

## Dominant vertical scales of gravity waves in the middle atmosphere observed with the MU radar and rocketsondes

Y. MURAYAMA,\* T. TSUDA,\* M. YAMAMOTO,\* T. NAKAMURA,\* T. SATO,† S. KATO\*  
and S. FUKAO\*

\*Radio Atmospheric Science Center, Kyoto University, Uji, Kyoto 611, Japan; †Department of Electrical Engineering II, Kyoto University, Kyoto 606, Japan

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**Abstract**—We have simultaneously observed wind motions in the altitude range of 5–90 km by means of the MU radar, rocketsondes and radiosondes. Dominant vertical scales of wind fluctuations due to gravity waves were 2–5 km in the lower stratosphere, about 5–15 km in the upper stratosphere and longer than 15 km in the mesosphere. The increase in the vertical scale with altitude is interpreted in terms of the saturation of upward propagating gravity waves. In the stratosphere, the observed vertical wavenumber spectra showed smaller amplitudes and more gradual slopes than the model values. Furthermore, the wind velocity variance in the stratosphere increases exponentially with an  $e$ -folding height of about 9 km, implying that the gravity waves were not fully saturated. On the other hand, the spectra in the upper stratosphere and mesosphere agreed fairly well with the model spectra. The variance in the mesosphere seems to cease increase of the wave amplitudes and agrees reasonably well with the model value.

### 1. INTRODUCTION

It is now widely accepted that mesoscale wind fluctuations in the middle atmosphere are caused by gravity waves which are mainly excited in the lower atmosphere due to, for example, the unstable behavior of a jet stream or various meteorological disturbances such as fronts, typhoons, thunder storms and cumulonimbus convections. Gravity waves propagate upward, transporting momentum and energy, generate shear or dynamical instabilities, and finally act to decelerate the mean flow in the mesosphere (e.g. LINDZEN, 1981; HOLTON, 1982; MATSUNO, 1982; FRITTS, 1984). Wind velocity profiles normally appear as the superposition of many waves with various vertical wavenumbers, which can best be examined by spectrum analysis. A linear saturation theory has been developed in order to explain the vertical wavenumber spectrum, which predicts that gravity waves with smaller vertical scales tend to become saturated due to convective or shear instabilities (DEWAN and GOOD, 1986; SMITH *et al.*, 1987).

Observations with MST radar have contributed to studies of the saturated gravity wave spectra in the middle atmosphere (e.g. FRITTS *et al.*, 1988; TSUDA *et al.*, 1989, 1990). MST radar is, however, unable to monitor wind velocity profiles in the upper stratosphere and lower mesosphere (normally between 25 and 60 km altitude), because the refractive index fluctuations are suppressed because of low atmospheric and electron densities. In order to complement vertical

profiles obtained on MST radar observations, we have carried out simultaneous observations of wind velocity profiles in the troposphere and middle atmosphere by means of the MU radar and rocketsonde. Routine radiosonde observations were also made for the comparison. We report in this paper the altitude variation of dominant vertical scales and vertical wavenumber spectra of wind velocity profiles, and their interpretation in terms of the saturated gravity wave theory.

### 2. SATURATED GRAVITY WAVE SPECTRUM

In this section, we briefly review the basic concept of gravity wave saturation developed by DEWAN and GOOD (1986) and SMITH *et al.* (1987). A model spectrum of gravity waves was introduced in order to describe fluctuations caused by a superposition of many waves (e.g. VAN ZANDT, 1985). For instance, a vertical wavenumber spectrum of horizontal wind velocity can be expressed as  $F_u(m/m_*) = C/[1 + (m/m_*)^t]$ , where  $m_*$  is the characteristic wavenumber,  $C$  is a constant and  $t$  is the asymptotic slope of the spectrum which we take to be 3 (DESAUBIES, 1976).

As illustrated in Fig. 1, the logarithmic spectral slope is  $-3$  for large wavenumbers, where gravity waves are thought to be saturated, while the spectrum is flat in the unsaturated range. Based on the linear theory of gravity waves, SMITH *et al.* (1987) quantitatively explained a saturated part of the spectrum

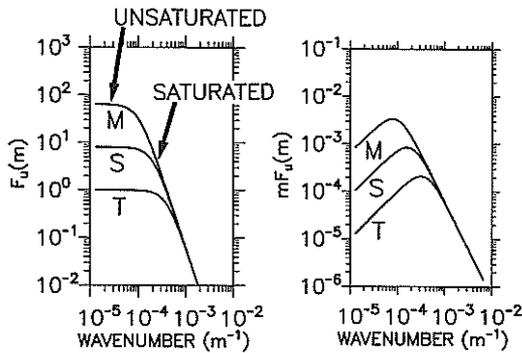


Fig. 1. A model vertical wavenumber spectrum of saturated gravity waves in normal (left) and energy content forms (right) (SMITH *et al.*, 1987). The portion of the spectrum with a slope of  $-3$  is caused by the saturated gravity waves, while the flat region is unsaturated.

and derived the saturated spectrum of horizontal wind fluctuations as

$$F_u^s(m) = \frac{N_b^2}{6m^3},$$

where  $N_b^2$  is the background Brunt-Väisälä frequency squared value.

We here schematically illustrate the underlying idea of the linear saturation theory. Figure 2 shows profiles

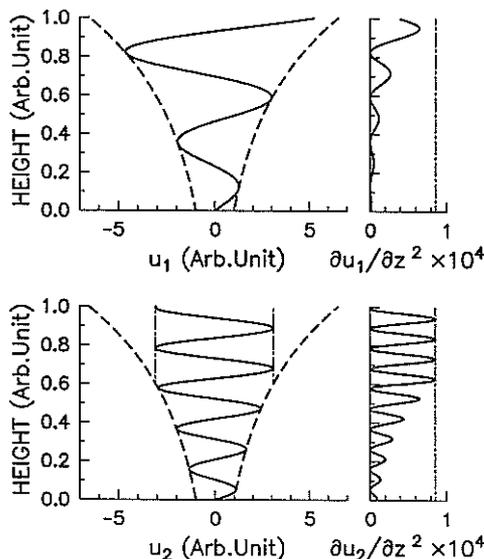


Fig. 2. Schematic profiles of wind velocity,  $u$  (left) and vertical wind shear squared,  $|\partial u/\partial z|^2$  (right) for gravity waves with long (top) and short (bottom) vertical wavelengths, respectively. The dashed curve shows an asymptotical growth of the wave amplitude due to the exponential decrease of the atmospheric density. Wave 2 is assumed to reach the saturation condition when  $|\partial u/\partial z|^2$  becomes as large as the value indicated by the vertical chained line.

of wind velocity,  $u$ , and vertical wind shear squared,  $|\partial u/\partial z|^2$ , for two monochromatic gravity waves indicated by subscript numbers, 1 and 2, which correspond to waves with small and large vertical wavenumbers,  $m_1$  and  $m_2$ , respectively. When gravity waves propagate upward, conserving their energy, both  $u_1$  and  $u_2$  increase similarly, in proportion to  $\exp(z/2H)$  where  $H$  is the density scale height in an isothermal atmosphere. On the other hand,  $|\partial u/\partial z|^2$  increases as  $[(1/2H)^2 + m^2]\exp(z/2H)$ . When  $1/2H$  is much smaller than  $m$ ,  $|\partial u/\partial z|^2$  becomes proportional to  $m^2$ , so that the shear intensity becomes much larger for the wave with  $m_2$  than that with  $m_1$ . An upward propagating wave becomes unstable when it reaches an altitude where the local stability becomes lower than some threshold. Since the atmospheric stability is inversely proportional to  $|\partial u/\partial z|^2$ , a wave with a larger  $m$  more easily becomes unstable. As a result, a wave with a larger  $m$  becomes unstable at lower altitudes and its amplitude remains a constant value. Thus, the spectrum amplitudes for large wavenumbers are limited to the saturated values, while those for unsaturated components increase as the altitude increases.

It is noteworthy that an energy content graph in Fig. 1 suggests that the wave component at the bend of the spectrum, which has just reached the saturation condition, exhibits the largest energy. In other words, such a component is most dominantly detected in the wind velocity profiles.

Figure 1 also predicts that the dominant vertical scale of gravity waves generally increases as they propagate upward. For example, if we assume an altitude difference of  $5H$  between the lower stratosphere and mesosphere, the wave amplitudes for an unsaturated component are expected to increase by a factor of  $\exp(5H/2H) = 12.2$ , that is, the spectral densities by a factor of  $\exp(5H/H)$ . In such a case, Fig. 1 suggests that the wavenumber corresponding to the bend of spectra decreases by a factor of  $\exp(5H/3H) = 5.3$ . These model predictions will be compared with the observed characteristics of gravity waves at the end of Section 4.

### 3. EXPERIMENTAL TECHNIQUES

A meteorological rocket, MT-135, is equipped with a temperature sensor, the values obtained being transmitted to a ground-based station by means of a transponder. The location of the descending sonde is determined with a ground-based radar and recorded on a magnetic tape every 100 ms. Horizontal wind velocities are determined from the time derivatives of horizontal displacements of the rocketsonde.

We have launched, in total, four meteorological rockets from the Kagoshima Space Center, Uchinoura, Japan ( $31^{\circ}15'N$ ,  $131^{\circ}05'E$ ), that is, at 1125 JST (0225 GMT) on 11 September 1985 (MT135-44), 1120 JST (0220 GMT) on 20 August 1986 (MT135-45), and 1100 and 1300 JST (0200 and 0400 GMT) on 9 September 1987 (MT135-47 and MT135-48). Note that MT135-47 and MT135-48 were successively launched with an interval of 2 h. Temperature profiles have successfully been obtained only with MT135-45 and MT135-48.

We also utilize wind velocity data collected by routine rocketsondes launched from Ryori ( $39^{\circ}02'N$ ,  $141^{\circ}50'E$ ) Meteorological Rocket Station, as well as radiosonde observations at the Kagoshima ( $31^{\circ}38'N$ ,  $130^{\circ}36'E$ ) and Sendai ( $38^{\circ}16'N$ ,  $140^{\circ}54'E$ ) Weather Stations. Radiosonde observations of temperature profiles in the lower stratosphere at Shionomisaki ( $33^{\circ}27'N$ ,  $135^{\circ}46'E$ ) Weather Station are used to determine the Brunt-Väisälä frequency squared values for the calculation of a model spectrum.

During the rocketsonde measurements we monitored the wind profiles in two separate height regions; the troposphere and lower stratosphere (5–22 km altitude), and the mesosphere (60–85 km) with height resolutions of 150 and 600 m, respectively, by means of the MU radar at Shigaraki, Japan ( $34^{\circ}51'N$ ,  $136^{\circ}06'E$ ). The antenna beam was sequentially steered

from the vertical to four oblique angles at  $10^{\circ}$  aligned to the east–west and north–south directions.

The locations of the above mentioned observations stations are indicated in Fig. 3.

#### 4. ALTITUDE VARIATION OF DOMINANT VERTICAL SCALES

In this section we present wind velocity profiles in the stratosphere and mesosphere obtained by combining observations with the MU radar, rocketsondes and radiosondes in order to study the altitude variation of dominant vertical scales of gravity waves.

The trajectories projected on a horizontal plane are plotted, in Fig. 4, for the two successive rocket experiments on 9 September 1987 (MT135-47 and MT135-48). Straight lines from the origin in the south–eastward direction correspond to the ascent of the rocket, while the wave-like curve shows the movement of the sonde ejected from the rocket. The mean wind direction was generally westward, which agrees with the basic features of the general circulation in the summer middle atmosphere. The wave-like structure was very similar for the two trajectories, exhibiting fluctuations with an apparent horizontal scale of about 8 km, which, however, cannot be interpreted as the horizontal wavelength, because they were not observed at a fixed altitude, the rocketsondes

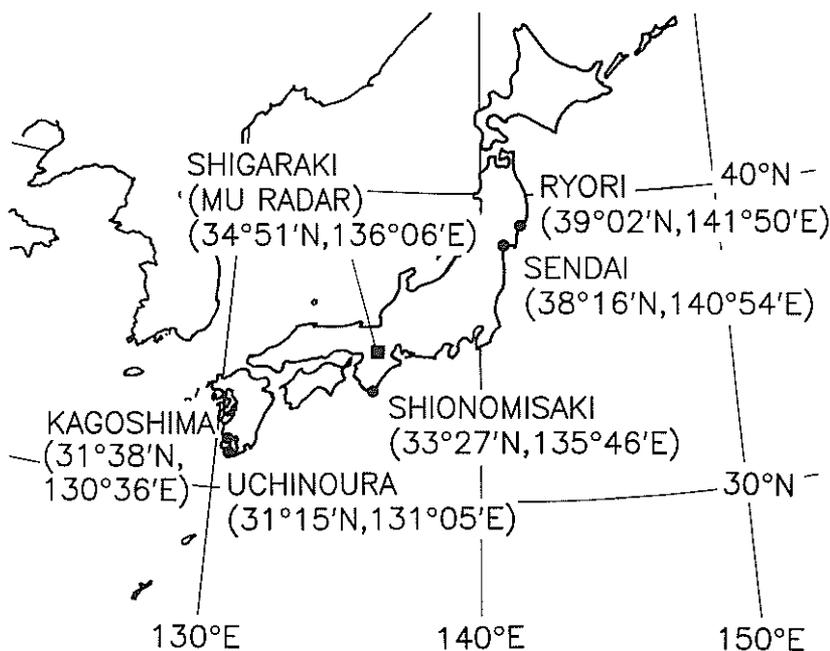


Fig. 3. Locations of observation stations used in this study (see the text).

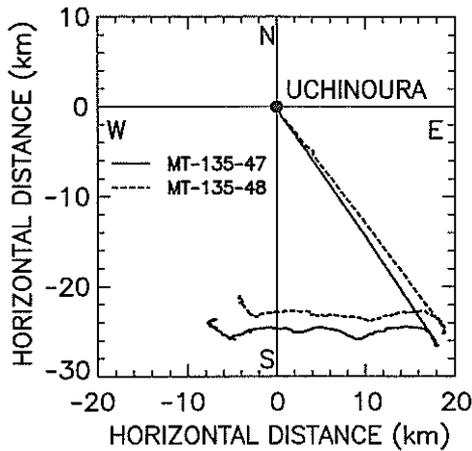


Fig. 4. The trajectories projected on a horizontal plane for a series of rocketsonde experiments (MT135-47 and 48) at the Kagoshima Space Center, Uchinoura, at 1100 and 1300 JST on 9 September 1987. A straight line shows the ascent of a rocketsonde, while a wave-like trajectory is detected during the descending motion of the rocketsonde ejected from the meteorological rocket.

descending from about 55 to 10 km during the horizontal movements. Trajectories in Fig. 4 show that a series of rocketsonde soundings were made in regions closely located with each other.

Figure 5 shows the vertical profiles of horizontal wind components detected by MT135-47 and 48, where estimation errors are indicated by a horizontal bar at altitudes higher than 45 km. Below 45 km the estimation error was not more than  $1 \text{ m s}^{-1}$  for the

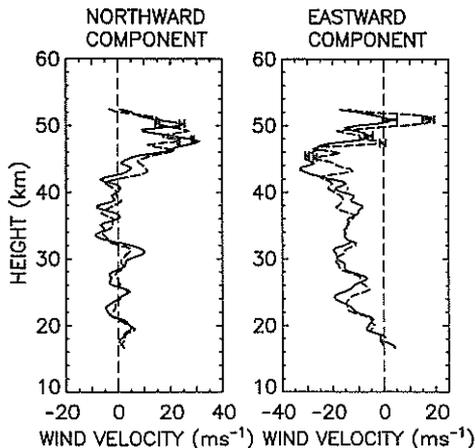


Fig. 5. Vertical profiles of northward (left) and eastward (right) wind velocity obtained with MT135-47 (solid line) and MT135-48 (dashed line) launched at 1100 and 1300 JST on 9 September 1987, respectively. The error bars are plotted at altitudes higher than 45 km.

eastward component, while it was smaller than  $0.3 \text{ m s}^{-1}$  for the northward component. The profiles demonstrate the similarity in vertical structures of the large-scale winds as well as wave-like structures with vertical scales of approximately 2 km at 17–22 km altitude, and of 4–5 km at 22–33 km for both the northward and eastward components. Since the spatial differences of those measurements were small as shown in Fig. 4, downward phase progressions of the wind motions can be interpreted as time evolution. Phase progressions are seen in almost all the height range, so that these wind fluctuations are recognized to be caused by systematic wind motions of atmospheric waves, possibly due to gravity waves.

Figure 6 shows a vertical profile of horizontal wind velocities observed on 11 September 1985 with MT135-44 rocketsondes at 14–56 km, with a routine rocketsonde at Ryori in the 20–57 km height region, with the MU radar in two separate regions, that is, at 5–25 and 60–90 km, and with routine radiosondes at Sendai and Kagoshima in the 0–25 km range. The estimation errors of the wind velocities were smaller than  $0.5$  and  $0.7 \text{ m s}^{-1}$  for the rocket sounding and the MU radar measurement, respectively, in the entire altitude ranges.

The zonal wind velocity profiles observed with the

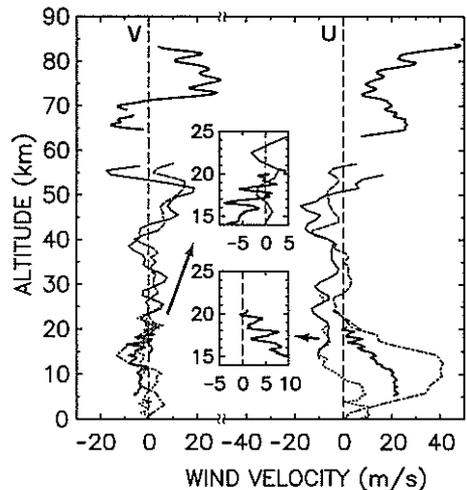


Fig. 6. Vertical profiles of northward (left) and eastward (right) wind velocity observed on 11 September 1985 with the MU radar (thick solid lines at 5–25 and 60–90 km), rocketsonde MT135-44 launched at 1125 JST (thin solid lines at 14–56 km), a rocketsonde launched from Ryori at 1101 JST (long dashed lines at 20–57 km), and radiosondes launched from Sendai (chained lines at 0–24 km) and Kagoshima (dotted lines at 0–26 km) at 0830 JST, respectively. The enlarged profiles are plotted for the MU radar and MT135-44 data in the height range of 14–25 km.

MU radar and radiosondes reflect the vertical and latitudinal structures of the jet stream in the troposphere and lower stratosphere. The jet stream has an eastward peak of  $40 \text{ m s}^{-1}$  centered at 12 km at Sendai ( $38^{\circ}16'N$ ), decreases in amplitude to about  $20 \text{ m s}^{-1}$  with a maximum at lower altitude over Shigaraki ( $34^{\circ}51'N$ ) and becomes as small as  $10 \text{ m s}^{-1}$  at Kagoshima ( $31^{\circ}38'N$ ), which indicates that the jet stream lay over the northern part of Japan in September.

Wind velocity perturbations in the troposphere seem to be suppressed due to a small  $N_b^2$ . In the lower stratosphere, the MU radar observations clearly show fluctuations in amplitudes and the vertical scale of about  $2\text{--}3 \text{ m s}^{-1}$  and 2 km, respectively, in both eastward and northward wind profiles. Note that the height resolution of the routine radiosondes is approximately 1 km, while that of the MU radar observations is 150 m in the lower atmosphere. Therefore, the fine structure of wind fluctuations detected with the MU radar is not correctly determined with radiosonde observations, the latter being biased toward the perturbations with vertical scales larger than the dominant components.

In the 17–35 km region the rocketsonde observations at Uchinoura shows predominance of a gravity wave with a vertical scale of approximately 5 km. There is clear correlation between the zonal and meridional components. The wave amplitude of the meridional wind increases from  $3 \text{ m s}^{-1}$  at 17 km to  $6 \text{ m s}^{-1}$  at around 30 km. The zonal component has a similar amplitude, but its growth is not obvious. The lowest part of the rocketsonde profile might also be affected by the poor height resolution. However, the dominant vertical scale of gravity waves at around 30 km altitude seems to be about 5 km.

At 35–55 km altitude, the rocketsonde observations show the overlapping of at least two waves with vertical scales of about 5 and 15 km, where the amplitude of the former is likely to be similar to that detected in the 17–35 km region, and for the latter it becomes as large as  $15 \text{ m s}^{-1}$ . The dominant vertical scale there seems to range from 5 to 15 km.

The wind velocity profile above 60 km observed with the MU radar shows a dominant wave with a vertical scale of approximately 15–20 km. The wave amplitude of the meridional component is about  $20 \text{ m s}^{-1}$ , which is larger than the zonal one ranging from 10 to  $12 \text{ m s}^{-1}$ . On the basis of a series of the MU radar observations, TSUDA *et al.* (1990) reported that the dominant wave is not repeated as a function of local time, so it is not due to atmospheric tides, but is likely a manifestation of upward propagating gravity waves. The mesospheric profile also consists

of smaller scale fluctuations with amplitudes ranging up to  $5 \text{ m s}^{-1}$ .

The wind velocity profile obtained by combining radar and rocket observations clearly shows that the vertical wavelength as well as the amplitude of the dominant gravity wave increases with altitude. The ratio of the dominant vertical scales between the mesospheric and lower stratospheric results was about 4–10, and the ratio of the amplitudes was about 5–10, which can be generally interpreted as the saturation of upward propagating gravity waves, as described in Section 2.

## 5. VERTICAL WAVENUMBER SPECTRA

Although dominant components of gravity waves are detected in each height region by using Fig. 6, the wind velocity profiles obviously include many waves with various vertical wavenumbers. In this section vertical wavenumber spectra of wind fluctuations are calculated from simultaneous MU radar and rocketsonde observations, and are compared with the model spectrum proposed by DEWAN and GOOD (1986) and SMITH *et al.* (1987).

Since the amplitudes of the model spectrum depend on the background values of  $N_b^2$ , we first determine  $N_b^2$  in each height region from the vertical derivative of temperature profiles observed with rocketsondes and radiosondes. However, simultaneous temperature measurements in the height range of the mesospheric MU radar observations were not performed, so we assumed  $N_b^2 = 3.35 \times 10^{-4}$  and  $3.45 \times 10^{-4} \text{ rad}^2 \text{ s}^{-2}$  in August and September at  $35^{\circ}N$ , respectively, on the basis of the CIRA 1986 model atmosphere.

Figure 7 shows both the temperature and  $N_b^2$  profiles observed with rocketsondes MT135-45 and 48 on 20 August 1986 and 9 September 1987, respectively, and those with radiosondes launched from Shionomisaki at 0830 JST on 20 August 1986 and 9 September 1987. The mean values of  $N_b^2$  on 20 August 1986 can be inferred to be  $6.81 \times 10^{-4}$ ,  $5.80 \times 10^{-4}$ ,  $5.43 \times 10^{-4}$  and  $4.58 \times 10^{-4} \text{ rad}^2 \text{ s}^{-2}$  in the height regions of 17–24, 20–30, 30–40 and 40–50 km, respectively. On 9 September 1987,  $N_b^2$  was  $6.25 \times 10^{-4}$ ,  $5.76 \times 10^{-4}$ ,  $5.34 \times 10^{-4}$  and  $4.96 \times 10^{-4} \text{ rad}^2 \text{ s}^{-2}$  in the respective four height ranges.

We present two sets of vertical wavenumber spectra for both zonal and meridional wind fluctuations in five ranges observed with the MU radar and rocketsondes, where the details of the spectral analysis are as described by TSUDA *et al.* (1989). Figure 8a shows spectra determined with rocketsonde MT135-45 and

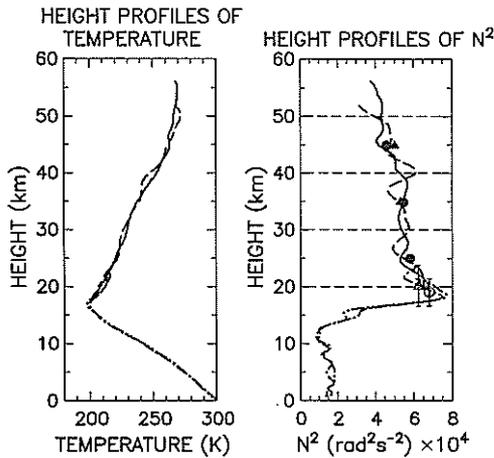


Fig. 7. The vertical profiles of temperature,  $T$  (K) (left), and Brunt-Väisälä frequency squared,  $N_b^2$  ( $\text{rad}^2 \text{s}^{-2}$ ) (right). Solid and dashed lines are the profiles at 17–55 km obtained from rocketsonde MT135-45 launched on 20 August 1986 and rocketsonde MT135-48 on 9 September 1987, respectively. The solid circles and triangles indicate the mean values of  $N_b^2$  at 20–30, 30–40 and 40–50 km altitudes for MT135-45 and MT135-48, respectively. The vertical bars with an open circle and triangle indicate  $N_b^2$  in the lower stratosphere determined from routine radiosonde soundings in the 0–21 and 0–24 km altitudes at Shionomisaki at 0830 JST on 20 August 1986 (chained) and 9 September 1987 (double-chained), respectively.

the MU radar at 1000–1200 JST on 20 August 1986, while Fig. 8b corresponds to the mean spectra of determinations with MT135-47 and 48, and the MU radar observations during 1000–1200 and 1200–1400 JST on 9 September 1987. The height ranges are (A) 17–24 km; (B) 20–30 km; (C) 30–40 km; (D) 40–50 km and (E) 65–85 km, where the MU radar observations give spectra in the lowest and highest altitude ranges. The straight lines in Fig. 8 are the corresponding model spectra  $F_i^*(m)$  calculated by using  $N_b^2$  in each height range plotted in Fig. 7. Note that in Fig. 8 vertical scales for spectra indicated as (A), (C) and (E) are shown on the left, while those for (B) and (D) are on the right, respectively. A 90% confidence interval for each spectrum is shown as a vertical bar in Fig. 8, which is calculated by using a method described by BLACKMAN and TUKEY (1958).

The spectra in region (A) are quite similar for the two observation periods; their logarithmic slopes were about  $-1.3$ , which are much smaller than the model prediction ( $-3$ ), furthermore, the observed spectral amplitudes were smaller than the model for  $m < 2 \times 10^{-3} \text{ m}^{-1}$  by a factor of roughly  $10^{-2}$ . On the other hand, the spectra in region (E) showed fairly good agreement with the model as to both slope and

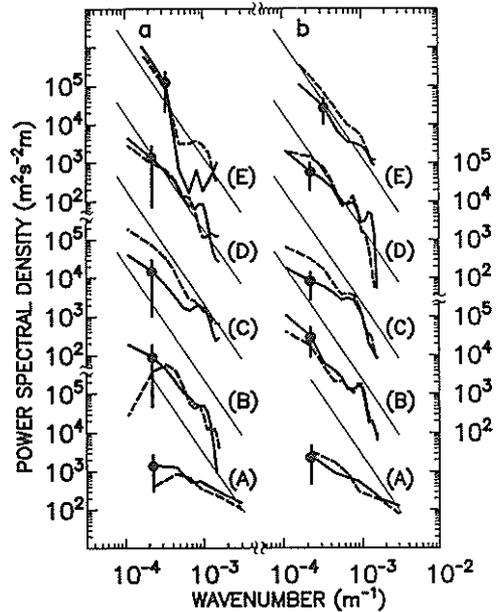


Fig. 8. Vertical wavenumber spectra of the meridional (dashed) and zonal (solid) wind fluctuations observed with the rocketsondes and the MU radar. The (a) left and (b) right spectra present the results obtained on 20 August 1986 with MT135-45, and on 9 September 1987 with MT135-47 and 48, respectively. The model spectrum is also plotted by using  $N_b^2$  shown in Fig. 7. Vertical axis for (A), (C) and (E) is plotted on the left, while that for (B) and (D) is shown on the right. A vertical bar indicates 90% confidence interval.

amplitudes, except for in the large wavenumber range of the zonal component in Fig. 8a.

So far we have shown the characteristics of gravity waves determined from the MU radar observations, that is, those in the mesosphere are well organized according to the saturated gravity wave theory, while in the lower stratosphere gravity waves are not fully saturated. The large change in the spectral shape between regions (A) and (E) can be interpolated from the rocketsonde results determined in the intermediate height ranges. In region (B), where the lower part of the height range overlaps with that in region (A), the spectral slopes are again more gradual than  $-3$  and the spectral amplitudes are considerably smaller than the model. The origin of spectral peaks near  $m = 1 \times 10^{-3} \text{ m}^{-1}$  in both Fig. 8a and b is not clearly understood, but they might be due to the poor height resolution of the rocketsonde sounding, because these enhancements are not indicated by the MU radar observations in region (A).

In region (C), the spectral slopes and amplitudes are still more gradual and smaller than the model, respectively, although the spectra approach the model

values. The entire shape of the spectra in region (*D*) becomes fairly close to the model curve; the slope is approximately  $-3$  and the amplitudes agree relatively well with the model except for the enhancement for  $m > 8 \times 10^{-4} \text{ m}^{-1}$ . It is noteworthy that the meridional component has larger amplitudes than the zonal one at a small wavenumber (large vertical scales).

The overall structures of these spectra are quite similar in each height region despite the difference in observation years, suggesting small year-to-year variation in the behavior of gravity waves. The spectral slope becomes steeper with a higher altitude, and reaches  $-3$ , which is predicted with the model at about 40–50 km, in the upper stratosphere.

The spectral amplitudes are generally smaller than the model in the lower stratosphere, although the discrepancy is relatively smaller at higher altitude regions, then the amplitudes become fairly close to the model values in the upper stratosphere and mesosphere. A bend in spectra at small wavenumbers, which is expected from the saturation theory in Fig. 2, seems to appear at low altitude regions; however, one is not recognized in the mesosphere within the observed wavenumber range.

The wind velocity variance,  $\overline{u'^2}$ , is calculated by integrating the spectra in the wavenumber range from  $m_1$  to  $m_2$  as

$$\overline{u'^2} = \int_{m_1}^{m_2} F_u(m) dm.$$

Note that the wavenumber ranges are not the same for five height regions, that is  $(m_1, m_2) = (2.3 \times 10^{-4} \text{ m}^{-1}, 3.2 \times 10^{-3} \text{ m}^{-1})$  and corresponding scales 4.4–0.3 km for the region (*A*) ( $1.1 \times 10^{-4} \text{ m}^{-1}, 1.6 \times 10^{-3} \text{ m}^{-1}$ ) and 9.0–0.6 km for (*B*), (*C*) and (*D*), and ( $1.7 \times 10^{-4} \text{ m}^{-1}, 1.5 \times 10^{-3} \text{ m}^{-1}$ ) and 6.0–0.7 km for (*E*). Now we define in the following the normalized variance,  $\alpha$ , as a ratio of the observed variance to a theoretical prediction assuming that gravity wave components for  $m_1 \leq m \leq m_2$  are saturated:

$$\alpha = \frac{\int_{m_1}^{m_2} F_u(m) dm}{\int_{m_1}^{m_2} F_u^s(m) dm} = \frac{\overline{u'^2}_{(\text{observation})}}{\overline{u'^2}_{(\text{theory})}}.$$

The mean value of  $\alpha$  between the meridional and zonal determinations is plotted in Fig. 9. When  $\alpha$  in each height region is close to unity, wind fluctuations can well be described by a saturated gravity wave theory.

$\alpha$  in region (*A*) was about  $1.5 \times 10^{-2}$  and  $2.5 \times 10^{-2}$  on 20 August 1986 and 9 September 1987, respectively. In region (*C*)  $\alpha$  ranged from 0.05 to 0.1.  $\alpha$  in region (*B*), which partially overlaps (*A*), is generally

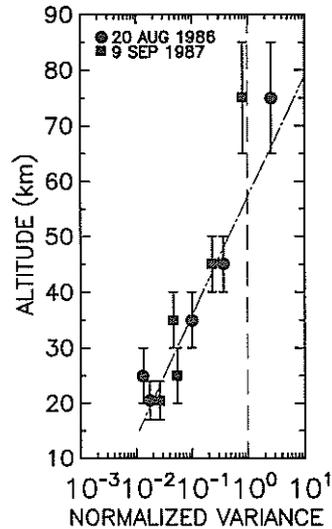


Fig. 9. Height variation of the normalized variance (see the text). The circles and squares denote the observations on 20 August 1986 and 9 September 1987, respectively, the vertical bars indicating the height regions. The chained line approximates the exponential growth of the normalized variance in the stratosphere.

in-between those in regions (*A*) and (*C*),  $\alpha$  increases to 0.2–0.4 in region (*D*), clearly showing the asymptotical growth of  $\alpha$ . While in region (*E*),  $\alpha$  on 9 September 1987 is close to unity, that on 20 August 1986 slightly exceeds unity. The overall height profile in Fig. 9 indicates the clear tendency that  $\alpha$  increases with altitude, and asymptotically approaches unity.

The normalized variance due to gravity waves in the stratosphere increase their amplitudes propagating upward and are hence considered not to be fully saturated. Their growths on 20 August 1986 and 9 September 1987 can be approximated by the chained line in Fig. 9 in the height region of 17–50 km. Assuming that the Brunt–Väisälä frequency is constant in the stratosphere, the  $e$ -folding height  $H_e$  of the exponential growth for the wind velocity variance with vertical scales smaller than about 10 km can be inferred as about 9 km. The values of  $\alpha$  in the mesosphere were close to unity, suggesting that the mesospheric wind fluctuations are well described by the saturated gravity wave model. It is clear that  $\alpha$  in the mesosphere was significantly smaller than the linearly extrapolated value based on the stratospheric results, implying that the gravity waves cease increase of wave amplitudes in the mesosphere, due to wave saturation.

## 6. CONCLUDING REMARKS

In this paper we have presented a case study on altitude variations of gravity wave characteristics in

the middle atmosphere as judged from observations with the MU radar and rocketsondes.

The wind velocity profiles observed on 11 September 1985 showed that the dominant vertical scales of gravity waves are 2–5, 5–15 and greater than 15 km in the lower and upper stratosphere, and mesosphere, respectively, which basically agree well with the predictions with the linear saturation theory. However, quantitative comparisons showed some discrepancies between the observations and the model, which might be explained by the fact that the gravity waves were not fully saturated in the lower stratosphere.

Vertical wavenumber spectra of wind velocity fluctuations observed on 20 August 1986 and 9 September 1987 were determined in five height regions of the middle atmosphere, and compared with the model. The spectra in the lower stratosphere had more gradual slopes than the model with considerably small spectral amplitudes. In the upper stratosphere and mesosphere, the model fairly well explains the observed characteristics.

The wind velocity variance normalized as to the model,  $\alpha$ , increases with altitude varying from 1/50 in the lower stratosphere to 1/5 in the upper stratosphere. The observed mesospheric variances agreed

well with the value predicted by the model and were smaller than the extrapolated value by using the exponential growth of the stratospheric variances. It can be thus suggested that gravity waves are not fully saturated in the lower stratosphere, while they are saturated in the mesosphere.

Long-term radiosonde observations have clarified that the activity of gravity waves with vertical scales of less than about 1 km in the lower stratosphere (15–30 km) shows a seasonal variation, such that they are well organized with the saturated gravity wave model in winter, while they are not necessarily saturated in summer (TSUDA *et al.*, 1991), which is basically consistent with the results described in this paper. We need to extend the seasonal range of the simultaneous observations using the MU radar and rocketsondes in order to elucidate the entire behavior of gravity waves in the middle atmosphere.

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