An Amplification Mechanism of Medium-Scale Tropopausal Waves

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ABSTRACT

The mechanism for the amplification of medium-scale tropopausal waves (with horizontal wavelengths of 2000–3000 km and wave periods of 1–2 days) that are basically neutral is investigated using a three-dimensional wave-activity flux. Distinctive upward wave-activity fluxes are found in the midtroposphere (400 hPa) upstream of the region where the medium-scale waves are dominant, that is, where the horizontal components of the wave-activity flux are large. These upward fluxes converge in the upper troposphere (300 hPa). This convergence is a major source of medium-scale waves, producing large values of the horizontal components of the wave-activity flux in the active regions.

Examination of the spatial structure of the heat flux shows that the upward fluxes are due to occasional baroclinic interactions between upper-level medium-scale waves and lower-tropospheric disturbances with the same temporal scales. The seasonal and geophysical variations of the waves depend mostly on the amount of baroclinic coupling with the lower disturbances.

1. Introduction

Upper-tropospheric disturbances have been discussed mainly in terms of their role in initiating surface cyclogenesis. Petterssen and Smeebye (1971) provided a concise classification of surface cyclogenesis, although its role had been recognized prior to their work. According to their definition, “type A” surface cyclogenesis is similar to the exponential growth of an unstable normal mode in a basic state with a jet as classically treated by Charney (1947) and Eady (1949), whereas “type B” surface cyclogenesis occurs when an upper-level precursor disturbance encounters a low-level baroclinic zone. The large contribution of upper-level troughs to cyclogenesis has been investigated in many studies, especially for extratropical cyclones that were remarkably developed, such as the Queen Elizabeth II storm in 1978 and the Presidents’ Day storm in 1979 (e.g., Uccellini et al. 1985; Uccellini 1986; Whitaker et al. 1988).

These upper-level troughs are often called “short waves,” because they have smaller horizontal scales than those of synoptic-scale cyclones. However, even if they are referred to as “waves,” it should be noted that these troughs are regarded as an anomaly of geopotential height or potential vorticity rather than as a wavy structure having several wavelengths.

Sanders (1988) made a statistical analysis of “mobile troughs.” In his study, mobile troughs were noticed as a southward displacement of the 5520-m geopotential height contour at 500 hPa in the Northern Hemisphere. Statistical analysis of data from nine cold seasons showed the following results. The troughs were long lived: they were maintained for 5 days on average and sometimes tracked for about 2 circuits around latitude circles. It was pointed out that the generation of troughs is encouraged by northwesterly background flow and occurs more (less) frequently than the extinction of troughs in mountainous (oceanic) areas. More recently, Lefevre and Nielsen-Gammon (1995) made a statistical analysis of mobile troughs over 20 yr in the Northern Hemisphere using a more objective method.

Several theoretical approaches have been used to explain the mobile troughs. One approach regards mobile troughs as neutral or slowly decaying modes. The theoretical work of Rivest et al. (1992) considered the mobile troughs as upper-level neutral modes of the Eady model including the stratosphere with a finite buoyancy frequency. They showed that the basic characteristics of the upper-level neutral modes are retained even if the static stability in the stratosphere is limited. In addition, Rivest and Farrell (1992) showed that for the basic state
with small constant planetary vorticity gradient, the upper mode can exist as a "quasi mode" that has a similar structure to a small decay rate. Another theoretical approach regards mobile troughs as unstable modes. Whitaker and Barcilon (1992a) noticed the variation in the structure of unstable modes under various basic states. Under a basic state assuming continental conditions with weak baroclinicity in the lower troposphere, strong static stability, and strong surface friction, the most unstable modes have a structure with large amplitudes near the tropopause. In contrast, under a basic state with strong baroclinicity in the lower troposphere, weak static stability and weak surface friction, as found over the ocean, the most unstable modes have a structure with large amplitudes near the surface. They proposed that upper precursor troughs and type B surface cyclogenesis are manifestations of the structures of these unstable modes modified under the different basic states. Furthermore, Whitaker and Barcilon (1992b) examined unstable modes and initial-value evolution for a zonally varying basic state. It was shown that a disturbance initiated within the continental basic state evolves into a wave packet with the amplitude maximized near the tropopause and the wave packet then undergoes a structural modification upon entering the low-level baroclinic zone, with amplitude maximized at the surface.

Recently, the temporal and spatial resolutions of operational objective analysis data and forecast models have been improved. These high-resolution data have enabled us to distinguish atmospheric phenomena with temporal scales of a few tens of hours from synoptic-scale phenomena. Consequently, it has become known that phenomena with smaller temporal scales have different characteristics to those with synoptic scales.

Advances in observation techniques have also promoted analyses of phenomena on smaller temporal scales. Combining these two advances, Sato et al. (1993) found "medium-scale waves" near the tropopause over Japan. They found an oscillation of the meridional wind with wave periods of around 20–30 h in the spring of 1990, using observational data from the mid- and upper atmosphere (MU) radar (a VHF clear air Doppler radar) with a temporal resolution of 3 min. Outputs from predictions made using the Japan Spectral Model (JSM) every 3 h revealed that the oscillation was due to disturbances with wavelengths of 2000–3000 km. The characteristics of the waves are summarized as follows: 1) the amplitudes are maximized around the tropopause at about 5° poleward of the midlatitude westerly jet axis and are evanescent both in the meridional and vertical directions; 2) meridional and vertical phase tilts are small; 3) a typical zonal wavelength, wave period, and zonal phase velocity are estimated statistically at 2100 km (about 23° in lon), 26 h and 22 m s⁻¹ (about 21° per day) respectively. Sato et al. called these waves medium-scale waves, following studies that treated disturbances with similar horizontal scales that develop in the lower troposphere. See Sato et al. (2000) for details about this terminology.

Subsequent studies have revealed other characteristics of the medium-scale waves, such as seasonal variations and global characteristics (Hirota et al. 1995; Yamamori et al. 1997; Sato et al. 2000). It has been shown that the medium-scale waves are active in three regions: the North Atlantic in winter, the North Pacific in spring, and the south Indian Ocean in autumn.

In regard to the maintenance mechanism of these tropopausal medium-scale waves, Sato et al. (1998) proposed that the medium-scale waves can be interpreted as a quasi-neutral mode that is trapped into a localized positive peak of the horizontal gradient of the quasigeostrophic potential vorticity (QGPV) near the midlatitude tropopause. Using the upper-air network data they showed that the gradient of the QGPV is maximized above and slightly poleward of the midlatitude tropopause because of a jump in static stability in the presence of the westerly shear of the mean wind. They analytically and numerically examined a one-dimensional problem in the vertical in the quasigeostrophic (QG) framework to elucidate the characteristics of the trapped mode, especially in terms of the dependence on parameters describing the background field. It was shown that the trapped mode has similar characteristics to those of the observed medium-scale waves.

On the basis of the above interpretation, Yamamori and Sato (1998) examined the structure of the large-scale field when the medium-scale waves are active, in terms of QGPV distribution, using data from both observation and operational analysis. It was found that the active periods and vertical locations of the medium-scale waves correspond to the periods and locations with extremely large positive values of the poleward gradient of the stretching vorticity, which are observed in the ridge phase of background synoptic-scale waves. These results support the theory of Sato et al. (1998).

It should be emphasized that an important question remains regarding the sources that sustain medium-scale waves in the real atmosphere, because the medium-scale waves lose their energy through dissipation processes such as viscous damping, even though the medium-scale waves are almost neutral.

The medium-scale waves around the midlatitude tropopause may be identical to the mobile troughs moving eastward in the upper troposphere, which can be precursor disturbances to cyclogenesis. However, the amplitudes of the medium-scale waves are small in the midtropospheric level (around 500 hPa) where mobile troughs examined by Sanders (1988) are located (Yamamori et al. 1997).

In parallel, the life cycles of the mobile troughs themselves have become a focus of attention, and have been analyzed using the improved analysis data and/or more sophisticated methods. These case studies have shown that cyclogenetic events differ from one another. For example, Nielsen-Gammon (1995) classified several
conceptual models for the formation of mobile troughs and showed that the main contributing factor was wave energy propagation from developing cutoff cyclones at the tropopause for a case using 3-h analysis data. Nielsen-Gammon and Lefevre (1996) examined a case that occurred over North America in winter, by evaluating each term of the advection of QGPV by geostrophic wind. They showed that downstream propagation of Rossby wave energy is significant in the initial stages of trough formation, and then mutual interaction between upper- and lower-level systems becomes a principal factor. Lackmann et al. (1997) analyzed in detail a life cycle of upper-tropospheric cyclogenetic precursors that contributed to the development of extratropical cyclones during the Experiment on Rapidly Intensifying Cyclones over the Atlantic (ERICA) intensive observation period (IOP; Hadlock and Kreitzberg, 1988). They depicted each stage of the progress toward surface cyclogensis over the whole troposphere. In the analyzed case, the first feature appeared in the upper troposphere as a lowered tropopause. The lowered tropopause then developed through the interaction with a midtropospheric front and finally resulted in surface cyclogensis.

In view of these advances, we examined the mechanisms for the amplification of the medium-scale waves and the causes of the observed seasonal and geophysical variations, with the aid of the concept of three-dimensional wave-activity flux.

This paper is organized as follows. First, the method and data used in this study are described in section 2. The results are then presented in section 3, and discussed in section 4. A summary and concluding remarks are given in section 5.

2. Methods of analysis and data description

A conservation relation with the form

\[ \frac{\partial A}{\partial t} + \nabla \cdot F = S + N_L \]  \hspace{1cm} (1)

holds in general for waves in a fluid (although whether the appropriate A and F can be defined for a system is another question), where A is the wave-activity density, F is the wave-activity flux, S represents nonconservative effects (zero for adiabatic and frictionless flow), and \( N_L \) is the nonlinear contribution (Andrews et al. 1987). In the case of steady linear, and conservative waves, the wave-activity flux is nondivergent \( (\nabla \cdot F = 0) \). It should be noted that the eddy energy is not conserved in a basic flow with horizontal and/or vertical shear even without nonconservative effects, and cannot be described in such a flux form.

The wave-activity flux can be expressed under the Wentzel–Kramers–Brillouin (WKB) limit of almost-plane waves:

\[ F = C \varepsilon A, \]  \hspace{1cm} (2)

where \( C \) is the group velocity. This relation is important in terms of both the interpretation and unique definition of A and F (Edmon et al. 1980). Using this relation, Eq. (1) is interpreted as that A integrated over a packet is conserved without the two terms on the right-hand side of Eq. (1).

The expressions for A and F depend on the definition of the basic state and on the governing equations. For example, the classical Eliassen–Palm flux is one of the wave-activity fluxes defined for eddies on the zonally uniform basic state. In this study, we used the three-dimensional wave-activity flux derived by Plumb (1986), which is defined for transient QG eddies on the time-mean flow.

The wave-activity flux satisfies an approximate conservation relation for QG transient eddies on the slowly varying time-mean flow. The expressions are as follows:

\[ A = \frac{1}{2} \frac{p \cos \phi}{p_s} \frac{q^2}{q} \left[ \nabla \cdot \mathbf{F} \right] \]  \hspace{1cm} (3)

\[ q = f + \frac{1}{a \cos \phi} \frac{\partial u}{\partial \lambda} - \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (\cos \phi u) \]

\[ + \frac{f R}{p \tilde{H}} \frac{\partial}{\partial t} \left( \frac{p}{N^2} \tilde{\varepsilon} \right) \]  \hspace{1cm} (4)

\[ M_R = \frac{p \cos \phi}{p_s} \times \left[ \frac{1}{\nabla \cdot \mathbf{F}} \left( \tilde{q}, \tilde{q}, 0 \right) \right] \]

\[ \times \begin{pmatrix} u'v' & -u'^2 & \frac{1}{N^2} \frac{R}{H} u'T^2 \\ v'^2 & -u'v' & \frac{1}{N^2} \frac{R}{H} v'T^2 \\ 0 & 0 & 0 \end{pmatrix} \]  \hspace{1cm} (5)

\[ S = \frac{p \cos \phi}{p_s} \frac{s_q'}{q'} \left[ \nabla \cdot \mathbf{F} \right] \]  \hspace{1cm} (6)

\[ \epsilon = \frac{1}{2} \left( u'^2 + v'^2 + \frac{1}{N^2} \left( \frac{R}{H} \right)^2 T'^2 \right) \]  \hspace{1cm} (7)

\[ \mathbf{F} = (F_x, F_y, F_z) = \mathbf{u} A + M_R, \]  \hspace{1cm} (8)

where \( q \) is the QGPV, \( u = (u, v) \) is the geostrophic wind velocity, T is the temperature, \( \phi \) is the latitude, \( \lambda \) is the longitude, \( p \) is the pressure, \( p_s = 1000 \) hPa, \( s \) represents the nonconservative sources and sinks of QGPV, \( H \) is a scale height, \( z = -H \ln(p/p_s) \) is the independent vertical variable, \( R \) is the gas constant, and \( N \) is the buoyancy frequency. An overbar represents a time-mean value and a prime indicates a deviation from the time mean. For further details of the wave-activity flux, see Plumb (1986). It is shown using the dispersion relation of Rossby waves that the A and F defined above satisfy Eq. (2).

The basic dataset consists of the 6-h European Centre for Medium-Range Weather Forecasts (ECMWF) op-
erational data of geopotential height, horizontal winds, and temperature at 10 pressure levels from 1000 to 100 hPa and 2.5° x 2.5° latitude–longitude grids for the 4 yr from 1990 to 1993. In order to extract the characteristics associated with the medium-scale waves, we calculated wave-activity fluxes (Eqs. (3)–(8)) using the high-pass-filtered values with periods shorter than 42 h, instead of simple deviations from the time-mean component. In this study, the time mean is taken over a period of 1 week. In the following section, statistical characteristics are examined for each season.

It should be noted that medium-scale waves are basically neutral and do not have a significant heat flux \( u' T' \sim 0 \). Thus it is considered that the vertical components of the wave-activity flux are related to other kinds of phenomena with similar timescales. We also assume that the rate of change of \( A \) is negligible when we consider seasonal-averaged characteristics. Thus it is convenient for the analysis of medium-scale waves to rewrite Eq. (1) as

\[
\nabla_h \cdot \mathbf{F} = -\frac{\partial F_z}{\partial z} + S + N_A \tag{9}
\]

and to treat the vertical convergence of the vertical components of the wave-activity flux as one of the source terms. As described later, among the three terms, we will pay greatest attention to the vertical convergence of the wave-activity flux.

3. Results

a. The distribution of the medium-scale waves

Previous studies used the mean square of the geopotential height or \( u' \) component as an index of the intensity of the medium-scale waves. Such quantities are good in the sense that they are not greatly affected by data noise, because their calculation does not require differentiation. In this study, we used the eddy potential enstrophy \( q^2/2 \), which is more appropriate for describing the intensity of medium-scale waves. This is because \( q^2 \) is an essential quantity in the QG system, as well as the horizontal gradient of the background QGVP. Wave activity density \( A \) is also an essential quantity, in the sense that Eq. (1) holds. However \( A \) is not suitable for diagnosing wave intensities because \( A \) is a measure of conservation when integrated over a packet, and does not always correspond to the wave intensity. Thus the distribution of the intensity of the medium-scale waves is discussed using the eddy potential enstrophy.

Figure 1 shows a polar-stereo projection map of the eddy potential enstrophy (shading) and the background QGVP (contours) at 300 hPa for each season and each hemisphere. In the Northern Hemisphere, large values of the eddy potential enstrophy are distributed within the longitudinal range extending from 150° to 60°E, especially in the Atlantic and Pacific regions in all seasons. Notable characteristic features of the eddy potential enstrophy in each region and season are as follows. Large values of the eddy potential enstrophy are seen over the east coast of North America and the North Atlantic, especially in winter (December–February (DJF)). In the Pacific region, the eddy potential enstrophy is large in the eastern Pacific (175°E–160°W) in autumn (SON) and winter, and in a more western part of the Pacific (150°E–160°W) in spring (MAM). Although the seasonal and longitudinal dependence is smaller in the Southern Hemisphere, large values are distributed in the longitude region of 30°–180°E in autumn (MAM). It should be noted also that contours of QGVP are distorted from latitude circles by the presence of stationary planetary waves, especially in the Northern Hemisphere. Large values of the eddy potential enstrophy are seen in the region of large horizontal gradient of QGVP, consistent with the theoretical work of Sato et al. (1998). These characteristics are similar to the distribution of the mean square of the geopotential height components of medium-scale waves given by Sato et al. (2000).

b. Characteristics of the wave-activity flux

The distributions of the wave-activity flux are shown in Fig. 2. Arrows indicate the horizontal component vectors of the wave-activity flux at 300 hPa in the upper troposphere. The horizontal component is predominantly eastward at this level in all regions and seasons. Of the two terms on the right-hand side of Eq. (8), the advective part \( \pi A \) is dominant at midlatitudes (not shown). The horizontal components are large around the east coast of North America and the north-central Pacific in winter, and over east Asia and the north-western Pacific in spring. In the Southern Hemisphere, the longitudinal dependence is smaller, similar to the potential enstrophy distribution. The regions where the horizontal components of the wave-activity flux are large coincide with the regions of large eddy potential enstrophy, although the dominant latitude is situated slightly equatorward of that of the potential enstrophy by about 5°. This occurs because the horizontal gradient of the background QGVP takes its maximum at the latitude where the potential enstrophy is dominant [see Eq. (3)].

We now turn to the question of what causes the large horizontal components of the wave-activity flux in the active regions mentioned above. The upward and downward components at 400 hPa (the next data level below 300 hPa) are shown in Fig. 2 by dotted and hatched areas, respectively. Large values of the upward flux component are distributed over the North Atlantic and North Pacific, and are largest in winter. These areas are located upstream of the regions where the horizontal components of the wave activity flux are large.

The zonal and vertical components of the wave-activity flux in a longitude–height section averaged over a latitudinal range of 30°–45°N, where the vertical com-
FIG. 1. Horizontal distributions of the eddy potential enstrophy (shading) and background QGPV (contours) at 300 hPa for each season and hemisphere. The left column shows the seasons (a) DJF, (c) MAM, (e) JJA, and (g) SON in the Northern Hemisphere, and the right column shows the seasons (b) DJF, (d) MAM, (f) JJA, and (h) SON in the Southern Hemisphere. The contour interval is $3.0 \times 10^{-9}$ s$^{-1}$. The outer latitude circle in each case shows 30°.
Fig. 2. Horizontal components of the wave-activity flux at 300 hPa (arrows) and vertical components at 400 hPa (tones). An arrow length of 1° corresponds approximately to a horizontal component of 5 m² s⁻².
component is significant, are shown in Fig. 3. In the longitudinal regions of the large upward component in Fig. 2, the upward fluxes are greatest at the 400-hPa level. The upward fluxes are small at 250 or 300 hPa, indicating that there is significant convergence of the vertical component. This implies that the convergence of the vertical component plays an important role in medium-scale waves having large activity.

Note that the large vertical component observed at 80°E and 500 hPa in Fig. 3c may be due to near-surface processes, because the elevation in the region is high.

Next, we investigated the relationship between vertical convergence and the magnitude of the horizontal components of the wave-activity flux. Figure 4 shows the difference between the vertical components of the wave-activity flux at 400 and 250 hPa, while Fig. 5 shows the magnitude of the horizontal components of the wave-activity flux at 300 hPa. Comparison of these figures reveals a clear tendency for large horizontal components to be distributed downstream of the area where strong convergence of the vertical component of the wave-activity flux is seen. In the Northern Hemisphere in winter, there is strong convergence of the vertical component at around 45°W, and large horizontal components at around 30°W. Such characteristics are also seen in the Pacific region in spring. Strong convergence is distributed in the region of 120°–180°E, whereas large horizontal components are located around the date line. Similarly, in the Southern Hemisphere convergence is strong in the region of 30°W–90°E, which is upstream of the region where the horizontal component is large over the Indian Ocean (30°–120°E).

Quantitatively, the convergence of the vertical components [the first term on the right-hand side of Eq. (9)] is insufficient to cover the divergence of the horizontal components [the left-hand side of Eq. (9)]. However, the values of these two terms are well correlated. We therefore judge that the convergence of the vertical components of the wave-activity flux is a major factor contributing to the horizontal divergence of the horizontal components.

c. Origins of the upward component of the wave-activity flux

The upward component of the wave-activity flux is due mainly to the poleward heat flux $v^T$ [see Eq. (5)]. In this section, temporal and spatial structures of $v^T$ are examined to elucidate the origin of the large upward flux. We examined $v'$ in order to easily see the structure of the waves, following previous studies (e.g., Sato et al. 1993). We found that the upward fluxes are due to the baroclinic interaction between the medium-scale waves and lower disturbances, as shown below.

Figure 6 shows a series of longitude–height sections of $v'$ and $v^T$ taken every 12 h from 0600 UTC 7 March
Fig. 4. As for Fig. 1, but for the difference of the vertical component of the wave-activity flux between 400 and 250 hPa. The contour interval is $2.0 \times 10^{-3}$ m$^2$ s$^{-2}$.
Fig. 5. As for Fig. 1, but for the magnitude of the horizontal components of the wave-activity flux at 300 hPa. The contour interval is 10 m² s⁻².
to 1800 UTC 9 March 1993. At 0600 UTC 7 March (Figs. 6a and 6b), there is a wave train around the longitudinal region of 125°–155°E. This wave train propagates eastward, and 12 h later is situated around the longitudinal region of 135°–165°E (Figs. 6c and 6d). The vertical tilt of the phase lines is not significant at these times. The corresponding \( v ' T' \) structure is wavelike in the zonal direction, with a wavelength half of that for \( v ' \). The net heat flux is small. This feature is consistent with the characteristics of medium-scale waves.

Twelve hours later, a strong positive heat flux appears at around 150°E (Fig. 6f) in the lower and midtroposphere. At the same time, the maxima of the northward wind are located at 150°E at 300 hPa and 155°E in the lower troposphere (Fig. 6e). The vertical phase lines of \( v ' \) are correspondingly tilted westward with height, reflecting the positive heat flux.

At 1800 UTC 8 March (in Figs. 6g and 6h), the coupling with the lower disturbances continues. The amplitude of \( v ' \) at 155°E increases compared with that at 145°E at 1800 UTC 7 March (Fig. 6c).

After a further 12 h, the upper-level northward wind maximum catches up with and passes over the lower disturbance (Fig. 6i). The baroclinic interaction almost ceases at 0600 UTC 9 March (Fig. 6j), and the upper disturbances recover their original almost-neutral structure (Fig. 6k and 6l). It is important that the amplitude at this stage is larger than that in Fig. 6c.
Fig. 7. Composite chart sequence of longitude–height sections of (left) $v'$ and (right) $v'T'$ at 40°N. The reference point is shown by $\times$. 
The case shown in Fig. 6 is not a special one. Figure 7 shows a composite sequence made using the following procedure. We took 28 cases when the time series of the 3-day low-pass-filtered \( \nu' T' \) averaged in the 400–600-hPa layer has a maximum value that is larger than 4 K m \( \text{s}^{-1} \), and when the time series of \( \nu' \) takes its positive (southerly) maximum at a longitude of 150°E in MAM seasons. A composite was made for time lags \( t \) of \(-18 \) to \(+30 \) h around the reference time.

Features similar to those seen in Fig. 6 are observed in Fig. 7. In the \( \nu' \) charts, it is clear that the amplitude after the coupling (at \( t = +24 \) h) is larger than the amplitude before (at \( t = -18 \) h). During the highest stage of the coupling (at \( t = -6 \) to \(+12 \) h), a positive heat flux extends through the whole troposphere. This result supports the robustness of the features observed in Fig. 6. From the results, it is likely that the medium-scale waves are amplified by occasional baroclinic coupling. It is worth noting that the feature in which the upper disturbance passes over the lower disturbance is different from the modal instability.

4. Discussion

a. Coupling with the lower-tropospheric disturbances

In the present analysis, we treated only the interaction of phenomena with timescales similar to those of the medium-scale waves. Interaction related to phenomena with other timescales must also be sources. The regions and seasons preferred by the medium-scale waves, listed in Sato et al. (2000), seem to be largely subject to growth by baroclinic coupling. The amount of baroclinic coupling will depend largely on the distribution of the lower-level disturbances. It has been proposed in previous studies that the oceanic basic state destabilized by a heat flux from the sea surface contributes to the development of the lower-level disturbances. For example, Mak (1998) examined unstable modes under the basic state with a sensible surface heat flux and found meso-\( \alpha \)-scale unstable modes that have a shallow and vertically tilted structure. It was shown that for certain heating profiles there exist unstable modes with medium scales in addition to meso-\( \alpha \)-scale unstable modes.

As mentioned in the introduction, past work on the interaction between upper- and lower-level disturbances has stressed the role of upper-level disturbances on the development of lower-level ones. However, there are also cases in which the lower-level disturbances do not amplify much, while the amplitude of the upper disturbances increases, as depicted in Fig. 6. The question of what factors determine whether the lower-level disturbances develop or not is an interesting problem.

b. Excitation processes

In the present study, only interactions among disturbances with timescales of 12–42 h are considered. The proposed growth mechanism presupposes that the amplitudes of the medium-scale waves are sufficiently large to interact with lower disturbances. Thus, it may be necessary to consider the mechanisms by which the medium-scale waves are excited to sufficient amplitudes. These mechanisms are likely to involve interaction related to phenomena with other timescales. For example, nonlinear effects that result from growing synoptic-scale cyclones may also contribute, as in the case outlined by Nielsen-Gammon (1995). Yamamori and Sato (1998) also pointed out that there are favorable phases of the background synoptic-scale flow pattern.

A possible candidate for the nonconservative source of the medium-scale waves over the western Pacific is diabatic heating associated with the convection around the Tibetan Plateau, which occurs diurnally (Kodama 1999, personal communication). It is likely that this heating is related to the generation of the medium-scale waves.

5. Summary and concluding remarks

In the present work we have made use of the three-dimensional wave-activity flux derived by Plumb (1986) to analyze the sources of medium-scale waves near the midlatitude tropopause. It was shown that there are prominent upward wave-activity fluxes in the midtroposphere upstream of the active region of the medium-scale waves. These upward fluxes converge in the upper troposphere (around 300 hPa), and this convergence serves as a major source for the medium-scale waves to attain large values of horizontal components of the wave-activity flux. It was also found that the upward fluxes are associated with baroclinic interaction with lower-tropospheric disturbances that have the same temporal scales. The present study provides a scenario in which the medium-scale waves are supplied with large wave-activity flux by occasional interaction with lower disturbances. The amplitudes of the medium-scale waves then become large downstream of the region where the interaction occurs. Figure 8 shows a schematic diagram describing these features. This schematic...
diagram resembles the concept of type B cyclogenesis. However, our results differ from type B cyclogenesis in that amplification of the upper disturbances (medium-scale waves) is the principal object in our diagram and is not always accompanied by the development of lower disturbances.

There are still various issues to be considered in order to depict the whole life cycle and actions of the medium-scale waves. One issue is that of excitation mechanisms. Another is the contribution of medium-scale waves, which have large amplitudes near the tropopause, to stratosphere–troposphere exchange. Further analyses from these viewpoints are necessary.

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