated with vertical stretching plays only a secondary role.

b. Review of existing theories on CCL formation

Keyser and Shapiro (1986, hereafter KS86) proposed a conceptual model leading to cutoff through the interaction between upper-level baroclinic wave and migratory jet streak when the three following conditions exist: 1) A confluent trough with a northwest–southeast-oriented axis (negative tilt), which is favorable for barotropic energy conversion from mean zonal kinetic energy (ZKE) into eddy kinetic energy (EKE); 2) a thermal trough that lags the height trough by $1/4$ wavelength, a condition favorable for baroclinic energy conversion from zonal mean available potential energy (ZAPE) into eddy available potential energy (EAPE) than EKE in an environment with north–south temperature gradient; and 3) an upstream jet streak to advect cold air from the north toward the base of the trough. These three processes are all favorable to the deepening of the trough and further development of disturbance (i.e., COL), which reaches maturity when the jet streak moves to the trough base (KS86).

Similar to the conceptual model of KS86 from the standpoints of energy conversion and thermal advection, Bell and Keyser (1993, hereafter BK93) partitioned PV into curvature PV that is associated with the trough itself and shear PV that accompanies the jet streak. When the upper-level jet streak moves from the west into the trough base, maximum shear vorticity becomes collocated with curvature vorticity. This causes the region of high PV to evolve from the north–south elongated shape into a quasi-circular shape and structure, and the COL forms at the base of the trough (BK93).
The KS86 and BK93 theories for COL due to the upper-level trough–jet streak interaction described above do not link the evolution of trough orientation with the scale of the resultant COL. Thorncroft et al. (1993, hereafter THM93) simulated the evolution of idealized upper-level baroclinic waves under different background westerly shear using dry models. For adiabatic and frictionless flow, PV is conserved following air motion \((dq/dt = 0)\) and its local tendency is determined by advection \((aq/\partial t = -\mathbf{V} \cdot \nabla q)\) on the isentrope (i.e., constant \(\theta\) surface), where a PV ridge corresponds to a trough on \(p\) level (Hoskins et al. 1985; Haynes and McIntyre 1987). For Northern Hemisphere (higher \(q\) to the north), consider a sinusoidal PV contour (on an isentrope) with anticyclonic horizontal shear in the background (i.e., maximum westerly wind to the north). Because of this shear, the initially north–south-oriented PV axis turns clockwise with time to become northeast–southwest oriented. As described in THM93, before the wind field can fully adjust to this new distribution of \(q\), stronger negative PV advection appears to the west of the PV ridge and causes the ridge to narrow. Thus, a relatively small COL results at the base of the PV ridge if cutoff occurs. On the contrary, the PV axis turns counterclockwise to a northwest–southeast alignment under cyclonic background shear (i.e., maximum westerlies to the south) and the PV ridge gradually widens because of a stronger positive PV advection to its east. As a result, a COL larger in size may form (THM93).

In analyzing an event of cyclogenesis near the tropopause, van Delden and Neggers (2003) suggested that the COL can form at the base of the trough via vertical stretching of vortex tubes when high-PV air moves southward and downwarp the isentrope and experiences a decrease in static stability to cause an increase in absolute vorticity \(\nabla q\) as required by PV conservation. Such a process is referred to as “unstable isentropic downgliding” by the authors.

Besides adiabatic processes, the effect of diabatic latent heating upstream (west) of the trough has also been investigated for events in Europe and Africa (e.g., Massacand et al. 2001; Knippertz and Martin 2007). It is found that the convection can amplify the upstream ridge and produce stronger northerly flow and PV advection, cause the trough to narrow and extend farther southward, and eventually aid to the subsequent cutoff (also Moore et al. 2008; Meier and Knippertz 2009).

For Palmer-type CCLs, in a study of TUTT related to the interaction between Hurricane Felix (1995) and the circumpolar vortex over the North Atlantic, Ferreira and Schubert (1999) found that the TUTT cell east of the TC formed likely through Rossby wave energy dispersion, and the overall evolution and its scale were in agreement with THM93. That is, the type of background horizontal shear is the primary factor to determine the evolution in the orientation of trough axis, the thinning/broadening of the trough, and the horizontal scale of the TUTT cell. More recently, Tsai et al. (2010) computed the filamentation time of upper-level trough using isentropic PV (IPV) equation on spherical coordinates, and applied their methods to diagnose the narrowing of the trough and the subsequent COL formation.

Because the strongest circulation of CCLs appears near the tropopause, the exchanges of ozone, pollutants, and air with contrasting moisture content, stability, and PV values between the troposphere and lower stratosphere often occur through three-dimensional advection, differential vorticity advection in the vertical, and transverse circulation associated with the CCL accompanying jet streaks (Erickson 1971; Hoskins et al. 1985; Bluestein 1993). Postel and Hitchman (1999) analyzed the climatology of RWB near the tropopause and found that this phenomenon is most frequent over the ocean in summer. Their case study further shows that the intensification of the upper-level PV ridge occurs when the Rossby wave extends southward and enters the critical layer prior to RWB, as anticipated by the linearized two-dimensional dispersion relationship (Postel and Hitchman 2001). That is, when the phase speed of the wave equals the environmental mean wind speed, the zonal and/or meridional wavenumber will approach infinity and influence the PV-ridge narrowing in either or both directions to possibly result in an RWB (i.e., COL formation).

c. Motivation of present study

From the above review of studies on the formation process of CCL/COL, it is obvious that the potential role played by diabatic processes downstream of the trough has received little attention. Over 3–7 May in the mei-yu season (May–June) of 1998, the northern-central Taiwan experienced afternoon showers in five consecutive days and the 3-h rainfall in Taipei reached 47 mm on 4 May. Synoptic examination indicated that there was no significant weather system at low levels to possibly cause a reduction in stability for such a lengthy period, except for a weakening mei-yu front near northern Taiwan on 3–4 May. In the upper troposphere, however, a midlatitude trough migrated from northwestern China toward the Yangtze River valley since 2 May, it gradually deepened as it moved southeastward, and a CCL finally formed near Taiwan from a cutoff process (to be shown in section 3). After 1200 UTC 5 May, the CCL turned northeastward and eventually merged into another midlatitude trough (not shown). Because of a lack of concurrent development at low levels, the formation of this CCL was clearly different from those appearing at
the mature stage (occlusion) of a classical extratropical cyclone (e.g., Reed 1990; Bluestein 1993, section 1.6; Carlson 1998, chapter 7). From the climatology of Bell and Bosart (1989), it is clear that COLs in southeastern China are rare events. In addition, as will be shown in section 4, the cutoff process in this case were also quite different from those in the past studies reviewed above in certain aspects, particularly in the role played by latent heat release from convection downstream. Thus, the present case deserves attention and can serve as one other example to illustrate the possible role of diabatic processes in leading to the cutoff (or RWB) in the upper troposphere.

The remainder of this paper is organized as follows: The data used and methodology of analysis, diagnosis, and model sensitivity test are described in section 2. The overview of the present CCL case is given in section 3. Further analysis of the cutoff process and comparison with existing theories, the effect of diabatic (condensation) heating from convection including model results, and the mechanism of the cutoff are presented and discussed in section 4. Finally, in section 5, concluding remarks are given.

2. Data and methodology

a. Data

The data used in this study include the following: the European Centre for Medium-Range Weather Forecasts (ECMWF) gridded analyses at 0.5° latitude/longitude resolution on 15 p levels (1000, 925, 850, 700, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, and 10 hPa) every 6 h, Japan Meteorological Agency (JMA) surface and upper-level weather maps, hourly cloud imageries at infrared (IR) and water vapor channels from the Geostationary Meteorological Satellite-5 (GMS-5) satellite, and surface and upper-level data collected during the South China Sea Monsoon Experiment (SCSMEX; Lau et al. 2000) covering our case period (2–7 May 1998). For rainfall distribution, Tropical Rainfall Measuring Mission (TRMM) 3B42 merged satellite rain rates at a resolution of 0.25° × 0.25° every 3 h are also employed (Huffman et al. 2007).

b. Analysis procedure

To analyze and discuss the formation and evolution of the CCL case in the upper troposphere, it is useful to examine the distribution of IPV (e.g., Bluestein 1992; Santurette and Georgiev 2005) because it is conserved for adiabatic and frictionless flow (Hoskins et al. 1985; Haynes and McIntyre 1987; THM93). In the upper troposphere, PV increases with height monotonously unless a tropopause fold exists. Thus, q is first obtained from Eq. (1) using ECMWF data then vertically interpolated onto a specific isentrope. Here, the 335-K isentrope is chosen for analysis and discussion based on Hoskins et al. (1985).

To diagnose the distribution of diabatic heating from condensation associated with convection, the apparent heat source \( Q_1 \) of Yanai et al. (1973) is used as

\[
\frac{Q_1}{c_p} = \frac{\partial T}{\partial t} + (\mathbf{V} \cdot \nabla T - \nabla \cdot \mathbf{V}) + \frac{\mathbf{\omega} \cdot \nabla}{c_p} \frac{\partial s}{\partial p}
\]

(2)

where \( c_p \) is the specific heat (1004 J kg\(^{-1}\) K\(^{-1}\)), \( T \) is temperature, \( \mathbf{V} \) is horizontal wind vector, \( \omega \) is \( p \) vertical velocity, \( s \) is the dry static energy \( (s = c_p T + gz) \), and the overbar denotes the areal mean. In the present study, the value of \( Q_1/c_p \) is computed using the ECMWF data with units of K day\(^{-1}\) but will be referred to as \( Q_1 \) for simplicity.

c. Numerical model and experiments

Except for the diagnosis mentioned above, the present study also employs the fifth-generation Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) Mesoscale Model (MM5, version 3; Grell et al. 1994) to perform numerical simulation and sensitivity test on the effect of latent heat release from convection on the formation of the CCL. The single model domain is set to 175 × 85 grid points with a horizontal grid spacing of 45 km (cf. Fig. 1) and 24 layers (with model top at 10 hPa), sufficient to reproduce the evolution of the synoptic-scale CCL. The physical options include the Dudhia (1989) simple ice for explicit cloud microphysics, the Grell (1993) convective parameterization scheme, a medium-range forecast (Hong and Pan 1996) for planetary boundary layer parameterization, and the Dudhia (1989) cloud-radiation scheme. Using the 6-h ECMWF 0.5° analyses as initial and lateral boundary conditions, the MM5 simulation starting from 0000 UTC 4 May (for 72 h) best captures the life cycle of the CCL and serves as the control experiment (designated as CTRL). Although MM5 offers nested grids at finer resolution and a variety of physical scheme combinations, extensive testing indicates that they do not produce significantly different results. To test the role played by latent heating in our case, a fake dry run (FDRY; readers are referred to http://www.mmm.ucar.edu/mm5/documents/mm5-desc-pdf/sec5.pdf) is carried out for comparison with the CTRL.

3. Case overview

During 3–7 May 1998, an afternoon shower occurred on five consecutive days over northern to central Taiwan, with peak 3-h amount all >20 mm (and ≥40 mm on 3 and 4 May; not shown). On JMA surface weather maps, a front moved across Taiwan slowly during 3–4 May
(Figs. 2a,b), but afterward only weak horizontal pressure gradient existed near Taiwan without significant disturbance over 5–7 May (Figs. 2c–e). Thus, the weak stability for such a long duration was unusual, especially over the three later days when Taiwan and its vicinity were within a weak synoptic environment.

Upper-level features, on the other hand, can often be revealed using satellite imageries at the water vapor channel (≈6.7 μm), which is sensitive to water vapor in the middle to upper troposphere (e.g., Tsai 1998; Santurette and Georgiev 2005). As shown in Fig. 3, the GMS-5 cloud imageries suggest that a dry slot extended southward to near 30°N, 105°E on 4 May and rotated cyclonically around the base of a trough and moved eastward to approach Taiwan on 5–6 May. Thus, the CCL is depicted. Note that there was abundant upper-level moisture between Taiwan and southwestern Japan (i.e., southeast of the CCL) over 4–6 May, so as to indicate deep convection there (Fig. 3). Upstream from the low, some convection was also present in central China, with a similar configuration to the cases reviewed in section 1b (e.g., Massacand et al. 2001; Knippertz and Martin 2007; Moore et al. 2008).

The ECMWF gridded analyses at 500 and 200 hPa at selected times during 2–7 May 1998 are presented in Figs. 4 and 5, respectively. At 1200 UTC 2 May, an upper-level midlatitude trough appeared near 45°N, 100°E with a north-northeast–south-southwest alignment (Figs. 4a, 5a) and deepened and moved into northern-central China on 3 May (Figs. 4b, 5b). With little tilt in the vertical, this upper-tropospheric trough weakened with height because it was associated with colder (warmer) temperatures at 500 (200) hPa. At 0000 UTC 4 May (Figs. 4c, 5c), the trough continued to advance southeastward and its section north of 35°N gradually rotated clockwise into a northeast–southwest orientation since 2 May. On the contrary, the trough axis farther south showed slight counterclockwise rotation at this time, in agreement with Fig. 3a. The 500-hPa CCL (5770 gpm) formed at 32°N, 113°E and the ridge to its immediate southeast grew stronger 12 h later (Fig. 4d), whereas a similar development also took place at 200 hPa near the eastern coast of China by 0000 UTC 5 May (Fig. 5d). After formation, the CCL was initially quite narrow in the east–west direction (≈800 km) and again possessed a cold core at 500 hPa but a warm core aloft at 200 hPa (Figs. 4e,d and 5c,d). Afterward, it gradually weakened and moved slowly southeastward at first to about 29°N, 118°E at 1200 UTC 5 May, when the centers at 500 and 200 hPa were collocated (Figs. 4e, 5e), then toward the northeast (Figs. 4f, 5e). At 1200 UTC 5 May, the ridge immediately downstream from the CCL remained clear. Compared to typical Palme´n-type CCLs such as those shown in Hoskins et al. (1985, their Fig. 5) and Bluestein (1993, his Fig. 1.143), the horizontal scale of about 1000 km of the present case (Figs. 4, 5) is relatively small.

In Fig. 1, the 500- and 200-hPa tracks of this Palme´n-type CCL are also depicted. After its formation around 1200 UTC 4 May, this CCL moved eastward at an average speed of 5.1 m s⁻¹ during its 3-day lifespan, in contrast to the mean speed of 4.1 m s⁻¹ toward the west obtained by Chen and Chou (1994) for Palmer-type CCL. This difference in movement is obviously due to their
different environment and background flow (i.e., westerly for Palmén-type CCLs but easterly for Palmer-type CCLs). Because the midlatitude upper-level westerly flow is typically stronger than the tropical easterly flow, the current CCL remained to travel faster than 4.1 m s$^{-1}$ even when it slowed down after cutoff (cf. Figs. 4–6).

The east–west vertical cross section of height and temperature anomalies ($z'$ and $T'$, where the mean is the 10-day average over 1–10 May 1998) and $\zeta$ through the 200-hPa CCL center indicates that at 0000 UTC 5 May (shortly after cutoff; cf. Fig. 5d) the CCL was evident from 700 hPa up to the lower stratosphere, with a maximum $\zeta$ at 300 hPa (not shown). The cold core ($T' < 0$) was maximized at 400–500 hPa and the warm core ($T' > 0$) at 100 hPa, consistent with Figs. 4 and 5. These structural characteristics were quite comparable to both

![Fig. 2. The JMA surface analysis of pressure (hPa; solid; at intervals of 4 hPa; 1000 and 1020 hPa thickened) at 0000 UTC (a) 3, (b) 4, (c) 5, (d) 6, and (e) 7 May 1998.](image)
Palmen– (e.g., Hoskins et al. 1985, their Figs. 8, 9; Bell and Bosart 1993; Bluestein 1993, his Fig. 1.140; van Delden and Neggers 2003) and Palmer-type CCLs (e.g., Kelley and Mock 1982; Chen and Chou 1994).

4. Cutoff process of the cold-core low

a. Description of cutoff and comparison with dry theories

In this subsection, further analysis of this CCL case is carried out on isobaric and isentropic surfaces, with a focus on the more detailed characteristics and evolution of the trough and CCL during the cutoff process. At 300 hPa, where the $\zeta$ of the CCL reached a maximum, the basic evolution of the trough–CCL system (Fig. 6), which was collocated with 500-hPa cold tongue with no tilt, was very similar to that shown in Figs. 4 and 5. On 3 May, when the (north-northeast–south-southwest oriented) trough at 300 hPa moved southeastward and deepened, it extended southward and narrowed as well (Figs. 6a,b). The trough was also accompanied by strong and strengthening northerly winds exceeding 30 m s$^{-1}$ over a vast area just to the west (upstream) of the trough axis and somewhat weaker southwesterly flow to the east (downstream). On 4 May (Figs. 6c,d), while extending farther into lower latitudes, the southernmost part of the trough gradually detached from the northern part to form the CCL/COL by 1200 UTC with a pair of local upper-level jet streaks. Since 3 May, the northerly jet streak to the west kept advecting cold and dry air toward the trough base (cf. Fig. 3a) and caused the trough to deepen. On 4 May, this jet weakened somewhat while the southwesterly streak intensified (Figs. 6c,d). As discussed earlier, the trough axis north of 35°N exhibited a clockwise rotation and traveled faster and completely detached from the CCL at 300 hPa by 0000 UTC 5 May (Fig. 6e). At this time, the ridge just east of the CCL showed clear intensification (cf. Fig. 5d). Twelve hours later (Fig. 6f), the northerly jet (still peaking over 25 m s$^{-1}$) moved from the west to south of the CCL, which gradually widened in the east–west direction (from about 700 to 1000 km) and assumed a more circular (axisymmetrical) shape. On the other hand, the southerly jet streak weakened significantly to below 20 m s$^{-1}$ on 5 May.

The evolution of the trough leading to the cutoff in the present case is also examined through PV distribution in

Fig. 3. GMS-5 water vapor channel (6.7 $\mu$m) cloud imageries at 0000 UTC (a) 4, (b) 5, and (c) 6 May 1998. Lighter (darker) tones represent more (less) water vapor.
Fig. 4. The 500-hPa ECMWF analyses of geopotential height (gpm; solid; at intervals of 20 gpm) and temperature (°C; dashed; at intervals of 1°C) at (a) 1200 UTC 2 May, (b) 1200 UTC 3 May, (c) 0000 UTC 4 May, (d) 1200 UTC 4 May, (e) 1200 UTC 5 May, and (f) 0000 UTC 7 May 1998. The thick dashed (dotted) lines denote trough (ridge).
FIG. 5. As in Fig. 4, but for 200-hPa geopotential height (gpm; solid; intervals of 50 gpm) and temperature (°C; dashed; intervals of 1°C) at (a) 1200 UTC 2 May, (b) 1200 UTC 3 May, (c) 0000 UTC 4 May, (d) 0000 UTC 5 May, (e) 1200 UTC 5 May, and (f) 1200 UTC 7 May 1998. The thick dashed (dotted) lines denote trough (ridge).
The IPV ridge (corresponding to a trough on p level) of interest was still near 42°N, 92°E over northwestern China (Fig. 7a). The westerly flow in the region was stronger near and north of 42°N, implying an anticyclonic horizontal background shear and that the ridge axis would rotate clockwise with time, as it did on 3 May (Fig. 7b, cf. Figs. 4–6). In addition to the ambient shear associated with a westerly jet to the north, the strong north-northwesterly winds upstream from the IPV axis also blew at large angles across the IPV contour at 0000 UTC 3 May (Fig. 7b), producing negative PV advection evident in Fig. 8a. Downstream from the ridge, the PV gradient (and positive advection) was weaker (Figs. 7b, 8a). With time, the difference in PV advection across the ridge (negative to the west and positive to the east) became larger, leading to a clear narrowing of the IPV ridge (and 250-hPa trough) through 0600 UTC 4 May (Figs. 7c, 8b–d).

In Figs. 6–8 (and also Figs. 4, 5), the evolution of the upper-level trough leading to the cutoff in the present case show clear differences from the upper-level trough–jet
Fig. 7. The IPV (PVU; 1 PVU = \(10^{-6} \text{ m}^2 \text{ K kg}^{-1} \text{ s}^{-1}\); solid lines; at intervals of 0.5 PVU) and horizontal wind vectors (m s\(^{-1}\); reference vector at bottom right) on the 335-K potential temperature \(\theta\) surface at (a) 0600 UTC 2 May and 0000 UTC (b) 3 and (c) 4 May 1998. Contour lines at 1.5 PVU are thickened, and thick dashed lines denote PV ridge of interest.
streak interaction model of KS86 and BK93 mainly in the three following aspects: 1) the orientation of the trough axis prior to the formation of the COL, 2) the phase difference in 500-hPa temperature and 300-hPa height troughs, and 3) the relative position of westerly jet streak and height trough. In KS86, as reviewed in section 1b, a northwest–southeast-aligned trough is favored for barotropic energy conversion from ZKE in the mean flow to EKE. However, the axis of the main trough in our case maintained more or less a northeast–southwest orientation throughout the cutoff process (Figs. 4–6), and corresponded to a condition of $u'v' > 0$, where $u'$ and $v'$ are the deviation from the areal mean (denoted by the overbar) of $u$ and $v$, respectively. Thus, the orientation of the trough axis was unfavorable and not ideal for the growth of the disturbance (i.e., COL) on the trough (e.g., Chen 1989; Hartmann 1994). Likewise, the in-phase structure of 300-hPa height and 500-hPa thermal troughs in our case (Figs. 4, 6) suggests no effective baroclinic energy conversion from ZAPE to EKE as well. Moreover, the position of migratory jet streaks relative to height trough (i.e., the relation between shear and curvature vorticity) is also different from the KS86/ BK93 model, which is quite similar to the trough evolution and cutoff of THM93 under a cyclonic horizontal background shear (their Figs. 12b, 13c). On the contrary, the trough in the present case lay to the south of the westerly jet stream during much of its lifespan (Figs. 6, 7). Apparently, this is the main reason for the differences in the characteristics of dynamic processes leading to the cutoff, and the overall environment in our case was not favorable for the formation of the COL from the perspective of dry dynamics based on KS86 and BK93. In other words, the trough–jet streak interaction process did not play a primary role in the cutoff of this particular case over southeastern China.
However, there are also similarities between the present case and the KS86/BK93 model, including 1) the process and characteristics of axisymmetrization in height and temperature fields when the jet streak moves to the trough base (Figs. 4–6; KS86) and 2) a northerly jet streak behind the trough to advect cold and high-PV air southward (and downward) and deepen the trough prior to the cutoff (Figs. 6–8; BK93). These similar characteristics were also evident in the present case.

Compared to KS86 and BK93, the trough (or IPV ridge) evolution prior to its distortion and the cutoff in the present case agrees more with the model of THM93 under anticyclonic background shear (their Figs. 12a, 13b). That is, the trough will narrow and its axis will rotate clockwise with time when it extends equatorward to the south of the westerly jet maximum. In our case, similar change in the orientation of the axis and a narrowing in the east–west horizontal scale of the trough also took place (Figs. 4–7). As for the cutoff, THM93 and Postel and Hitchman (1999) suggest that it may occur if the base of the trough can extend far south into the lower latitudes with weak background westerlies. On the other hand, the amplifying ridge immediately downstream of the trough on 4–5 May was characterized by anticyclonic circulation (Fig. 7c) and small PV values with negative PV advection (Figs. 8b–d). In the next subsection, the amplification of this ridge and its impact on the cutoff are further investigated.

b. Development of downstream ridge and the role of latent heating

Near the time of the cutoff, the upper-level ridge immediately east/southeast of the narrowing trough can be seen to intensify as early as 1200 UTC 4 May (Figs. 4–6). On the 335-K θ surface, this is reflected as relatively low (near zero) PV values (cf. Fig. 8d). Figure 9 shows height, PV, and ageostrophic winds at 250 hPa as well as GMS-5 IR blackbody brightness (cloud top) temperature \( T_{BB} \) distribution. Over 4–5 May, high clouds with low \( T_{BB} \) values (reaching \(-56^\circ\) to \(-64^\circ\)C) occupied the region between Taiwan and southwestern Japan constantly (Figs. 9a–d). Associated with the active convection were ageostrophic winds (at 250 hPa) blowing from the southeast toward the northwest (reaching 20 m s\(^{-1}\) at 0600 UTC; Fig. 9b), which is consistent with the gradual amplification of the ridge (also Figs. 4–6). Meanwhile, the upper-level trough (as depicted by \( q > 1.5 \) PVU, 1 PVU = \(10^{-6} \text{ K m}^2\text{ kg}^{-1}\text{ s}^{-1}\)) approached from the west and pressed almost against the convective region near the coast of China (25°–30°N) since 0600 UTC (Fig. 9). The time–height section of apparent heat source \( Q_1 \), obtained using Eq. (2) over the rectangular box in Fig. 9a indicates significant heating below 300 hPa to the northeast of Taiwan (i.e., in the region occupied by the upper-level ridge) throughout the period from 1800 UTC 3 May to 0000 UTC 5 May (Fig. 10). On 4 May, the strongest heating occurred at 0600 UTC, reaching nearly 6 K day\(^{-1}\), with maximum upward motion (\(-0.15 \text{ Pa s}^{-1}\)) at 300 hPa. This result is similar to that of tropical cloud clusters obtained by Yanai et al. (1973) and suggests a close linkage between the amplification of the upper-level ridge and the persistent condensational heating in the same region.

Figure 11 shows the 3-h rainfall at the three sites, Naze, Naha, and Ishigakijima (cf. Figs. 1, 9a for locations), to the east-northeast of Taiwan over 3–6 May 1998 and confirms convective rain in the region during this period, in agreement with Figs. 9 and 10. The 3-h rainfall reached 19 mm at 1200 UTC 4 May at Naha and 20 mm at 0000 UTC 5 May at Naze. The layer of 250–400 hPa at Naze, at Naha, and over the rectangular box (cf. Fig. 9a) from ECMWF data all show a tendency to thicken for a day or two (by 20 gpm at most) since 0000 UTC 4 May, except at Ishigakijima, which is likely too far south. In addition to the three island sites shown above, the TRMM satellite-derived rain rates (Huffman et al. 2007) also indicate convective rain (reaching 8–14 mm h\(^{-1}\)) inside the same box region over 4–5 May (not shown).

Thus, over the time period leading to the cutoff, active and persistent convection is shown to result in an amplification of the upper-level ridge immediately downstream (to the east-southeast) of the approaching trough, with simultaneously thickening in the middle to upper layers and diverging ageostrophic wind aloft (Figs. 9–11). These evidences indicate that the latent heating was effective to induce the geostrophic adjustment process (cf. Figs. 4–6; e.g., Carlson 1998, 15–19; Holton 2004). Next, the impact of this ridge to the cutoff process is further assessed through sensitivity test using the MM5. The test includes two runs (starting at 0000 UTC 4 May 1998) for comparison, CTRL (full physics) and FDRY (no latent heat), as described in section 2c.

The 250-hPa (geopotential) height simulated by CTRL is first compared with the ECMWF analysis in Fig. 12. As in Fig. 9, the observed CCL (dashed) cut off from the trough at 0000 UTC 5 May (Fig. 12b) at 250 hPa with an evolution similar to those in Figs. 4 and 5. After cutoff, the low also showed clear axisymmetrization during the next 12 h (Fig. 12c) then moved slowly toward the northeast and weakened (Figs. 12d–f). In CTRL, the 250-hPa height patterns (solid) are generally similar and the narrowing of the trough before cutoff is also evident (Figs. 12a,b). A low with closed contour at the trough base first appears at 0900 UTC 5 May (not shown), whereas the trough south of 33°N also shows clear signs...
to become detached from the midlatitude trough at 1200 UTC (Fig. 12c). At 0000 UTC 6 May, the COL appears near 26°N, 118°E and remains close to Taiwan afterward, so it is to the south-southwest of the observed location with a position error of some 400–500 km (Figs. 12d–f). At 1200 UTC 6 May, the COL in CTRL also becomes more axisymmetrical (Fig. 12e). Besides that the low cuts off later (roughly by 12 h) and is more to the south in the CTRL, other discernible differences include that the downstream ridge is stronger and extends farther toward the northwest and the trough travels faster after cutoff in the ECMWF analysis (Figs. 12b–e). Despite the differences, the MM5 CTRL captures some important features and processes in this case: 1) the clockwise rotation in the axis of the midlatitude trough, 2) the narrowing of the trough leading to the cutoff, 3) the amplification of the upper-level ridge east of the forming COL (Figs. 12b–d), and most importantly 4) the detachment of the trough segment south of about 33°N and the formation of the COL. Because the cutoff process is reproduced by the CTRL, it can be used for further comparison with the result of FDRY, although model errors clearly exist in Fig. 12.

When the latent heating effects are removed, the results of FDRY run (Fig. 13) exhibit several distinct differences from CTRL: 1) the 250-hPa height is lower important features and processes in this case: 1) the clockwise rotation in the axis of the midlatitude trough, 2) the narrowing of the trough leading to the cutoff, 3) the amplification of the upper-level ridge east of the forming COL (Figs. 12b–d), and most importantly 4) the detachment of the trough segment south of about 33°N and the formation of the COL. Because the cutoff process is reproduced by the CTRL, it can be used for further comparison with the result of FDRY, although model errors clearly exist in Fig. 12.

When the latent heating effects are removed, the results of FDRY run (Fig. 13) exhibit several distinct differences from CTRL: 1) the 250-hPa height is lower
without condensational heating, especially over regions of large cloud coverage (cf. Fig. 9); 2) the trough ceases to narrow after 0000 UTC 5 May and even starts to broaden 12 h later; 3) the downstream ridge exhibits no apparent amplification; 4) the southern segment of the trough keeps traveling eastward; and most importantly 5) the cutoff does not occur and the 250-hPa COL does not form (Fig. 13). This is in sharp contrast to CTRL, in

FIG. 10. The time–height section of $Q_1$ (K day$^{-1}$; at intervals of 1 K day$^{-1}$) as computed using Eq. (2) and omega vertical velocity $\omega$ (10$^{-1}$ Pa s$^{-1}$; shading; scale shown at bottom) from ECMWF analysis averaged inside the rectangular boxed depicted in Fig. 9a from 1800 UTC 3 May to 0000 UTC 5 May 1998.

FIG. 11. The 250–400-hPa thickness (gpm; curves with symbols; scale on left) and 3-h rainfall (mm; bars; scale on right) from the three sounding sites, including Naze (46909), Naha (46936), and Ishigakijima (46918), every 12 h from 0000 UTC 3 May to 0000 UTC 6 May 1998. For 250–400-hPa thickness, the curve with the crosses is from ECMWF data averaged over the rectangular box shown in Fig. 9a. For rainfall, white (left), black (middle), and gray (right) bars are for 46909, 46936, and 46918, respectively, and their locations are marked in Figs. 1 and 9a.
which the southern trough segment slows down and becomes detached from the midlatitude trough near 33°N after 0900 UTC 5 May.

Figure 14 presents the difference in 250-hPa height $\Delta z$ between CTRL and FDRY (former minus latter) and the maximum reflectivity in the vertical in CTRL. The locations of convection in CTRL, as judged from the reflectivity, are in good agreement with the GMS-5 IR $T_{BB}$ observation (cf. Figs. 9a,b). Over the region with active convection (>30 dBZ) near Taiwan and to the southwest of Kyushu, Japan (i.e., the area occupied by the upper-level ridge; cf. Figs. 4–6, 9, 13), the positive height difference ($\Delta z > 0$) is clear since 1200 UTC 4 May and enlarges with time (Fig. 14). By 1200 UTC 5 May,
the peak difference has exceeded 100 gpm and is just downstream from the trough in CTRL (thick dashed line) and further reaches 130 gpm northeast of Taiwan at 0000 UTC 6 May (Figs. 14c,d). Thus, as in the diagnosis presented earlier, the MM5 model results also show that the amplification of the downstream ridge northeast of Taiwan was due to diabatic latent heat release; further, the cutoff does not occur without this heating effect in the FDRY experiment (Fig. 13). This suggests that the diabatic heating associated with condensation downstream was a key factor to determine whether the upper-level trough would distort enough to cause separation of its southern segment and the cutoff. Although convection also existed upstream from the trough (Figs. 3, 9), it
was farther away and the height difference produced is quite small in Fig. 14. Thus, diabatic heating upstream found important in some events (e.g., Massacand et al. 2001; Knippertz and Martin 2007) only played a minor role in the present case.

c. Dynamical aspect from Rossby wave breaking

Over the region of the upper-level ridge (manifested as a low-PV feature; cf. Figs. 7, 8d), two processes from the convection can distort PV contours and affect the formation of the CCL, in our case from a kinematic viewpoint. The first (direct effect) is the negative PV transport near the tropopause by the upward motion associated with convection (Davis et al. 1993; Stoelinga 2003), whereas the second (indirect effect) is the negative PV advection caused by the divergent ageostrophic flow. From a dynamical perspective, the cutoff process from a midlatitude trough can be viewed as RWB, which is explained by Postel and Hitchman (2001) through the theory of critical layer. Consider the linearized Rossby wave dispersion equation that is simplified for zonal direction only (Postel and Hitchman 2001; Holton 2004, 215–217),

\[
\frac{c_x}{u} = -\frac{\partial \overline{q}}{\partial y} \frac{1}{k^2 + l^2},
\]

where \(c_x\) is the zonal phase speed of the Rossby wave; \(u\) is zonal wind speed; \(q\) is (geostrophic) PV; \(k\) and \(l\) are the wavenumber in zonal and meridional direction, respectively; and here the overbar denotes east–west average. Note that both \(\pi\) and \(\overline{q}\) vary with latitude. When the Rossby wave enters a critical layer, which means its phase speed \(c_x\) equals the background wind speed \(u\), causing the lhs of Eq. (3) to be zero. Thus, for Eq. (3) to remain valid, either \(\partial \overline{q}/\partial y\) needs to be 0 or \(k\) and/or \(l\) approaches infinity. Because north–south gradient in background geostrophic PV often exists in the upper troposphere (i.e., \(\partial \overline{q}/\partial y \neq 0\); cf. Figs. 7, 9), this condition inevitably cause \(k\) and/or \(l\) to approach infinity and the

Fig. 14. The 250-hPa geopotential height difference (gpm; solid; intervals of 20 gpm) between CTRL and FDRY experiments (CTRL minus FDRY) and maximum reflectivity in the vertical (dBZ; dotted; intervals of 10 dBZ) in the CTRL run every 12 h from (a) 1200 UTC 4 May to (d) 0000 UTC 6 May 1998. The thick dashed lines denote the trough, and a thick open circle in (d) marks the center of the COL in CTRL.
rapid narrowing of the wave until RWB and the formation of the COL. Typically, nonplanetary Rossby waves in midlatitudes move eastward at a phase speed slower than the background westerly flow (Holton 2004, p. 217). When the wave extends south into lower latitudes, its phase speed and background wind speed both reduce. However, if $u$ decreases faster, it is possible for the lhs of Eq. (3) to approach zero (Postel and Hitchman 2001). In the present case, the southward extension of the trough clearly took place. Moreover, the sensitivity test results indicate that the RWB does not happen (i.e., the COL does not form), although narrowing of the trough occurs if the latent heating effect is absent. The difference in 250-hPa wind fields (Fig. 15; CTRL minus FDRY) exhibit an anticyclonic circulation over the East China Sea since 0000 UTC 5 May, in agreement with Fig. 14. Near and east of Taiwan, intensifying southeasterly to easterly winds appear and can reach 10–30 m s$^{-1}$ over 21°–27°N, 120°–130°E (Figs. 15b–d), just downstream and southeast of the approaching trough (cf. Figs. 13b–d). Along the trough, large positive differences in $\zeta$ also exist prior to 1200 UTC 5 May and form a center near 28°N at 0000 UTC 6 May, signifying the development of the COL (Fig. 15; cf. Fig. 14).

Figure 16 shows the zonal mean westerly wind speed $u$ and estimated phase speed of the trough $c_x$ at 250 hPa in CTRL during 4–5 May 1998 (cf. Fig. 12). Since 1200 UTC 4 May, the trough south of 33°N moves at about 3–7 m s$^{-1}$ (thick lines), and the mean background flow (thin lines) over the same latitudes generally increases northward and is above 10 m s$^{-1}$ prior to and at 0000 UTC 5 May (Fig. 16). Since 0600 UTC 5 May, however, a significant reduction in mean flow speed occurs south of 30°N, to a minimum of only 4 m s$^{-1}$ near 27°N at 1200 UTC and clearly in response to the easterly flow induced by latent heating (cf. Figs. 14, 15), and the condition of a critical layer (Postel and Hitchman 2001) is undoubtedly met prior to 1200 UTC. This is consistent with the
occurrence of RWB and the formation of the COL in the present case. In FDRY, the zonal mean flow is stronger south of 30°N than in CTRL, with a minimum of 9–10 m s⁻¹ near 32°N during 0000–1200 UTC 5 May (not shown). This minimum also approaches cₚ as the trough in FDRY exhibits significant narrowing near 32°N as well (cf. Figs. 13a–c). Although the trough evolution in FDRY is also similar to the anticyclonic type of THM93 based on dry dynamics, further narrowing ceases after 1200 UTC and an eventual cutoff does not occur as mentioned (cf. Fig. 13).

Using sensitivity test in numerical models, partitioned PV integration, and piecewise PV inversion technique, Stoelinga (1996) diagnosed the role of nonconserved processes (condensational heating and friction) in the evolution of a midlatitude, baroclinic surface cyclone. He suggested that the convection significantly enhanced the divergent winds and caused the eastward-moving upper-level trough to slow down. In his case, the reduction in speed allowed the trough aloft to maintain a favorable phase difference with the surface cyclone (i.e., former lagging the latter by 1/4 wavelength) and thus the continuous development of the low-level cyclone. In the present case, negative PV anomaly q' (where the mean is the time average over 1–10 May as mentioned in section 3) also appeared in upper levels (maximized at 200 hPa) on the southwest–northeast vertical cross section through the convection (Fig. 17; cf. Fig. 9a for location). This PV and the ageostrophic wind patterns caused by convection (cf. Figs. 9a,b) together would lead to negative PV advection in upper levels (e.g., Davis et al. 1993; Stoelinga 2003) and cause a reduction in eastward-moving speed of the trough as discussed above, much like the trough evolution described by Stoelinga (1996) from a kinematic perspective.

FIG. 16. The distribution of zonally averaged (105°–140°E) westerly wind component u (m s⁻¹; thin lines) at 1200 UTC 4 May (dotted) and 0000 (dashed), 0600 (dot–dot–dashed), and 1200 UTC 5 May (solid) and the estimated zonal phase speed of the trough cₚ (m s⁻¹; thick lines) for 12-h periods of 1200 UTC 4 May–0000 UTC 5 May 1998 (dotted), 0000–1200 UTC 5 May 1998 (dashed), and 1200 UTC 5 May–0000 UTC 6 May 1998 (solid) in CTRL (cf. Fig. 12).

FIG. 17. Vertical cross section of PV anomaly q' (PVU; intervals of 0.1 PVU) along the southwest–northeast-oriented dotted line in Fig. 9a at (a) 0000 and (b) 0600 UTC 4 May 1998. The mean is defined as the 10-day average over 1–10 May. The arrows at the bottom mark that position of high cloud top (low T₉H) along the section at the time in Figs. 9a,b.
5. Conclusions

Relatively rare during the mei-yu season over southeastern China, a CCL and the cutoff process leading to its formation in May 1998 are studied. Compared with conceptual models of KS86 and BK93 for midlatitude upper-level cyclogenesis through migratory jet streak–trough interaction based on dry dynamics, the distinct differences in our case (Figs. 4–7) suggest that the present CCL/COL did not form through such a mechanism. Although the overall evolution of the upper-level trough and the resultant scale of the low (Figs. 7, 8) are in agreement with the THM93 model of interaction between trough and mean westerly flow with anticyclonic background shear, this idealized model does not offer an explanation for the cutoff.

Over the time period leading to the cutoff, evidences are presented to show that active and persistent convection occurred immediately east/southeast of the trough (Figs. 3, 9) and induced divergent ageostrophic flow near the tropopause, an increase in mid-to-upper-level thickness due to effective heating, and concurrent amplification of the upper-level ridge in the region (Figs. 9–11). The impact of latent heating is further assessed through sensitivity tests using the MM5 model, where the cutoff is successfully reproduced in the CTRL (Fig. 12). In the contrasting FDRY run without latent heating effects (Fig. 13), the upper-level trough in the present case is not distorted enough and the COL does not form as observed and in the CTRL. The development of the negative PV anomaly aloft (Figs. 8d, 15, 17) was accompanied by the amplification of ridge downstream from the approaching trough (Figs. 4–6, 9, 14) and can be contributed by the negative PV transport either vertically from lower levels by the convection (direct effect) or horizontally by the induced divergent ageostrophic flow in the upper troposphere (indirect effect; Figs. 10, 11, 14). The negative PV advection can explain the appearance of east/southeast flow just downstream of the approaching trough (Fig. 15) and the subsequent distortion of the trough (with clockwise/counterclockwise rotation of northern/southern trough axis) and cutoff (Figs. 4–6, 9) from a kinematic viewpoint. The cutoff and the formation of the COL can also be explained by the theory of RWB and critical layer through the Rossby wave dispersion equation from a dynamical perspective (Fig. 16). Nonetheless, both the above arguments indicate the key role played by the convection in leading to the cutoff in our case.

Thus, although the present case evolved in a way more similar to the model of THM93 compared to KS86 and BK93, it indeed still exhibited different characteristics with the inclusion of moist processes (i.e., latent heat release). Despite the absence of a dominant mei-yu front during our case period (Fig. 2), as compared to midlatitudes, the abundant moisture over the subtropics near Taiwan in the mei-yu season can often allow a greater impact from organized convection under suitable conditions (e.g., Chen et al. 2003, 2008). Finally, it is noted that the present case can provide one other example for the possible role of diabatic heating on the cutoff process, whereas such effects downstream of the trough near southeastern China were not previously seen in the literature. In the future, studies on the impact of downstream diabatic heating to CCLs/COLs in other seasons (such as winter) over East Asia are also worthwhile.

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