A Case Study of the Formation of an Embedded-Areal-Type Cloud Band over Kanto Plain

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Abstract

A narrow cloud band developed over Kanto Plain on 22 April 1992 in association with the passage of an extratropical cyclone. Its unique features include the rapid formation and appearance of an arc-shaped solid cloud wall, rather than a line of a number of isolated convective cells. It evolved along the southern edge of a much broader weak rain area in a manner similar to the embedded areal type as identified by Bluestein and Jain (1985).

A mesoscale analysis of conventional observation data reveals that a local cold air mass propagated southeastward from the northwestern part of Kanto Plain. It was driven by the horizontal pressure gradient force associated with the eastward-propagating cyclone. The cloud band was found to be forced to form when and where the cold surge collided with the synoptic-scale southwesterly flow which was prevailing over the southern part of Kanto Plain.

It is further revealed that the cold surge was a leading part of a low-level mesoscale air mass which was termed as the “forerunner”. It was so named since it existed in front of the synoptic-scale northwesterly flow and beneath the southwesterly flow ahead of the upper-level trough.

1. Introduction

This paper presents the result of a mesoscale analysis of a cloud band that developed over Kanto Plain on 22 April 1992. As shown in Fig. 1, Kanto Plain is located to the east of the Central Mountain Region of Honsyu. It is bounded by mountain ranges to the west and north and is surrounded by the Pacific Ocean to the south and east. It has been known that a wide variety of mesoscale meteorological events occurs over Kanto Plain because of this unique terrain feature (*e.g.* Ogura *et al.*, 1991).

The cloud band under discussion developed around 1600 JST when an extratropical cyclone passed over the Tokyo area. Hereafter, all references to time are in JST (Japan Standard Time; 0900 JST is 0000 UTC). Two aspects of the cloud band attracted the attention of the authors. One is its fast pace of formation process; it took less than 30 minutes for the cloud band to reach the mature stage. The other is its structure at the mature stage, as viewed on the radar screen; it looked more

like a solid arc-shaped cloud wall rather than a number of active convective cells organized to form a line structure.

The objective of the present analysis is to identify sub-synoptic-scale processes that led to the formation of the cloud band. All data used in the analysis came from the operational observations of Japan Meteorological Agency, except for data from a wind profiler, a Doppler radar, and a meteorological tower of the Meteorological Research Institute (MRI). The operational data include rawinsonde upper-air sounding data, AMeDAS (Automated Meteorological Data Acquisition System) surface data, meteorological traces recorded at a couple of tens of surface meteorological stations and weather surveillance radar data.

2. Synoptic conditions

Figure 2 shows the surface weather chart at 0900 JST 22 April 1992, approximately 7 hours prior to the formation of the cloud band. A developing extratropical cyclone with a central pressure of 996 hPa was located about 550 km to the southwest

Fig. 1. The map of the central part of Honshu including Kanto Plane with selected station identifications: M = Maebashi, T = Tokyo, T/T = Tsukuba/Tateno. The lines are topographic height contours drawn every 200 m. The weak shaded areas are higher than 1000 m and the dark shaded areas are higher than 2000 m.

of Tokyo at this time. Subsequently it moved east-northeastward to pass the Tokyo metropolitan area around 1600 JST. As will be described later, the formation and structure of the cloud band was strongly modulated by the evolution of the sub-synoptic-scale environment associated with the movement of the cyclone.

In addition, there was another extratropical cyclone located to the west of Hokkaido in Fig. 2. Located between these two cyclones, it was rainy almost all over Honshu.

Figure 3 is the 500 hPa weather chart at the same time as Fig. 2. The major feature is a pressure trough located over the western part of Japan Sea and to the west of the surface cyclone. With the approach of the pressure trough, the wind speed at Tateno increased, as shown in the wind hodograph at Tateno 1500 JST (Fig. 4). There was a low-level jet centered around 800 hPa with a maximum wind speed of 42 m s^{-1} . Southwesterly winds prevailed

all over the troposphere ahead the trough. The wind speed at 900 hPa was also as strong as 21 m s^{-1} .

Figure 5 shows the vertical distributions of potential temperature (θ), equivalent potential temperature (θ_e), and the saturated θ_e (θ_e^*) as observed at Tateno 0900 JST. A convective mixed layer was present in the layer below 930 hPa, without a distinct inversion layer above it. There was a conditionally unstable layer below 850 hPa and the lifted condensation level for an air parcel near the ground surface was about 900 hPa. However, the atmosphere was so dry that there was no level of free convection for an air parcel near the surface, indicating no positive CAPE present at 0900 JST. We shall return to this point later.

3. Initiation and development of the cloud band

Figure 6 shows the radar echo distributions observed by the Tokyo radar. At 1300 JST (Fig. 6a),

0900JST 22 APRIL 1992

surface

Fig. 2. The surface weather chart at 0900 JST 22 April 1992.

0900JST 22 APRIL 1992

500 hPa

Fig. 3. The 500 hPa weather chart at 0900 JST 22 April 1992. Solid lines are height (m) and dashed lines are temperature (°C).

the western and southern parts of Kanto Plain were covered with wide-spread rain areas which had been advancing toward east. At 1400 JST (Fig. 6b), the entire Kanto Plain was covered with weak to mod-

erate radar echoes. Here, attention may be focused on echoes aligned roughly along the line connecting Tokyo and Tsukuba/Tateno (see Fig. 1 for their locations). Subsequently these echoes expanded in

Fig. 4. The wind hodograph at Tateno at 1500 JST 22 April 1992. The heights are shown in units of 100 hPa, except for S—ground surface.

Fig. 5. The vertical profiles of potential temperature (θ), equivalent potential temperature (θ_e) and the saturated equivalent potential temperature (θ_e^*) at Tateno 0900 JST 22 April 1992.

size and intensified and by 1530 JST (Fig. 6c), a broad, moderately strong rain band had formed with spotty strong radar echoes. Meanwhile, radar echoes over other parts of Kanto Plain had weakened to the level of weak intensity. It was the southern edge of this broad rainband where the cloud band under discussion started to form around 1545 JST (no diagram shown). At 1600 JST (Fig. 6d), the line structure was clearly recognizable. Once formed, the cloud band started to propagate toward south-

east, keeping its arc-shape (Figs. 6e and 6f). It is noted that there was another shorter line of clouds which was connected to the northern end of the cloud band. Further, we shall later refer to the fact that no radar echoes were present in the wide area behind the cloud band (Figs. 6e and 6f).

The observation of the cloud band by the MRI Doppler radar began at 1649 JST. By that time, the cloud band had reached its mature stage (Figs. 6e and 6f). Assuming that result of the Doppler velocity observation may be reported separately, only the distribution of radar reflectivity as viewed from the MRI Doppler radar is shown in Fig. 7. The broad rain area bounded by the 20 dBZ contour line extended from southwest to northeast, with a width of 80 km. Along its southeast edge was located the cloud band under consideration. Its outstanding feature was that the echo band bounded by the 35 dBZ contour line extends like a solid arc-shaped wall, 15 km in width and 140 km in length. A number of towers were dotted inside the wall with an interval of approximately 20 km. The result of the radar observation with various elevation angles indicated that the maximum cloud top height was 7 km.

Analyzing the reflectivity data of an 11-year period from the National Severe Storms Laboratory's radar in Norman, Oklahoma, Bluestein and Jain (1985) identified four distinct kinds of severe, mesoscale convective-line development. They are broken line, back building, broken areal and embedded areal. The embedded-areal formation occurs when a convective line appears within a larger area of weaker, stratiform precipitation. It occurred least frequently among the four kinds: out of 40 cases analyzed, only 5 cases were embedded-areal formation. Another noteworthy feature of embedded-areal formation is its environment: the averaged CAPE was only $1340 \text{ m}^2\text{s}^{-2}$, whereas CAPEs for all other 3 kinds of formation exceeded $2090 \text{ m}^2\text{s}^{-2}$. The embedded-areal formation process takes place on a time scale of 30~90 minutes and is not repeated, unlike the broken-line and broken-areal formation.

The characteristics of the present line formation fit to those of embedded areal formation. In particular, we note that the upper air soundings indicated no positive CAPE present in our case, when measured about 7 hours prior to the formation of the cloud band. This suggests that the cloud band formed when the near-saturated air within the broad area of relatively weak precipitation was forced to ascend. We shall investigate possible forcing mechanism(s) in the next section.

4. Mesoscale environmental conditions

Figure 8 shows the distributions of surface wind and temperature, as viewed from AMeDAS at 1300 JST, 3 hours before the formation of the cloud

Fig. 6. CAPPI displays of the radar echo distributions at a height of about 2 km at selected times on 22 April 1992 (Tokyo radar).

band. Strong southerly winds prevailed over Kanto Plain, ahead of the approaching cyclone center. The temperature distribution was also nearly homogeneous over the entire Kanto Plain. No local front was seen in the Kanto Plain region.

One hour later (Fig. 9a), however, northwesterly or northerly winds appeared over the northwestern part of Kanto Plain. A distinct convergence line developed between the northerly winds to the north and the southerly winds to the south. The convergence line then propagated toward south-southeast (Fig. 9b), and the position of the convergence line was almost the same as that of the cloud band at 1600 JST (Fig. 9c). Also note that the northwesterly wind speed just north of the convergence line was substantially increased, while the southerly wind remained as strong as before. Thus, it may be concluded that cloud band formed because of the forced ascending motion along the convergence line.

This raises the next question why the surface northerly wind developed. Firstly, the wind profiler

data, which will be shown later in Fig. 18, does not show any northerly wind present at any level in the troposphere in the period covered in Fig. 9. The development of the surface northerly wind was therefore not the consequence of the downward transport of the upper-level momentum. Secondly, the surface northerly wind was a local flow, not on a synoptic scale, confined mainly to the northern part of Kanto Plain.

We speculate that the northerly wind under discussion was a low-level cold flow driven mainly by the mesoscale horizontal pressure gradient force over Kanto Plain. Figure 10 depicts the evolution of the observed surface pressure distribution (adjusted to the sea-surface level) at one-hour intervals. At 1300 JST (Fig. 10a), the center of the synoptic scale cyclone was located at the northern end of Izu Peninsula. In addition, there was a local low near Maebashi (see Fig. 1 for its location). However, in the next two hours, the local low disappeared and was replaced by a local high, whereas the cyclone

Fig. 7. The radar reflectivity distribution in units of dBZ measured by the Doppler radar of the Meteorological Research Institute (MRI), Tsukuba, at 1649 JST 22 April 1992. Elevation angle was 1.5° and range circles are in 25 km intervals. The contour lines are drawn at 5 dBZ. The weak shaded areas indicate that the reflectivity is higher than 35 dBZ and dark shaded areas higher than 40 dBZ.

deepened as much as 4 hPa while it moved eastward (Figs. 10b and 10c). Consequently, the strong north-south pressure gradient had been established over the central part of Kanto Plain by 1500 JST.

This evolution may best be seen in Fig. 11 where temporal variations of surface pressure at Maebashi and Tokyo are plotted. It indicates that the pressure at Tokyo was higher than that at Maebashi until 1320 JST. Thereafter, not only the difference was reversed in sign, but also the difference increased in magnitude. The maximum pressure gradient force was 5 hPa/100 km, which corresponds to the increase of 14 m s^{-1} of velocity in one hour, when applied to a non-viscous air parcel.

The next question is what could be the cause whereby a local high developed near Maebashi around 1500 JST. According to the meteorological traces from Maebashi (Fig. 12), the surface pressure continued to decrease until 1330 JST, responding to the approach of the cyclone center. However, it increased as much as 2~3 hPa during the following 2

hours.

We note that the pressure rise occurred in the period of time when Maebashi received a substantial amount of rainfall. During that period, the temperature decreased while the dew point temperature increased. In fact, the mixing ratio was 8.3 g/kg at 1200 JST when a southwesterly wind was blowing. It increased to 9.3 g/kg at 1400 JST with a northwesterly wind.

We further note that the northwestern part of Kanto Plain, including the Maebashi area, was covered with wide-spread radar echoes at 1300 JST (Fig. 6a).

Given the terrain feature that the Maebashi area is close to and surrounded by the mountain ranges to the north as well as to the west (Fig. 1), it is speculated that the local high over the Maebashi area was caused by accumulation of evaporatively cooled air near the ground surface. In their investigation of the Nagasaki heavy precipitation event, Nagata and Ogura (1991) concluded from the sensitivity study

Fig. 8. The distributions of surface wind and temperature at 1300 JST 22 April 1992 measured by AMeDAS. A wind barb is equal to 1 m s^{-1} and a pennant is 5 m s^{-1} . Temperature is converted to the sea surface level with the assumed temperature lapse rate of $0.5^\circ\text{C}/100 \text{ m}$ and isotherms are in 1°C intervals. The heavy dashed line denotes wind shear line.

Fig. 9. The distributions of surface winds measured by AMeDAS at 1400, 1500 and 1600 JST 22 April 1992. The heavy dashed lines denote wind shear lines.

of their modeling result that the evaporative cooling played an essential role in the formation of the local front and low-level jet. Later, in Figs. 13 and 14, further evidence will be presented to show the close association between temperature fall and rainfall.

The cold air thus generated surged toward the south and southeast. Its leading edge reached Tokyo at 1540 JST, as marked by sudden changes in many variables such as temperature, dew point temperature, pressure and wind direction (Fig. 13).

Fig. 10. The distributions of observed surface pressure converted to the sea surface level measured at 1300, 1400 and 1500 JST 22 April 1992.

Summarizing, it is tentatively concluded that the cold air pool generated in the northern part of Kanto Plain by evaporative cooling was accelerated to move toward south and southeast by the low-level horizontal pressure gradient associated both with the passage of the developing cyclone center in the south and the formation of the local high in the north. Thus, the cold air surge depicted above is more complex than a simple density current in that no pressure gradient is considered in a less dense fluid. This may account for the strong northwest-

Fig. 11. Comparison between the time series of surface pressure at Maebashi and Tokyo.

erly wind discussed earlier in relation to Fig. 9c. The cloud band formed, as described earlier, when the cold air surge collided with the warm southerly flow and the warm air was forced to ascend.

From this point of view, it may not be accidental that the cloud band formed inside the pre-existing broad precipitation area (embedded areal). This is because the lower and middle layers of the atmosphere were near saturation and therefore rendered less resistance to the forcing.

5. Meteorological conditions at Tsukuba/Tateno

The Tsukuba/Tateno area is special in the sense that not only Tateno is one of the operational upper-air sounding sites in Japan, but also MRI located in Tsukuba houses many facilities such as a wind profiler, two Doppler radars and a meteorological tower. We examined data from some of these facilities closely, in order to gain further information of the environment surrounding the cloud band.

First, we note from Figs. 6e and 6f that the Tsukuba/Tateno area was located close to the northern end of the cloud band. Figure 14 shows the meteorological traces from the Tateno aerologi-

Maebashi

22-23 APRIL 1992

Fig. 12. Meteorological traces from the Maebashi meteorological station on 22-23 April 1992. The pressure and the temperature are in situ measured values. The horizontal bars at the bottom of the trace of the wind direction indicate that the wind direction should be referred to the direction inside the parentheses.

cal observatory. The temporal variations of meteorological variables associated with the passage of the cold air surge are qualitatively similar to those observed in Tokyo (Fig. 13). Quantitatively, the cold

air arrived at 1610 JST, and the temporal variations of temperature and pressure were more gentle. Nonetheless, the duration times of gusty wind and the pressure variation were about 2 hours, the same

Tokyo

22-23 APRIL 1992

Fig. 13. Same as Fig. 12, but for the Tokyo meteorological station.

Tateno

22-23 APRIL 1992

Fig. 14. Same as Fig. 12, but for the Tateno meteorological station.

Fig. 15. The evolution of 1-minute average wind measured at selected heights of the MRI's meteorological tower on 22 April 1992.

as in Tokyo.

The Meteorological Tower is 213 m high above the ground and it is equipped with sensors for wind, temperature and humidity at various altitudes. Consistent with the surface observation (Fig. 14), both the wind direction and temperature changed almost instantaneously at 1610 JST (Figs. 15 and 16). The fact that the abrupt changes occurred almost without time lag at all altitudes indicates that the leading edge of the cold air surge stood almost vertically.

Figure 15 further shows that the cold air surge lasted for 2 hours from 1600 through 1800 JST and it was a divergent flow; the wind direction changed almost 180° , starting with west, turning to north and ending with east. Although a divergent flow is common in the downburst, the life span of the downburst is much shorter than 2 hours in general. For example, the average life span of downbursts on the ground, as observed in Colorado, USA, is 15 minutes (Wilson *et al.*, 1984).

Furthermore, a close examination of Fig. 15 reveals that a convergent flow of the duration time of 2 hours preceded the divergent flow. Thus, there was a couplet of convergence and divergence across the front of the cold air surge. At present, we do not know what mechanism gave rise to this couplet.

Figure 17 shows winds above the 500 m height, as measured by the MRI wind profiler in 6 minutes and 250 m intervals. Evidently, the air having the northerly wind component invaded beneath the layer of southerly wind with a wedge-like shape. The boundary surface between the northerly and southerly winds had a hump at its leading front, just as the density current does. Unfortunately, vertical temperature profile data are not available. It may however be assumed that the height of the bound-

Fig. 16. Same as Fig. 15, but for temperature.

ary surface was proportional, or at least related, to the thickness of the surging cold air. This assumption is based upon the fact that the boundary surface dipped during the period between 1712 and 1736 JST and this dip corresponded to the surface pressure dips shown in both Figs. 13 and 14. Note that these pressure dips were not accompanied by the temperature rise and consequently must have been associated with the shallowness of the cold air. No definite explanation can be offered at present as to the cause of the dip, except for that the upper boundary of the density current often exhibits such a variation owing to the growing Kelvin-Helmholtz waves (*e.g.* Droegemeier and Wilhelmson, 1987).

Returning to Fig. 14, we note that there was the second burst of wind centered around 2430 JST. The nature of this burst may be investigated from Fig. 18 which shows the 1-hour-averaged wind profiler data covering the time period from 0900 April 22 through 0700 JST April 23. The most conspicuous feature found in Fig. 18 may be the variation associated with the passage of an upper-level trough; wind direction changed from south-southwest to northwest. As is the case with developing baroclinic waves in general, the vertical axis of the trough tilted west-

Fig. 17. Time series of wind profiles measured by the MRI's wind profiler at 6-minutes intervals from 1536 through 1748 JST 22 April 1992. The heavy dashed line denotes the boundary between the low-level northerly and mid- and upper-level southerly flows. A short wind barb is equal to 5 m s^{-1} , a long wind barb is 10 m s^{-1} , and a pennant is 50 m s^{-1} .

Fig. 18. Time series of 1-hour average wind profiles measured by the MRI's wind profiler from 0900 JST 22 April through 0700 JST 23 April 1992. (For example, the 1500 JST wind means the wind averaged from 1500 to 1600 JST.) The heavy dashed line denotes the boundary between the northerly and southerly flows (as in Fig. 17), the thin dashed line a synoptic scale trough and the dotted line the rear of the "forerunner". A short wind barb is equal to 5 m s^{-1} , a long wind barb is 10 m s^{-1} , and a pennant is 50 m s^{-1} .

Fig. 19. Time-height cross-section of the vertical velocity measured by the MRI's wind profiler in 0.1 m s^{-1} intervals from 0900 JST 22 April through 0800 JST 23 April 1992. The solid and dashed lines are for the ascending and descending motions, respectively. The shaded areas indicate negative values.

ward with height.

The identity of the trough, however, was lost in the layer below 2 km. Instead, a mesoscale region of northerly flow was present ahead of the synoptic-scale northwesterly region, extending from 1600 through 2430 JST beneath the southwesterly flow with the thickness of 2~2.5 km. For convenience of the description, this region will be referred to as the "forerunner" in this paper. A part of the forerunner was described as the cold surge in Fig. 17. The second burst of the northwesterly wind referred to just above marked the boundary between the forerunner and the synoptic-scale northwesterly region. With the arrival of the latter, the nearly saturated surface air inside the forerunner was wiped out.

The forerunner can be identified also in the time-height cross-section of the vertical velocity measured by the wind profiler (Fig. 19). In drawing the diagram, vertical velocities exceeding 0.7 m s^{-1} were eliminated with the assumption that they were contaminated by the precipitation. Note in Fig. 19 that the descending motion of -0.1 m s^{-1} in magnitude was present in the layer below 2–3 km during the period of the forerunner passage. The period of the low-level descending motion ended around 2300 JST when the low-level ascending motion took place for

a brief period of time.

The descending motion associated with of the forerunner can be confirmed in Fig. 20 which compares the vertical temperature profiles at 0900 against 2100 JST. Note that the observations were made before and after the arrival of the forerunner, respectively. As expected, temperature in the lowest layer (below 500 m) decreased with the arrival of the forerunner. On the other hand, the temperature did not change in the layer above 1500 m. In between, temperature at 2100 JST was significantly higher than that at 0900 JST. For the following reasons, the warming in the layer between 500 and 1500 m was concluded to have been caused by subsidence of the air; (a) the lapse rate in that layer was close to that of dry adiabat, (b) there was a distinct inversion at the 700 m level, (c) as shown in Table 1, relative humidity was reduced from 63 % to 32 % at 900 hPa and from 70 % to 41 % at 925 hPa, whereas it increased from 40 % to 47 % at 850 hPa and from 67 % to 98 % at the ground surface.

The rapid clearing of radar echoes behind the cloud band was noted earlier in connection with Fig. 6. This may be accounted for by the subsidence in the forerunner.

To our best knowledge, no report has been published which described any phenomena similar to

Fig. 20. The vertical distribution of temperature measured at 0900 JST and 2100 JST 22 April 1992 at Tateno. The solid line labelled at Γ_d denotes the dry adiabatic lapse rate.

Table 1. Relative humidity (%) at selected levels observed at Tateno at 0900 and 2100 JST 22 April 1992.

	0900	2100
850 hPa	40	47
900 hPa	63	32
925 hPa	70	41
surface	67	98

the forerunner identified in this paper. At moment, we do not know where the air inside the forerunner came from and, in more general terms, whether or not the forerunner developed only over Kanto Plain because of its unique terrain features.

6. Summary and conclusions

Around the time when the center of an extra-tropical cyclone passed over Tokyo, a cloud band developed along the southern edge of a broad rain band over Kanto Plain, in the process similar to that classified as the embedded areal type by Bluestein and Jain (1985). The cloud band was characterized as an arc-shaped solid cloud wall of 5~10 km in width, rather than a group of active cumulonimbi

organized to form a line. The maximum cloud top height was at most 7 km. The environmental atmosphere observed 7 hours before the formation of the cloud band was stably stratified without positive CAPE.

A mesoscale analysis of surface environmental conditions was made using all available data including AMeDAS data and traces from a couple of tens of surface meteorological stations. It is found that a cold surge originated from the northern part of Kanto Plain and that it was driven by the horizontal pressure gradient mainly associated with the passage of the center of the developing cyclone. The cloud band formed when and where the southward-propagating cold surge collided with the southerly flow prevailing ahead of the cyclone center. The possible source of the cold air is unknown, though evaporative cooling of air over the mountains is suggested.

It is further revealed from an analysis of the wind profiler and meteorological tower data that the cold surge was a part of the low-level mesoscale air mass which was coined as the forerunner. The forerunner was present in front of the synoptic-scale northwesterly flow and beneath the southwesterly flow which prevailed ahead of an upper-level trough. Its thickness was 2~2.5 km and there were evidences that the air inside the forerunner was subsiding, forming an inversion layer at the 700 m. Obviously, further studies are needed to understand both the source of air that constitute the forerunner and the forerunner's underlying physics.

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関東平野における Embedded-Areal-Type 降雨帯の形成の事例解析

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1992 年 4 月 22 日、温帯低気圧の通過に伴い関東平野で細長いメソスケールの降雨帯が発生した。その主な特徴は、弧状の連続した壁雲の急速な発達により降雨帯が形成された点で、他の例に見られるような多くの孤立対流セルの並んだ降雨帯とは異なるものであった。この降雨帯は、それ自身に比べてはるかに広く弱い降雨域の南端に発達したもので、Bluestein and Jain (1985) の分類では “embedded-areal-type” に属するものであった。

気象庁定常観測データのメソ解析から関東平野の北西部分から局地的な寒冷気塊が南西に流出したことが明らかになった。それは東進する温帯低気圧に伴う水平気圧傾度により強制されたものであった。この寒気流が関東平野南部の総観規模場の南西風との間に収束を起こし、これにより降雨帯が形成された。

さらにこの寒気流は、「前駆寒冷気流」と名付けられた下層のメソスケールの寒冷気塊の先端部であることが明らかになった。このように名付けられたのは、それが総観規模場の北西風の前方、上層トラフの前の南西風の下に存在したからである。