Mesoscale Cyclogenesis in Winter Monsoon Air Streams : Quasi-geostrophic Baroclinic Instability as a Mechanism of the Cyclogenesis off the West Coast of Hokkaido Island, Japan

By Kazuhisa Tsuboki

Ocean Research Institute, University of Tokyo, Tokyo 164, Japan

and

Gorow Wakahama

Institute of Low Temperature Science, Hokkaido University, Sapporo 060, Japan

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Abstract

The mechanism of mesoscale cyclogenesis in winter monsoon air streams is studied by data analyses of observations and linear instability analysis. Mesoscale cyclones, with a diameter of 200–700 km, occasionally develop off the west coast of Hokkaido, Japan. Based on the analysis of satellite images, the mesoscale cyclones are classified into two types according to their horizontal scale: Type I (200–300 km in diameter) and Type II (500–700 km). The observational studies suggest that the baroclinicity in the lower troposphere is important for the mesoscale cyclogenesis. The quasi-geostrophic equations are used to investigate an instability problem of a baroclinic flow with both the zonal and meridional components. The basic flow of the model expresses the essential elements of the observed wind profiles. The linear instability analysis showed that two unstable modes are found: Mode I with a wavelength of 200–300 km and Mode II with 500–700 km. The former is caused by the vertical shear of the meridional wind component and the latter mainly by that of the zonal wind component. The comparisons between theoretical and observational aspects indicate that Mode I and Mode II can account for the characteristic properties of Type I and Type II, respectively. The energetics shows that both Mode I and Mode II are maintained by the increase of the eddy available potential energy and its conversion into the eddy kinetic energy. Based on the data analyses of observations and the linear instability analysis, we concluded that the mesoscale cyclones are baroclinic instability disturbances developing in a particular baroclinic flow.

1. Introduction

Satellite images occasionally show occurrences of a mesoscale vortex in winter monsoon air streams off the west coasts of Hokkaido and Sakhalin. The vortex has a characteristic horizontal diameter of 200–700 km. Since the mesoscale vortex is accompanied by a lower surface pressure and a cyclonic low-level wind circulation, it is referred to as a mesoscale cyclone. In the late 1940s, surface observations revealed an occurrence of the mesoscale cyclone along the west coast of Hokkaido and the cyclone has been studied as a cause of local severe snow storms along the coastal regions.

Okabayashi (1969) noticed a satellite image in which a mesoscale cyclone is accompanied by a

broad single cloud band, namely a convergence band cloud. The convergence band cloud is a distinctive characteristic feature of this mesoscale cyclone. Recently, Doppler radar observations revealed the radar echo and kinematic structures of the convergence band cloud accompanying a mesoscale cyclone (Fujiyoshi *et al.*, 1988a, 1988b; Tsuboki *et al.*, 1989). Kobayashi *et al.* (1989) used radar to find a detailed structures of a convergence band cloud and a mesoscale vortex of several tens kilometers in diameter.

In recent years, polar lows which are a mesoscale cyclone in polar air streams have been studied by many authors. We consider that the mesoscale cyclone of the present study is a particular type of polar low, because the cyclone is accompanied by the convergence band cloud and the cyclogenesis is

localized in the Hokkaido region. Two types of instability have been considered to be important for the development mechanism of polar lows: baroclinic instability (Harrold and Browning, 1969; Mansfield, 1974; Duncan, 1977; Reed and Duncan, 1987) and conditional instability of the second kind (CISK) (Rasmussen, 1979; Bratseth, 1985; Økland, 1987). Duncan (1977) used a quasi-geostrophic numerical model with a small static stability near the earth's surface and found unstable modes which have similar dimensions to polar lows. Reed (1979) concluded that polar lows are primarily a baroclinic phenomenon while CISK and barotropic instability cannot be ruled out as possible forcings. Ninomiya (1989) studied the polar low and comma cloud over the Japan Sea (45–35°N) and the northwestern Pacific Ocean (55–40°N) in winter and found that they develop in a zone 500–1000 km north of the major polar frontal zone, where baroclinicity in the lower-middle troposphere is strong. Ninomiya (1991) also showed a detailed description of a polar low developed over the east coast of the Asian Continent and concluded that the baroclinicity between the continent and ocean is important for the development of the polar low.

The mesoscale cyclone in the Hokkaido region is usually generated when the synoptic pressure gradient is small and the northwesterly monsoon wind is light. Harimaya (1970) studied a mesoscale cyclone and found that it was generated when an upper cold air mass passed the northern part of Hokkaido. The upper cold advection and the light wind may enhance the radiative cooling over the islands of Hokkaido and Sakhalin. Okabayashi and Satomi (1971) found that the mesoscale cyclone is generated along a discontinuous line of temperature which is formed between the radiationally cooled air over the land and the warmer over-sea air. Tsuboki *et al.* (1989) observed a significant temperature drop associated with the passage of a convergence band cloud. These previous studies suggest that baroclinicity in the lower troposphere may be important in the genesis of the mesoscale cyclone. It is, however, difficult to explain the small horizontal scale (200–700 km) and the rapid growth rate ($2\text{--}4\text{ day}^{-1}$) of the mesoscale cyclone by a simple application of the conventional linear baroclinic instability theory (Charney, 1947; Eady, 1949). Mansfield (1974), using the Eady model with baroclinicity confined to below a lid at 1.6 km, showed that if baroclinic waves are shallow, then the wavelength of the most unstable wave is ~ 600 km; this is similar to the observed wavelength of polar lows. The application of this method, however, hardly explains the rapid growth rate of the mesoscale cyclone observed off the west coast of Hokkaido.

The mechanism of the cyclogenesis and the structure of the mesoscale cyclone in the Hokkaido region

are still unknown, because the scale of the cyclone is smaller than the rawinsonde network and larger than the data processing range of a radar. In addition, the cyclogenesis occurs rapidly over the sea and the time-variation of the mesoscale cyclone is also rapid.

The purpose of this research is to clarify the structure and mechanism of the mesoscale cyclone which is generated in the winter monsoon air streams off the west coasts of Hokkaido and Sakhalin. This paper will show characteristics of the mesoscale cyclone by data analysis and we will give an explanation for the scale selection of the cyclogenesis. The previous studies (*e.g.*, Okabayashi and Satomi, 1971; Yamaguchi and Magono, 1974) have focused on the horizontal shear along the convergence band cloud. We will, on the contrary, point out the importance of the vertical shear which is related to the baroclinicity in the lower troposphere. Using the quasi-geostrophic equations in pressure coordinates, we make a linear instability analysis of a characteristic baroclinic state which is observed along the west coast of Hokkaido in the winter monsoon season. Section 2 presents classification of the mesoscale cyclones and descriptions of two typical cases based on analysis of satellite images. Section 3 describes the basic states of the flow. The model formulation is given in Section 4. Theoretical results are given in Section 5 and Section 6. The mechanism of the cyclogenesis is discussed in Section 7, and the results are summarized in Section 8.

2. Case studies

a. Data

A geographical map of the Japan Islands and East Asia is presented in Fig. 1. We used 3-hourly satellite images from the geostationary meteorological satellite of Japan during the period from 1978 to 1986 in winter season (December–March), the surface weather maps at 09 and 21 JST (Japan Standard Time), and the 850, 700, and 500 hPa upper-air maps at 21 JST in Daily Weather Maps of the Japan Meteorological Agency (J.M.A.). Rawinsonde data at Wakkanai (45.42°N, 141.68°E) and Sapporo (43.05°N, 141.33°E) supplied by the J.M.A. were also used to study characteristics of the basic flow and the stratification.

b. Classification of mesoscale cyclones

The mesoscale cyclones developed most frequently in January and February. The frequency of the cyclogenesis varied greatly from year to year. Based on the satellite images, we chose 49 distinctive mesoscale cyclones which developed off the west coasts of Hokkaido and Sakhalin during the period from 1978 to 1986. Small vortices of which horizontal diameter is smaller than 100 km were excluded in this study.

Fig. 1. Geographic map of East Asia and the Japan Islands. The light shaded areas indicate regions higher than 500 m and the dark shaded areas indicate regions higher than 1000 m.

Horizontal diameters of the cyclones were estimated from the satellite image. The histogram (Fig. 2) of the horizontal diameter shows that two peaks are significant; one is a diameter of 200–300 km and the other 500–700 km. We refer to the mesoscale cyclone with the former horizontal scale as Type I and to that with the latter scale as Type II.

Type I cyclones usually did not appear in 850, 700, and 500 hPa upper-air maps and sometimes were not analyzed in the surface maps. This suggests that Type I cyclones were shallow disturbances which were confined to below 850 hPa. In contrast, Type II appeared in every level of weather maps. This suggests that Type II cyclones were rather deep disturbances which extended to ~500 hPa. Type II cyclones are distinguished from Type I according to their horizontal scale and vertical extension. The lifetime of Type I and of Type II were 0.4–1.6 days and 1.0–3.0 days, respectively.

The locations of cyclogenesis of Type I (Fig. 3a) were localized over the Japan Sea within ~150 km from the west coasts of Hokkaido and Sakhalin. On the other hand, the mesoscale cyclones of Type II (Fig. 3b) were generated at a distance from the coasts. Three cases of Type I (denoted by open circles) moved toward the south or southwest and developed into Type II (denoted by closed circles).

The following sections describe characteristics of typical mesoscale cyclones of Type I and of Type II.

c. Mesoscale cyclone of Type I

A typical mesoscale cyclone of Type I was observed off the west coast of Hokkaido on 18 January 1983. The visible image from satellite for 12 JST 18

Fig. 2. Histogram of the horizontal diameter of the mesoscale cyclones observed off the west coasts of Hokkaido and Sakhalin by satellite.

January 1983 (Fig. 4) shows the typical cloud pattern of a mesoscale cyclone of Type I: a cloud vortex with a convergence band cloud. The mesoscale cyclone, with a diameter of ~250 km, centered ~100 km to the west of Hokkaido, and the convergence band cloud extended along the west coasts ~500 km to the north of the cyclone and ~200 km to the south. The weather was clear to the east of the cyclone and the convergence band cloud; this enhanced the radiative cooling over the islands of Hokkaido and Sakhalin. In contrast, the air mass to the west of the disturbances was supplied heat and moisture from the sea surface and cumulus clouds developed over the Japan Sea.

The surface weather map (Fig. 5a) shows that a synoptic scale low was located ~2000 km to the ENE of Hokkaido and that a polar air outbreak occurred over Hokkaido and its surroundings before the mesoscale cyclogenesis. On the west coast of Hokkaido, the pressure gradient was rather small and the prevailing northwesterly monsoon wind was light. The 850 hPa map (Fig. 5b) shows that an intense cold advection occurred in the west of the synoptic low. The extension of the cold air mass in the 850 hPa map, *i.e.*, an intense cold advection, was a characteristic condition of temperature field for the mesoscale cyclogenesis.

The locations of the cyclone center (Fig. 6) were derived from satellite images. The mesoscale cyclone was generated at ~60 km to the northwest of Wakkanai at 01 JST 18 January and moved southward along the west coast of Hokkaido. The cyclone dissipated at 18 JST 18 January. The lifetime of the cyclone was ~0.7 days. The averaged U_p and V_p were respectively -1.8 and -3.0 m s^{-1} during the period from 01 to 12 JST 18 January, where U_p is

Fig. 3. The locations of the mesoscale cyclogenesis observed by satellite during the period from 1978 to 1986: (a) Type I and (b) Type II, where the open circles in (b) indicate the points of cyclogenesis of Type I mesoscale cyclones which rapidly developed into Type II (closed circles connected with the open circles by arrows).

Fig. 4. Visible satellite image of the mesoscale cyclone of Type I at 12 JST 18 January 1983.

the X- (zonal) component of the phase velocity and V_p is the Y- (meridional) component. The phase velocity was approximately equal to the horizontal wind velocity between 1000 and 900 hPa observed by Wakkanai rawinsonde at 21 JST 17 January 1983. The mesoscale cyclones of Type I usually moved southward. The movement of disturbances of Type I was usually dominated by the flow below 900 hPa.

Figure 7 shows a time-height cross section of the equivalent potential temperature and winds derived from rawinsondes launched from Wakkanai on 17–19 January 1983. A mixed-layer developed from the surface to ~ 750 hPa and a deep stable layer was present above the level during the development of the cyclone. Winds were backing (counterclockwise) below the stable layer. These suggest that the mesoscale cyclone was a shallow disturbance below the stable layer. Before the cyclogenesis, the wind below ~ 900 hPa was northerly at 21 JST 17 January and was backing below ~ 750 hPa. During the development of the cyclone, the backing with height is a characteristic feature of winds: easterly below 900 hPa, northerly or northwesterly between 900 hPa and 800 hPa, and northwesterly or westerly above 800 hPa.

d. Mesoscale cyclone of Type II

A typical mesoscale cyclone of Type II was observed during the period from 18 to 19 January 1981. A satellite image (Fig. 8) shows a cloud pattern of a

Fig. 5. (a) Surface weather map showing pressure (solid lines, labeled in hPa), and (b) 850 hPa map at 21 JST 17 January 1983, showing heights of the pressure surfaces (solid lines, labeled in meters) and isotherms (dashed lines, labeled in °C).

Fig. 6. Movement of the center of the Type I cyclone.

mesoscale cyclone of Type II. The cyclone was in its mature stage and centered ~100 km from the west coast of Hokkaido. A vortex cloud, with a diameter of 500–700 km, developed. An intensely cloudy region surrounded the cyclone center. The horizontal scale is significantly larger than that of Type I.

The movement of the mesoscale cyclone was dominated by the flow below 900 hPa in the early stage. The cyclone moved to the SSW until 06 JST 19 January, then changed direction and moved to the SSE. The vortex cloud pattern disappeared at 18 JST 19 January, when it moved over the southern part of Hokkaido. The lifetime of the cyclone was about 1.8 days. The mesoscale cyclones of Type II usually move toward the southwest in the early stage.

Figure 9 shows the 500 hPa weather map at 21 JST 18 January 1981 when the cyclone was fully developed. The cyclone on the 500 hPa map was located ~500 km to the WNW of the surface mesoscale cyclone. The disturbance extended to ~500 hPa, i.e., the mesoscale cyclone of Type II was a relatively deep disturbance in comparison with the disturbance of Type I.

The cyclone was generated ~150 km to the northwest of Wakkanai at 03 JST 18 January 1981. Figure 10 shows the averaged wind profile derived from rawinsonde observations at Wakkanai during the period from 21 JST 17 January to 15 JST 18 January. The zonal component has a significant vertical shear between 800 and 900 hPa. The vertical shear is considered to be important for the genesis of the Type II cyclone.

The rawinsonde data at Wakkanai and at Sapporo give the meridional temperature gradient along a meridian of 141°E. The profile of the meridional temperature gradient at 21 JST 17 January 1981 (Fig. 11) shows that the gradient below ~600 hPa was steep and has a maximum peak at 850 hPa

Fig. 7. Time-height cross section of equivalent potential temperature and winds derived from serial rawinsondes launched from Wakkanai during the period from 17 to 19 January 1983. The interval of the isotherms is 1K. The thick solid lines are drawn in every 10K. Winds are plotted according to the standard meteorological convention; a half barb indicates 5 knots, a full barb 10 knots, and a flag 50 knots. The horizontal bar at the bottom of the figure indicates the period of appearance of the mesoscale cyclone in satellite images.

($-1.8^{\circ}\text{C}/100\text{ km}$), before the mesoscale cyclone of Type II was generated. This steep meridional temperature gradient was due to an intense cold advection. Such intense cold advection is also an important condition for the cyclogenesis of Type II.

The maximum peak of the temperature gradient caused the maximum vertical shear of the zonal component of the horizontal wind at the same level. The vertical shear of the zonal component and the meridional temperature gradient suggest that baroclinicity is important for the cyclogenesis of Type II.

3. Vertical profiles of the basic flow components

The two case studies in the previous section showed that a cold advection is an important condition for mesoscale cyclogenesis and that vertical shear is strong during the genesis of the mesoscale cyclones in the Hokkaido region. The results of the

Fig. 8. Visible satellite image of the mesoscale cyclone of Type II at 12 JST 19 January 1981.

Fig. 9. The 500 hPa map at 21 JST 18 January 1981, showing heights of the pressure surfaces (solid lines, labeled in meters) and isotherms (dashed lines, labeled in $^{\circ}\text{C}$).

Fig. 10. Averaged vertical profiles of the horizontal wind velocity derived from rawinsondes launched from Wakkanai during the period from 21 JST 17 January 1981 to 15 JST 18 January 1981. The thick line represents the averaged zonal component and the thin line the averaged meridional component.

Fig. 12. Averaged vertical profiles of the horizontal wind velocity derived from rawinsondes launched from Wakkanai. The wind data used for the average were observed within 6 or 12 hours before and after occurrences of the 33 mesoscale cyclones in winter seasons from 1981 to 1986. The thick line represents the averaged zonal component and the thin line the averaged meridional component.

Fig. 11. Vertical profile of the meridional temperature gradient ($^{\circ}\text{C}/100\text{ km}$) between Wakkanai and Sapporo at 21 JST 17 January 1981, before the mesoscale cyclone of Type II was generated. Negative value means the temperature of Wakkanai is colder than that of Sapporo. Winds at Wakkanai (abbreviated as WAK) and Sapporo (SPR) are also shown.

observational studies suggest that baroclinicity in the lower troposphere is important for the genesis of the mesoscale cyclones. In this section, we will present vertical profiles of a basic flow.

Satellite images were used to find when mesoscale cyclones occurred off the west coasts of Hokkaido

and Sakhalin in winter season from 1981 to 1986. We averaged wind data observed by rawinsondes at Wakkanai within 6 or 12 hours before and after the occurrences of 33 mesoscale cyclones, in order to obtain a characteristic profile of the basic flow for the mesoscale cyclogenesis. Type II cyclone sometimes occurs with Type I cyclone. Three cases of Type II cyclone developed from Type I cyclone (see Fig. 3b). Therefore, it is difficult to separate the basic state of Type II from that of Type I by the analysis of the wind data observed at every 6 hours. Since the west coasts of Hokkaido and Sakhalin are approximately parallel to a meridian, we choose horizontal coordinates x and y as eastward (zonal direction) and northward (meridional direction), respectively.

The averaged horizontal wind during developments of mesoscale cyclones (Fig. 12) is characterized by the backing with height below ~ 700 hPa. Since disturbances are shallow in their early stage, we focus on the vertical shear below 500 hPa. The zonal component below 500 hPa is characterized by strong shear between 900 and 700 hPa. The meridional component has strong shear below ~ 850 hPa with weak shear above that level.

We use the following analytic functions of pressure to express the essential elements of the basic flow (Fig. 13):

$$\bar{u}(p) = U_0 - U_1 \tanh\left(\frac{P_I - p}{P_H}\right), \quad (1)$$

Fig. 13. Vertical profiles of the idealized basic flow for the linear instability analysis. The thick line represents a zonal component and the thin line a meridional component.

$$\bar{v}(p) = V_0 \left\{ 1 + \epsilon \exp \left[-B \left(\frac{P_1 - p}{P_0} \right)^2 \right] \right\}, \quad (2)$$

where

$$P_1 = P_0 \left[1 + (2B)^{-\frac{1}{2}} \right],$$

$$\epsilon = -\exp \left[B \left(\frac{P_1 - P_0}{P_0} \right)^2 \right].$$

For the zonal component $\bar{u}$, P_I is the pressure of the inflection point in the vertical profile of $\bar{u}$, U_0 is the value of $\bar{u}$ at P_I , U_1 is one-half of the increase in $\bar{u}$ across the shear layer, and P_H is one-half of the thickness of the shear layer. For the meridional component $\bar{v}$, V_0 gives the magnitude of the meridional basic flow, P_0 is the pressure at the surface (assumed to be 1000 hPa), P_1 is the pressure of the inflection point in the vertical profile of $\bar{v}$, ϵ adjusts $\bar{v}$ to zero at P_0 , and B gives the strength of the low-level shear; the shear becomes strong and the thickness of the shear layer decreases when B increases. In the present study, we assume $U_0 = 0$ m s⁻¹, $U_1 = 6$ m s⁻¹, $P_I = 850$ hPa, $P_H = 120$ hPa, $V_0 = -8$ m s⁻¹, and $B = 80$. We will examine the dependence of the most unstable wave on these parameters in Section 5, because it is difficult to determine fixed values of these parameters from the data analysis of observation.

4. Formulation

When the mesoscale cyclone is generated, the prevailing wind is light: usually less than 10 m s⁻¹. The characteristic Rossby number, Ro is defined as

$$Ro = \frac{U}{f_0 L}. \quad (3)$$

If we use the Coriolis parameter, $f_0 = 1.0 \times 10^{-4} \text{ s}^{-1}$ (latitude = 45°N), characteristic horizontal scale $L = 300$ km, and $U = 8$ m s⁻¹, then $Ro = 0.27$. Based on a rawinsonde analysis, the mesoscale cyclone is characterized by the Richardson number $Ri \sim 10$. While the Rossby number of the mesoscale cyclone is somewhat larger than that of a synoptic scale midlatitude cyclone, $Ro^2 Ri$ is still on the order of unity. Therefore, we expect that the qualitative discussion would not be changed by the nongeostrophic effect and that the quasi-geostrophic system is useful to describe the characteristic properties of the mesoscale cyclogenesis.

The non-dimensional quasi-geostrophic potential vorticity equation for the perturbation can be written:

$$\begin{aligned} \frac{D}{Dt} \left[\nabla^2 \phi + \frac{1}{Ro^2} \frac{\partial}{\partial p} \left(\frac{1}{S_N} \frac{\partial \phi}{\partial p} \right) \right] \\ - \frac{1}{Ro^2} \left[\frac{\partial \phi}{\partial x} \frac{d}{dp} \left(\frac{1}{S_N} \frac{d\bar{u}}{dp} \right) + \frac{\partial \phi}{\partial y} \frac{d}{dp} \left(\frac{1}{S_N} \frac{d\bar{v}}{dp} \right) \right] = 0, \end{aligned} \quad (4)$$

where

$$\frac{D}{Dt} = \frac{\partial}{\partial t} + \bar{u} \frac{\partial}{\partial x} + \bar{v} \frac{\partial}{\partial y},$$

and the non-dimensional thermodynamic energy equation for the perturbation can be written:

$$\frac{D}{Dt} \frac{\partial \phi}{\partial p} - \left(\frac{\partial \phi}{\partial x} \frac{d\bar{u}}{dp} + \frac{\partial \phi}{\partial y} \frac{d\bar{v}}{dp} \right) + Ro S_N \omega = 0, \quad (5)$$

where ϕ is the geopotential and S_N is a non-dimensional static stability parameter in isobaric coordinates defined as

$$S_N = -\frac{P_0^2}{U^2} \alpha \frac{\partial \ln \theta}{\partial p}, \quad (6)$$

in the conventional notation (see Appendix). Since we consider mesoscale phenomena, the β -effect may be negligible.

The lower and upper boundary conditions in isobaric coordinates are

$$\omega = 0 \quad \text{at} \quad p = 1, 0. \quad (7)$$

The normal mode solution is

$$\phi = \text{Re} \{ \Phi(p) \exp[i(kx + ly - \nu t)] \}, \quad (8)$$

where ν and Φ are respectively eigenvalue and eigenfunction of this problem and could be complex. The “Re” denotes the real part.

We solve the eigenvalue problem numerically using the finite difference approximation. In this investigation, the number of grid points is 200 and we employ the matrix method to solve the eigenvalue problem.

Fig. 14. Growth rate diagram of unstable wave; abscissa represents the magnitude of non-dimensional wavenumber vector and ordinate represents its direction. The parameters of the basic flow are as follows; $U_0=0\text{ m s}^{-1}$, $U_1=6\text{ m s}^{-1}$, $P_l=850\text{ hPa}$, $P_H=120\text{ hPa}$, $V_0=-8\text{ m s}^{-1}$, and $B=80$.

The baroclinic instability wave is maintained by two processes; the increase of eddy available potential energy due to the transport of the sensible heat, and its conversion into eddy kinematic energy due to ascending warm air and descending cold air (Lorenz, 1955). The time changes of the eddy kinetic energy KE and of the eddy available potential energy PE are given by

$$\frac{\partial}{\partial t}[\overline{KE}] = -[\overline{\omega\alpha}], \quad (9)$$

$$\frac{\partial}{\partial t}[\overline{PE}] = [\overline{\omega\alpha}] + \frac{1}{RoS_N} \left(\left[\frac{d\bar{v}}{dp} \overline{u\alpha} \right] - \left[\frac{d\bar{u}}{dp} \overline{v\alpha} \right] \right), \quad (10)$$

where the overbar denotes horizontal integration over one wavelength with the cyclic boundary condition and the square brackets denote the vertical integration with the boundary condition given by (7). These energy conversion processes are characterized by the release of potential energy $-[\overline{\omega\alpha}]$, the meridional heat transport $[\overline{v\alpha}]$, and the zonal heat transport $[\overline{u\alpha}]$.

5. Growth rate and phase velocity

We now examine the instability properties of the basic flow. The static stability parameter is assumed to be constant: $S_N=60$. Since the problem contains both the zonal and meridional components, the growth rate and phase velocity depend on both the zonal and meridional components of the wavenumber vector $\mathbf{k}$.

Figure 14 shows a growth rate diagram of an unstable wave; the abscissa represents the magnitude

Fig. 15. As in Fig. 14 except for $\bar{u}=0$.

of the non-dimensional wavenumber vector $|\mathbf{k}|$ and the ordinate represents its direction θ_N . The growth rate diagram has two significant maximum peaks; one is $\sim 2.4\text{ day}^{-1}$ which is located at $|\mathbf{k}|=8$ and $\theta_N=\pi/2$, and the other is $\sim 2\text{ day}^{-1}$ at $|\mathbf{k}|=2.7$ and $\theta_N=11/12\pi$. The corresponding dimensional wavelength of the former unstable wave is $\sim 240\text{ km}$ and that of the latter unstable wave is $\sim 700\text{ km}$. The unstable wave with the former wavelength and that with the latter wavelength are referred to as Mode I and as Mode II, respectively.

In order to examine the role of the meridional component of the basic flow $\bar{v}$, we simply set the zonal component $\bar{u}$ to zero. Then, the growth rate diagram has only one maximum peak (2.4 day^{-1}) at $|\mathbf{k}|=7.5$ and $\theta_N=\pi/2$ which corresponds to Mode I (Fig. 15). This indicates that Mode I is mainly caused by the shear of the meridional component, *i.e.*, by the zonal temperature gradient. Figure 15 also shows that this type of shear has an unstable mode for almost all wavelengths and directions of wavenumber vector.

Figure 16 shows dependence of the growth rate, the phase speed, and the wavenumber of the maximum peak of Fig. 15, *i.e.*, the peak of Mode I on the parameters B and V_0 of the meridional component $\bar{v}$. When the parameter B increases from 40 to 100, the wavenumber increases linearly from 5.6 to 9.0. The corresponding dimensional wavelength changes from 340 km to 210 km. The growth rate gradually increases with increase of B . If we take this parameter range of B , Mode I is comparable with Type I with respect to the horizontal scale and growth rate. In contrast, the growth rate changes drastically with change of the parameter V_0 . Therefore, V_0 should be considerable in order that the growth rate of Mode I may account for the rapid development of Type I. The wavenumber is independent of V_0 . These results indicate that the values of B and V_0 given in Section 3 could be appropriate for the comparison

Fig. 16. Variations of the growth rate of the most unstable wave (closed circles), the corresponding phase velocity (solid triangle), and the corresponding wavenumber (solid squares), (a) with the parameter B where $V_0 = -8 \text{ m s}^{-1}$, and (b) with the parameter V_0 where $B = 80$. The zonal component $\bar{u} = 0$.

between the observed mesoscale cyclones and the most unstable waves. Figure 16 indicates that the wavelength of Mode I is dependent on B and is independent of V_0 ; this means that the wavelength is proportional to the thickness of the low-level shear layer. The growth rate increases when the low-level shear becomes strong. The phase speed is almost independent of B , while it increases linearly with increase of V_0 .

We also set the meridional basic flow component $\bar{v}$ to zero and examine the role of the zonal basic flow $\bar{u}$. In this case, the growth rate diagram (Fig. 17) has one maximum peak at $|\mathbf{k}| = 2.7$ and $\theta_N = 0$ or π . The growth rate is $\sim 2.0 \text{ day}^{-1}$; this is the same as that of Mode II in Fig. 14. Since the growth rate, the horizontal scale, and vertical structure (not shown) of this maximum peak are the same as those of Mode II, this maximum peak is essentially the same as that of Mode II. Therefore, Mode II is mainly caused by the shear of the zonal component, *i.e.*, the meridional temperature gradient and is modified by the meridional component of the basic flow. This type of shear has a short-wave cut-off as the Eady model has.

We examine the dependence of the maximum peak of Fig. 17, *i.e.*, the peak of Mode II on the parameters U_0 , U_1 , and P_H of the zonal basic flow (Fig. 18). When the parameter P_H increases from 90 hPa to 180 hPa, the wavenumber decreases from 3.0 to 1.9. The corresponding dimensional wavelength increases from 630 km to 990 km. If P_H takes a parameter range of 90 hPa $\sim$ 120 hPa, then the wavelength is 630 km $\sim$ 700 km; which is comparable with the horizontal wavelength of Type II. The growth rate is $\sim 2 \text{ day}^{-1}$ when P_H is in the parameter range. The growth rate increases linearly with the parameter U_1 . Therefore, U_1 should be considerable in order that the growth rate of Mode II may account for the rapid

Fig. 17. As in Fig. 14 except for $\bar{v} = 0$.

development of Type II. The wavenumber is independent of U_1 . These results indicate that the values of P_H and U_1 given in Section 3 could be appropriate for the comparison between the observed mesoscale cyclones and the most unstable waves. Figure 18a also shows that the wavelength increases with the thickness of the shear layer and the growth rate increases with strength of the shear. The phase speed increases with P_H ; this means that the steering level increases with the thickness of the shear layer. Figure 18b shows that the phase speed increases linearly with U_1 . Since the vertical shear of the zonal basic flow is independent of the parameter U_0 , the growth rate and the wavenumber are independent of U_0 , while the phase speed is proportional to U_0 (figure is not shown).

6. Dynamic structure and energy conversions

Figure 19 shows the vertical structure of Mode I. The direction of wavenumber vector θ_N is $\pi/2$, there-

Fig. 18. Variations of the growth rate of the most unstable wave (closed circles), the corresponding phase velocity (solid triangle), and the corresponding wavenumber (solid squares), (a) with the parameter P_H where $U_1=6\text{ m s}^{-1}$, and (b) with the parameter U_1 where $P_H=120\text{ hPa}$. The meridional component $\bar{v}=0$.

fore, the vertical plane is directed northward. The perturbation geopotential (Fig. 19a) has a maximum amplitude at the ground level. The disturbance is very shallow and is confined to a layer below 850 hPa. The trough line tilts northward with height. The vertical motion (Fig. 19b) is downward on the north side of the trough where the temperature perturbation (Fig. 19c) is negative, and is upward on the south side where the temperature perturbation is positive. The agreement of the phase between the vertical motion and the perturbation temperature ($\omega < 0$ and $\alpha > 0$, and $\omega > 0$ and $\alpha < 0$) indicates that the unstable wave transports heat upward and the eddy available potential energy is converted into eddy kinetic energy. Recalling the relation between the geopotential and the geostrophic wind, the wind is easterly on the north side of the trough where the temperature perturbation is negative and the wind is westerly on the south side where the temperature perturbation is positive.

Figure 20 shows the vertical profiles of energy conversion rate. The profile of the zonal heat transport $\bar{u}\bar{\alpha}$ has maximum at the ground level and decreases to 850 hPa. Since $d\bar{v}/dp$ is positive or zero throughout the atmosphere in this case, the eastward heat transport increases the eddy available potential energy. The meridional heat transport is zero, because the direction of wavenumber vector is $\pi/2$, i.e., $k=0$. The profile of $-\bar{v}\bar{\alpha}$ has a positive maximum peak at ~950 hPa and largely positive below ~800 hPa. Therefore, the energy source of Mode I is the zonal temperature gradient, i.e., the vertical shear of the meridional component. This is consistent with the conclusion that Mode I is mainly caused by the vertical shear of the meridional component of the basic flow (see Section 5).

Figure 21 shows the vertical structure of Mode II.

The direction of wavenumber vector θ_N is $11/12\pi$, therefore, the vertical plane is directed approximately toward the WNW. In comparison with Mode I, Mode II is a relatively deep disturbance. The perturbation geopotential (Fig. 21a) has a maximum amplitude at the ground level and extends from the ground level to 400 hPa. The trough line tilts toward the WNW with height. The vertical motion (Fig. 21b) is downward on the WNW side of the trough where the temperature perturbation (Fig. 21c) is negative, and is upward on the ESE side of the trough where the temperature perturbation is positive. Therefore, the unstable wave transports heat upward and the eddy available potential energy is converted into eddy kinetic energy. The horizontal perturbation flow on the WNW side of the trough has both easterly and northerly components, which is a cold advection, and the horizontal perturbation flow on the ESE side has both westerly and southerly components, which is a warm advection.

Both the zonal heat transport $\bar{u}\bar{\alpha}$ and the meridional heat transport $\bar{v}\bar{\alpha}$ have positive maximum peaks at ~850 hPa and vanish at ~600 hPa (Fig. 22). Since $d\bar{v}/dp$ is positive and $d\bar{u}/dp$ is negative throughout the atmosphere, both the northward and eastward heat transports increase the eddy available potential energy. $\bar{v}\bar{\alpha}$ is significantly larger than $\bar{u}\bar{\alpha}$; this indicates that the main energy source of Mode II is the meridional temperature gradient. This is also consistent with our conclusion that Mode II is mainly caused by vertical shear of the zonal component of the basic flow and modified by that of the meridional component. The profile of $-\bar{v}\bar{\alpha}$ has a positive maximum peak at ~850 hPa and is positive below ~700 hPa.

The static stability in the lower levels of the tro-

Fig. 19. Vertical structure of the disturbance of Mode I; (a) perturbation geopotential (solid lines indicate positive and dashed lines negative). (b) perturbation vertical velocity (solid lines indicate upward motion and dashed lines downward motion). (c) perturbation specific volume (solid lines indicate positive, *i.e.*, warm, and dashed lines negative, *i.e.*, cold). Values in the figures are normalized by the maximum value.

posphere is small owing to the mixing caused by a sensible heat supply from the sea surface. In the real atmosphere, the non-dimensional static stability parameter S_N which is derived from rawinsonde data at Wakkanai increases gradually from the surface to ~ 500 hPa, and increases drastically

Fig. 20. Vertical profiles of energy conversion rate of Mode I; unit is arbitrary.

above ~ 500 hPa during winter season. We also used the linear model to investigate the effect of the stratification of the atmosphere on the unstable waves of Mode I and Mode II. Since Mode I is a shallow disturbance, the effect of high static stability in the upper levels is small. Consequently, the vertical structure of the unstable wave and the profiles of energy conversions are almost the same even if we take into account the effect of stratification. On the other hand, the effect of stratification slightly modifies the structure, growth rate and phase velocity of Mode II, because Mode II is relatively deep disturbance in comparison with Mode I. The characteristic properties are, however, essentially the same as those of the case excluding the stratification.

7. Discussions

a. Upper cold advection and baroclinicity in the lower troposphere

The synoptic analysis showed that an intense cold advection is a necessary condition for the genesis of the mesoscale cyclone of Type I and Type II. When the mesoscale cyclone develops, the weather over the islands of Hokkaido and Sakhalin is usually clear. Both the cold advection and radiative cooling decrease the air temperature over the lands of Hokkaido and Sakhalin, while the air mass over the Japan Sea is warmed owing to the heat supply from the warm sea surface. As a result, the zonal temperature gradient becomes strong and a shallow baroclinic layer develops.

The sea surface temperature of the Okhotsk Sea is colder than that of the Japan Sea. In addition, the Okhotsk Sea is usually covered with sea ice. The sea ice in the Okhotsk Sea reduces the heat flux from the sea surface significantly and may enhance the zonal temperature gradient.

The baroclinic layer due to the differential heating between Hokkaido/Sakhalin and the Japan Sea is shallow and is most likely an available energy source

Fig. 21. As in Fig. 19 except for Mode II.

for the cyclogenesis of Type I. The mesoscale cyclone of Type I and the convergence band cloud developed along the boundary of the colder air over the land and the warmer over-sea air. Based on an observation using two research vessels, Sasaki and Satomura (1991) shows a significant temperature difference in the lower troposphere between the east and west sides of a convergence band cloud.

The cold advection may also intensify the meridional temperature gradient and the vertical shear of the zonal component. The baroclinic layer due to the cold advection is relatively deeper than that caused by the differential heating between Hokkaido/Sakhalin and the Japan Sea. This may be important energy source for the cyclogenesis of Type II. The rawinsonde observations at Wakkanai

Fig. 22. As in Fig. 20 except for Mode II.

and Sapporo showed a significant meridional temperature gradient when Type II cyclone was generated.

The vertical profile of the meridional component of the basic flow given by (2) represents the shallow baroclinic layer due to the differential heating between Hokkaido/Sakhalin and the Japan Sea. The derivative of the meridional component, $d\bar{v}/dp$, has a maximum at the lower boundary and decreases to zero at ~ 800 hPa. On the other hand, the vertical profile of the zonal component given by (1) represents the relatively deep baroclinic layer due to a cold advection in the Hokkaido region. The derivative of the zonal component, $d\bar{u}/dp$, has a maximum at 850 hPa and decreases to zero at ~ 500 hPa. The profile of $d\bar{u}/dp$ is similar to that of the observed meridional temperature gradient below 600 hPa which is shown in Fig. 11.

The cold advection is often associated with an upper low or upper trough. Suginaka (1964) pointed out that a cold low at 500 hPa is important to the cyclogenesis. Kono and Magono (1967) also found a mesoscale cyclone formed below the upper cold low. The dynamic relation between the mesoscale cyclogenesis and the upper cold low is beyond the present paper and remains for a future work.

b. Comparisons between observational and theoretical aspects of the mesoscale cyclones

The comparison of the horizontal scale between the theory and the observation indicates that Mode I and Mode II respectively correspond to Type I and Type II. Some other characteristic properties of Mode I agree well with those of Type I. The growth rate of Mode I ($\sim 2.4 \text{ day}^{-1}$) can account for the rapid development of Type I. The direction of wavenumber vector of Mode I is approximately $\pi/2$ and its phase velocity is negative, therefore, Mode I moves southward; this can account for the southward movement of Type I. The observation shows that the phase velocity of Type I was approximately

equal to the horizontal wind velocity between 1000 and 900 hPa. The theory predicts that the steering level of Mode I is located at ~ 950 hPa.

The locations of cyclogenesis of Type I are within ~ 150 km from the west coasts where the zonal temperature gradient is intense between the warmer Japan Sea and colder lands of Hokkaido and Sakhalin. The rawinsonde observation at Wakkanai during the occurrence of the mesoscale cyclogenesis showed that the meridional wind component had a strong shear below ~ 850 hPa, which was caused by the zonal temperature gradient. We, therefore, infer that the zonal temperature gradient, *i.e.*, the vertical shear of the meridional component is important for the cyclogenesis of Type I. The theoretical results indicate that Mode I is mainly caused by the vertical shear of the meridional component, *i.e.*, the zonal temperature gradient. The energetics of Mode I also indicates that the disturbance is developed by the increase of eddy available potential energy due to the eastward transport of the sensible heat and its conversion into eddy kinetic energy.

The analysis of Wakkanai rawinsonde data shows that the mesoscale cyclone of Type I was a shallow disturbance. The dynamic structure of Mode I which is revealed by the linear theory also shows that the disturbance is shallow and confined to below ~ 850 hPa. The linear theory also predicts that there are the easterly component and cold advection in the north of the trough of Mode I and the westerly component and warm advection in the south. Tsuboki *et al.* (1989) observed a significant shift of the surface wind and temperature drop with a passage of the convergence band cloud which was accompanied by a mesoscale cyclone. Harimaya (1970) showed the surface wind was northeasterly to the north of the mesoscale cyclone and northwesterly to the south. These observational results are consistent with the structure of the disturbance which is predicted by the linear theory.

Characteristic properties of Mode II also can account for the behavior of Type II. The growth rate of Mode II ($\sim 2.0 \text{ day}^{-1}$) can account for the rapid development of Type II. The locations of cyclogenesis of Type II suggest that the meridional temperature gradient is more important than the zonal temperature gradient for the cyclogenesis of Type II. The case study in Section 2d shows that the meridional temperature gradient was intense when the Type II cyclone was generated. The linear theory shows that Mode II is mainly caused by the shear of the zonal component, *i.e.*, the meridional temperature gradient and is modified by the shear of the meridional component of the basic flow. The energetics of Mode II indicates that the disturbance is developed by the increase of eddy available potential energy due to the eastward and northward transport of the sensible heat and its conversion into eddy ki-

netic energy. The northward heat transport is significantly greater than the eastward heat transport. Therefore, the main energy source of Mode II is the meridional temperature gradient.

Type II is relatively deep disturbance and appears even in 500 hPa weather map. The case study in Section 2d shows that the upper-air low on the 500 hPa map was located to the WNW of the surface mesoscale cyclone of Type II. The dynamic structure of Mode II which is revealed by the linear theory also shows that the disturbance is deep and extends to ~ 400 hPa. The trough line of Mode II tilts toward the WNW with height.

c. Cumulus heating

Cumulus heating is most likely important forcing for the rapid development of the mesoscale cyclones, because the mesoscale cyclones and the convergence band cloud consist of cumulus convections over the warm Japan Sea and large amounts of latent heat must be released in the disturbances. Sardie and Warner (1985) used a numerical model to find that moist baroclinicity and CISK are both important in the genesis of the Atlantic and Pacific polar lows. By an analytic theory of quasi-geostrophic baroclinic instability with a parametrization of heating, Mak (1982) showed that as heating intensity increases the growth rate of the most unstable wave significantly increases and its wavelength decreases. This suggests that the latent heat release may be important for the rapid development and small horizontal scale of the mesoscale cyclone.

8. Summary and conclusions

When the northwesterly winter monsoon wind is light, mesoscale cyclones occasionally develop off the west coasts of Hokkaido and Sakhalin. They have a diameter of 200–700 km and are accompanied by a convergence band cloud. We studied the mechanism of the mesoscale cyclogenesis by data analyses of observations and a linear instability analysis.

Based on the analysis of satellite images, we classified the mesoscale cyclones into two types according to their horizontal scale: Type I with a diameter of 200–300 km and Type II with a diameter of 500–700 km. The analyses of rawinsonde data and weather maps showed that the upper cold advection is a necessary condition for the mesoscale cyclogenesis. The cold advection increases the meridional temperature gradients and also enhances the radiative cooling over the lands of Hokkaido and Sakhalin. The zonal temperature gradient is formed between warmer air over the Japan Sea and colder air over the lands. The locations of cyclogenesis of Type I are over the Japan Sea within ~ 150 km from the west coasts of Hokkaido and Sakhalin, where the zonal temperature gradient is intense. The vertical shear of the meridional component is strong below 850 hPa when

Type I is generated. These suggest that the vertical shear of the meridional component, *i.e.*, the zonal temperature gradient is important for the cyclogenesis of Type I. On the other hand, the locations of the cyclogenesis of Type II are at a distance from the west coasts. When Type II develops, the vertical shear of the zonal component is strong. The meridional temperature gradient is primarily important for the cyclogenesis of Type II. The results of the observational studies indicate that the baroclinicity in the lower troposphere is important for the genesis of the mesoscale cyclones.

Using the quasi-geostrophic equations, we investigated the instability problem of a baroclinic flow with both the zonal and meridional components. The profiles of the basic flow are defined as characteristic profiles of winds which were derived from rawinsondes launched from Wakkanai when the mesoscale cyclones were generated. The linear analysis showed that two unstable modes are found: Mode I with a wavelength of 200–300 km and Mode II with 500–700 km. The former is caused by the vertical shear of the meridional component and the latter mainly by that of the zonal component. Mode I is a shallow disturbance which is confined to below 850 hPa and Mode II is a relatively deep disturbance which extends over 500 hPa. The comparisons between theoretical and observational aspects indicate that Mode I and Mode II can account for the characteristic properties of Type I and Type II, respectively: their scale selection, rapid development, and phase velocities. The vertical structures and energetics show that both Mode I and Mode II are developed by the increase of eddy available potential energy and its conversion into eddy kinetic energy. Therefore, the theoretical results also indicate that baroclinic instability is an important mechanism of the mesoscale cyclogenesis.

Based on the data analyses of observations and the linear instability analysis, we concluded that the mesoscale cyclones off the west coast of Hokkaido are baroclinic instability disturbances developing in a particular baroclinic flow.

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Appendix

List of Symbols

i	Square root of minus one
$\mathbf{k}$	Horizontal wavenumber vector $\mathbf{k}=(k, l)$
k	Zonal wavenumber
l	Meridional wavenumber
p	Pressure
t	Time
u	x component of velocity (eastward)
v	y component of velocity (northward)
L	Horizontal length scale
U	Horizontal velocity scale
P_0	Pressure at the ground level; $P_0 \equiv 1000$ hPa
S_N	Non-dimensional static stability parameter
α	Specific volume
θ	Potential temperature
θ_N	Direction of wavenumber vector in polar coordinates
ν	Angular frequency of wave
ϕ	Geopotential
ω	Vertical wind velocity in isobaric coordinates
Φ	Amplitude of the geopotential: eigenfunction
∇^2	Two-dimensional Laplacian operator

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冬期季節風の寒気流中に発生するメソスケール低気圧：北海道西岸小低気圧の発生メカニズムに対する準地衡風傾圧不安定論の適用

坪木和久

(東京大学海洋研究所)

若濱五郎

(北海道大学低温科学研究所)

冬期季節風の寒気流中に発生するメソスケールの低気圧（小低気圧）のメカニズムをデータ解析と線形論を用いて調べた。北海道や樺太の西岸沖では水平直径が 200–700 km の小低気圧がしばしば発生する。衛星画像の解析に基づいて、これらの小低気圧を水平スケールの異なる二つのタイプに分類した。一つは水平直径が 200–300 km のスケールを持つ低気圧、もう一つは 500–700 km のスケールを持つ低気圧である。前者を Type I、後者を Type II と呼ぶ。小低気圧発生時には、北海道の東にある総観規模低気圧の西側で強い寒気の移流がみられる。この寒気移流は南北方向の水平温度傾度を強めるだけでなく、北海道の陸上の放射冷却を強化する。この放射冷却により、南北に延びる海岸線に沿って、相対的に暖かい日本海と陸との間に強い東西温度傾度が対流圏下層で形成される。小低気圧はこの水平温度傾度の最も強いところに発生しており、発生時に観測される風は特徴的な鉛直シアを持つ。データ解析により、北海道西岸沖の小低気圧の発生には、下層の傾圧性が重要であることが示唆された。

準地衡風方程式系を用いて、小低気圧発生時にみられる特徴的な鉛直シアを持った流れの場の線形安定

性を調べた。その結果、不安定波の成長率には二つのピークが現れることがわかった。一つは波長が 200–300 km の不安定波で Mode I と言う。もう一つは 500–700 km の不安定波で Mode II と言う。前者は南北風のシアーに起因し、後者は主として東西風のシアーによって発生する。観測と理論を比較すると Mode I と Mode II はそれぞれ Type I と Type II の小低気圧の特徴をよく説明する。エネルギー変換を調べると、Mode I と Mode II は有効位置エネルギーの増大とその運動エネルギーへの変換によって発達することがわかった。本研究により、この小低気圧の発生において傾圧不安定が重要な働きをしていることが示された。