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Internal Gravity Wave Selection in the Upper Troposphere and Lower Stratosphere Observed by the MU Radar: Preliminary Results

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Abstract—Marked wavelike variations of the lower stratospheric wind observed on 7–10 May, 1985 by an MST radar in Japan (by the MU radar) are analyzed assuming that they are induced by monochromatic internal inertio-gravity waves. These variations are mainly composed of two modes (periods: 22 and 44 hours), both of which have zonal phase velocities (C_X) slower than the mean westerly wind ($\bar{u}$). A statistical analysis of the zonal phase velocity shows that $C_X \lesssim \bar{u}$ above and $C_X \sim \bar{u}$ below the tropopause jet stream, which is considered to be a vivid proof of wave selection due to the tropospheric mean flow and upward wave emission from the tropopause jet. A comparison between the MU radar results and routine meteorological observations leads to the conclusion that the marked waves appear when the jet stream takes a maximum wind speed.

Key words: MST radar observation, internal gravity waves, monochromatic analysis, wave-mean flow interaction.

1. *Introduction*

Observational studies of stratospheric (inertio-) gravity waves can be classified into two categories. One is the computation of a continuous spectrum from the observational data; some universal shapes of spectral functions have been found and fitted to models based either on two-dimensional turbulence theory (*e.g.*, GAGE and NASTROM, 1985) or else on the superposition of gravity waves (*e.g.*, VAN-ZANDT, 1985). Such studies have become possible after the recent progress of the MST radar technique which has provided us continuous high-resolution wind data in time and altitude.

The other category is the analysis of monochromatic wave parameters from a selected or filtered portion of observational data. This category was originated by

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the pioneering study of SAWYER (1961) using several rawinsondes, and followed by many studies in the last decade using rawinsondes (THOMPSON, 1978; CADET and TEITELBAUM, 1979), instrumented large balloons (SIDI and BARAT, 1986), rocketsondes (HIROTA and NIKI, 1985) and MST radars (MAEKAWA *et al.*, 1984; HIROTA and NIKI, 1986). It is more straightforward to compare the observations and theories in the second rather than the first category, because most of the theories of the middle-atmospheric gravity waves assume slowly-varying quasi-monochromatic waves (see, *e.g.*, FRITTS, 1984).

In the present paper we show some examples of the second category, using an MST radar called the MU radar which is located at Shigaraki, Japan (34.86°N, 136.10°E; *cf.*, FUKAO *et al.*, 1985a,b, 1988, 1989). In Section 2 we report two marked wavelike structures observed by the MU radar on 7–10 May, 1985, and analyze them by assuming a theoretical structure for monochromatic gravity waves. In Section 3 we perform a statistical analysis on the gravity-wave phase velocities for the same observational data. Section 4 discusses the results, including comparison with the routine meteorological observations.

2. *Monochromatic Analysis*

Figure 1 shows the temporal-vertical distributions of zonal and meridional wind perturbations from the average profiles during the observational period. We find a predominance of the following two wave modes in the lower stratosphere (14–24 km): one has a period (τ) of about 22 hours and a vertical wavelength (λ_z) of 2–3 km, and the other is $\tau \sim 44$ hours and $\lambda_z \sim 1$ km. Hereafter these two predominant modes are called the 22- and 44-hour modes. It should be noted that the amplitudes of these modes are not constant throughout the observational period and altitude range; they are particularly strong during 8–9 May at altitudes higher than the tropopause jet stream (~ 12 km). Also note that the inertial period τ_i ($= 2\pi/f$; f is the Coriolis parameter or inertial frequency) is about 21 hours at the latitude of the MU radar (34.85°N).

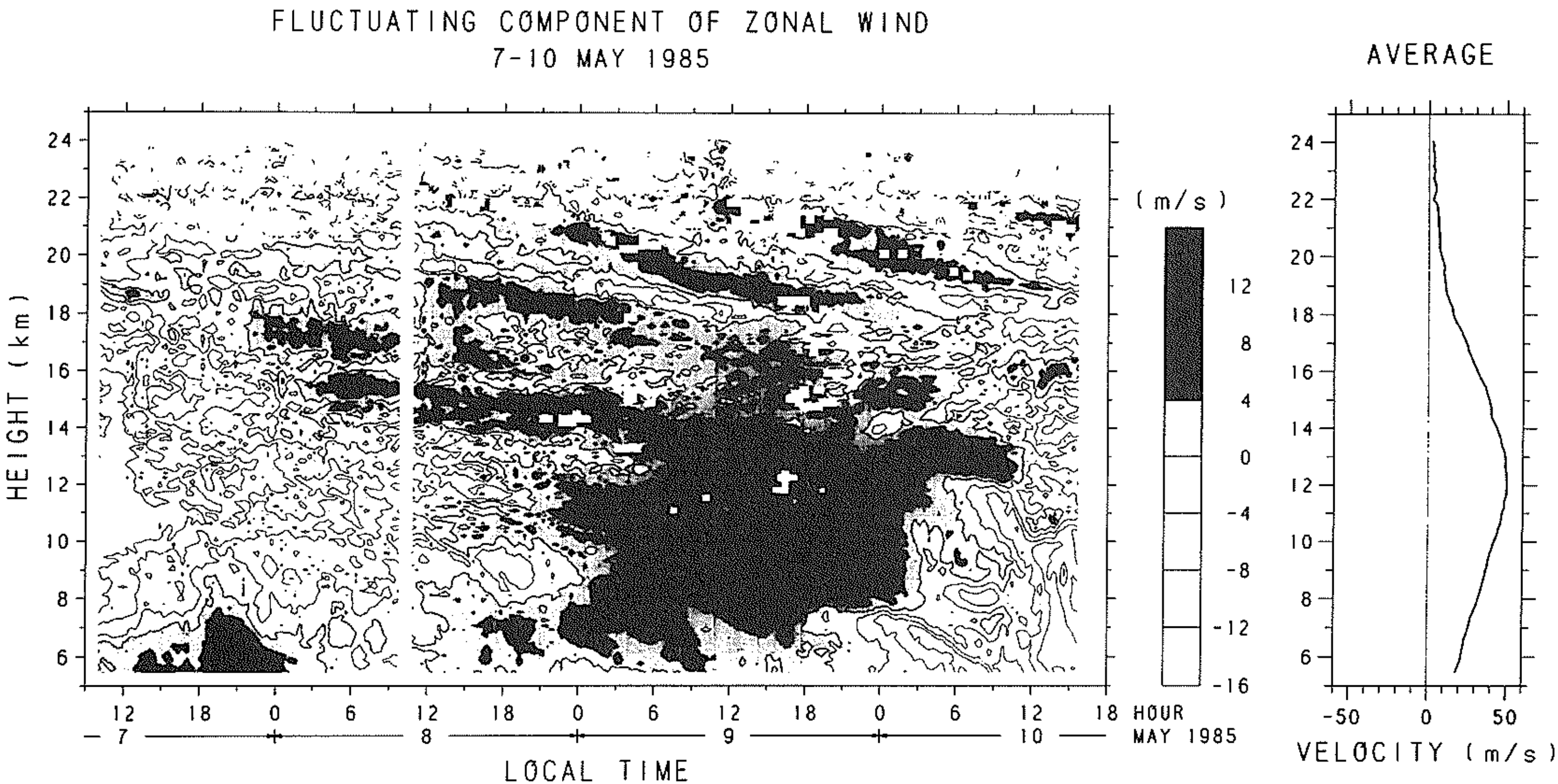
In order to examine the structures of the 22- and 44-hour modes we made 30-min averages in time and then filtered out 1.5–4.5 and 0.6–1.6 km bands in vertical wavelength as shown in Figures 2a and b, respectively. It can be clearly confirmed in this filtered data that the marked wave modes are particularly strong during 8–9 May at altitudes higher than the tropopause jet stream. We can also find that the phases propagate downward whereas the strong amplitudes proceed upward, which suggests an upward energy radiation in a shape of wavepacket as described in many foregoing studies (*e.g.*, CADET and TEITELBAUM, 1979; MAEKAWA *et al.*, 1984; HIROTA and NIKI, 1985, 1986).

For each filtered data set we analyzed hodographs (*cf.*, HIROTA and NIKI, 1985, 1986) and determined the wave parameters such as U'_H (amplitude of horizontal wind variation), $\hat{\tau}$ (intrinsic period), λ_H (horizontal wavelength), α (direction of horizontal wavenumber vector), C_X (zonal phase velocity) and E (kinetic energy density) assuming a monochromatic inertio-gravity wave. The results for the 22- and 44-hour modes shown in Figures 2a and b are plotted in Figures 3a and b, respectively. It should be noted that the parameters shown here are averaged in time throughout the observational period and smoothed in the vertical by an approximately 2 km running mean. First of all, the energy densities of both modes are concentrated in a layer between 10–17 km altitude. In this layer, variations of $\hat{\tau}$, λ_H, α and C_X are small enough to use the assumption of a monochromatic wave for these analyses.

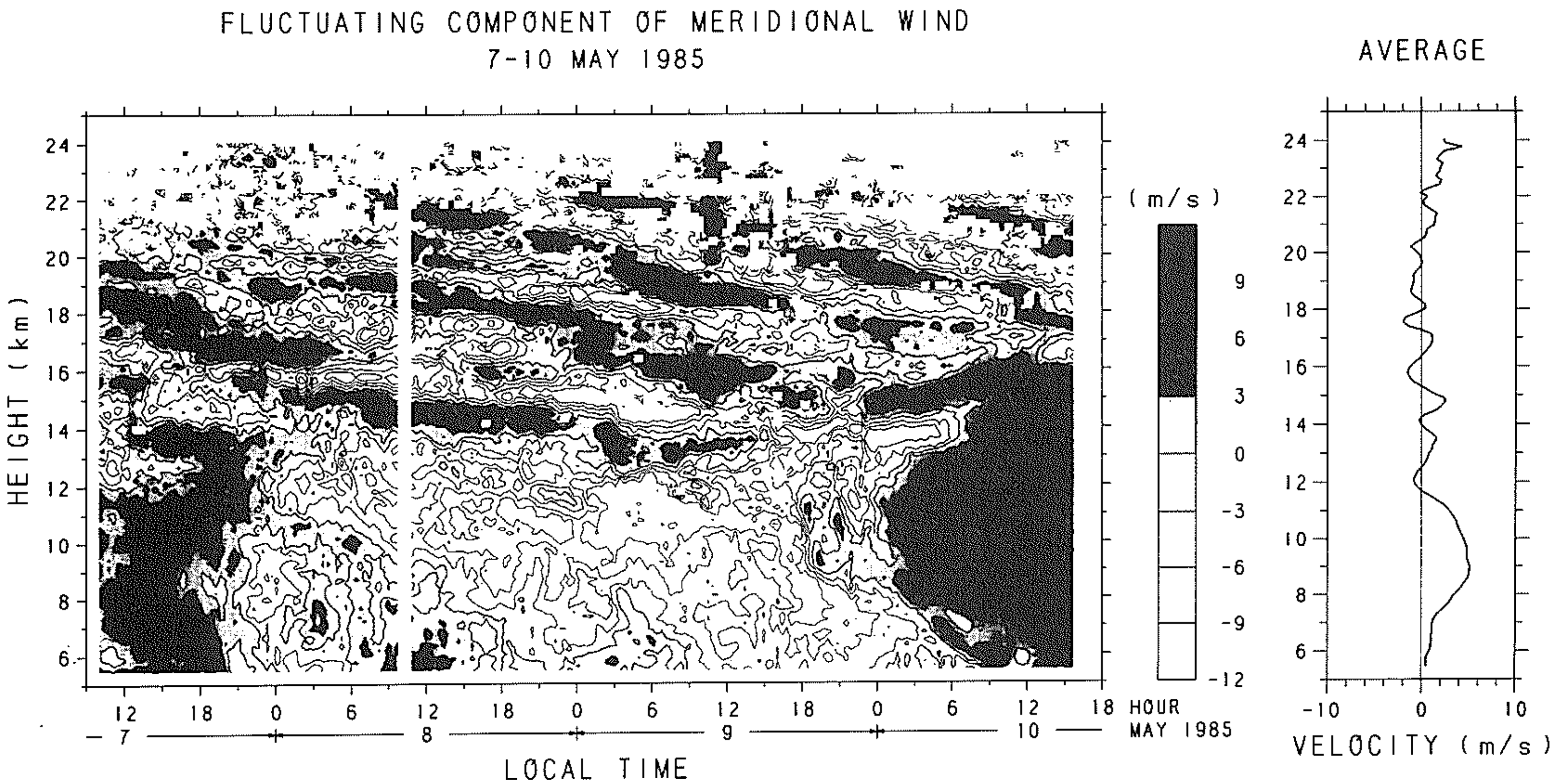
Both waves have almost southward wavenumber vectors in the whole altitude range observed. The wavefronts are aligned almost zonally and so the zonal phase velocities are significantly larger than the meridional phase velocities. However, the horizontal wavelengths are different ($\sim$200 and $\sim$50 km, respectively). For the 22-hour mode in the 10–17 km layer, we find from Figure 3 that $C_X \sim 7$ m/s and that the zonal wavelength (λ_X) is $\sim$600 km. For the 44-hour mode in the 10–17 km layer, we have $C_X \sim 3$ m/s and $\lambda_X \sim 500$ km. Parameters U'_H, $\hat{\tau}$, λ_H, C_X and E of the 44-hour mode are smaller than those of the 22-hour mode.

The intrinsic periods of the 22 and 44 hour modes are estimated as 7 and 5 hours, respectively, so that we have $\hat{\tau}/\tau_I = 0.33$ and 0.24 for the respective periods. These values of $\hat{\tau}/\tau_I$ are close to the most predominant ones (0.3–0.4) in the middle- and upper-stratospheric rocketsonde data (HIROTA and NIKI, 1985), and are out of the inertio-gravity wave critical layer ($\hat{\tau}/\tau_I \gtrsim 1/\sqrt{2}$) predicted by YAMANAKA (1985). The modulation of such a predominant mode by the earth's rotation (or the inertial effect due to the Coriolis force) is considerable for the wave polarization (hence the hodograph becomes an ellipse), but it is not so effective for the wave dispersion (hence the ellipse is much elongated in the wavenumber vector direction). Therefore, we can describe the vertical propagation of the marked wave modes using the simple noninertial wave theory in a valid approximation.

In the simple theory, the critical level is defined by an altitude at which $\bar{u} = C_X$, and the breaking level by the local convective instability is an altitude at which $U'_H \simeq |C_X - \bar{u}|$ (see FRITTS, 1984). Here $\bar{u}$ is the mean zonal wind shown on the right-hand side of Figure 1a. The mean meridional wind in the present observational period is much weaker than $\bar{u}$ as shown in Figure 1b. Using typical values in the 10–17 km layer, the critical and breaking levels for the 22-hour mode are estimated as $\sim$20 and $\sim$19 km altitudes, respectively. For the 44-hour mode they are $\sim$21 and $\sim$20 km, respectively. Remembering that the results shown in Figure 3 are smoothed, we consider that local wavebreaking may occur at altitudes somewhat lower than the breaking level estimated above. Accordingly, the saturation of U'_H and decrease of E above the 10–17 km layer can be interpreted as the result of wavebreaking. This interpretation is consistent with self-acceleration of C_X above the layer (*cf.* TANAKA, 1986). There may be some waves newly emitted as a


FLUCTUATING COMPONENT OF ZONAL WIND
7-10 MAY 1985
AVERAGE
HEIGHT (km)
(m/s)
12
8
4
0
-4
-8
-12
-16
HOUR
MAY 1985
LOCAL TIME
VELOCITY (m/s)
-50
0
50

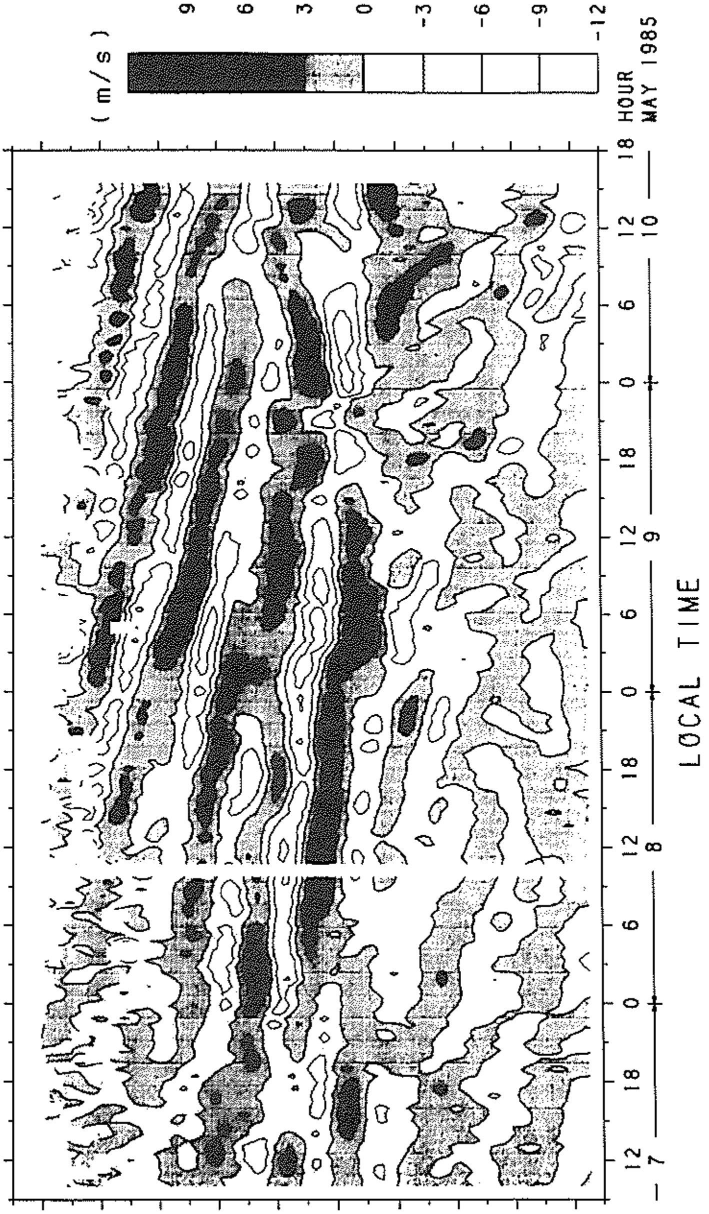
(a)



(b)

Figure 1

Time-altitude cross-sections of (a) zonal and (b) meridional wind perturbations relative to the mean wind averaged over the whole observational period shown on the right-hand side. Contours are drawn at 4 m/s intervals.


(m/s)
9
6
3
0
-3
-6
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-12
HOUR
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HEIGHT (km)
24
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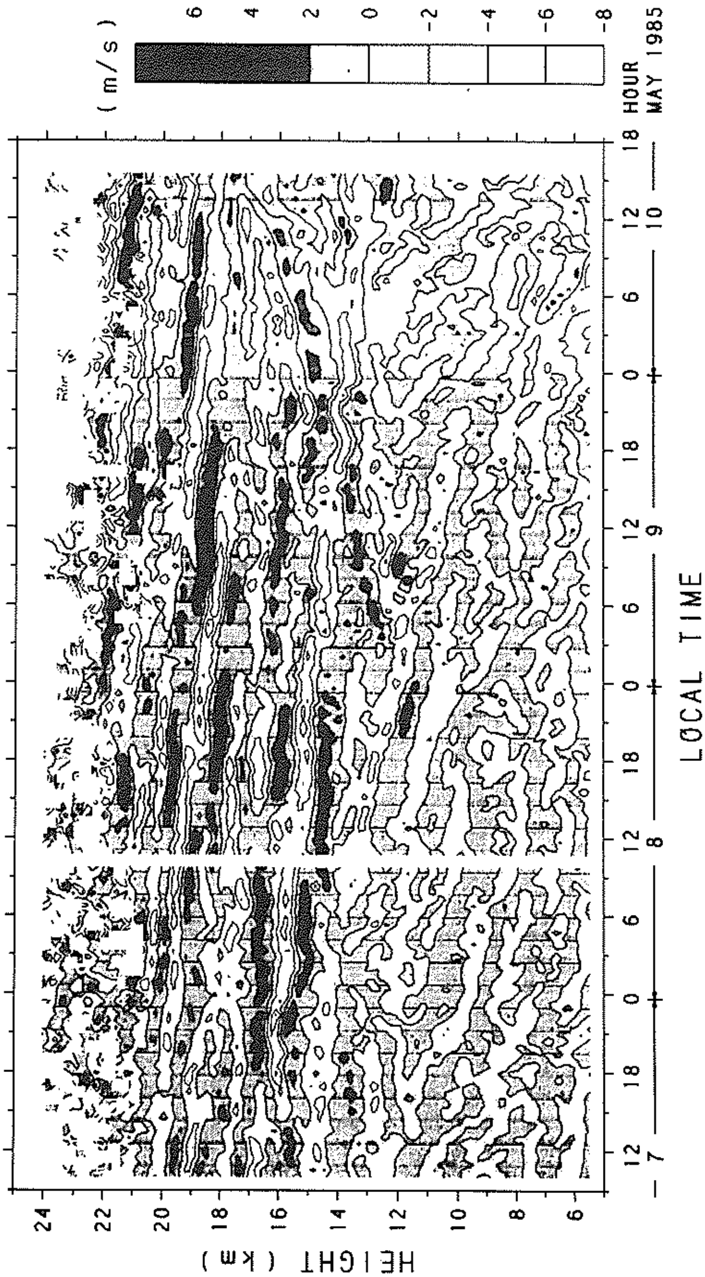
10

(a)



(b)

Figure 2

Time-altitude cross-sections of 30-min averaged, (a) 1.5–4.5 km and (b) 0.6–1.6 km band-passed meridional wind perturbations. Contours of (a) and (b) are drawn at 3 and 2 m/s intervals, respectively.

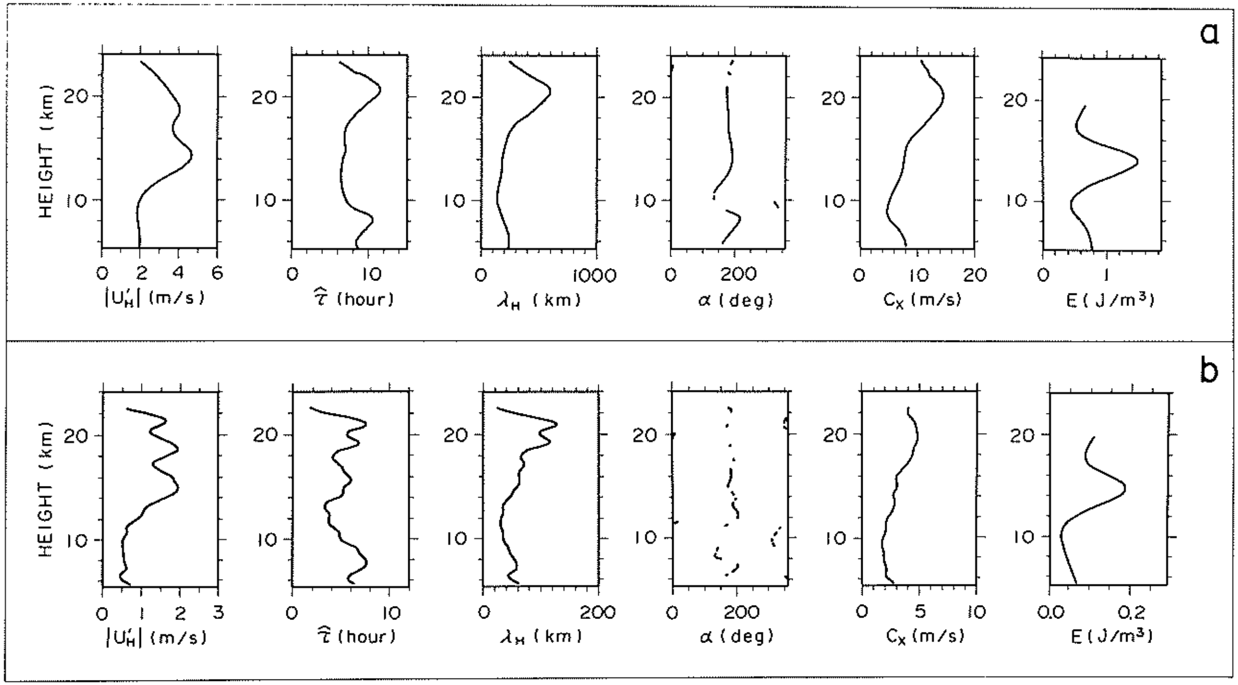


Figure 3
Vertical profiles of analyzed monochromatic wave parameters for (a) 22-hour and (b) 44-hour modes. U'_H: the horizontal velocity amplitude; $\hat{\tau}$: the intrinsic period: λ_H: the horizontal wavelength; α: the horizontal wavenumber vector direction; C_X: the zonal phase velocity; and E: the energy density.

result of wavebreaking (*cf.* YAMANAKA, 1985; ZHU and HOLTON, 1987). It should be noted that $C_X < \bar{u}$ is satisfied for both modes within the 10–17 km layer.

The echo power distribution (not shown) suggests a turbulence layer structure approximately along wavefronts of the 22- and 44-hour modes. These resemble turbulence layers observed by many MST radar observations (*e.g.*, SATO and WOODMAN, 1982; WOODMAN and RASTOGI, 1984; SATO *et al.*, 1985). Theoretical studies predict that the turbulence layers made by breaking internal gravity waves might be very thin and require a higher vertical resolution for their detection (TANAKA, 1983; YAMANAKA and TANAKA, 1984a,b). Such a thin structure has been found in some *in situ* observations (*e.g.*, BARAT, 1982, 1983; YAMANAKA and TANAKA, 1984a,c, 1985; YAMANAKA *et al.*, 1985; SIDI and BARAT, 1986). In order to decide whether the turbulence found by the echo power data is generated directly by the predominant wave modes seen in the Doppler-frequency data, we are going to analyze both types of data in detail in a subsequent paper.

3. *Statistical Analysis of Wave Phase Velocities*

In the preceding section we have studied monochromatic wave parameters averaged throughout the observational period. However, as noted in Figures 1 and 2, two predominant modes are amplified in particular during 8–9 May. We consider that activities and wave parameters of the predominant modes might vary with time, and that some other modes might be activated. Among the various kinds of wave parameters, the zonal phase velocity C_X is the most important quantity to compare the observational evidence with the wave-mean flow interaction theory (*cf.*, FRITTS, 1984). In this section we perform a statistical study on the variability of C_X during the observational period in each altitude.

We take the 1.5–4.5 km band data set of which the meridional component wind has been shown in Figure 1a. Figure 4 shows the histograms of C_X and analyzed from this data set for each altitude. The mean value of C_X at each altitude has been plotted in a panel of Figure 3a, that is, C_X of the 22-hour mode discussed in Section 2. The thick curve in Figure 4 is the mean zonal wind $\bar{u}$ transcribed from the right-hand side panel of Figure 1a.

We find clearly from Figure 4 that $C_X \lesssim \bar{u}$ throughout the observational altitude range and period. This feature is consistent with the wave selection mechanism by the mean flow as predicted theoretically by MATSUNO (1982). This mechanism is essentially explained by the fact that the upward propagating waves should be trapped under their critical levels. Since the wave breaks near the critical levels, each wave becomes predominant at an altitude where $C_X = \bar{u} + a$ and a is a nonzero unknown parameter. This implies that $\hat{\tau}[= \lambda_X/(C_X - \bar{u})]$ is almost constant, if a and λ_X are almost constant. Therefore, $\hat{\tau}/\tau$ becomes almost constant for a latitude (see the preceding section and also HIROTA and NIKI, 1985), when a and λ_X are almost constant.

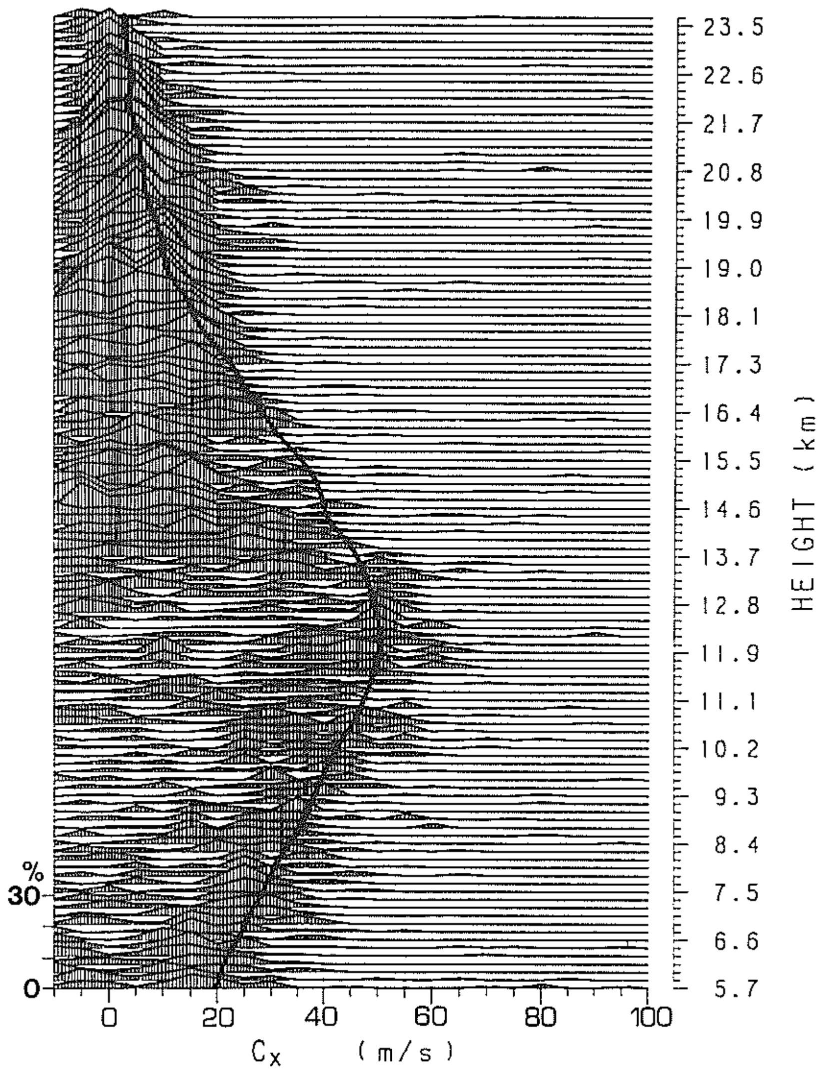


Figure 4

Vertical profile of histograms of the zonal phase velocity analyzed from the 30-min averaged, 1.5–4.5 km band-passed data set. The histogram scale is correct for the lowest spectral curve. The thick curve is the mean zonal flow profile.

However, we notice in Figure 4 that the values of C_X are concentrated near $\bar{u}$ at altitudes lower than the peak of the tropopause jet stream (the tropospheric side), and that it is spread in a broad range less than $\bar{u}$ above the jet peak (the stratospheric side). The feature of the tropospheric side can be easily understood by the selection mechanism of the predominant mode at each altitude, but the feature of the stratospheric side cannot be explained by assuming that all the waves are

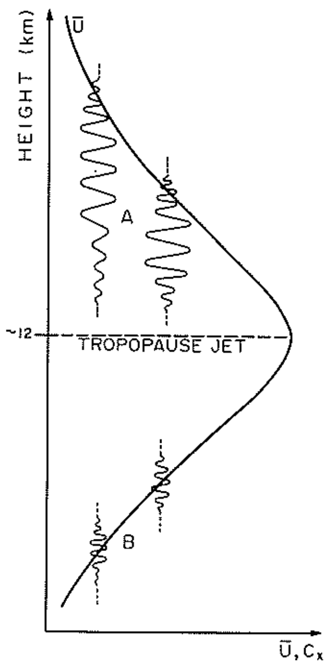


Figure 5
A tentative model of the wave phase-velocity selection. See text for details.

generated near the ground. Although topographic stationary waves are not detected by single-station observations, most of them are considered to be trapped in the lower troposphere unless C_X varies greatly during the propagation (*cf.*, TANAKA, 1986). Therefore, most of stratospheric gravity waves appear to be generated at altitudes higher than the jet stream peak.

Figure 5 shows a tentative model based on the present observations: "A" denotes the stratospheric waves generated mainly at the tropopause jet stream, and "B" indicates the tropospheric waves which have considerable amplitudes only near their critical levels. The upward wave emission from the jet stream has been described by several authors (*e.g.*, CADET and TEITELBAUM, 1979; HIROTA and NIKI, 1986), but at least in the present observations downward wave emission from the jet as described by Hirota and Niki is not clearly seen. Mechanisms of the two wave-generation sources will be left as an important problem for future study (*cf.*, FUKAO *et al.*, 1989).

4. *Further Discussions and Conclusions*

In Section 3, we have obtained a model to explain why the two modes described in Section 2 can be predominant at the observed altitude range. However, we have not explained why such marked waves appeared so strongly on 8–9 May. Although the wave generation mechanism is a problem of future studies, here we point out an interesting feature related possibly to the wave generation, based on routine meteorological observations. Figure 6 shows that a wind-speed maximum passed over the MU radar on 8–9 May. If the tropopause jet stream can generate the marked waves by some instability or adjustment mechanism, such wave generation most probably appears near the maximum-speed portion of the jet stream (*cf.*, MASTRANTONIO *et al.*, 1976; CADET and TEITELBAUM, 1979).

Furthermore, Figure 6 shows some wavy patterns at the particular times when the marked waves were observed by the MU radar. Pressure-surface analyses, on

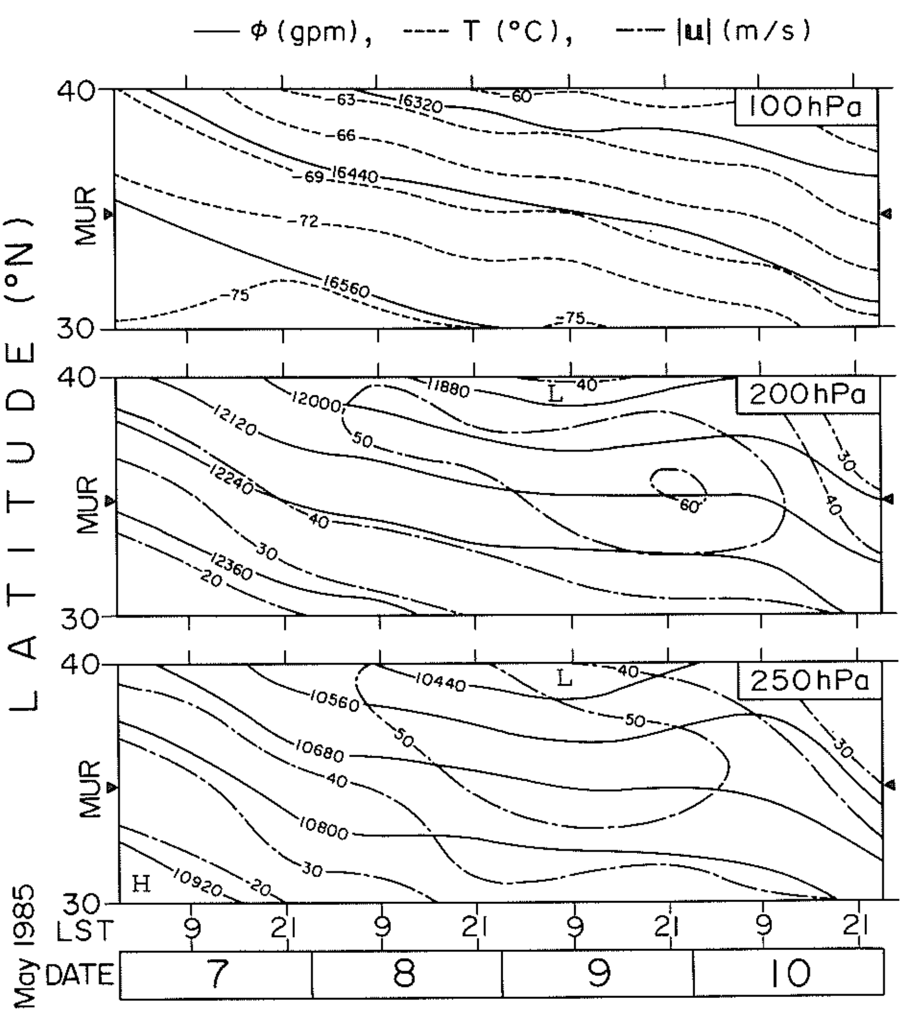


Figure 6
Time-latitude cross-sections of geopotentials (ϕ), temperatures (T) and wind speed ($|\mathbf{u}|$) over 100-, 200- and 250-hPa levels, based on JMA (1983, 1985a,b,c).

which Figure 6 is based, are objective using an optimum interpolation method (JMA, 1983). Thus we consider that detection of long-wavelength stratospheric gravity waves by the routine observations and analyses is not impossible, although the time resolution is 12 hours. Such a study is very important not only for estimation of horizontal wavelengths and horizontal dimensions of wavepackets, which cannot be directly measured by the single-stationed radar observations, but also to illustrate the utility of the MST radar data to assimilate and initialize numerical weather predictions.

In summary, the observed internal gravity waves seem to be selected essentially by the mechanism of MATSUNO (1982). They are predominant above the peak altitude of the tropopause jet stream. They may break and decelerate the mean flow in the lower stratosphere as considered by TANAKA and YAMANAKA (1985), but we cannot find any evidence that they are generated directly by the surface orography. In the present observation most of the tropospheric waves have phase velocities around the mean flow. If both stationary and travelling waves are generated near the ground by some mechanisms, most of them would be trapped in the tropospheric altitudes and could not arrive in the tropopause altitude.

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