Orographic Effects on Cold Surges and Lee-Cyclogenesis as Revealed by a Numerical Experiment

Part II. Transient Aspects

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Abstract

A numerical experiment using a 10-layer hemispheric dry model produced prominent low-level cold surges and subsequent lee-cyclogenesis on the east and southeast sides of a hypothetical facsimile of the Tibetan Plateau. These phenomena were preceded by a sequence of events that occurred along the western, northern, and eastern periphery of the prescribed mountain.

(1) When a major anticyclone approaches the western end of the mountain, cold air advected from the north accumulates in the lower troposphere due to the barrier effect of the east-west elongated mountain.

(2) The cold air pool is then advected rapidly eastward along the northern periphery of the mountain by the prevailing low-level westerly jet. Associated with this cold air pool is the development of a small-scale edge anticyclone trapped below 700 mb near the north-eastern side of the mountain.

(3) The clockwise (southward) movement of the edge anticyclone is substantially reduced east of the mountain. However, it is still faster than the movement of the edge cyclone which has already passed the eastern periphery, and became nearly stagnant at the southeast corner of the mountain. The phase speed difference between the edge anticyclone to the north and the edge cyclone to the south, results in a sudden increase in the pressure gradient with prominent subgeostrophic northerlies east of the mountain. These subgeostrophic northerlies are responsible for the strong advection of cold air from the north. Here, the advected cold air pool takes a form similar to a gravity-type shock (cold) front.

(4) The edge cyclone begins to shift eastward by the prevailing southwesterlies near the southeast corner of the mountain. The subsequent phase is the amalgamation of the edge cyclone with a major trough which has travelled north of the mountain along about 50°N. Also occurring during this merging phase is the eastward extension of cold front from the southeastern corner of the mountain to as far downstream as eastern China. The merged lee-cyclone intensifies into a major mid-latitude system due to the active baroclinic conversion process occurring around the extended cold front. To the rear of this intensified lee-cyclone are widespread northerly cold surges, which penetrate southward to the South China Sea.

1. Introduction

During the winter, the circulation over East Asia is governed by a dominant high pressure

system over Siberia, which is primarily maintained by radiative cooling. There are frequent monsoonal cold surges bursting out of Siberia which spread over the China continent, and propagate southward as far as the South China Sea, Chang *et al.* (1979) investigated the rela-

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tionship between cold surges and equatorial disturbances developing over the South China Sea. Chang and Lau (1980) also discussed the role of cold surges as it applies to the interaction between the mid- and low-latitudes, and the east-west circulation in the equatorial region.

It is worth mentioning that cold surges are unique to East Asia. This region is also characterized by strong lee-cyclogenesis in the downstream of the Tibetan Plateau (Chung *et al.*, 1976). Frequently in other regions; such as, eastern North America, lee-cyclogenesis occurs, but no strong cold surges are associated with it. Thus, the question arises as to what mechanism is responsible for this unique phenomenon over East Asia.

Recently Lim and Chang (1981) and Lau and Lim (1982) made very informative numerical experiments on cold surges and associated tropical motions. They used shallow water equations on the equatorial β -plane, and examined the response to prescribed mass sources in the mid-latitudes, which they regarded as separated cold air pools from the Siberian high. Interestingly, subgeostrophic northerly surges occur south of the prescribed mass sources and propagate equatorward. The surges reach the Southern Hemisphere summer monsoon region (5°S – 10°S), and eventually induce equatorial Kelvin waves, mixed-Rossby gravity waves, and westward propagating Rossby waves. These waves contribute to the formation of the east-west overturning (Walker circulation), and to the local Hadley circulation. Thus, their studies provide insight on the role of cold surges upon lateral (north-south) and equatorial (east-west) interaction.

However, a fundamental problem still remains unsolved. What mechanisms are responsible for the initial mass sources, and the imbalance between pressure and wind fields? An inspection of daily synoptic maps indicates that the surface high pressure system over East Asia is not steady, but fluctuates substantially in its location and intensity. Frequently, a small-scale high pressure system is separated from the major Siberian high. In fact, the standard deviation of short-period surface pressure fluctuations during winter is quite large over East Asia (Blackmon *et al.*, 1977).

Murakami (1981b and c) investigated the orographic effect of the Tibetan Plateau on transient disturbances in East Asia. He showed that at the northeast corner of the Tibetan Plateau

edge disturbances frequently form, and propagate southward along the eastern periphery of the Tibetan Plateau. Fig. 1 shows an observed map at 850 mb for 12 GMT on March 13, 1979. An anticyclonic cell is located at 37°N , 102°E . To the south and southeast of this high pressure cell are strong northerly surges and cold air advection, forming an NE-SW extended zone of sharp temperature gradient from the southeast corner of the Tibetan Plateau to eastern China and possibly beyond. Note also the existence of an edge cyclone near the southeastern end of the Tibetan Plateau and a major cyclonic system north of the mountain around 50°N , 110°E .

Over the China continent and the East China Sea, strong lee-cyclogenesis often occurs due to amalgamation of the orographically induced edge cyclone and a major mid-latitude cyclonic system, which has propagated eastward north of the Tibetan Plateau across Mongolia and Siberia.

Thus, the frequent occurrence of cold surges and lee-cyclogenesis, unique to East Asia, is probably related to the orographic influence of the Tibetan Plateau, the world's largest massif.

Fig. 1 Geopotential height analysis at 850 mb for 1200 GMT on 13 March 1979 (extracted from the Daily Weather Map of the Japan Meteorological Agency). Solid lines are for geopotential height at 30 m intervals, while dashed lines are temperatures at 6°C intervals.

This possibility will be examined in the Part II of this study. More specifically, we will investigate, by a numerical experiment, the detailed structures of orographically induced edge disturbances, cold surges, and lee-cyclones, and then clarify the mechanisms responsible for their generation.

Another problem to be investigated in Part II is the effect of the Tibetan Plateau on upper tropospheric disturbances when they pass over the mountain region. Staff Members, Academia Sinica (1958a) showed that during winter: (1) transient disturbances in the upper troposphere split to the north and south of the Tibetan Plateau when they approach the mountain; and (2) the northern cell moves much faster to the east than the corresponding southern one, resulting in a pronounced meridional tilt of the axis. These phenomena were confirmed by Murakami (1981b and c) using FGGE level IIIa data. The mechanism responsible for the phase speed difference of upper tropospheric disturbances will be investigated in Section 3.

2. Model and numerical integration

The model used here is identical to the model in Part I of this study (Nakamura and Murakami, 1983). The reader is directed to that paper for a complete development of the model, as only a brief outline will be presented here.

A hemispheric dry model which is similar to an older version of the UCLA general circulation model (Arakawa, 1972) and modified by Nakamura (1978), is used. The model has 10 layers in the vertical and 2.5 degree resolution in latitude-longitude domain. Physics included are: convective adjustment, vertical momentum mixing, and surface drag. Radiative process is simplified to the form of Newtonian cooling, by which the atmospheric temperature is restored to the zonally symmetric, observed winter mean temperature. An elliptical-shaped mountain is introduced as a hypothetical facsimile of the Tibetan Plateau.

Time integration is performed until day 72. The 16-day period from day 21 to 36 is chosen for our detailed analysis of unique disturbances which passed through the mountain region. For convenience, the date within this selected period will hereafter be signified as in the range 0–15 days, rather than 21–36 days. Computed results are stored every 3 hours and interpolated from sigma coordinate to pressure levels. The area of

analysis is confined from 55°E to 135°E and 20°N to 60°N with the mountain being approximately at its center. All variables are separated into two parts: 16-day means ($\bar{}$), and departures from the mean (δ). Thus, transient quantities are defined at 3-hour intervals.

3. Phase propagation and intensity changes of upper troughs and ridges passing through the mountain region

Figures 2 and 3 show latitude-longitude cross-sections of transient geopotential perturbations at 300 mb from hour 12 on day 5 to hour 12 on day 8. When NE–SW extended trough and ridges approach the western end of the mountain, they usually split into two cells. The northern cell shifts northeastward to the north of the mountain and then propagates eastward along about 50°N. The southern part moves slowly eastward along 25°N–30°N, reaching the southwest corner of the mountain where its intensity is reduced. These features are very similar to those of observation (Staff members, Academia Sinica, 1958b, and Murakami, 1981b and c). Since the northern cyclonic (anticyclonic) cell

Fig. 2 Transient geopotential height fields at 300 m. Contour interval is 25 m. Full (dashed) lines represent positive (negative) departures from the 16-day mean values. An elliptical curve indicates the topographic height contour of 3 km. Top: At hour 12 on day 5. Bottom: At hour 12 on day 6.

Fig. 3 As in Fig. 2, except for hour 12 on the 7th (top) and hour 12 on the 8th (bottom). The trajectories and corresponding intensities (m) of negative (low) pressure cells, plotted every 12 hours from day 3 to day 9, are also shown in the bottom diagram.

propagates eastward faster than the corresponding southern cell, the result is a change in the tilt of the trough (ridge) from a NE-SW to an ENE-WSW orientation when it leaves the high elevation zone. This sharp ENE-WSW tilt in the trough and ridge axes results in pronounced northeastward eddy momentum fluxes in the lee of the mountain. These momentum fluxes are directed downgradient of the 16-day mean winds and, thus, transient eddies feed their kinetic energy to 16-day mean westerlies [Figs. 11 (top) and 13 (top) of Part I].

Fig. 3 (bottom) demonstrates the changes in intensity and phase speed of the northern and southern cyclonic cells along the trajectory. The intensity of the northern cell becomes weak when approaching the mountain region, but it reintensifies markedly after leaving the mountain. This reintensification is related to low-level cold surges and cyclogenesis in the lee of the mountain. This point will be elaborated on later.

Fig. 3 (bottom) also shows that the phase speed of the northern cell is about 10° long/day or 8.4 ms^{-1} , while that of the southern cell is about 6° long/day or 6.8 ms^{-1} . In contrast, the 16-day mean zonal winds at 300 mb increase

considerably from about 20 ms^{-1} at 45°N to approximately 30 ms^{-1} at 25°N . Here arises a question as to why the phase propagation of 300 mb perturbations south of the mountain is much slower than that of the northern counterpart. This problem may be addressed by estimating the value of each term in the vorticity equation. Following Murakami (1981b and c), the vorticity equation for ζ' perturbations may be written as:

$$\frac{\partial \zeta'}{\partial t} = A + B + C + D + E + F + R, \quad (1)$$

where

$$A = -\left(\bar{u} \frac{\partial \zeta'}{\partial x} + \bar{v} \frac{\partial \zeta'}{\partial y} + \bar{\omega} \frac{\partial \bar{\zeta}}{\partial p}\right), \quad (2)$$

$$B = -\left(u' \frac{\partial \bar{\zeta}}{\partial x} + v' \left(\beta + \frac{\partial \bar{\zeta}}{\partial y}\right) + \omega' \frac{\partial \bar{\zeta}}{\partial p}\right), \quad (3)$$

$$C = -\left(u' \frac{\partial \zeta'}{\partial x} + v' \frac{\partial \zeta'}{\partial y} + \omega' \frac{\partial \zeta'}{\partial p}\right), \quad (4)$$

$$D = -((f + \bar{\zeta})\delta)', \quad (5)$$

$$E = -\left(\frac{\partial \omega}{\partial x} \frac{\partial v}{\partial p} - \frac{\partial \omega}{\partial y} \frac{\partial u}{\partial p}\right), \quad (6)$$

$$F = \frac{\partial}{\partial p} \left(k_M \frac{\partial \zeta'}{\partial p}\right), \quad (7)$$

where K_M denotes the vertical momentum mixing coefficient. The other variable are listed in Appendix.

The A term represents the advection of ζ' -perturbations by the 16-day mean ($\bar{u}$, $\bar{v}$, $\bar{\omega}$) wind fields and the B term denotes the advection of mean absolute vorticity ($f + \bar{\zeta}$) due to the transient components in of winds (u' , v' , ω'). The C term indicates the ζ' advection due to transient components of winds, while the D term represents the divergence effect and the E term corresponds to the twisting effect associated with transient perturbations. The F term represents frictional dissipation of vorticity.

The A , B , C , D , E , and F terms are measured by utilizing transient quantities at time n . Thus, the sum of these terms at time n should be equal to the finite difference tendency term, i.e., $(\zeta'_{n+1} - \zeta'_{n-1})/(2\Delta t)$ where Δt denotes a time increment of 5 minutes in the present numerical integration. Unfortunately, computed transient quantities are stored only at 3-hour intervals. This inhibits an exact measurement of the $\partial \zeta'/\partial t$ tendency term at time n , therefore, it is approximated by $\delta \zeta'/\delta t$ where $\delta \zeta'$ represents the difference in ζ' perturbations during the 6-hour

time interval ($\delta t=6$ hours). Consequently, there is an imbalance between this approximated $\partial \zeta' / \partial t$ value and the sum of A , B , C , D , E , and F terms. In eq. (1), R signifies this imbalance and is evaluated as a residual. The residual R thus estimated is substantially large in the lower troposphere, in particular near the mountain. This is due to the occurrence of gravity-type, short-period (less than 6 hours) oscillations near the mountain. In comparison, R is quite small in the upper troposphere above about 500 mb.

Computed results for each term in eq. (1) at 300 mb along 45°N and 25°N at hour 00 on the 6th are summarized in Figs. 4 and 5, respectively. Both the friction term F and the residual term R are small when compared to the other quantities, consequently, they are not reproduced here. At 45°N (Fig. 4), an approximate in-phase relationship exists between the $\partial \zeta' / \partial t$ tendency term and the A amplitude, the difference between these two terms should be compensated for by other factors. Interestingly, the B factor is inversely correlated with the A term; namely, the B term effectively reduces the excessive advection effect A . The sum of the nonlinear advection term C and divergence term D also con-

Fig. 4 Longitudinal profiles of transient vorticity ζ' (10^{-5} s^{-1}) and the terms (10^{-10} s^{-2}) in eq. (1), measured at 300 mb for hour 00 on day 6 along 45°N . Top: ζ' and $\partial \zeta' / \partial t$. Middle: A and B . Bottom: C , D , and E .

Fig. 5 As in Fig. 4, except for 25°N .

tributes significantly to the reduction of the advection effect A . In comparison, the E term contributes little to the determination of the phase speed and amplitude changes of 300 mb vorticity perturbations at 45°N .

At 25°N (Fig. 5), the amplitude of ζ' perturbations is slightly larger than the corresponding amplitude at 45°N . Yet, the amplitude of $\partial \zeta' / \partial t$ tendency at 25°N is nearly equal to, or even smaller than, the counterpart at 45°N . This is due to the fact that the phase speed of ζ' perturbations at 25°N is much slower than that at 45°N . The smallness of $\partial \zeta' / \partial t$ tendency at 25°N results in a large difference between $\partial \zeta' / \partial t$ and A terms. The B term contributes the most to balance the required $(\partial \zeta' / \partial t - A)$ difference. In fact, the B term at 25°N is approximately twice as large as that at 45°N . This can be explained as follows: $v'(\beta + \partial \bar{\zeta} / \partial y)$ represents the largest contribution to the B term and the cumulative β effect, i.e., $(\beta + \partial \bar{\zeta} / \partial y)$ is much more predominant at 25°N (south of the $\bar{u}$ jet; $\partial \bar{\zeta} / \partial y > 0$) than the corresponding value at 45°N (north of the $\bar{u}$ jet; $\partial \bar{\zeta} / \partial y \leq 0$). Accordingly, the cumulative β effect has a strong influence on the phase propagation of transient ζ' perturbations north and south of the mountain. The difference in the phase speed between these two regions is associated with the structure of the 16-day mean wind fields, which are primarily determined by the mechanical effect of the prescribed mountain. Compared with the B term, the nonlinear advection term C and the diver-

gence effect D are less significant at 25°N . The term D includes the divergence effect associated with forced vertical motions due to topography. This divergence effect is customarily signified as the "topographically-induced β -effect." The smallness of the D term in Fig. 5 indicates that the topographically-induced β -effect is of secondary importance for the phase propagation at 25°N .

Staff members, Academia Sinica (1958a) investigated wave propagation over and around the Tibetan Plateau, and noted a large difference between phase propagations north and south of this massive mountain. Based on a barotropic potential vorticity equation, they emphasized the importance of the topographically-induced β -effect. On the other hand, by computing each term in the vorticity equation for transient disturbances at 200 mb along 45°N and 25°N during the 1978-79 winter, Murakami (1981b and c) found a large difference in the cumulative $(\beta + \partial \zeta / \partial y)$ effect between the two

latitudes, which favors the slow (fast) phase speed to the south (north) of the Tibetan Plateau. Thus, the present numerical experiment reconfirms Murakami's observational findings.

4. Edge disturbances, lee-cyclogenesis, and cold surges in the lower troposphere

Compared to the upper troposphere, the nature of wave propagation in the lower troposphere is more complicated. This can be seen in Figs. 6 and 7, which depict latitude-longitude cross-sections of geopotential perturbations at 850 mb from hour 12 on day 5 to hour 12 on day 8, respectively. On day 5, an anticyclonic cell at 45°N , 65°E propagates northeastward approximately following the zero contour height of the mountain (not shown). On day 6, this cell reaches 50°N , 78°E and at the same time, a small-scale anticyclonic system is induced on the northeast side of the mountain. Between day 6 and 8, this orographically-induced, anticyclonic edge disturbance moves clockwise along the eastern boundary of the mountain, reaching the southeastern end (35°N , 100°E) at hour 12 on the 8th (Fig. 7, bottom).

In Fig. 6 (top), there are also two cyclonic systems; one is a major trough system at 55°N ,

Fig. 6 Transient geopotential height fields at 850 mb. Contour interval is 20 m. Full (dashed) lines represent positive (negative) departures, respectively. An elliptical curve indicates the topographic height contour of 1.5 km. Top: At hour 12 on day 5. Bottom: At hour 12 on day 6. Trajectory and corresponding intensity of a major high pressure cell, plotted every 12 hour from day 5 to day 8, are also shown in the top diagram.

Fig. 7 As in Fig. 6, except for hour 12 on day 7 (top) and hour 12 on day 8 (bottom). Trajectory and intensity of a major low pressure cell are shown in the bottom diagram.

90°E and the other a small-scale, orographically-induced, edge cyclonic system east of the mountain at 35°N, 100°E. From day 6 to 7, these cyclonic systems merge and intensify, and by day 8 become a single, pronounced cyclone, i.e., lee-cyclogenesis. The trajectory of the major cyclonic system with its intensity changes is shown in Fig. 7 (bottom).

Fig. 8 shows longitude-height cross-sections of geopotential perturbations along 35°N. At hour 12 on day 4 (Fig. 8, top), a major 300 mb trough system located at 72°E approaches the mountain zone and a small-scale, low-level cyclonic system is induced at the eastern periphery of the mountain. On day 6 (Fig. 8, bottom), the major trough system has moved over the mountain and the small-scale, orographically-induced cyclonic cell is confined to below 700 mb. This orographically-induced, edge cyclone moves eastward and merges vertically with the major trough by day 8 (not shown).

Figs. 9–11 are longitude-latitude cross-sections of wind, geopotential, and temperature fields at 850 mb east of the mountain. At hour 12 on

day 6 (Fig. 9, bottom) a small-scale edge anticyclone is generated near the northeast corner of the mountain. Interestingly, this high pressure cell is embedded within a zone of orographically-induced anticyclonic shear (not within an anticyclonic circulation) of the 850 mb wind field. South of this edge anticyclone, strong non-geostrophic northerlies are crossing the contour heights from high to low. These sub-geostrophic northerlies are responsible for strong advection of cold air from the north. In Fig. 10 (top-right), note the pronounced temperature gradient near the southeastern periphery of the mountain. Between hour 00 and 12 on day 7, this area with steep temperature gradient is displaced southward along the mountain. On this southeast side of the mountain, the temperature perturbations (Fig. 10, bottom-right) exhibit characteristics similar to those in observed cold fronts. In the present numerical experiment, a cold front becomes first established in association with the development of a low-level edge anticyclone which is excited by the mechanical (not thermal) effect of the mountain. The cold front further intensifies due to the subsequent enhancement of cold northerly surges south of the edge cyclone. Between day 7 and 8, this cold front extends northeastward from the southeastern periphery of the mountain to as far downstream as around 40°N, 120°E (Fig. 11, bottom-right). Associated with this NE–SW extended cold front are strong northerly surges covering an extensive region to the rear of the major lee-cyclone centered around 45°N, 125°E (Fig. 11, bottom-left). In Section 7, there will be further discussion pertaining to the relationship between the development of this NE–SW extended cold front and the strong cyclogenesis occurring in the lee of the mountain.

5. Low-level temperature perturbations associated with edge disturbances along the mountain

To investigate the nature of the edge disturbances which move along the mountain periphery, an elliptical grid which parallels the mountain's 1.5 km contour (but 400 km outward from it) was selected. Fig. 12 shows time sections of the 850 mb temperature and 200 mb geopotential fields along the elliptical grid. In Fig. 12 (left), the low temperature (dashed, negative) system at 850 mb first becomes established near the western end of the mountain around

Fig. 8 Longitude-height sections of transient geopotential height along 35°N. Contour interval is 20 m. Solid (dashed) lines are positive (negative), respectively. Stippled area denotes the mountain. Top: At hour 12 on day 4. Bottom: At hour 00 on day 6.

Fig. 9 Longitude-latitude sections of transient winds, geopotential height, and temperature fields at 850 mb east of the mountain. Left: Wind and geopotential height fields. Winds (barbs) are in units of 5 ms^{-1} . Interval of geopotential height is 20 m. Solid (dashed) lines are positive (negative), respectively. Right: Temperature fields at intervals of 4°C . Right (left)-slanted hatching denotes area of northerly (southerly) wind components exceeding 5 ms^{-1} . Top: At hour 00 on day 6. Bottom: At hour 12 on day 6.

gridpoints 1 and 2 at about hour 12 on day 5. This low temperature system then splits, with the northern section traveling clockwise from gridpoints 1 and 2 (west), via gridpoints 7 to 13 (north and east), to as far as gridpoints 18 and 19 (south) around the mountain. Thus, Fig. 12 (left) reconfirms the clockwise movement of orographically induced temperature perturbations at 850 mb. The southern section of the low temperature system moves from gridpoints 1 and 2 counterclockwise to gridpoints 18 and 19 where it merges with the northern section and leaves the gridpoint region sometime between hour 12 on the 10th and hour 12 on the 11th. The low temperature system dissipates immediately after leaving the gridpoint region (refer to Fig. 21 of Part I).

Fig. 12 (right) merely reflects the eastward propagation of the NE–SW oriented 200 mb

troughs and ridges across the mountain (refer to Fig. 2 and 3). In Fig. 12 (right), the low pressure cell (dashed, negative) at 200 mb first enters the gridpoint region at gridpoints 3 and 4 (rather than 1 and 2) at about hour 00 on day 5. This is associated with the NE–SW tilt of the 200 mb trough, when it reaches the western end of the mountain (Fig. 2, top). In Fig. 12 (right), note that after hour 00 on day 5, the northern section of the cyclonic system propagates clockwise half-way around the gridpoint region from gridpoints 3 and 4 (west-northwest), via gridpoint 7 (north), to gridpoints 14 and 15 (east-southeast). Conversely, the southern section travels counterclockwise half-way around the gridpoint region from gridpoints 3 and 4 to gridpoints 14 and 15, there merging with the northern section. This counterclockwise movement from gridpoints 3 and 4 to gridpoints 14 and 15 actually cor-

Fig. 10 As in Fig. 9, except for hour 00 (top) and hour 12 (bottom) on day 7.

responds to the eastward phase propagation of the NE–SW oriented 200 mb trough south of the mountain.

In Fig. 12 (left), the 850 mb temperature southeast of the mountain (gridpoints 14 and 15) decreases sharply at about hours 00–12 on the 7th. This can be more clearly seen in Fig. 13 (top), which shows a temperature drop of about 15°C occurring between hour 21 on day 6 and hour 09 on the 7th for 30°N, 102.5°E. As noted earlier (Fig. 10, top), this rapid temperature decrease is associated with the southward passage of a frontal-type system and cold air advection by northerly surges. In Fig. 13 (top), note that the temperature drop begins nearly simultaneously with the changes in v' winds from southerly (positive) to northerly (negative). To further confirm this point, we look at the thermodynamic equation for transient T' -perturbations which may be written as:

$$\frac{\partial T'}{\partial t} = I + J + K, \quad (8)$$

where

$$I = -\left(\bar{u} \frac{\partial T'}{\partial x} + \bar{v} \frac{\partial T'}{\partial y} + \bar{\omega} \left(\frac{\partial}{\partial p} - \frac{\kappa}{p} \right) T'\right), \quad (9)$$

$$J = -\left(u' \frac{\partial \bar{T}}{\partial x} + v' \frac{\partial \bar{T}}{\partial y} + \omega' \left(\frac{\partial}{\partial p} - \frac{\kappa}{p} \right) \bar{T}\right), \quad (10)$$

$$K = -\left(u' \frac{\partial T'}{\partial x} + v' \frac{\partial T'}{\partial y} + \omega' \left(\frac{\partial}{\partial p} - \frac{\kappa}{p} \right) T'\right). \quad (11)$$

Here, the I term represents the advection of T' -perturbation by the 16-day mean wind fields, while the J term denotes the advection of 16-day mean T fields by transient components of winds. The K term, which is nonlinear with respect to transient perturbation quantities, represents the advection of T' -perturbations by transient components of the wind fields.

As in the ζ' vorticity equation (1), the $\partial T'/\partial t$ tendency term is approximated by the centered finite difference scheme using T' values at 3-hour intervals. For brevity, the residual due to the imbalance between the approximate tendency term and the sum of I , J , and K terms, is

Fig. 11 As in Fig. 9, except for hour 00 (top) and hour 12 (bottom) on day 8.

Fig. 12 Elliptical grid along the mountain (lower center box) and time sections of temperature at 850 mb (left) and geopotential height at 200 mb (right) between day 0 and day 15 along the elliptical grid points. Temperature contour interval is 4°C and geopotential height interval is 20 m. Solid (dashed) lines are positive (negative), respectively. Arrows indicate the lines of maximum and minimum.

Fig. 13 Time changes of temperature, wind, and vertical p -velocity for 850 mb level at 30°N , 102.5°E . Grid position is shown by dot in lower right-hand box. Top: T' and ω' . Middle: u' and v' . Bottom: I , J , K terms in eq. (8). Solid line denotes I term; dashed line, J term; and solid line with cross, K term.

omitted in eq. (8). (We are aware of the importance of the residual in the thermodynamic equation. This point, which is beyond the scope of the present study, will be reported elsewhere.) The diabatic heating (Newtonian cooling) effect is also disregarded in eq. (8).

The bottom portion of Fig. 13 depicts the computed results of each term in eq. (8) at 3-hour intervals for 30°N , 102.5°E . The K factor is predominantly negative between hour 00 and 06 on day 7. Hence, the K factor (cold air advection due to non-geostrophic northerly surges) is primarily responsible for the sharp temperature decrease southeast of the mountain.

It should be noted that the K term merely represents the advection of T' -perturbations from one location to another by the transient wind fields, and there is still the question of how and where such low T' -perturbations are generated. In Fig. 12 (left), note that the low T' cell first intensified near gridpoints 2 and 3 at about hour 06 on the 5th. These low T' -perturbations propagate clockwise to the east of the mountain, while continuing to maintain their intensity. Fig. 14 (top) demonstrates how the initial tempera-

Fig. 14 Time changes of 850 mb temperature and terms in eq. (8), measured at 40°N , 77.5°E (top) and 47.5°N , 90°E (bottom). Right-hand box shows grid position.

ture decrease occurs near the northwestern end of the mountain at 40°N , 77.5°E . It shows that the J effect is the dominant factor for the temperature decrease between hour 00 on day 4 and hour 12 on day 5. During this period, northerly transient winds which carry low T temperatures from the north are the biggest contributor to the T' decrease. At hour 12 on day 5, a high pressure cell at 850 mb is located near 43°N , 66°E (Fig. 6, top). East of this high pressure cell is a strong southward advection of the 16-day mean, high-latitude temperature fields by the transient components of the northerly winds. Due to the mechanical blocking effect of the mountain, the cold air pool generated by transient northerly winds cannot move southward. Rather, it is forced to move eastward by the 16-day mean westerlies at 850 mb, which prevail along the northern periphery of the mountain. This can be confirmed in Fig. 14 (bottom) which indicates that at 47.5°N , 90°E , the largest contributor to the temperature decrease is the I factor. That is, the temperature decrease is primarily associated with the eastward advection of low T' -perturbations by the 16-day mean westerly winds prevailing to the north of the mountain (refer to Fig. 2 of Part I). The J and K effects also contribute to the temperature decrease at gridpoint 47.5°N , 90°E , however, not

to the extent of the I effect.

6. Low-level vorticity perturbations associated with edge disturbances along the mountain

In the previous sections we emphasized that non-geostrophic northerly surges are primarily responsible for rapid 850 mb temperature decreases near the southeast corner of the mountain. This section investigates the mechanisms through which these strong northerly surges intensify along the eastern periphery of the mountain.

Figs. 15 and 16 present the vertical profiles of transient vorticity perturbations along the elliptical grid from hour 00 on day 5 to hour 00 on day 8. In Fig. 15 (top), a pronounced cyclonic center is located near 300 mb at gridpoint 3. This 300 mb cyclonic cell propagates clockwise around the mountain reaching gridpoint 10 on the 7th. The cell continues to propagate clockwise until hour 00 on day 8 when it leaves the gridpoint region. Similarly, an anticyclonic cell at 250 mb located near gridpoint 18 on the 5th moves counterclockwise to gridpoint 15 on day 7 and leaves the gridpoint region sometime

Fig. 15 Vertical sections of transient vorticity at the elliptical grid around the mountain. Contour interval is $1 \times 10^{-5} \text{ s}^{-1}$. Hatching denotes area of anticyclonic vorticity. Top: At hour 00 on day 5. Bottom: At hour 00 on day 6.

Fig. 16. As in Fig. 15, except at hour 00 on day 7 (top) and hour 00 on day 8 (bottom).

between day 7 and 8. Thus, these figures reconfirm the movement of upper tropospheric disturbances with respect to the elliptical grid as shown earlier in Fig. 12 (right).

Let us examine vorticity perturbations in the lower troposphere. On day 5 (Fig. 15, top), a weak anticyclonic cell at 850 mb is located near gridpoint 8 and a cyclonic cell trapped below 700 mb is centered at gridpoint 12. Between days 5 and 6, the low-level edge anticyclone makes a rapid movement along the mountain to gridpoint 10, while the low-level edge cyclone moves much slower from gridpoint 12 to 13. Then from day 6 to 7, the clockwise movement of the edge anticyclone is substantially reduced, however, it is still slightly faster than the movement of the edge cyclone which has become nearly stagnant around gridpoints 14–15. Interestingly, this period also coincides with the development of the non-geostrophic (sub-geostrophic) northerly winds between the orographically-induced cyclonic and anticyclonic edge disturbances east of the mountain (Figs. 9 and 10). It appears that the phase difference between the edge anticyclone to the north and the cyclonic edge cell to the south results in an imbalance of the pressure fields with the associated wind fields. It is then hypothesized that the sub-geostrophic

northerly surges are induced through a pressure-to-wind adjustment process.

To further investigate these phase speed changes, a time-section of vorticity perturbations at 850 mb is presented in Fig. 17 (top-left). It reveals significant changes in phase speed of 850 mb vorticity perturbations at around gridpoint 11. The clockwise phase speed is much faster to the north of the mountain (gridpoints 6 to 11) than to the east and south of the mountain ((gridpoints 12 to 18).

Computed results for some of the terms in the ζ' -vorticity equation (1) are shown in Figs.

Fig. 17 Time sections of transient vorticity and terms in eq. (1) at 850 mb along elliptical grid points north and east of the mountain. Top-left: vorticity at intervals of 10^{-5} s^{-1} . Solid (dashed) lines denote cyclonic (anticyclonic) vorticity, respectively. Arrows indicate lines of maximum and minimum. Top-right: Vorticity tendency at intervals of $2 \times 10^{-10} \text{ s}^{-2}$. Solid (dashed) lines are positive (negative), respectively. Hatching denotes areas of greater than $2 \times 10^{-10} \text{ s}^{-2}$. Bottom-left: Term A. Bottom-right: Term D. Intervals and hatching are the same as those of top-right.

17 and 18, respectively. Between gridpoints 6 and 11 the $\partial\zeta'/\partial t$ tendency is well correlated with the advection effect A . The divergence effect D is also correlated with the $\partial\zeta'/\partial t$ term, but with much smaller amplitudes than the A term. Since substantially strong 16-day mean westerlies prevail at 850 mb to the north of the mountain (Fig. 2 of Part I), the advection term A contributes the most to the $\partial\zeta'/\partial t$ tendency between gridpoints 6 to 11.

In comparison, the advection term A becomes less significant to the east and south of the mountain (gridpoints 12 to 18), where the 16-day mean winds at 850 mb are generally weak. Over this region, the $\partial\zeta'/\partial t$ tendency term does not show systematic phase propagation, but a pair of large positive and negative $\partial\zeta'/\partial t$ values are seen around gridpoints 13 to 15 during the period from hour 00 on the 6th to hour 00 on day 7. This pattern reflects the development, decay, and subsequent eastward propagation of an edge cyclone away from the elliptical grid region (refer to Figs. 9 and 10). It appears that the divergence term D , i.e., $D = -((f + \zeta)\delta)'$, contributes the most to the development of the edge cyclone around hour 12 on day 6 (Fig. 17, bottom-right). The friction term F , which is mainly due to surface friction, is a major contributor to the decay of the edge cyclonic cell at around hour 00 on day 7 (Fig. 18, top-right).

In Fig. 17 (top-left), note the systematic clockwise propagation of an anticyclonic ($\zeta' < 0$) edge disturbance from gridpoint 11 at hour 12 on day 6 to gridpoint 14 at hour 12 on day 9. However, the $\partial\zeta'/\partial t$ pattern (Fig. 17, top-right) does not well reflect this clockwise (southward) propagation of the edge anticyclone. On the other hand, the $(\partial\zeta'/\partial t - \partial(K_M \partial\zeta'/\partial p)/\partial p)$ pattern shown in Fig. 18 (bottom-left) more clearly indicates this southward propagation of the edge anticyclone. Fig. 17 (bottom-right) exhibits a distinct band of large negative D values from about gridpoint 11 at hour 12 on the 6th to gridpoint 13 at hour 00 on day 9. This divergence, which is induced by the southward acceleration of non-geostrophic northerlies ahead of the edge anticyclone (see Fig. 10, top-left), favors the southward (clockwise) propagation of the anticyclone cell along the eastern periphery of the mountain. It also appears that the nonlinear advection term C' (Fig. 18, top-left) is as important as the D effect for the clockwise propagation of the edge anticyclone. Near the eastern periphery of the

Fig. 18 Time sections of terms in eq. (1) at intervals of $2 \times 10^{-10} \text{ s}^{-2}$ along elliptical grid points at 850 m. Hatching denotes area of greater than $2 \times 10^{-10} \text{ s}^{-2}$. Top-left: Term C. Top-right: Term F. Bottom-Vorticity tendency minus term F. Bottom-right: Residual.

mountain, the twisting term E (not shown) occasionally becomes substantial, and contributes significantly to the changes in intensity and phase speed of the edge disturbances.

In Fig. 18 (bottom-right), of particular interest is the residual term R , which happens to be of the same order of magnitude as the major terms, such as, the A and D terms. For example, a large negative R at hour 15 on day 6 around gridpoint 14 nearly offsets the large positive divergence effect D . Likewise, a dominant positive R near gridpoint 10 at hour 00 on the 10th approximately cancels the large negative A advection. The residual R is subject to some computational errors in the interpolation of transient quantities from the sigma coordinate to pressure levels, and also in the evaluation of each term in eq. (1) from a finite difference scheme. How-

ever, these computational errors are expected to be small and random with regard to space and time. Nevertheless, in Fig. 18 (bottom-right), we see that R is quite large in magnitude with well-organized features with respect to time and grid-point. Here, recall that the $\partial \zeta' / \partial t$ tendency term was evaluated using a 6-hour interval finite difference scheme; namely, it represents an average rate of ζ' change during the 6-hour period. Consequently, the approximated $\partial \zeta' / \partial t$ tendency term does not balance with the sum of A , B , C , D , and E terms, which were computed with instantaneous transient variables at 3-hour intervals. The R term represents this imbalance and is evaluated as a residual in eq. (1). Thus, it is hypothesized that R can be regarded as the Reynolds' stress effect ($u' \zeta'$, etc.) due to fluctuations with periods less than 6 hours. The present numerical integration scheme with a time increment of 5 minutes is not free from gravity-type oscillations. In particular, short-period gravity waves are frequently excited in the immediate vicinity of the mountain. It is highly probable that these gravity waves play an important role in phase propagation and the changes in intensity of ζ' vorticity perturbations associated with edge disturbances around the mountain. An extensive study is now in progress to measure an exact contribution due to gravity-type, short-period oscillations. An attempt will also be made to estimate computational errors due to conversion from σ to p . Computed results will be reported elsewhere.

Next, the vertical structural features of edge disturbances are examined. Fig. 19 (top) depicts the existence of a pair of pronounced anticyclonic and cyclonic ζ' vorticity cells at grid-points 11 and 14, respectively, with their maximum at 900 mb. [Also refer to Fig. 9 (bottom-left) for the corresponding vorticity pair at 850 mb along the eastern periphery of the mountain.] Associated with the anticyclonic ζ' cell is a predominance of divergence in the shallow layer below about 800 mb, and a strong convergence above that level (Fig. 19, bottom). The immediate effect of these divergence and convergence fields is to intensify (weaken) the anticyclonic cell below (above) 800 mb. Similarly, the cyclonic vorticity system located near grid-point 14 tends to develop (weaken) in the lower (upper) portion due to the associated convergence (divergence). Through these dynamical processes, both the anticyclonic and cyclonic edge systems

Fig. 19 Vertical sections of vorticity at intervals of 10^{-5} s^{-1} (top) and divergence at intervals of $2 \times 10^{-6} \text{ s}^{-1}$ (bottom) along elliptical grid points at hour 15 on day 6. Solid (dashed) lines are positive (negative). Stippling indicates the mountain.

become predominant (trapped) only in the shallow layer below 800 mb after hour 15 on day 6.

In Fig. 20 (top), the high geopotential cell centered at 900 mb near gridpoint 11 takes a form similar to a gravity-type shock front with strong cross-isobar flows and updrafts to its south. The horizontal component of winds is directed southward, crossing the geopotential height from high to low. This indicates strong southward $v'\phi'$ geopotential energy fluxes due to nongeostrophic northerlies. As shown in Part I, these geopotential energy fluxes are primarily responsible for the development and/or maintenance of edge disturbances along the eastern periphery of the mountain. Yet another feature of interest in Fig. 20 is the association of northerlies (southerlies) with low (high) temperature below about 800 mb northward (southward) of gridpoint 13, implying northward sensible heat fluxes due to edge disturbances. Between gridpoints 10 and 12 (13 and 15), upward motions are associated with cold (warm) temperature. Further to the south from gridpoint 16 to 18, one finds the association of descending motions with warm temperature. When averaged from gridpoint 6 to 18, the product between ω' and T' is slightly positive. This is congruent with the finding in Part I that the baroclinic conversion process contributes little to the development of edge disturbances trapped below about 800 mb.

Fig. 20 Vertical sections of geopotential height, winds, and temperature along elliptical grid points at hour 15 on day 6. Top: Geopotential height at an interval of 10 m. Vectors are composed of horizontal velocity components parallel to 1.5 km contour of the mountain with unit vectors of 10 ms^{-1} and vertical p -velocity with unit vectors of $-10^{-3} \text{ mbs}^{-1}$. Bottom: Temperature at intervals of 2°C . Solid (dashed) lines are positive (negative), respectively.

7. Lee-cyclogenesis

The next problem to be investigated is the lee-cyclogenesis that occurred between day 6 and day 9. Lee-cyclogenesis indicates a clear distinction between the initial phase and the subsequent phase. The initial phase was characterized by the clockwise propagation of a low-level, edge cyclonic cell along the eastern periphery of the mountain, reaching gridpoint 13 at hour 00 on the 6th (Fig. 9, top-left). Between hour 00 and 15 on day 6, this cyclonic cell intensifies due to strong low-level convergence south of the cold northerly surges. This can be confirmed more clearly in Fig. 19, which exhibits the strong association of low-level cyclonic vorticity with existing convergence below 800 mb. After reaching gridpoint 15 at hour 21 on day 6, the edge cyclone begins to shift eastward by the 16-day mean southwesterlies near the southeast corner of the mountain. However, the vorticity perturbation in the form of cyclonic shear (not circulation) keeps propagating clockwise along the southern periphery of the mountain and reaches gridpoint 16 on day 9 (not shown).

The subsequent phase is the amalgamation of

the edge cyclone with a major trough in the lee of the mountain. In Fig. 9 (bottom-left), a major cyclone is located at 55°N , 100°E , while the edge cyclone is centered near 34°N , 104°E .² The two cyclones become merged by hour 00 on day 8 (Fig. 11, top-left), and intensify into a major mid-latitude cyclone at 850 mb, reaching maximum intensity at hour 00 on day 9. Refer also to Fig. 3 (bottom) for the changes in intensity along the trajectory of the major cyclonic system at 300 mb.

The two cyclonic systems also become coherent in the vertical between days 6 and 8. In Fig. 8 (bottom), note the existence of two separate cyclonic cells. The major cell is centered at 300 mb to the west of the mountain around 82°E , while the orographically-induced edge cell is located at 850 mb east of the mountain near 100°E . Vertical merging of these two cyclonic cells occurs around hour 12 on day 7, when the upper trough has already passed through the mountain region and the low-level edge cyclone has moved eastward away from the eastern end of the mountain (not shown).

Also occurring during this merging period is the establishment of a distinct NE–SW zone of strong temperature gradient (Fig. 11, bottom-right), which was initially generated by cold surges east of the mountain on day 7 (Fig. 10, top-right). Between hour 00 and 12 on the 8th, cold northerly surges to the rear of the merged cyclone contribute to the eastward extension of the cold front, which in turn contributes to further intensification of the cyclone. The intensified cyclone tends to enlarge the region of strong northerly surges and the associated cold front further eastward. Thus, via these mutual interaction processes, the merged lee-cyclone intensifies into a major mid-latitude system over northern China (45°N , 125°E) at hour 00 on the 9th. Associated with this lee-cyclone are prominent northerly surges penetrating as far

Fig. 21 Vertical sections of temperature, winds, and divergence at hour 12 on day 8 along the line A–O–B in Fig. 11 (bottom-right). Top: Temperature at intervals of 2°C . Vectors are composed of horizontal velocity component parallel to the line A–O–B with unit vectors of 10 m s^{-1} and vertical p -velocity with unit vectors of 10^{-3} mb s^{-1} . Bottom: Divergence at intervals of $2 \times 10^{-6}\text{ s}^{-1}$. Solid (dashed) lines are positive (negative), respectively.

south as 20°N and possibly beyond to the South China Sea (Fig. 11, bottom-left).

In Fig. 21, we investigate the vertical structural features of the cold front. It shows the distribution of (u', ω') wind vectors, temperature, and divergence at hour 12 on day 8 along the A–B axis, which is nearly perpendicular to the cold front (see Fig. 11, bottom). Fourteen gridpoints numbered from -7 to 6 are chosen along the A–B line. Gridpoint 0 coincides with the cold front's maximum temperature gradient at 950 mb. The grid interval is approximately 200 km. Between gridpoints -7 and 0 , there exists a deep layer of cold air with the lowest temperature (less than -12°C) at 950 mb occurring near gridpoint -4 . Associated with this deep cold air dome are pronounced descending motions throughout the lower troposphere, and a deep layer of divergence extending from 600 to 950 mb (Fig. 21, bottom). This deep divergence layer is capped by a convergence layer overhead at about 500 mb (not shown). Thus, the vertical scale of the divergence fields associated with the cold front is on the order of 1000 mb. In comparison, the corresponding vertical scale for edge disturbances is only 200 mb

² The geopotential heights and wind fields in Fig. 9 (bottom-left) are, in many respects, similar to those in the observed map shown in Fig. 1. Both maps exhibit the existence of a major cyclonic system north of the mountain, an edge anticyclonic (cyclonic) cell near the northeastern (southeastern) corner of the mountain, and strong non-geostrophic northerlies between these two edge systems. Perhaps, this confirms that the present numerical model is adequate to investigate the nature of edge disturbances and associated lee-cyclogenesis.

or less (see Fig. 19, bottom). In Fig. 21 (top), one also notes the strong association of ascending motions with high temperature perturbations throughout the lower troposphere southward of gridpoint 1. This ascending warm air south of gridpoint 1 and the descending cold air north of gridpoint 0 are indicative of active baroclinic conversion processes occurring around the cold front. This further indicates the importance of baroclinic processes for the development of a lee-cyclone. However, as mentioned earlier, the baroclinic processes are insignificant in the development of edge disturbances.

Next, an attempt is made to more firmly establish the importance of baroclinic processes for the development of a lee-cyclone. Following Murakami (1981b), the problem is approached by applying an area averaged, enstrophy equation as follows:

$$\left[\frac{\partial Z}{\partial t} \right] = L + M + N + P + Q, \quad (12)$$

where $Z = \zeta'^2/2$, and $[]$ denotes an area mean averaged over 20° longitude by 20° latitude. The other quantities are defined as:

$$L = - \left[\frac{\partial u Z}{\partial x} + \frac{\partial v Z \cos \phi}{\cos \phi \partial y} + \frac{\partial \omega Z}{\partial p} \right], \quad (13)$$

$$M = - \left[u' \zeta' \frac{\partial \bar{\zeta}}{\partial x} + v' \zeta' \left(\beta + \frac{\partial \bar{\zeta}}{\partial y} \right) + \omega' \zeta' \frac{\partial \bar{\zeta}}{\partial p} \right], \quad (14)$$

$$N = - \left[\zeta' \left(\frac{\partial \omega}{\partial x} \frac{\partial v}{\partial p} - \frac{\partial \omega}{\partial y} \frac{\partial u}{\partial p} \right) \right]', \quad (15)$$

$$P = - [f \zeta' \delta' + \zeta' (\zeta \delta)'], \quad (16)$$

$$Q = \left[\zeta' \frac{\partial}{\partial p} \left(K_M \frac{\partial \zeta'}{\partial p} \right) + \zeta' R \right]. \quad (17)$$

The term L represents the flux divergence of enstrophy across the boundaries, while M indicates the rate of transfer of enstrophy between the 16-day mean absolute vorticity fields and transient vorticity perturbations. N expresses the contribution due to the twisting effect, and P represents the divergence effect. The term Q

denotes the net effect of friction and the residual R in eq. (1).

Table 1 summarizes the computed results of the terms in eq. (12) made with respect to the trajectory of the 600 mb vorticity center from hour 12 on day 7 to hour 12 on day 8. The 600 mb level was chosen because the changes in $[Z]$ were largest at that level when compared to the corresponding values at other levels. In Table 1, $[Z]$ increases from 123 units to 186 units during the 24 hour period. The divergence effect P represents the main contributor to the $[Z]$ increase. Conversely, the terms L , M , and N do not appear to be very important to the increase in $[Z]$ enstrophy. A positive sign in L indicates the import of enstrophy across the boundaries of our region. However, this import is much smaller than the enstrophy production due to divergence effect P . M is negative and substantial, implying that transient disturbances lose their enstrophy via nonlinear interactions with the 16-day mean absolute vorticity fields. The twisting effect N also makes a significant contribution to the reduction of $[Z]$ enstrophy. Since the frictional effect and the residual are both small at 600 mb, Q contributes little to the balance of $[Z]$ enstrophy.

8. Concluding remarks

Our numerical experiment using a 10-layer hemispheric dry model investigated the mechanisms responsible for cold surges and cyclogenesis in the lee of a hypothetical facsimile of the Tibetan Plateau. These phenomena were preceded by the development of a small-scale edge anticyclone trapped in a shallow layer near the northeast corner of the mountain. Key factors for the excitement of this edge anticyclone are as follows:

(a) An east-west elongated mountain barrier blocking southward penetration of cold air from the higher latitudes,

(b) Prevailing strong 16-day mean westerlies

Table 1 Computed results for the terms in eq. (12) at 600 mb, averaged over a region extending 10° north and south, and 10° to the east and west of the cyclonic center located at 45°N , 110°E (45°N , 122.5°E) at hour 12 on day 7(8). Units are 10^{-12} s^{-2} for Z and 10^{-17} s^{-3} for the terms in eq. (12).

Location	Time	$[Z]$	$[\partial Z / \partial t]$	L	M	N	P	Q
44N, 110E	12 : 00 7th	123	91	47	-22	-32	84	14
45N, 122.5E	12 : 00 8th	186	52	20	-87	-41	179	-19
	Difference	65	-39	-27	-65	-9	95	-33

at 850 mb along the northern border of the mountain (Fig. 2 of Part I),

(c) Almost stagnant 16-day mean winds at 850 mb to the east of the mountain, and

(d) Clockwise propagation of orographically induced waves in the Northern Hemisphere, such as, Kelvin waves and topographic Rossby waves, which are excited through topographically forced vertical motions.

Factors (b) and (c) are the result of the orographic effect on the 16-day mean wind fields, which in turn affect the phase propagation of transient disturbances along the mountain. Thus, the orographic effect is two-fold. In comparison, factors (a) and (d) represent direct mechanical effects of the mountain upon transient disturbances.

To summarize, the main results of Part II are as follows:

(1) When a major anticyclone approaches the western end of the mountain, northerly winds in front of the anticyclone advect cold air from the higher latitudes, which forms a shallow cold air pool along northern border of the mountain due to its barrier effect (a).

(2) This shallow cold air is transported rapidly eastward along the northern border by the prevailing low-level westerly jet (b). The tendency for clockwise propagation of topographic waves (d), also contributes to some extent to the rapid eastward movement.

(3) At the northeast corner of the mountain, the propagation speed of the cold air pool suddenly drops due to the weakening of the 16-day mean winds (c). Concurrently, a small-scale high pressure cell is induced near the northeastern edge of the mountain.

(4) This high pressure edge cell takes a form similar to a gravity-type shock front with strong subgeostrophic, cross-isobar flows and updrafts to its south. These subgeostrophic northerlies contribute to the southward advection (propagation) of the anticyclonic edge system. The subgeostrophic northerlies are also responsible for the strong advection of cold air, forming a cold front ahead (south) of the edge anticyclone. In this manner, the edge anticyclone maintains its gravity (Kelvin wave)-type structure while propagating slowly southward along the eastern periphery of the mountain.

(5) The reason why strong subgeostrophic northerlies intensify east of the mountain can be

explained as follows: Although the southward (clockwise) movement of the edge anticyclone is very slow, it is still faster than the movement of the preceding edge cyclone which has already passed the eastern periphery, and has become nearly stagnant at the southeast corner of the mountain. A rapid reduction in distance between these two edge systems results in a sudden increase in the pressure gradient, and a subsequent intensification of cross-isobar northerly flows.

(6) The edge anticyclone, as it propagates southward along the eastern periphery of the mountain, predominates only in the shallow layer below about 800 mb. The association of anticyclonic vorticity with horizontal divergence below 800 mb favors the maintenance (trapping) of the edge anticyclone below that level. In contrast, above 800 mb, the edge anticyclone tends to dissipate due to the associated convergence. The baroclinic conversion process is not capable of generating edge disturbances.

(7) The edge cyclone begins to shift eastward by the prevailing 16-day mean southwesterlies near the southeast corner of the mountain. The subsequent phase is the amalgamation of the edge cyclone with a major trough which has travelled to the north of the mountain along about 50°N. The two cyclonic systems also become coherent in the vertical.

(8) The merging phase is also characterized by the eastward extension of the cold front from the southeastern periphery of the mountain to as far downstream as eastern China. Recall that the cold front was initially enhanced only in the immediate vicinity of the mountain by subgeostrophic northerlies south of the edge anticyclone. The extended cold front contributes to the further intensification of the merged cyclone. The intensified cyclone tends to enlarge the region of strong northerly cold surges and the associated cold front further eastward. Through these mutually interacting baroclinic processes, the merged lee-cyclone intensifies into a major mid-latitude system over northern China. Namely, the baroclinic conversion process is the major contributor to lee-cyclogenesis. In this numerical experiment, the baroclinicity was enhanced through the mechanical (not thermal) effect of the mountain.

(9) To the rear of the merged lee-cyclone are widespread northerly cold surges bursting out of northern China, which propagate as far south as 15°N and possibly beyond. These cold surges

are responsible for the northward transport of sensible heat.

Let us next consider the question, why cold surges and associated lee-cyclogenesis are peculiar to the region east of the Tibetan Plateau. The barrier effect (a) is necessary to form and maintain a large cold air pool to the north of the mountain. The configuration of the Tibetan Plateau is favorable for this effect because of its east-west elongated extension. This is not the case for the Rockies whose east-west extension is less than 20 degrees of longitude. Conversely, the north-south extension of the Rocky Mountains is much longer than that of the Tibetan Plateau. A substantial part of the Tibetan massif projects upward to almost the 500 mb level, while the Rockies extend to slightly above 700 mb. These significant geographical contrasts may result in substantial differences between the winter mean circulations and transient disturbances over and around the North American and Asiatic continents.

In this experiment, a cold front and associated subgeostrophic northerlies are first generated in the area of weak 16-day mean winds east of the mountain (c). As shown by Murakami (1981a), the winter mean westerly flows tend to go around, rather than over, the high Tibetan Plateau and a stagnant area is formed in the downstream between the northern and southern jet branches. On the other hand, the westerly flows tend to go over the Rockies, which are elongated north-south with an average elevation of about 2 km. Thus, there is only a limited stagnant zone to the east of the Rockies (Lyons and Murakami, 1981; Murakami and Ho, 1981). Therefore, the Tibetan Plateau, being the world's largest massif, and having a huge cold air pool to the north, is probably the only mountain complex which can produce strong cold surges and subsequent lee-cyclogenesis.

Within the scope of the present study, a 10-layer representation with a 2.5° latitude-longitude resolution appears to be adequate for investigating the orographic influence upon transient disturbances. However, a higher degree of vertical and horizontal resolution is certainly desirable for further study of orographically-induced disturbances which are small-scaled and trapped in the shallow layer. There are a few other shortcomings. The domain of the model is hemispheric and does not include the heat contrast between the continent to the north and

the ocean to the south. Therefore, in its interaction with the equatorial circulation, the cold surges do not penetrate to the equatorial region, but remain north of about 15°N . Further numerical simulation with realistic physical processes is needed.

Murakami (1981b and c) suggested that surface friction may be important in the generation of small-scale edge disturbances at the northeast corner of the mountain, through a mechanism similar to the Kármán vortex behind obstacles (Hirota and Miyakoda, 1965). Since the present model does not properly include the Ekman surface boundary layer effect, such a possibility cannot be examined by this numerical study. Furthermore, the cold surges in the present experiment are much more exaggerated than those in the real atmosphere. A temperature drop of 15°C in front of the cold front (Fig. 13) is too large compared to observed cold surges. Northerly subgeostrophic winds (15 ms^{-1}) are also too strong. Thus, it is necessary to include the Ekman boundary layer in the model for the realistic simulation of cold surges, as well as, for the examination of Kármán vortex mechanism due to the Tibetan Plateau.

It is fair to say that the present study fails to provide enough information on the large residual in the vorticity equation (refer to Fig. 18). This residual is related to an imbalance between the 6-hour mean tendency and the instantaneous (10-minute) time change rate. It was hypothesized that the residual can be regarded as the Reynold's stress effect due to fluctuations with periods less than 6 hours. This hypothesis requires further clarification, since the instantaneous vorticity changes are associated not only with synoptic-type low frequency oscillations, but also with unwanted high frequency noises. However, it remains unclear as to exactly what portion of the instantaneous vorticity change is due to unwanted noises. An appropriate time interval to eliminate noises and obtain physically meaningful Reynold's stress, is still unknown.

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Appendix: List of Symbols

- λ, ϕ : longitude and latitude
 p : pressure
 dx, dy : $a \cos \phi d\lambda$ and $ad\phi$; a , radius of earth
 ∇ : $\left(\frac{\partial}{\partial x}, \frac{\partial}{\partial y} \right)$
 u, v : zonal and meridional winds
 ω : vertical p -velocity
 T : temperature
 Φ : geopotential height
 ζ : vorticity
 δ : divergence
 κ : R/c_p ; R , gas constant and c_p heat capacity of the air at constant pressure
 f : Coriolis parameter
 β : df/dy
 K_M : vertical momentum mixing coefficient
 $(\overline{\quad})$: 16-day time mean
 $(\quad)'$: departure from 16-day mean
 Z : enstrophy; $\zeta^2/2$

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数値実験から見たコールド・サージとリー・サイクロジェネシスに 及ぼす山岳の影響

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10層の半球乾燥大気モデルを用いた数値実験で、チベット高原の仮想模型の東側と南東側における顕著な下層の寒波吹き出し（コールド・サージ）と、それに伴うリー・サイクロジェネシスを作り出した。これらの現象に先行して、山の北・西・東の境界で一連の現象が起る。

(1) 山の西端に親の高気圧が接近すると、北から寒気が運ばれ、東西に細長い山の障壁効果によって対流圏下部に堆積する。

(2) 寒気の溜りは次に、下層の卓越する偏西風ジェットによって、山の北側斜面を東へ運ばれる。この寒気に伴なって山の北東端に 700 mb 以下に閉じ込められた小規模な高気圧が発達する。

(3) この小高気圧の時計回り（南向き）の移動は、山の東側でかなり遅くなる。しかし、その速度は、東の境界を先に通過し南東端でほとんど停止してしまった小低気圧よりも、まだ速い。この北の小高気圧と南の小低気圧の伝播速度の違いにより、気圧傾度が急激に増加して顕著な亜地衡風の北風を生じる。この亜地衡風の北風は、寒気の北からの移流を強める。ここで寒気塊は重力波のショック型（寒冷）前線に似た形状になる。

(4) 山の南東端で、南西の卓越風により小低気圧は山から離れ東へと移動をはじめる。次の段階でこの小低気圧は、山の北約 50°N に沿って移動してきた親のトラフと合併する。この併合過程と同時に、寒冷前線が山の南東端から下流の中国東部まで、東方に延びる。合併したリー・サイクロンは、拡大した寒冷前線の付近の活発な傾圧変換によって、中緯度の非常に良く発達したシステムとなる。低気圧の後面には寒波の吹き出しが広範囲に拡がり、南支那海まで南下する。