Observations of Tropospheric Temperature Fluctuations with the MU Radar–RASS

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ABSTRACT

Applying the RASS (radio acoustic sounding system) technique to the MU (middle and upper atmosphere) radar, profiles of both temperature and wind velocity were observed every 90 s in the height range of about 1.5–7.0 km, with a height resolution of 300 m, for about 40 h on 6–8 August 1990. The temperature profiles obtained with RASS agreed well with the virtual temperature derived from radiosonde sounding, where the mean difference between the temperature values was approximately 0.3°C. The observed frequency spectra above about 2.5-km altitude, having an asymptotic slope of −1 and approximately 0 for temperature and vertical wind velocity fluctuations, respectively, were reasonably consistent with a model spectrum of gravity waves. But, below 2.5 km, low-frequency components were conspicuously enhanced, especially for vertical wind velocity, presumably affected by convection. Wavelike temperature fluctuations with a dominant period of 6–8 h clearly showed downward phase progression and a π/2 phase lag between temperature and vertical wind velocity. In addition, short-period components were also recognizable for both temperature and vertical wind velocity fluctuations. However, for low-frequency components, which were sometimes enhanced at the lowest altitudes of the observation range, the time variations of temperature and vertical wind velocity were in phase. The covariance between temperature and vertical wind velocity was also determined, and heat flux profiles were further estimated. Although a major part of the fluctuations above 2.5 km could be explained by gravity waves, those below 2.5-km altitude seemed to be due to effects of convective motions in the planetary boundary layer.

1. Introduction

A radio acoustic sounding system (RASS) has been developed as a ground-based remote-sensing technique for measurements of atmospheric temperature (e.g., Marshall et al. 1972). RASS utilizes the physical principle that sound speed \( c_s \) is related to atmospheric temperature \( T \) as follows,

\[
T = \left( \frac{c_s}{k_b} \right)^2.
\]

(1)

The constant \( k_b \) is defined as

\[
k_b = \left( \frac{\gamma R}{M} \right)^{1/2},
\]

(2)

where \( \gamma, R, \) and \( M \) are the specific heat ratio, gas constant, and mean molecular weight of the atmosphere, respectively. The apparent sound speed observed with a Doppler radar corresponds to the sum of the true sound speed \( c_s \) and the radial movement of the ambient atmosphere; therefore, the apparent velocity must be compensated for the radial wind velocity, which should be independently determined, before calculating temperature by using the relation (1).

The value of \( k_b \) becomes 20.046 for a dry atmosphere, but it slightly varies depending on the humidity, that is, when water vapor is included in the atmosphere, the mean molecular weight is slightly reduced from the value for dry air, so \( k_b \) becomes larger. However, since \( \gamma \) is also a weak function of the humidity, the actual dependence of \( k_b \) on the humidity cannot be simply expressed. Note that the virtual temperature can be obtained through a RASS experiment when the value for a dry atmosphere is used in Eq. (1).

To receive an intense RASS echo the Bragg condition should be satisfied between the scale of refractive-index fluctuation produced by acoustic waves and the radar wavelength. Since the acoustic wavelength varies depending on the background temperature and radial wind velocity, the transmitted acoustic signal must be designed to consist of various frequency components.
with an appropriate frequency distribution so as to satisfy the Bragg condition in a large-altitude range (e.g., Adachi et al. 1993).

It was reported that the height range of RASS measurements can be limited by the geometrical configuration between the antenna beam of a radar and the shape of acoustic wave fronts (Masuda 1988). That is, in the case of a monostatic radar the antenna beam must be steered into directions that are normal to acoustic wave fronts, for the RASS echoes to be reflected toward the radar.

RASS observations were limited to the planetary boundary layer (e.g., Marshall et al. 1972; Bonino et al. 1981; Peters et al. 1985) until Matsuura et al. (1986) successfully measured temperature profiles in the troposphere and lower stratosphere by applying RASS to the MU (middle and upper atmosphere) radar, operated at 46.5 MHz, in Shigaraki, Japan (34°51’N, 136°06’E, 370 m above sea level). RASS experiments with the MU radar (hereafter referred to as the MU radar–RASS) are characterized by coverage of a large altitude range, owing to the fact that the attenuation of acoustic waves due to turbulence is negligible in this acoustic frequency range (about 100 Hz). Furthermore, the antenna steerability of the MU radar makes it easier to track the specular reflection region of RASS echoes. RASS observations have now been extensively conducted with an MST (mesosphere–stratosphere–troposphere) radar or a wind profiler for simultaneous profiling of both wind velocity and temperature in the lower atmosphere (e.g., Tsuda et al. 1989; May et al. 1990; Ecklund et al. 1990).

We study in this paper fluctuations of both temperature and vertical wind velocity observed with the MU radar–RASS at 1.5–7.0 km in the troposphere on 6–8 August 1990. In section 2 we describe the MU radar–RASS equipment as well as the observation procedure in detail. Section 3 is devoted to the comparison of RASS temperature profiles with radiosonde results in order to determine the accuracy of the RASS measurements. We illustrate in section 4 characteristics of the fluctuations by analyzing frequency spectra, the cross-correlation function, and heat flux profiles, and further discuss interpretation of the observed behavior in terms of gravity waves and convection.

2. Experimental setup

The RASS hardware used in this study consists of a high-power acoustic transmitter and the MU radar, which is a pulse Doppler radar operated at 46.5 MHz with a peak and average transmitting power of 1 MW and 50 kW, respectively. A general description of the hardware system as well as a fundamental observation procedure were reported previously by Tsuda et al. (1989). We have recently employed a woofer speaker as an acoustic transmitter, which is commercially available (SW-46W-UL2 produced by TOA Co., Ltd.). Because the speaker should be protected from rain, it was installed in a small box about 50 cm above the ground with the speaker face down. Therefore, acoustic waves are transmitted downward, reflected off the ground and then propagated upward. The sound pressure level of an acoustic pulse is 5 W m⁻² at 1 m above the speaker. We can simultaneously operate several speakers, but during the experiment described in this paper we used only one speaker located near the center of the MU radar antenna.

Now we describe the procedure involving the MU radar–RASS experiment in detail. During a single measurement of a Doppler spectrum of RASS echoes, a series of (eight or nine) short acoustic pulses were successively transmitted by changing the frequency, as described in Table 1, in order to roughly satisfy the Bragg condition at different heights (Tsuda et al. 1989; Adachi et al. 1993). Each acoustic pulse consisted of 40 acoustic wave cycles corresponding to a pulse duration of about 0.4 s. The interval between successive pulse transmissions was 3 s, which was large enough so that two RASS echoes with different frequencies would not be received in a single radar sample volume. The MU radar observations of RASS echoes were delayed by several to a few tens of seconds after the transmission of acoustic pulses—that is, until they reached the observation altitudes (Tsuda et al. 1989).

In advance of the RASS experiment, we computed the propagation characteristics of acoustic wave fronts by means of a two-dimensional ray tracing in order to predict the location of echoing regions (Masuda 1988). Since the mean winds were fairly small and their directions were relatively stable during the experiment, as described in a later section, the effective backscatter region was expected to appear in the vertical direction. Therefore, we steered the antenna beam of the MU radar into the vertical direction. In addition, three oblique beams were used to observe horizontal wind velocity, as summarized in Table 2. Although temperature profiles were determined in all the four beam directions, we mainly use, in this paper, the results ob-

<table>
<thead>
<tr>
<th>Observation period</th>
<th>Number of acoustic pulses</th>
<th>Minimum frequency (Hz)</th>
<th>Maximum frequency (Hz)</th>
<th>Step of frequency shift (Hz)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1942 LT 6 August–1215 LT 7 August</td>
<td>8</td>
<td>99.4</td>
<td>107.1</td>
<td>1.1</td>
</tr>
<tr>
<td>1218 LT 7 August–1829 LT 7 August</td>
<td>8</td>
<td>99.0</td>
<td>106.7</td>
<td>1.1</td>
</tr>
<tr>
<td>1930 LT 7 August–1154 LT 8 August</td>
<td>9</td>
<td>97.2</td>
<td>107.6</td>
<td>1.3</td>
</tr>
</tbody>
</table>
Table 2. Azimuth and zenith angles of the three oblique antenna beams. Azimuth angle is measured clockwise from the north.

<table>
<thead>
<tr>
<th>Observation period</th>
<th>Beam 1</th>
<th>Beam 2</th>
<th>Beam 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>1942 LT 6 August-0852 LT 7 August</td>
<td>210°, 10°</td>
<td>300°, 10°</td>
<td>300°, 5°</td>
</tr>
<tr>
<td>1018 LT 7 August-1829 LT 7 August</td>
<td>270°, 10°</td>
<td>0°, 10°</td>
<td>20°, 10°</td>
</tr>
<tr>
<td>1930 LT 7 August-2054 LT 7 August</td>
<td>95°, 10°</td>
<td>5°, 10°</td>
<td>5°, 15°</td>
</tr>
<tr>
<td>2100 LT 7 August-2345 LT 7 August</td>
<td>95°, 10°</td>
<td>20°, 15°</td>
<td>5°, 15°</td>
</tr>
<tr>
<td>0009 LT 8 August-0054 LT 8 August</td>
<td>95°, 10°</td>
<td>5°, 10°</td>
<td>5°, 15°</td>
</tr>
<tr>
<td>0100 LT 8 August-0109 LT 8 August</td>
<td>95°, 10°</td>
<td>20°, 15°</td>
<td>5°, 15°</td>
</tr>
<tr>
<td>0112 LT 8 August-0154 LT 8 August</td>
<td>95°, 10°</td>
<td>5°, 10°</td>
<td>5°, 15°</td>
</tr>
<tr>
<td>0200 LT 8 August-1154 LT 8 August</td>
<td>95°, 10°</td>
<td>5°, 10°</td>
<td>20°, 10°</td>
</tr>
</tbody>
</table>

The width of a transmitted radar pulse was 2 μs without pulse compression and the interpulse period (IPP) for radar pulse transmissions was 400 μs. The RASS echoes were sampled at 32 range gates starting from 1.42-km altitude above sea level, with a range resolution of 300 m. Time series were constructed over 1024 data points, which were further transformed into a frequency spectrum by means of a fast Fourier transform (FFT), where the number of FFT points was made as large as possible to obtain good frequency resolution.

Since the expected maximum frequency shift due to the sound velocity in RASS echo is as large as about 320 m s⁻¹ (about 102 Hz), the data sample interval must be as short as about 5 ms. Thus, the number of coherent integrations during this experiment was limited to two in each antenna direction, which gives an effective sample rate of 3.2 ms, corresponding to a maximum Doppler frequency shift of 156.25 Hz (or maximum sound speed of 503.9 m s⁻¹).

It took 3.28 s to obtain a single-frequency spectrum of RASS echoes. In between successive RASS measurements, which were repeated every 90 s, the radial wind velocity was measured for about 40 s using the normal MST radar technique with 24 coherent integrations, and 1024 FFT points.

We adopted the moment method to estimate the mean Doppler frequency shift of spectra for both RASS and MST echoes, which can be converted into the apparent sound speed and radial wind velocity, respectively. By linearly interpolating the wind fields detected before and after each RASS measurement, we estimated the radial wind velocity in the antenna beam direction for the RASS measurements. The true sound speed was determined by subtracting the radial wind velocity from the apparent sound speed, and was subsequently transformed into the atmospheric temperature. Note that Angevine et al. (1994) proposed a simultaneous measurement of both acoustic and radial wind velocities in order to improve estimation of the true sound speed. However, we did not find a significant error in the temperature and heat flux determination by interpolating the vertical wind measurements, which is reported in section 4e.

Profiles of both temperature and wind velocity were observed for about 40 h continuously, from 1942 LT 6 August to 1155 LT 8 August 1990. During the RASS experiment, six conventional radiosondes were launched to obtain reference temperature profiles at 2026 LT 6 August at 0202, 0841, 1403, and 2020 LT on 7 August, and at 0208 LT 8 August.

3. Comparison between RASS and radiosonde measurements

We first investigate the accuracy of the RASS temperature profiles by comparing them with the results measured with radiosondes. Figure 1 shows a mean temperature profile, averaged for 40 h, for the RASS observations on 6–8 August 1990. Average temperature profiles of the six radiosonde soundings are also plotted in Fig. 1, where solid and dashed lines correspond to the atmospheric temperature and virtual temperature, respectively. A profile of relative humidity is also plotted in Fig. 1, which indicates that the humidity was fairly high below 1.5 km in altitude, decreased linearly from about 70% at 1.5 km to 35% at 2 km, and then gradually decreased to about 10% at 6 km except for an enhancement near 3 km. The virtual temperature was generally higher than the atmospheric temperature, there being a difference of up to 2°C at the lowest altitudes, but they became almost the same above about 4 km, where the humidity is less than about 20%.

The profile observed with the MU radar–RASS agreed fairly well with the virtual temperature in the entire height range. The difference in the determined temperatures between RASS and radiosondes shown in Fig. 1 was less than about 0.2°C, although a relatively large discrepancy of up to 0.5°C was occasionally detected, mainly in the bottom and top regions of the observed height range. Since the time variations of temperature fluctuations were great below 2-km altitude, as described in later sections, the difference between the two sets of profiles seemed to be due to the sparse sampling of the radiosonde soundings.

To investigate the agreement of instantaneous temperature profiles, we compare in Fig. 2 individual radiosonde measurements with the RASS temperature profiles, where the time interval for averaging RASS profiles was limited to about 2 h. The radiosonde pro-
Fig. 1. Comparison of mean temperature profiles obtained with RASS and radiosondes. Solid and dashed lines in the left panel indicate the atmospheric and virtual temperature profiles observed with radiosondes, while the crosses indicate the RASS determination. The discrepancy between temperature determinations with RASS and radiosondes is plotted in the center panel, where a horizontal bar corresponds to the standard deviation within the entire observation period of 40 h. The mean humidity profile observed with radiosondes is shown in the right panel.

Profiles suggest that variations of temperature profiles between day and night were the largest below about 2 km, indicating the effects of the diurnal temperature variations near the ground. The deviations were suppressed near 2.5-km altitude. The temperature gradient in the region above 2 km was approximately 6°C km⁻¹. The overall structures of the observed profiles were quite similar to each other between the radiosonde and RASS measurements, although fine vertical structures were smoothed out for the RASS observations because of the 300-m resolution.

Profiles of the standard deviation of the RASS determinations are also plotted in Fig. 2, which seemed to consist of both measurement error and temperature fluctuations with time scales shorter than 2 h. The standard deviation values ranged from 0.1° to 0.4°C, the mean value being approximately 0.2°C. Except for the results collected at 1945–2143 LT, the standard

Fig. 2. Comparison of individual radiosonde temperature profiles with the RASS determinations, averaged over about 2 h centered near the launch times of the corresponding radiosondes (left). Note that the RASS profiles are shifted toward right by 5°C. The standard deviation of temperature fluctuations within 2 h for the six observation periods are also shown (right).
deviation values were small near 2.5-km altitude, where the temperature profiles themselves also showed good agreement among six profiles, which indicates that the temperature perturbations were suppressed there. The standard deviation was relatively large below 2.5 km, reflecting large time variations of temperature fields in the planetary boundary layer caused by the surface temperature variations. Above about 5.5 km the standard deviation values generally increased with altitudes, although rather gradually.

In Fig. 3, the difference in the temperature values between radiosonde and RASS measurements is presented for each pair of profiles shown in Fig. 2, and the standard deviation for the RASS results is also indicated by a horizontal bar. No systematic bias was recognized in the profiles of the differences. The best agreement was detected for the fourth and sixth profiles, where almost all of the discrepancy was within the standard deviation of the RASS measurements, while the largest difference was found for the first profile, which showed the maximum deviation of up to 1°C. For the other three cases the difference was normally within 0.5°C. The mean values of the discrepancy were 0.5°, 0.3°, 0.3°, 0.2°, 0.35°, and 0.2°C for the profiles shown in Fig. 3. The RASS profiles after averaging for 2 h showed a difference of approximately 0.3°C, in comparison with the simultaneous radiosonde measurements. Since a part of the disagreement is due to the horizontal variations of temperature fields, the actual measurement accuracy of the RASS profiles seems to be better than 0.3°C.

4. Temperature fluctuations

We now present the characteristics of the temperature and vertical wind velocity fluctuations in the troposphere revealed by means of the MU radar–RASS observations conducted on 6–8 August 1990. Investigation of a surface weather map showed that two weak low pressure centers existed east and south of the Japanese mainland, but the meteorological conditions were fairly calm during the experiment. The observation period coincided with the hottest period in summer 1990, but the surface temperature exceeded 35°C during the daytime.

Before describing the fluctuating components, we present in Fig. 4 the mean profiles of wind velocity for the eastward, northward, and vertical components derived from the MU radar observations made during the RASS experiment. Figure 4 also shows the profile of the standard deviation of each wind velocity component during the observation period of 40 h. Note that the values are magnified by ten times for the vertical component.

The direction of the horizontal wind velocity was south-southeastward below about 2.5 km and rotated toward the south above 3.5 km. The amplitudes of the horizontal winds were 3 m s⁻¹ near the bottom of the observation height range, decreased to 1 m s⁻¹ at 3–3.5 km and then increased to 5 m s⁻¹ at 5–6-km altitude. The vertical wind velocity was generally downward in the entire height range, although the amplitudes were as small as −0.01 m s⁻¹ except for the relatively large value of −0.04 m s⁻¹ below 2 km.

The standard deviation of the horizontal wind velocity was fairly large in the bottom region and it became minimum at 2.5–3.5 km, which coincided with the region of the small temperature perturbations depicted in Fig. 3. Then it increased rapidly above 3.5- and 2.5-km altitudes for zonal and meridional components, respectively, the profiles being almost the same above 5 km for both components. The standard deviation of the vertical wind velocity was about 0.2 m s⁻¹ at the lowest altitude, decreased to 0.1 m s⁻¹ near 2.5 km, and became fairly constant in the upper height region.

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**Fig. 3.** Differences between temperature determinations with RASS and radiosondes derived from the profiles in Fig. 2, where the standard deviation for RASS is also indicated by the horizontal bar.

**Fig. 4.** Profiles of mean wind velocity and standard deviation for the eastward (solid line), northward (dotted line), and vertical (dotted-dash line) components observed with the MU radar during the RASS experiment on 6–8 August 1990. Note, however, that the vertical wind velocity and its standard deviation value were magnified by ten times.
It can be suggested that the behavior of the wind velocity fluctuations differed between the regions below and above 2.5-km altitude. Perturbations accompanied by large vertical motions probably caused by convection seemed to appear in the lower region, while fluctuations caused by upward-propagating gravity waves occurred in the upper height range.

\[a. \text{Time-height section of temperature fluctuations}\]

Figure 5 shows time variations of the fluctuating components of both temperature, \(T'\), and vertical wind velocity, \(w'\), at each altitude, where the observed values were averaged over three consecutive measurements, resulting in a time resolution of 4.5 min. The behavior

\[\text{FIG. 5. Time variations of temperature (top) and vertical wind velocity (bottom) fluctuations determined with the MU radar-RASS on 6-8 August 1990. Thin horizontal line indicates the altitude, scaled on the left side, and also corresponds to the zero values at each altitude. The scale on the right side shows deviation from the 40-h mean.}\]
of $T'$ revealed the superposition of many waves with different frequencies, producing a complicated structure. However, it can be recognized that relatively long period components with periods ranging from 6 to 10 h were dominant, showing downward phase progressions. The amplitudes of such long-period components sometimes exceeded 1°C, especially at the lowest several altitudes, for instance, in the first one-third of the observation period. Moreover, short-period fluctuations were frequently superimposed on the time series of $T'$, although their amplitudes were relatively smaller than those of the long-period components.

On the other hand, fluctuations of $w'$ mainly consisted of short-period components. Bursts of such short-period fluctuations continued for 30 min to several hours, and their phases were clearly aligned to the vertical direction for a height range that sometimes encompassed to the entire observation region.

It is noteworthy that large perturbations with periods shorter than about 30 min were conspicuously enhanced for both $T'$ and $w'$, their amplitudes being up to 1°C and 0.5 m s$^{-1}$, respectively, at 1.5–2.0-km altitude for about 5 h from 1100 to 1600 LT 7 August. Corresponding fluctuations were also clearly recognized for the zonal and meridional wind velocity components with amplitudes of up to 3 m s$^{-1}$, although the results are not illustrated here.

b. Frequency spectra

Using the time series of $T'$ and $w'$ shown in Fig. 5, we analyzed the frequency spectra, $F_T(\omega)$ and $F_w(\omega)$, at nine altitudes, as shown in Fig. 6, in which the results are plotted at every 300 m below 3 km. Since the time series was processed with a moving average over three points, the resultant time resolution is 4.5 min (or $3.7 \times 10^{-3}$ Hz in frequency), and therefore the spectral amplitudes in Fig. 6 are reduced for $\omega$ larger than approximately $2 \times 10^{-3}$ Hz.

First, we investigate the behavior of $F_w(\omega)$ shown in Fig. 6b, which showed a clear change in spectral shape with altitudes. In the lowest three height ranges, from 1.42 to 2.02 km, the spectral slope was approximately $-2/3$. The slope decreased at 2.32 km and was
almost zero at 2.62 km. Furthermore, the spectral slope was basically zero at higher altitudes, being consistent with other radar observations under weak mean wind conditions (e.g., Ecklund et al. 1986). Therefore, the characteristics of $w'$ fluctuations were quite different between the two height regions of below 2.32 km and above 2.62 km, such that the relatively low frequency components ($\omega \leq 10^{-4}$ Hz) were enhanced in the lower region, while they were reduced at the higher altitudes.

A model spectrum based on the gravity wave theory (e.g., VanZandt 1982) agreed reasonably well with the observed spectra above 2.62 km. However, this model is not applicable to an interpretation of the results in the lower region. By taking into account the Doppler shift effects, a negative slope of $F_\omega(\omega)$ can theoretically be predicted (Fritts and VanZandt 1987), but these effects seemed to be negligible because of the weak mean winds as shown in Fig. 4. To interpret the observed behavior of $F_\omega(\omega)$ in the lower region, it is necessary to include dynamical processes other than gravity waves, which should have a red spectrum (an approximate slope of $-2/3$) for $\omega \leq 10^{-4}$ Hz. Such motions can most probably be explained by convection that is actively generated within and near the planetary boundary layer, but its direct effect seemed to be confined to below about 2.5 km during this experiment.

The changes in spectral shapes detected for $F_\omega(\omega)$ were not entirely clear for $F_T(\omega)$ plotted in Fig. 6a, showing a spectral slope of approximately $-5/3$ at all altitudes. However, if the observed $F_T(\omega)$ is compared with the reference line with a model slope of $-5/3$ plotted in Fig. 6a, height variations of the spectral amplitudes can be recognized. That is, below 2.32 km the amplitudes were close to or slightly exceeding the reference line, but they became smaller than that at 2.62 and 2.92 km, where the decreases seemed to be more significant for $\omega \leq 7 \times 10^{-5}$ Hz. Note that components for $\omega \geq 4 \times 10^{-4}$ Hz were also enhanced at the 1.42-, 1.72-, and 2.02-km altitudes.

The slope of $F_T(\omega)$ can be approximated as $-5/3$ at 3.82 and 5.02 km, and it became slightly steeper at 6.52 km. The model spectrum of gravity waves predicts that $F_T(\omega)$ is proportional to a spectrum for the vertical displacement, which has the same shape as the hori-
horizontal wind velocity, having a slope of $-5/3$ to $-2$ (e.g., Fritts et al. 1990). Therefore, the observed spectral slope of $F_T(\omega)$ was again consistent with the gravity-wave model.

Detailed comparisons show that spectral spikes superimposed on the asymptotic slope coincided fairly well between $F_T(\omega)$ and $F_w(\omega)$, indicating that the activity was closely associated with these variables. It is also interesting to note that at 1.42-, 1.72-, and 2.02-km altitude the general shape as well as the detailed structure was quite similar between $F_T(\omega)$ and $F_w(\omega)$ for $\omega \approx 2 \times 10^{-4}$ Hz.

To summarize, both $F_T(\omega)$ and $F_w(\omega)$ observed above 2.62 km agreed fairly well with a model spectrum predicted on the basis of the gravity-wave theory, that is, their spectral slopes were estimated to be about

Fig. 7. Contour plot of time–height sections of temperature (top) and vertical wind velocity (bottom) fluctuations after processing with a low-pass filter with a cut-off period of 1 h. The scale on the right side indicates deviation from the 40-h mean.
\(-\sqrt{3}\) and nearly 0, respectively. The spectra below 2.32 km exhibited enhancements of low-frequency components that were not consistent with the gravity wave characteristics. The enhancements seemed to be caused by other mechanisms, such as convection that is generated within the planetary boundary layer.

c. Long-period components

To extract long-period components of fluctuations, we applied a low-pass filter with a cutoff period of 1 h to the time series shown in Fig. 5; the results are presented in Fig. 7 as contour plots for \(T'\) and \(w'\) fluctuations.

From Fig. 7 it can be clearly recognized that \(T'\) variations showed wavelike structures with a downward phase progression, their vertical scales being about 1.5 km. Disturbances of \(w'\) shown in Fig. 7 had fairly large amplitudes below 2.5 km in the observation period from 1100 LT 7 August to 0600 LT 8 August, being consistent with the discussion concerning \(F_{w'(\omega)}\) in Fig. 6. Associated large \(T'\) fluctuations can also be seen in the same time-height region. Furthermore, the horizontal wind velocity fluctuations also included large perturbations, which was the main cause of their large standard deviation values below 2.5 km in Fig. 4.

We now investigate in more detail the behavior of fluctuations during two periods: (i) from 1945 LT 6 August to 0739 LT 7 August, and (ii) from 1032 to 1826 LT 7 August. A wavelike structure is recognized during the first period, and disturbance at low altitudes is dominant in the latter.

During period (i), \(T'\) in Fig. 7 shows a wave with downward phase progression whose wave period can be estimated from Fig. 5 to be about 8 h near 2-km altitude, which became shorter, 6–7 h, at higher altitudes. The amplitude of the wave was about 1°C at 1.72-km altitude and became slightly smaller at higher altitudes.

Figure 8 shows normalized cross-correlation function (CCF) between \(T'\) and \(w'\) at several altitudes, where the positive lag indicates that the phase of \(T'\) was delayed compared to that of \(w'\). Although the wave activity was not clearly seen for \(w'\), CCF plotted in Fig. 8 showed a component with a period of about 8 h. Moreover, CCF during period (i) showed a broad peak centered at \(-3\) and \(-2\) h at 1.42 and 1.72 km, respectively, indicating that the phase of \(T'\) preceded that of \(w'\) by about \(\pi/2\) for the 8-h wave component, which can be explained by a dispersion relation of a linear gravity wave theory. At 3.82 km, the CCF showed similar variation to those in the lower region, although the maximum amplitude was small and the wave period became as short as 6 h.

On the other hand, the CCF in Fig. 8, corresponding to observation period (ii), indicates that \(T'\) and \(w'\) perturbations were in phase at 1.72- and 2.02-km altitude.

This phenomenon can be explained if a convective motion uplifted the warm air near the ground to higher altitudes through vertical motion. The CCF, however, became rather similar to the result in period (i) at 4.12 km, giving the phase delay for \(w'\) relative to \(T'\).

d. Short-period fluctuations

In Fig. 5, we observed that \(T'\) fluctuations with periods of 10–20 min were superimposed on the long-period components. Figure 9 shows short-period fluctuations of \(T'\) and \(w'\), after applying a high-pass filter having a cutoff at 30 min, for observations from 0000 to 0858 LT 7 August 1990.

Quite large variations of the short-period fluctuations can be recognized for both \(T'\) and \(w'\). During the first one-third of the observation interval, \(w'\) showed fairly sinusoidal fluctuations with a period of about 10 min, being close to the Brunt–Väisälä frequency, and the
correlations among the three altitudes were quite good. In the rest of the observation period there were also isolated events—for instance, at 0450 or around 0630 LT. All these $w'$ fluctuations were in phase in the vertical direction, which is consistent with the characteristics of short-period gravity waves.

The activity of $T'$ fluctuations generally resembled that of $w'$, that is, corresponding $T'$ oscillations with similar wave periods were seen during the three periods mentioned above for $w'$ perturbations. However, the coherence of the $T'$ oscillations between different altitudes was less significant. Detailed comparison shows that $T'$ and $w'$ were not necessarily in phase, but $T'$ often preceded $w'$ by about one-fourth of the wave cycle, which can most easily be recognized for the enhanced activity centered at 0630 LT.

Since the temperature was delineated on the basis of the sound speed measurements with the vertical beam, fluctuations of $w'$ could be reflected on $T'$ variations, being in phase (or out of phase), if the compensation of the radial wind velocity for the apparent sound velocity were not appropriate. It is noteworthy, however, that this error was not significantly seen in our results, because the time variations plotted in Fig. 9 frequently showed phase delays between $T'$ and $w'$.

e. Heat flux profile

By taking advantage of the high time resolution of both $T'$ and $w'$ measurements, a profile of heat flux $Q$ was calculated by using the relation

$$ Q = C_p \rho T'w', $$

where $C_p$ and $\rho$ are the specific heat under constant pressure and the air density, respectively. The values of $\rho$ were calculated by means of the equation of state using the results of six radiosonde soundings, and $C_p$ is assumed to be a constant (1004 J K$^{-1}$ kg$^{-1}$).

We first calculated CCF between $T'$ and $w'$ with the maximum time lag of 3 h, and then estimated $T'w'$ from the CCF value at zero lag. It is noteworthy that CCF, for example, as shown in Fig. 8, did not have a negative spike at zero lag, suggesting that there was no systematic error in the heat flux determination caused by poor resolution of the vertical wind velocity measurement (Peters et al. 1985). We determined the value of $T'w'$ at each altitude by shifting the center of the time series every 30 min. Then, profiles of $Q$ were averaged over about 8 h and are presented in Fig. 10 for six observation periods that partially overlap each other.

The values of $Q$ were normally as small as 2–3 W m$^{-2}$ above about 2 km. The observed small values of the heat flux are consistent with the explanation that fluctuations in that height range were mainly due to gravity waves that are not effective in transporting heat flux.

![Fig. 10. Profiles of heat flux averaged over about 8 h during the six observation periods.](image-url)
During the first three observation periods in Fig. 10, \( Q \) was fairly small, even below 2 km. On the other hand, in the latter three periods, corresponding to the disturbed conditions, as shown in Fig. 7, the values of \( Q \) became significantly enhanced below 2 km as large as 15–25 W m\(^{-2}\). In particular, the \( Q \) profile obtained from 1130 to 2008 LT 7 August showed large positive values at the bottom of the observed height range, which decreased rapidly to almost zero at 2 km. It is suggested that the upward heat transfer from the ground to the troposphere was large during the daytime due to convective motions.

Peters et al. (1985) conducted the observation of the planetary boundary layer with sodar and RASS on 13 November 1983 and reported that the values of \( Q \) were 37 ± 5, 39 ± 11, 46 ± 12, and −20 ± 29 W m\(^{-2}\) at the 75-, 113-, 150-, and 188-m altitudes, respectively. Angervin et al. (1993b) also determined upward heat flux exceeding 100 W m\(^{-2}\) below about 500 m. The amplitudes of \( Q \) in Fig. 10 were much smaller than these results, which might indicate that the large heat flux values are normally confined in the planetary boundary layer, although these results may not be directly compared because of various differences in the measurement systems and meteorological conditions.

Figure 11 shows a contour plot of heat flux, based on determinations of \( Q \) every 30 min at each altitude. During the first one-third of the observation period, the values of \( Q \) were small in the entire height range, showing fairly calm conditions. But, a disturbed structure with large positive and negative \( Q \) values appeared in the latter period. The \( Q \) values were largely positive below 2-km altitude from 1300 to 1500 LT 7 August but became negative from 1700 LT 7 August to 0100 LT 8 August, and then again were positive until 0500 LT 8 August. Thus, the observed characteristics of the time variations of \( Q \) values may not be simply explained. The effects of the large \( Q \) values seemed to extend into the region above 2 km, although the amplitudes of \( Q \) were relatively small.

5. Concluding remarks

We have reported in this paper preliminary results of the MU radar–RASS observations of time–height variations of temperature profiles in the troposphere at 1.5–7 km, which were continued for about 40 h on 6–8 August under fairly calm meteorological conditions. The RASS temperature profiles averaged for about 2 h, being similar to a launch interval of a normal radiosonde measurement, agreed fairly well with the virtual temperature deduced with radiosondes, with a discrepancy of about 0.3°C. The temperature and vertical wind velocity fluctuations were analyzed with a time resolution of 4.5 min. The characteristics of the \( T' \) and \( w' \) fluctuations were greatly different in the regions below and above 2.5 km. Above 2.5 km, frequency spectra of \( T' \) had a slope of about −5/3, while those of \( w' \) were flat, which suggests that the fluctuations can be described in terms of atmospheric gravity waves. On the other hand, low-frequency vertical wind components were dominant below 2.5 km.

A wave component with a period of 6–8 h showed a \( \pi/2 \) phase lag between temperature and vertical wind velocity, which again agreed with the prediction based on the gravity-wave theory. However, the fluctuations

![Fig. 11. Contour plot of heat flux using results determined every 30 min from CCF for 3-h time series.](image-url)
of temperature and vertical wind velocity near the planetary boundary layer were in phase, suggesting that mechanisms other than gravity waves, such as convection, could become dominant there. Short-period components were also recognized for both $T'$ and $w'$ fluctuations, having a period close to the Brunt–Väisälä frequency.

To summarize, the major part of the fluctuations above 2.5 km were explained by gravity waves, but at the lower altitudes convective motions with relatively long periods were enhanced during the middle of the observation period. A relatively large heat flux was observed during the daytime, coinciding with large disturbances of $T'$ and $w'$ fluctuations. Based on these results, we conclude here that the MU radar–RASS has a potential that can greatly contribute to studies of tropospheric dynamics by allowing accurate observation of the fine time–height structure of wind velocity and temperature fluctuations.

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