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Horizontal variations of gravity wave activities in the lower stratosphere over Japan: A case study in the Baiu season 1991

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Vertical wavenumber spectra are analyzed by using temperature, zonal wind and meridional wind data obtained with MU radar and radiosondes at the MU observatory in Shigaraki and with operational rawinsondes at five stations over Japan. The temperature and meridional wind spectral power densities of the dominant gravity wave increase with a decrease of latitude, whereas the zonal wind spectral power seems to have a maximum near the tropopausal jet stream axis. An analysis technique to estimate a characteristic value of intrinsic frequency from the wind and temperature vertical-wavenumber spectra is proposed. The different meridional distributions of spectral power densities for temperature, zonal wind and meridional wind possibly can be explained by a latitudinal distribution of the characteristic intrinsic frequency.

1. Introduction

Recent progress of observational studies mainly using the MST radar techniques have suggested that (inertio-) gravity wave activities may have considerable horizontal (mainly meridional) variabilities (see, e.g., Sato, 1994; Tsuda *et al.*, 1994). In order to extend the observational coverage, applications of the operational network data of rocketsondes in the upper stratosphere (Hirota, 1984; Hirota and Niki, 1985; Eckermann *et al.*, 1994) and rawin/radiosondes in the lower stratosphere (Kitamura and Hirota, 1989; Allen and Vincent, 1995; Yamanaka *et al.*, 1996) have been proposed. We have also analyzed a data set of rawinsondes launched from a research vessel which sailed for 14°S–25°N over the western Pacific (Ogino *et al.*, 1995). These studies suggest two major tropospheric sources of such horizontal variabilities: mid-latitude synoptic-scale baroclinic disturbances (including jet stream and front systems) and tropical convection systems. In this study we focus ourselves into the former category, in particular into a case of typical activation of the subtropical baroclinic zone during Baiu (early summer rain) season. We analyze the same data set as we used in Yamanaka *et al.* (1996) in more detail, and try to study variabilities of dominant waves with respect to latitudes and background atmospheric conditions.

2. Data

We carried out continuous observations of the troposphere and the lower stratosphere over Japan for three weeks (17 June–8 July 1991) during Baiu season in 1991 (Yamanaka and Fukao, 1994), by using a VHF Doppler radar (the MU radar) at Shigaraki (35°N) from which we obtained three components of wind data (Shibagaki *et al.*, 1996). Radiosondes were also launched at 6 or 12 hour intervals during the observation campaign period (except for the first two days) at the radar site, and we obtained profiles of temperature, humidity and pressure. We also obtained the JMA (Japan Meteorological Agency) operational rawinsonde data set (09 and 21 LT) of temperature and horizontal wind. In this paper, we use the data of zonal wind (u), meridional wind (v) and temperature (T).

The horizontal wind observations of the JMA rawinsondes are based on tracking of the balloon location every 1 min, following the WMO (World Meteorological Organization) guidelines. However, after 40 min from the launch time, the calculation of horizontal wind is based on a horizontal displacement of the balloon every 4 min, considering the accuracies of balloon azimuth and elevation angles. Since the balloon ascending speed is normally ~6 m/s, the routine observation procedure mentioned above corresponds to taking wind data by an approximately 1.5-km running average at altitudes upper than ~ 15 km. By this running average the amplitude of a wave-perturbed quantity, say, zonal wind perturbation $u(z)$, associated with a sinusoidal wave component of vertical wavelength λ_z becomes

$$\left|\frac{1}{1.5\ \mathrm{km}}\int_{z-1.5\mathrm{km}/2}^{z+1.5\mathrm{km}/2} u(\zeta)\, d\zeta / u(z)\right| \approx \frac{1}{\pi}\frac{\lambda_z}{1.5\mathrm{km}}\left|\sin\left(\pi\frac{1.5\mathrm{km}}{\lambda_z}\right)\right|$$

times the true value. This consideration suggests that the amplitude and spectrum of the dominant inertio-gravity waves (vertical wavelength $\lambda_z \approx$ 2 km) analyzed from the JMA rawinsonde data become ~1/3 and ~1/9 of their true values, respectively. Considering also that the accuracy of rawinsonde wind is ~0.5 m/s, we expect that the dominant gravity

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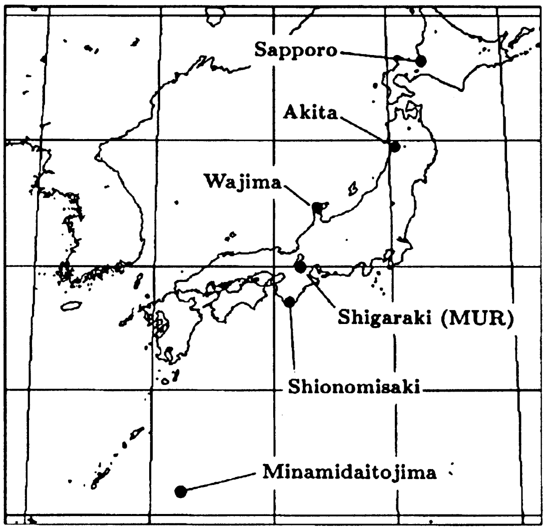


Fig. 1. Location of the MU radar (34.85°N, 136.01°E) and the five JMA rawinsonde stations used in this case study: Sapporo (43.01°N, 141.33°E), Akita (39.72°N, 140.01°E), Wajima (37.38°N, 136.90°E), Shionomisaki (33.45°N, 135.77°E) and Minamidaitojima (25.83°N, 131.23°E).

waves can be detected even in the rawinsonde data, if the true amplitude is larger than $\sim 0.5 \times 3 \approx 1.5$ m/s.

The data of the MU radar and radiosondes at the radar site and those of rawinsondes at five JMA stations (23–43°N) shown in Fig. 1 were combined in order to analyze meridional variations of gravity wave activities in the lower stratosphere. Radiosonde and JMA rawinsonde data sets have been linearly interpolated for the same altitudes (at 147.62 m interval) as observed by the MU radar. The MU radar data were averaged for 60 min and the profiles of 09 and 21 LT were chosen from all of them in order to compare with the 12 hourly JMA rawinsonde data.

In our previous study (Yamanaka *et al.*, 1996) vertical wavenumber spectra and hodographs have been analyzed from the MU radar wind data in an altitude range of 15–20 km, and it has been suggested that the dominant component of gravity waves had a vertical wavelength of ~2 km and an intrinsic period of ~9 h as mean values for the three weeks. These dominant wind fluctuations were accompanied with temperature fluctuations which were frequently regarded as multiple structures of the tropopause. The later concept was originated by Bjerknes and Palmén (1937), and was connected with wave fluctuations first by Danielsen (1959) and Kitaoka (1963). In this paper we mainly discuss horizontal variabilities of such dominant components of inertio-gravity waves.

3. Vertical Wavenumber Spectra of Wind and Temperature

We have analyzed vertical wavenumber spectra of wind and temperature for all the profiles obtained by the MU radar, radiosondes at the radar site and the JMA rawinsondes. Here we demonstrate overall features which appear in temperature and horizontal wind spectra for the 15–19 km altitude range, averaged for three weeks at each stations (see Fig. 2). The rawinsonde wind spectra plotted here have been multiplied with an attenuation factor of the 1.5 km-running average as noted in Section 2 (see Yamanaka *et al.*, 1996, for more detailed discussions).

Theoretically, if the total spectrum has been assumed to be separable in vertical wavenumber m $(= 1/\lambda_z)$, intrinsic frequency $\hat{\omega}$ and azimuthal direction ϕ and is normalized concerning azimuthal spectrum, then the power spectral density $F_q(m)$ of the vertical wavenumber spectrum (given by integrations of the total spectrum over all $\hat{\omega}$ and all ϕ) has a common shape for a physical quantity q (one of zonal wind u, meridional wind v and temperature T) (Garrett and Munk, 1975; VanZandt, 1982). As such a common shape, many observations suggest

$$F_q\,(m) \propto \begin{Bmatrix} m^0 \\ m^{-3} \end{Bmatrix} \qquad \text{for} \qquad m \begin{Bmatrix} \ll \\ \gg \end{Bmatrix} m_* \tag{1}$$

(e.g., Endlich *et al.*, 1969; Smith *et al.*, 1985; Yamanaka, 1985; Fritts *et al.*, 1988; Shibata *et al.*, 1988; Tsuda *et al.*, 1989; VanZandt and Fritts, 1989), which is confirmed also in this study as shown in Fig. 2. (1) implies that most of energy is contained around the so-called characteristic vertical wavenumber m_*, and that waves with $m \approx m_* \approx 1/(2\text{ km})$ are quasi-monochromatically dominant in vertical profiles (see Yamanaka *et al.*, 1996).

Then the ratio between wind and temperature spectra is given by

$$\frac{F_u(m) + F_v(m)}{F_T(m)} = \frac{\displaystyle\int_f^N \left[1 + \left(f/\hat{\omega}\right)^2\right]\left[1 - \left(\hat{\omega}/N\right)^2\right] \cdot B\left(\hat{\omega}\right) d\hat{\omega}}{\left(\overline{T}N/g\right)^2 \displaystyle\int_f^N \left[1 - \left(f/\hat{\omega}\right)^2\right] \cdot B\left(\hat{\omega}\right) d\hat{\omega}}, \tag{2}$$

where g is gravity acceleration, f is the Coriolis parameter, N is the Brunt-Väisälä frequency, $\overline{T}$ is a background temperature and $B(\hat{\omega})$ is the frequency spectral component. If we assume the frequency spectrum with a continuous shape such that

$$B\left(\hat{\omega}\right) \propto \hat{\omega}^{-p} \quad \text{for} \quad f < \hat{\omega} < N \quad \text{and} \quad p > 1,\, f \ll N, \tag{3}$$

the ratio (2) becomes

$$\frac{F_u(m) + F_v(m)}{\left(g/\overline{T}N\right)^2 F_T(m)} \approx p \tag{4}$$

(VanZandt, 1985; Smith *et al.*, 1987; Fritts and VanZandt, 1993).

Figure 3 shows the ratio $[F_u(m) + F_v(m)]/[(g/\overline{T}N)^2 F_T(m)]$ obtained from Fig. 2. It is found that the average value of the ratio for $m \approx m_*$ becomes around 2, from which (4) implies that $p \approx 2$. This value has been suggested by some studies in ocean (Garrett and Munk, 1975) and atmosphere (Yamanaka, 1985; Fukao *et al.*, 1989; Muraoka *et al.*, 1990), and is not so far from a universal value $p \approx 5/3$ for slopes of ground-based frequency spectra (Balsley and Carter, 1982; Larsen *et al.*, 1982; Gage and Nastrom, 1985). It must be noted that the ground-based

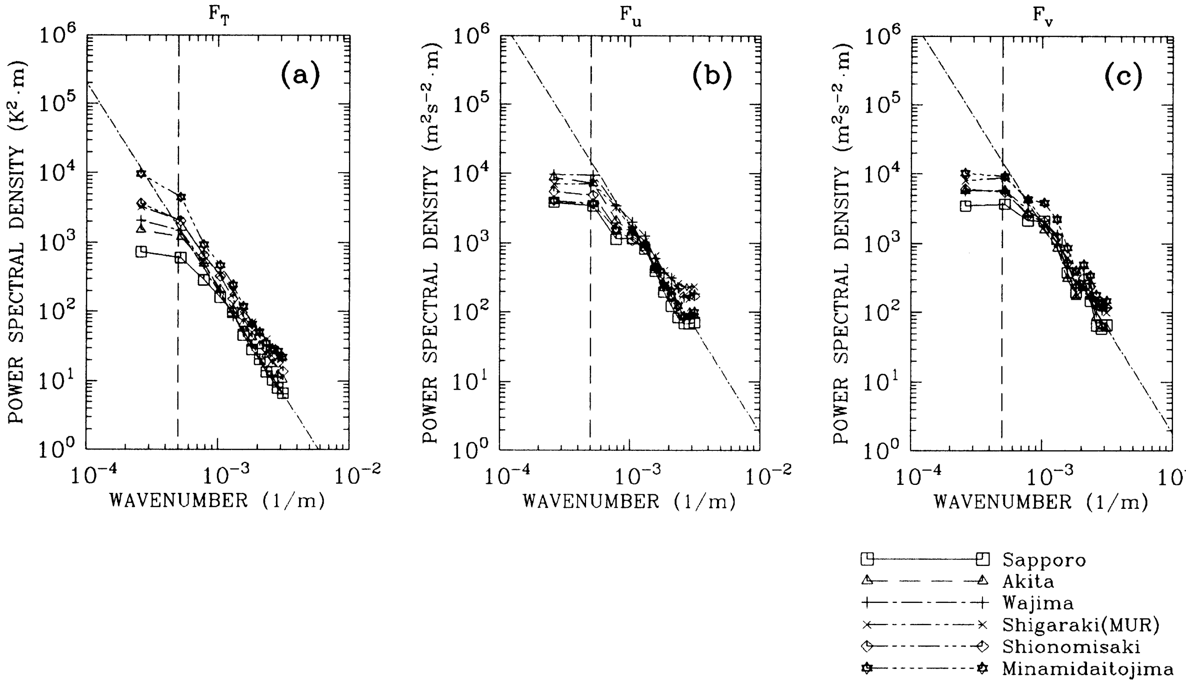


Fig. 2. Mean vertical wavenumber spectra of (a) temperature $F_T(m)$, (b) zonal wind $F_u(m)$ and (c) meridional wind $F_v(m)$ for 15–19 km altitude range, averaged for three weeks (17 June–8 July 1991). Broken vertical line indicates a vertical wavelength 2 km in each panel.

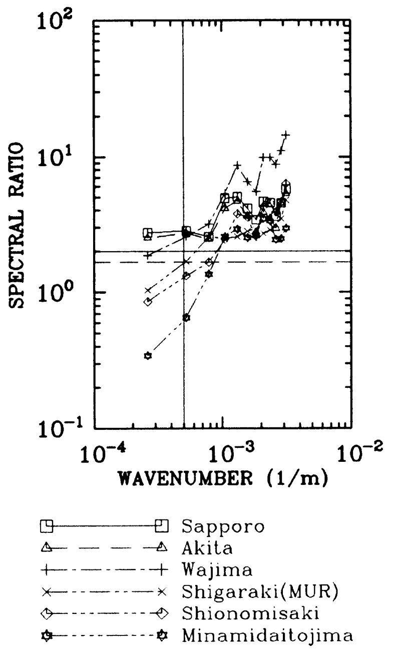


Fig. 3. The spectral ratio $[F_u(m) + F_v(m)] / [(g/\overline{T}N)^2 F_T(m)]$ obtained from Fig. 2. Solid and broken horizontal line indicate the value of 5/3 and 2, respectively, which were suggested as a slope of frequency spectra p in some earlier studies. Solid vertical line indicates a vertical wavelength of 2 km.

frequency ω is in general Doppler-shifted from the intrinsic frequency $\hat{\omega} = \omega - \boldsymbol{k} \cdot \bar{\boldsymbol{u}}$, where $\boldsymbol{k}$ and $\bar{\boldsymbol{u}}$ are horizontal wavenumber and background flow vectors. In Fig. 3 the spectral ratios for m's higher (lower) than m_* seem in general to have p's larger (smaller) than 2, which suggest that the frequency spectra for such non-dominant modes may be different from the so-called universal shape, if the assumptions used in derivation of (4) are correct.

Although horizontal and temporal intervals of the operational rawinsonde observations ($\geq$300 km and $\geq$0.5 day) are quite critical to depict a single wave structure (horizontal wavelength: 10^2–10^3 km; cf. Yamanaka *et al.*, 1996), we may extract horizontal distributions of power spectral densities $F_T(m_*)$, $F_u(m_*)$ and $F_v(m_*)$ for the dominant gravity waves from such observations. For example, it is found from the rawinsonde data along the Japanese Archipelago shown in Fig. 1 that $F_T(m_*)$ and $F_v(m_*)$ increases with decreasing latitude, whereas $F_u(m_*)$ seems to have a maximum near the subtropical tropopausal jet stream axis (Fig. 4(a)). Figure 4(b) shows that for dominant waves with $m_* \approx 1/(2\text{ km})$ the spectral ratio increases (p increases) with a decrease of latitude. These results suggest that the horizontal distributions of gravity wave activity may be different between u, v and T, and that $B(\hat{\omega})$ also has a horizontal dependence.

A difference between the maximum latitudes of wind and temperature fluctuations with vertical wavelengths of 1–10 km, was pointed out by Hirota (1984), based on the rocketsonde data in the upper stratosphere higher than ~30 km over the world. Similar results are found over an altitude range higher than the peak of jet stream in Kitamura and Hirota (1989) who used the JMA rawinsonde data in another year, although they considered that the zonal wind amplitude maximum might be due to sharp peak of the jet stream, and also suggested existense of another maximum common for temperature and zonal wind amplitudes at the north edge of the tropopausal jet stream. Recently, tropical maxima of $F_T(m_*)$ are demonstrated by Allen and Vincent (1995) based on the

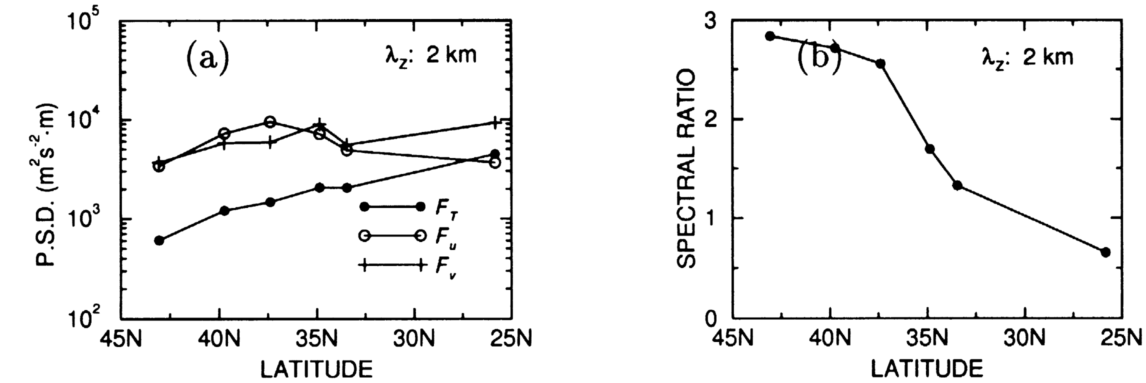


Fig. 4. Meridional variations of (a) power spectral densities of temperature $F_T(m_*)$, zonal wind $F_u(m_*)$ and meridional wind $F_v(m_*)$ for the 2 km wavelength component, and of (b) the spectral ratio $[F_u(m_*) + F_v(m_*)]/[(g/\overline{T}N)^2 F_T(m_*)]$ obtained from Fig. 2 for the 2 km wavelength component.

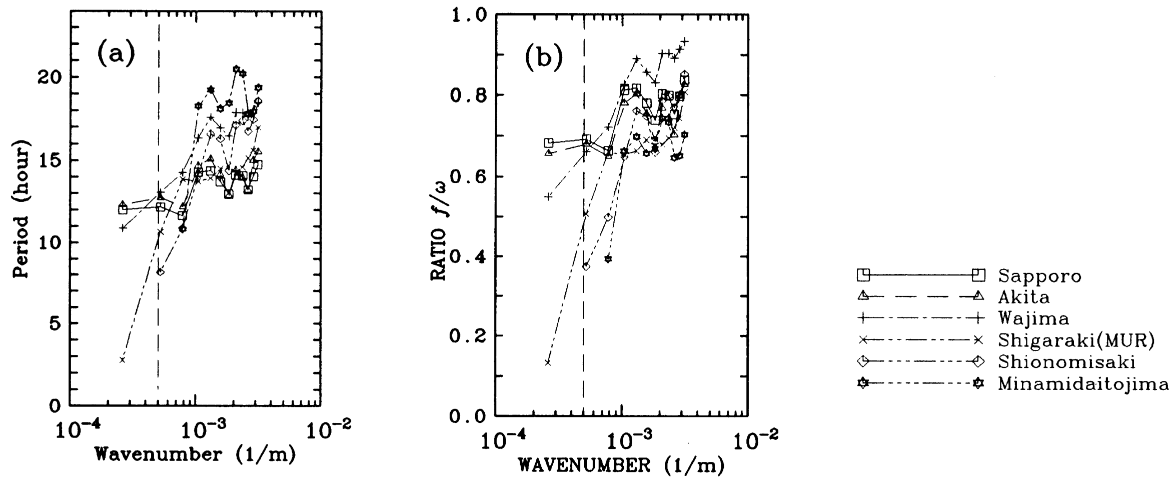


Fig. 5. (a) The intrinsic period $1/\hat{\omega}(m)$ and (b) the ratio $f/\hat{\omega}(m)$ obtained by applying observed values $F_T(m)$, $F_u(m)$, $F_v(m)$, $\overline{T}$ and N to the relationship (6) at each stations. Broken vertical line indicates a vertical wavelength 2 km in each panel.

ABM (Australian Bureau of Meteorology) radiosonde data during 1990–1993, and those of $F_T(m_*)$, $F_u(m_*)$ and $F_v(m_*)$ are also suggested by our analysis of cross-equatorial rawinsonde data in the lower stratosphere (Ogino *et al.*, 1995).

4. Variability of Characteristic Intrinsic Period

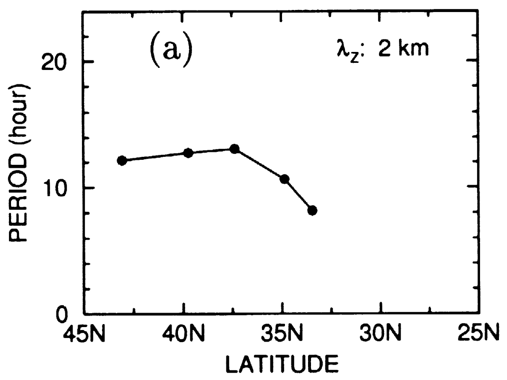
In the previous section we have found that vertical-wavenumber spectral densities for the dominant gravity waves in the mid-latitudes have different meridional variations between temperature and wind disturbances. Since we have found that the parameter p also has a meridional variation, we shall abandon the 'universal' continuous shape (3). Instead, if we hypothesize as an approximation that the monochromatically dominant waves near the characteristic wavenumber $m \approx m_*$ have a line-shape frequency spectrum:

$$B\left(\hat{\omega}\right) \propto \delta\left(\hat{\omega} - \hat{\omega}_*\right), \tag{5}$$

where $\delta(\,)$ is the Dirac delta function, then the spectral ratio (2) becomes

$$\frac{F_u(m_*) + F_v(m_*)}{F_T(m_*)} \approx \left(\frac{g}{\overline{T}N}\right)^2 \left(1 + \frac{f^2}{\hat{\omega}_*^2}\right) \left(1 - \frac{f^2}{\hat{\omega}_*^2}\right)^{-1}, \tag{6}$$

which is not constant but dependent on several parameters (f, $\hat{\omega}_*$, N and $\overline{T}$) with different horizontal variations from each other. In the observations used in the present case study, $\overline{T}$ and N have been given by averaged values (function of time) for 15–19 km altitude range in each vertical profiles, and $\hat{\omega}_*$ is given by an averaged value [1/(9 h)] obtained by hodographs analysis of the MU radar wind data throughout the observational period (see Yamanaka *et al.*, 1996 for details). Then we find that $F_T(m)$ based on radiosonde data at the radar site (shown in Fig. 2(a)) is roughly consistent



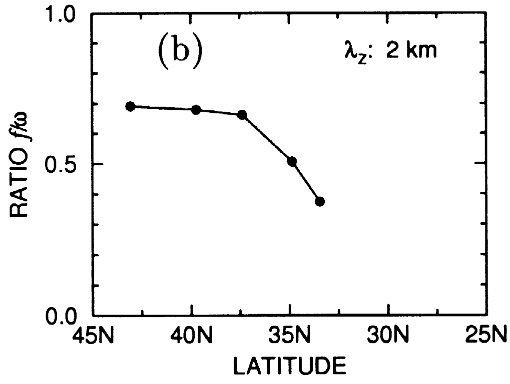


Fig. 6. Meridional variation of intrinsic period $1/\hat{\omega}\,(m_*)$ and (b) the ratio $f/\hat{\omega}\,(m_*)$ obtained by using relationship (6) for the 2 km wavelength component.

with $F_{u,v}(m)$ calculated from the MU radar data (shown in Fig. 2(b) and (c)).

The hypothesis (5) implies that $\hat{\omega}(m_*) \approx \hat{\omega}_*$, but does not deny variations of $\hat{\omega}$ for m and for physical space. For example, if observed wave field is made up with a superposition of finite number of monochromatic waves, a component of such waves may be a set of fixed values of $\hat{\omega}$, m and $\boldsymbol{k}$. In such cases a value of $\hat{\omega}$ is uniquely given for a value of m, that is, $\hat{\omega}$ becomes a function $\hat{\omega}(m)$ of m. Then we shall try to estimate $\hat{\omega}(m)$ which is necessary to explain the difference between observed $F_T(m)$ and $F_{u,v}(m)$. Figure 5 shows intrinsic period $1/\hat{\omega}(m)$ and a ratio $f/\hat{\omega}(m)$ obtained by applying observed values $F_T(m)$, $F_u(m)$, $F_v(m)$, $\overline{T}$ and N to the relationship (6) at each stations. There is a tendency that the intrinsic period becomes longer for waves with shorter vertical wavelengths; $|f/\hat{\omega}|$ becomes smaller for smaller m. This feature is qualitatively consistent with that predicted for an inertio-gravity wave near the critical level ($|\hat{\omega}| \to |f|$), and also not inconsistent with the pioneering results ($|f/\hat{\omega}| \approx$ 0.2–0.4) by Hirota and Niki (1985) for waves with somewhat longer vertical wavelengths ($\lambda_z \approx 1/m \sim 10$ km) under a monochromatic wave assumption (see also Eckermann and Hocking, 1989, for a criticism on this assumption). Figure 6 shows meridional variations of $1/\hat{\omega}(m_*)$ and $f/\hat{\omega}(m_*)$ for dominant gravity waves ($\lambda_z \approx 1/m_* \approx 2$ km). We can find that $1/\hat{\omega}(m_*)$ becomes larger near the jet stream axis, and this may explain differences among the meridional variations of $F_T(m_*)$, $F_u(m_*)$ and $F_v(m_*)$. These meridional variation characteristics are quite interesting to discuss the wave source problem.

5. Consideration on Contaminations of Background Field

It must be considered that the background field also varies with altitude (and time), such as temperature minimum and wind velocity maximum near the tropopause, which may contaminate the vertical wavenumber spectra and their meridional variations (cf. Kitamura and Hirota, 1989; Allen and Vincent, 1995). Figure 7 shows spectra $F_{T'/\overline{T}}(m)$ for the normalized temperature $T'/\overline{T} = (T - \overline{T})/\overline{T}$, where $\overline{T}$ is a background temperature (mean temperature for whole observational period is adopted in this study), and Fig. 8(a) shows

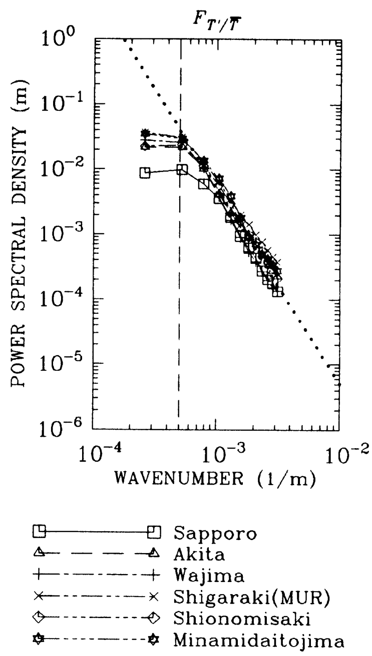


Fig. 7. Mean vertical wavenumber spectra of normalized temperature $F_{T'/\overline{T}}\,(m)$ averaged for the whole observational period (17 June–8 July 1991) at each stations. The 2 km vertical wavelength is indicated by a vertical broken line.

a meridional variation of $F_{T'/\overline{T}}(m_*)$ for the dominant vertical wavelength component ($\lambda_z \approx 2$ km). We can find that $F_{T'/\overline{T}}(m_*)$ increases with decreasing latitude as have been shown for $F_T(m_*)$ (Fig. 2(a)) but the meridional difference of $F_{T'/\overline{T}}(m_*)$ is gentler than that of $F_T(m_*)$. This difference is considered due to contamination of the background temperature minimum in $F_T(m)$. Using the normalized temperature spectra $F_{T'/\overline{T}}$, the spectral ratio (6) becomes

$$\frac{F_u(m_*) + F_v(m_*)}{F_{T'/\overline{T}}(m_*)} \approx \left(\frac{g}{N}\right)^2 \left(1 + \frac{f^2}{\hat{\omega}_*^2}\right)\left(1 - \frac{f^2}{\hat{\omega}_*^2}\right)^{-1}. \quad (7)$$

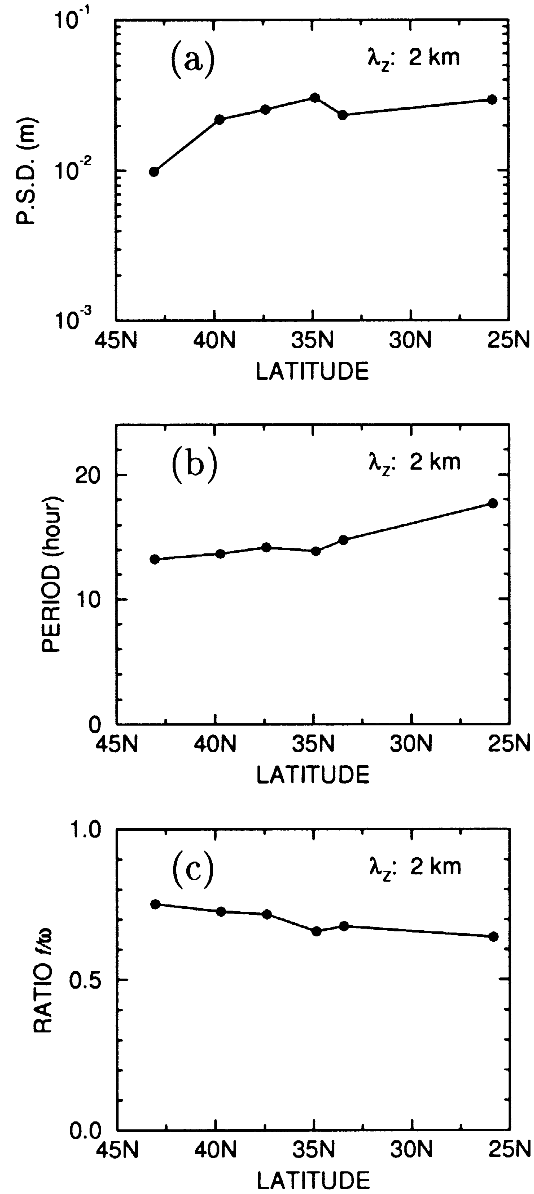


Fig. 8. Meridional variations of (a) power spectral density of normalized temperature $F_{T'/\overline{T}}\,(m_*)$, (b) the intrinsic period $1/\hat{\omega}\,(m_*)$ and (c) the ratio $f/\hat{\omega}\,(m_*)$ obtained by using relationship (7) for the 2 km wavelength component.

From this relationship, meridional variations of the characteristic intrinsic period $1/\hat{\omega}\,(m_*)$ and the ratio $f/\hat{\omega}\,(m_*)$ for the dominant waves ($\lambda_z \approx$ 2 km) are re-estimated as shown in Fig. 8(b) and (c), respectively. The results are different from the results in the previous section. $1/\hat{\omega}\,(m_*)$ is larger in low-latitudes, and $f/\hat{\omega}\,(m_*)$ does not show a considerable meridional variation. This tendency was similar to that suggested by Hirota and Niki (1985) for waves in the upper stratosphere and mesosphere, although their results are somewhat smaller than the present results probably because they analyzed longer vertical wavelengths.

The removal of contaminations due to the background field is dependent on how we define the basic field. If the vertical profile of basic temperature or wind has a sufficiently sharp minimum or maximum, then the contamination may become quite large. However, the basic field must be stable (that is, the gradient of temperature and wind must be sufficiently small) and in the geostrophic and hydrostatic equilibria. Therefore, if an observed profile has quite sharp tropopause or jet stream peak, such a sharpness is not due to the basic field itself but due to contribution of transient ageostrophic components, or inertio-gravity waves.

6. Conclusions

In this paper we have analyzed wind and temperature fluctuations in the lower part (15–19 km) of the stratosphere under an assumption that they are associated with internal inertio-gravity waves (see Yamanaka *et al.*, 1996). We have shown that the dominant modes of such fluctuations have horizontal (meridional) variabilities. This result itself is not new (Hirota, 1984; Kitamura and Hirota, 1989; Allen and Vincent, 1995; Ogino *et al.*, 1995), but here we used an objective analysis to estimate the characteristic intrinsic frequency from vertical wavenumber spectra and have proposed a candidate to explain the different meridional variations between wind and temperature fluctuations. More detailed studies using another technique (cf. Shimomai *et al.*, 1996, 1997) and other data of different seasons are strongly expected to clarify why these variabilities appear. Answers to this question may explain also the generation mechanisms of the middle-atmospheric inertio-gravity waves.

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