High-Resolution Radio Acoustic Sounding System Observations and Analysis up to 20 km

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ABSTRACT

The first campaign-based measurements of virtual temperature in the upper-troposphere and lower-stratosphere (UTLS) region were made with the middle- and upper-atmosphere (MU) radar radio acoustic sounding system (RASS) during 4 days in August 1995. This dataset was examined in order to study high-frequency changes in the stability below 20 km, but especially in the UTLS region. Calculations of the WMO tropopause and cold-point tropopause heights showed the latter to be \( (1.0 \pm 0.6) \) km higher, where \( 0.6 \) km is the standard deviation. A diurnal cycle of temperature and wind dominated the spectra, which was identified as the diurnal solar tide. Its phase maximum occurred in the afternoon between 5 and 15 km and showed upward energy propagation above this height. Changes in the UTLS kinetic energy dissipation rate \( \epsilon \) showed significant high-frequency fluctuations embedded within layers that persisted for at least 1 day. Relative to the WMO tropopause height, the median \( \epsilon \) increased from \( (0.5 \pm 0.1) \times 10^{-3} \) m\(^2\) s\(^{-3}\) in the upper troposphere to \( (0.7 \pm 0.1) \times 10^{-3} \) m\(^2\) s\(^{-3}\) in the lower stratosphere.

1. Introduction

The radio acoustic sounding system (RASS) has been developed and used with a variety of radars located around the world for over 20 yr for the purpose of obtaining high-resolution profiles of tropospheric virtual temperature \( T_v \) (Matuura et al. 1986; May et al. 1989; Tsuda et al. 1994). The RASS technique uses acoustic speakers to generate artificial fluctuations whose Doppler velocities are then measured by collocated radar. The \( T_v \) measurement can then be obtained because it is a function of acoustic velocity \( c_a \). In practice, \( T_v \) is generally only obtained in the lower range gates at which the radar can also obtain wind measurements. Limitations with radar power, the absence of full radar beam steering, or strong background winds distorting the acoustic wave act to restrict the maximum height at which \( T_v \) can be retrieved.

Despite the height restrictions, RASS has been used to increase understanding of tropospheric phenomena such as cold fronts (Neiman et al. 1992), boundary layer heat fluxes (Angevine et al. 1993), humidity profiles (Furumoto and Tsuda 2001), fog inversions (Bonino et al. 1981), and gravity waves (Tsuda et al. 1994; Yamamoto et al. 1996). Recent experiments with the very high-frequency (VHF) middle- and upper-atmosphere (MU) radar RASS in Shigaraki, Japan (located at 34.85°N, 136.10°E; 370 m MSL), have shown fluctuations in turbulent parameters, such as the turbulent spectral width \( \sigma_t \) and kinetic energy dissipation rate \( \epsilon \) over short time scales (Hermawan and Tsuda 1999; Furumoto and Tsuda 2001; Alexander et al. 2007).

The \( T_v \) is related to the acoustic velocity \( c_a \) and the vertical wind velocity \( w \) by (e.g., Tsuda et al. 1994)

\[
T_v = \frac{(c_a - w)^2}{k_h},
\]

where \( k_h \) is composed of the specific heat ratio \( \gamma \), the gas constant \( R \), and the mean molecular weight of the atmosphere \( M \),

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

For a dry atmosphere, \( k_h = 20.046 \), but it varies slightly depending upon the humidity. The addition of atmospheric water vapor results in a slight reduction of \( M \), but because \( \gamma \) is also a weak function of humidity, it is
not possible to simply determine the dependence of $k_n$ on humidity. The largest difference between $T_n$ and the dry atmospheric temperature $T$ is below 5 km, because that is where most of the water vapor is located.

The largest RASS $T_n$ error is not associated with the humidity, but it occurs if nonsimultaneous measurements of $c_a$ and $w$ are made. The derived $T_n$ must include the $w$ correction, otherwise it can be affected by $\sim 0.7$ K (Moran and Strauch 1994). On short time scales $w$ varies, thus requiring the simultaneous observation of $c_a$ and $w$. Increases in computational power since the late 1980s have enabled simultaneous measurements of $c_a$ and $w$ (May et al. 1989; Angevine and Ecklund 1994), thus removing the need to interpolate coarse-resolution $w$, and so allowing an accurate estimation of $T_n$. Higher-order error sources (of the order of a few tenths of a kelvin only in extreme cases) include vertical temperature gradients (Takahashi et al. 1988), turbulence (Latasitis 1992), temperature retrieval approximations (Angevine and Ecklund 1994), and range errors resulting from gradients in radar reflectivity (Görsdorf and Lehmann 2000).

The tropopause can be defined in a variety of ways, such as dynamically (using potential vorticity), chemically (ozone concentration), thermally [by analyzing the lapse rate, i.e., the World Meteorological Organization (WMO) tropopause], or by finding the coldest temperature in a vertical profile (cold-point tropopause). The extratropical tropopause closely corresponds to the 2 PVU (1 PVU $= 10^{-6}$ m$^2$ s$^{-1}$ K kg$^{-1}$) constant potential vorticity surface (Holton et al. 1995). The choice of tropopause affects the value of averaged parameters, such as $N^2$ in the midlatitudes (Birner et al. 2002). Previous work with MU radar RASS campaign data showed the clearest separation of troposphere–stratosphere air during turbulent mixing with the use of the WMO tropopause (Alexander and Tsuda 2007). Therefore, the tropopause was calculated by using the WMO definition, which states that the tropopause occurs at the lowest altitude where the lapse rate decreases to 2 K km$^{-1}$ or less and that the average lapse rate at levels within 2 km above the tropopause does not exceed 2 K km$^{-1}$ (WMO 1957).

During August 1995, a unique 4-day experiment was conducted with the MU radar RASS, which for the first time enabled high-temporal-resolution measurements of $T_n$ to be made in the upper-troposphere and lower-stratosphere (UTLS) region. This experiment was conducted during a time of weak winds and daily convection, allowing an extensive thermodynamic picture of the summertime atmosphere below 20 km at Shigaraki to be constructed. Using these data, an increased understanding of high-frequency changes in the Brunt–Väisälä frequency $N$, tropopause height, and turbulence parameters was possible. The dataset also allowed an analysis of the interaction of temperature and wind with $N$ and a study of the diurnal tide. Section 2 describes the datasets and data collection method as well as comparisons with collocated radiosondes. The full temperature and stability datasets are discussed in section 3a, with a detailed examination of the tropopause height following in section 3b. Large-scale processes and the diurnal tide are discussed in sections 3c and 3d, respectively. Tropopause-relative turbulence parameters are examined in detail in section 3e. The results obtained are discussed in section 4 and conclusions are presented in section 5.

2. Datasets and background conditions

A synoptic-scale map taken from National Centers for Environmental Prediction (NCEP) data (Kalnay et al. 1996) at 0000 UTC 4 August 1995 is shown in Fig. 1. This time corresponds to 0900 LT 4 August, near the middle of the RASS campaign period. The tropopause height above the MU radar at Shigaraki at this time was between 105 and 110 hPa, corresponding to a height of about 17 km. NCEP maps from other days during the campaign showed that this high-tropopause region moved away from Shigaraki, so that by 6 August, the tropopause height had lowered to about 115 hPa. The 200-hPa geopotential height is also shown in Fig. 1, showing that this pressure level was higher over southern Japan than northern Japan at 0000 UTC 4 August. Analysis of other NCEP synoptic-scale maps showed that the 200-hPa geopotential height always exceeded 12.5 km above Shigaraki.

Data were collected with the MU radar RASS for 4 days, 2–6 August 1995. Twenty acoustic sources were located around the MU radar. During one cycle of data collection, two wind and three RASS observations were made. These were combined to give a temporal resolution of 3.6 min for three wind components and virtual temperature. Winds were observed between 2 and 24 km, while $T_n$ was observed between 2 and 20 km.

The wind components were obtained from the Doppler radial velocities measured by five beams, with one pointing toward the zenith, while the other four were at 6° off zenith in the cardinal directions. The pulse length used was 300 m, with the data oversampled at 150-m vertical intervals. Radar beam steering was used to follow the RASS acoustic spot by applying a ray tracing of acoustic waves (Masuda 1988). As such, the following five beam directions were used for the RASS experi-
ment (azimuth angle, zenith angle): (315°, 10°), (330°, 10°), (300°, 14°), (315°, 14°), and (330°, 14°). The MU radar was operated in RASS mode over three height intervals (1.20–10.65, 4.80–14.25, and 10.80–20.25 km). Full observational parameters are described in Table 1. To satisfy the Bragg condition and obtain an RASS echo across a wide range of heights, the RASS acoustic speakers were frequency swept in different bands in these three height regions, as detailed in Table 2. The standard wind observation mode parameters of the MU radar are provided in Table 3.

The mean winds during the campaign are shown in Fig. 2. The magnitude of the zonal wind |\(u|\) was less than 7 m s\(^{-1}\) at all altitudes; \(u\) was positive eastward below 15 km and westward above 15 km. The meridional wind \(v\) was southward at all altitudes, with a maximum of \(-9\) m s\(^{-1}\) at 15 km. These light winds assisted in the ability to retrieve \(T_v\) profiles up to 20 km by not distorting the acoustic signal to the extent that strong winds would have.

During the campaign, radiosondes were launched from the MU radar site about once every 6 h, with 17 launched in total. The vertical radiosonde resolution of 30 m was interpolated to match the 150-m vertical resolution of the radar. The radiosondes took approximately 70 min to ascend to 20 km. To compare the RASS and radiosonde data, RASS \(T_v\) data were extracted from the time of each launch until 70 min after the launch of that flight. Because of RASS UTLS data gaps, only 9 of the 17 radiosondes were used to perform direct comparisons. These RASS data were then used to produce a median profile for the entire campaign. This median RASS profile is shown with the median \(T_v\) calculated from radiosonde measurements of tempera-

### Table 1. MU radar RASS observational parameters. Interpulse period (IPP) and fast Fourier transform (FFT).

<table>
<thead>
<tr>
<th>Observation mode</th>
<th>Low</th>
<th>Middle</th>
<th>High</th>
</tr>
</thead>
<tbody>
<tr>
<td>Height range (km)</td>
<td>1.20–10.65</td>
<td>4.80–14.25</td>
<td>10.80–20.25</td>
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<tr>
<td>IPP ((\mu)s)</td>
<td>412</td>
<td>404</td>
<td>416</td>
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<td>Subpulse length ((\mu)s)</td>
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<td>1</td>
<td>1</td>
</tr>
<tr>
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<td>28</td>
<td>24</td>
</tr>
<tr>
<td>No. of FFT points</td>
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<td>128</td>
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<tr>
<td>Incoherent integration</td>
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<td>5</td>
<td>5</td>
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<tr>
<td>Observation duration (s)</td>
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### Table 2. Acoustic speaker parameters.

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<th>Transmitter</th>
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<th>Pneumatic transducer</th>
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<tbody>
<tr>
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<td>CW</td>
<td>FM chirped pulse</td>
</tr>
<tr>
<td>No. of pulses</td>
<td>—</td>
<td>4</td>
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<tr>
<td>Pulse interval (s)</td>
<td>—</td>
<td>10</td>
</tr>
<tr>
<td>Repetition period (s)</td>
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<td>—</td>
</tr>
<tr>
<td>Pulse length (s)</td>
<td>—</td>
<td>0.4 (1.20–10.65 km)</td>
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<tr>
<td>Chirped frequency range (Hz)</td>
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<td>85–115 (1.20–10.65 km)</td>
</tr>
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</table>

FIG. 1. Background height of the tropopause (shaded) and the 200-hPa geopotential height (lines; km) taken from NCEP reanalysis at 0000 UTC 4 Aug. The location of the MU radar is marked by the cross.
Table 3. Clear-air echo observational parameters.

<table>
<thead>
<tr>
<th>Observation mode</th>
<th>Low</th>
<th>High</th>
</tr>
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<tbody>
<tr>
<td>Height range</td>
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<td>1</td>
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<tr>
<td>Coherent integration</td>
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<td>No. of FFT points</td>
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<td>Incoherent integration</td>
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<td>1</td>
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<tr>
<td>Observation duration (s)</td>
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<td>53</td>
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structure. The differences between averaging data in height coordinates and tropopause-relative coordinates in relation to turbulence parameters will be discussed further in section 3e.

3. Analyses and results

a. Temperature and static stability structure

The perturbations \( T' \) from the campaign mean \( T \) at each height are shown in Fig. 4. Downward progression of phase fronts is visible in the troposphere below 10 km with periods of greater than 2 days. Shorter-period waves of less than 12 h are visible in the UTLS region. The RASS and radiosonde tropopause are marked for reference and will be discussed below. Profiles of \( u' \) and \( v' \) are not shown, but revealed UTLS downward phase-shifted inertio-gravity waves with vertical wavelengths of about 2 km and periods on the order of 1 day.

The full 3.6-min-resolution \( N^2 \) is shown in Fig. 5. Many layers of enhanced stability can be observed in the troposphere below 10 km. Two main layers are visible at 5 and 8 km, which showed a descending motion during the campaign. Periods of convection, which are defined here as times where vertical wind variances exceeded 0.01 m² s⁻², occurred each day throughout the campaign period. Convective activity, duration, and height are recorded in Table 4. Surface rainfall was only recorded on 2 August. The effects of convection can be seen in the \( N^2 \) oscillations during each afternoon. The convection on 3–5 August disturbed but did not destroy these enhanced \( N^2 \) layers. The 5-km layer strengthened in stability during each night after the effects of convection ceased. Enhanced kinetic and potential energies were observed during each afternoon (not shown). The convection with the largest total energy occurred
on 2 August, but in all cases the total energy was less than 3 J kg\(^{-1}\) above 12 km. From 10 to 14 km, the atmosphere was very unstable, with \(N^2\) often negative. The UTLS region above 14 km was characterized by many stable, long lasting layers. Despite the data gaps, these layers can be traced for periods of several days. The radiosonde and RASS WMO tropopause are in close agreement except for the 0300 LT 5 August.

**FIG. 3.** (a) The mean RASS and radiosonde \(T_v\) with (b) the RASS minus radiosonde difference, and (c) the mean RASS and radiosonde \(N^2\) profiles. In each of these panels, the horizontal solid line is the RASS WMO tropopause height, while the horizontal dashed line is the radiosonde WMO tropopause height. (d) The corresponding tropopause-relative \(T_v\) profiles, (e) the \(T_v\) difference, and (f) the \(N^2\) across the tropopause are shown. The solid lines in (a), (c), (d), (f) are the RASS data, while the dashed lines in (a), (c), (d), (f) are the radiosonde data. The short horizontal lines in (c), (d), and (f) mark the std error in the data at each height.

**FIG. 4.** RASS virtual temperature perturbations from the mean at each height. RASS and radiosonde WMO tropopauses are marked by the crosses and red circles, respectively; x-axis tick mark represents midnight local time. Missing data are marked white.
launch. On 2 August, an upward movement of the tropopause was evident, with a break in tropopause height occurring at about the time of the second radiosonde launch. Around 0000 LT 4 August, the tropopause was flat and at 17.5-km altitude, before descending to 16.5 km by 0000 LT 5 August. On 5 August, the tropopause height showed a downward movement with a 12-h period. The tropopause is discussed in more detail below.

b. Tropopause height

The WMO tropopause heights and the cold-point tropopause heights obtained from the individual 3.6-min-resolution RASS profiles are shown in Fig. 6a. The campaign mean cold-point tropopause occurred at (17.7 ± 0.5) km, where the standard deviation was (0.5) km. Note that the cold-point tropopause is defined as the height of the coldest temperature. The cold-point tropopause was (1.0 ± 0.6) km higher than the WMO tropopause during the campaign, where (0.6) km refers to the standard deviation from the mean. There are occasions when no WMO tropopause height was determinable because the definition required at least 2 km of data above the suspected tropopause.

The tropopause structure can be broken into the following three parts: the first is on 2 August, the second is on 3 August and early 4 August, and the third is from 1200 LT 4 August onward. The WMO tropopause on 2 August moved upward in height, from 16.5 to 17.0 km. The cold-point tropopause showed a matching increase from 17.5 to 18.0 km and, as illustrated in Fig. 6b, resulted in a difference of about 1 km between the two tropopauses.

The next period of available data occurred late on 3 August and early on 4 August, when both tropopauses showed a flat structure and coincided in height at 17.5–18.0 km. Results from the radiosondes in Fig. 5 suggest a period of flat tropopause extending back to the 0500 LT 3 August launch, but higher-frequency changes in tropopause height may also have existed.

After 1200 LT 4 August, the structure of the WMO tropopause changed markedly as it decreased in altitude from 18.0 to 16.5 km. On 5 August, it exhibited two downward movements each of 12-h duration, covering heights from 17.2 to 16.2 km. The WMO tropopause once again jumped up to 17.5 km shortly before 0000 LT 6 August. During this time, the cold-point tropopause was between 0.5 and 2 km above the WMO tropopause, except at the times of the height discontinuities at 1200 LT 5 August and 0000 LT 6 August.

![Fig. 5. The $N^2$ profile during the 4-day campaign. The radiosonde WMO tropopause heights are marked by the red circles, while the RASS WMO tropopause heights are marked by the crosses.](image-url)
where the two tropopauses were coincident. The $T_e^*$ in Fig. 4 showed downward phase–propagating gravity waves on the order of 12 h on 5 August. Large, positive perturbations at 16 km were associated with these tropopause height discontinuities.

The tropospheric diurnal convective cycle resulted in the emission of gravity waves (which will be illustrated in section 3c), but these were of small amplitude each day and it is not clear that they influenced the tropopause height structure. The $N^2$ time–height plot in Fig. 5 showed many stable layers, which fluctuated in height but lasted for a day or longer in both the upper troposphere and lower stratosphere. Because the WMO tropopause followed these layers, the changes in the WMO tropopause height may be related to effects from more distant sources rather than the diurnal convection.

Model output or reanalysis data can be at too coarse a resolution to distinguish between the WMO and cold-point tropopause in certain situations (Highwood and Hoskins 1998). Observationally, the cold-point tropopause is mainly used when studying the tropical atmosphere, because the troposphere and stratosphere are nearly convectively and radiatively adjusted (Highwood and Hoskins 1998). Because of barocline instabilities, the midlatitudes are not in convective or radiative equilibrium; therefore, the dynamical WMO tropopause definition is more useful (Holton et al. 1995; Birner 2006). However, the very cold and high tropopause in summertime above Shigaraki ($\sim 35^\circ$N) is closer in structure to the tropical regions. The cold-point tropopause was often higher than the WMO tropopause in tropical northern Australia (Selkirk 1993), which agrees qualitatively with the results presented here.

c. Large-scale features

The area–preserved spectra of wind and virtual temperature are shown in four height regions in Fig. 7. In the lower and middle troposphere (3–7 and 7–11 km, respectively), the diurnal component dominated the $u$, $v$, and $T_e$ spectra. The horizontal wind spectral slopes changed for periods of less than about 4 h, and show a similar slope to the vertical wind spectrum in this frequency range. This is probably a result of the daily generation of convective gravity waves with high frequencies. The upper troposphere (11–15 km) corresponded to the region of low or negative $N^2$. The spectral slopes of the horizontal wind components in the upper troposphere are greater than that below 11 km. In the 15–19-km UTLS region, the $w$ spectrum for periods greater than 1 h is at noticeably lower energies than at lower altitudes. This will be related to convective effects below.

The 4–24-h bandpass-filtered $T_e^*$ time and height variations are shown in Fig. 8. Four-hour low-pass-filtered $N^2$ data in the troposphere below 11 km are overplotted. Because $T_e^*$ is proportional to $N$, some of the disturbances are related to changes in the stability. This is apparent around the 5- and 8-km synoptic descending $N^2$ layers. Below 5 km, a strong diurnal component is evident. Positive $T_e^*$ occurred in the afternoon.
and evening of each day and showed a slight downward phase propagation. A phase change in $T''$ is evident at the 5-km enhanced stability layer. A diurnal cycle in $T''$ of amplitude $\sim 0.5^\circ C$ is apparent in the upper troposphere. The maximum positive perturbation in the upper troposphere coincided with the maximum $T''$ below 5 km. No phase tilting is evident in the upper-troposphere fluctuations, implying a standing wave with neither upward nor downward energy propagation. Larger-amplitude downward phase–propagating subdiurnal stratospheric perturbations were visible on 5 August.

The same bandpass filtering is applied to the zonal wind, with the results shown in Fig. 9. The diurnal cycle is clearly visible below 10 km, showing an upward progression of the phase fronts. Positive $u'$ occurred in the morning of each day below the 5-km $N^2$ layer. Nocturnal increases in the strength of the 5-km $N^2$ layer often corresponded to a positive $u'$. The maximum eastward and westward $u'$ perturbations were confined between the 5- and 8-km layer and also showed a descending motion. Downward phase propagation of $u'$ was evident above 11 km. The bandpass-filtered meridional wind component is not illustrated here because it showed similar characteristics to the zonal component. Upward phase shifting below 10 km was also observed. Maximum $v'$ perturbations were also confined between the descending 5- and 8-km enhanced $N^2$ layers.

The observed diurnal cycle of convection produced gravity waves, which were most easily observed in the vertical wind velocity data. The 2–12-h bandpass-filtered $w''$ are shown in Fig. 10. Convection below 5 km is clearly visible from midday each day. Gravity waves propagated upward toward the UTLS region. The majority of gravity waves dissipated in the unstable region between 11 and 14 km. By the time that the waves reached the tropopause, their amplitudes were less than 0.1 m s$^{-1}$.

d. Diurnal tide

In the lower atmosphere, especially between 3 and 9 km, the diurnal tide is mainly a result of water vapor absorbing sunlight, thus exciting it (Riggin et al. 2002). Latent heat release from convection is also responsible

![Fig. 7. Wind and temperature spectra in energy content form, averaged over (a)–(d) 3–7, (e)–(h) 7–11, (i)–(l) 11–15, and (m)–(p) 15–19 km. Note that the vertical scales are different between each parameter.](image-url)
for some of the diurnal tide, with larger amplitudes over the continents than the oceans (Hamilton 1981). Because of longitudinal inhomogeneities in water vapor distribution, the diurnal tide is not consistent and is divided into two components: the migrating (sun synchronous and with a zonal wavenumber of one) and the nonmigrating (Williams and Avery 1996) tide. The nonmigrating diurnal tide is observed here because the long wavelength and low amplitude of the migrating component are not resolvable in the limited-height RASS dataset. During the MU radar RASS campaign, a clear diurnal cycle of convection was noticed throughout the troposphere. Diurnal cycles in $T_v$, $u$, and $v$ were previously documented in section 3c and dominated the spectral plots of Fig. 7.

For the 3 days where 24 h of data existed (3–5 August), the amplitude and phase of the diurnal tide were extracted. The method of extraction generally follows

**Fig. 8.** The 4-24-h bandpass-filtered $T_v'$; RASS WMO tropopause heights are marked by the crosses. The line contours show the 4-h low-pass-filtered $N^2 (10^{-3} \times \text{rad}^2 \text{s}^{-2})$ values in the troposphere. Missing data are marked white.

**Fig. 9.** The 4-24-h bandpass-filtered $u'$; RASS WMO tropopause heights are marked by the crosses. The line contours show the 4-h low-pass-filtered $N^2 (10^{-3} \times \text{rad}^2 \text{s}^{-2})$ values in the troposphere. Missing data are marked white.
that of Riggin et al. (2002). First, the raw dataset was interpolated to 1-h resolution. A 36-h high-pass filter was used to remove low-frequency components, including the multiday tropospheric $T'_v$ observed in Fig. 4. To reduce noise, a 3-km low-pass vertical filter was then applied. After the daily composite from the 3-day-filtered dataset was formed and any residual mean removed, a sine wave with a 24-h period was fitted to the daily composite from which amplitude and phase were extracted. Results are illustrated in Fig. 11. The large diurnal amplitude of $u'$ and $v'$ between 6 and 10 km is visible in Fig. 11a. In the region above 15 km, $v'$ had an amplitude of up to 2.0 m s$^{-1}$, while the $u'$ amplitude was significantly smaller. The $T'_v$ oscillations were less than 0.5°C between 5 and 15 km, but slightly larger above and below this region.

The phase is shown in Fig. 11b and indicates the local time of maximum positive perturbation. The $u'$ maxi-

![Figure 10](image10.png)

**Fig. 10.** The 2–12-h bandpass-filtered $w'$, showing the convectively generated gravity waves; RASS WMO tropopause heights are marked by the crosses. The line contours show the 4-h low-pass-filtered $N^2 (10^{-3} \times \text{rad}^2 \text{s}^{-2})$ values in the troposphere. Missing data are marked white.

![Figure 11](image11.png)

**Fig. 11.** The 3-day average of the diurnal tide’s (a) amplitude and (b) phase. In each case, the solid line marks $u$, the dashed line $v$, and the dash-dot line $T'_v$. 

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mum phase occurred about 5 h (i.e., about one-quarter of a wave cycle) after the $u'$ maximum in the 6–11-km region. The $u'$ and $T_o'$ phases showed jumps of 12 h between 5 and 6 km, which were coincident with the enhanced $N^2$ layer at this level. Between 9 and 15 km, the maximum $T_o'$ occurred in the midafternoon period and showed a standing wave motion. Above about 14 km, all three phases from $u'$, $v'$, and $T_o'$ showed the expected downward phase progression, indicating upward energy propagation. Note that the upward energy propagation started below the tropopause height, but corresponded with the height of the lowest of the UTLS stable layers observed in Fig. 5.

e. Tropopause-relative turbulence parameters

The kinetic energy dissipation rate $\epsilon$ can be calculated using different formulas (Hocking 1999), including the Labitt approach

$$\epsilon = 0.79 \frac{\sigma_r^3}{L_r(z)} c,$$  

(5)

where $L_r(z)$ is the horizontal distance covered by the radar beam at a certain altitude $z$ and $c$ is a correction factor very close to unity. The half-power half-width Doppler spectral width resulting from turbulence is given by $\sigma_r$. The buoyancy-scale approach of calculating $\epsilon$ is given by

$$\epsilon = A^{-3/2} N \sigma_r^2,$$  

(6)

where $A = 1.6$ is a Kolmogorov constant. Because of the narrow beamwidth ($3.6^\circ$) and short pulse length (300 m) of the MU radar, the method of calculating $\epsilon$ via Eq. (6) was adopted, because it was accurate to 10%. The theoretical background of calculating $\epsilon$ is fully described by Hocking (1999).

Effects resulting from beam and wind shear broadening were removed from the observed spectral width in order to obtain $\sigma_r$ (Hocking 1985). A study of $\sigma_r$ (not shown) revealed larger values in the afternoon and evening of each day than in the morning, which corresponded with the diurnal convective cycle. Spectral widths of up to 1.2 m s$^{-1}$ were measured between 10 and 13 km on 2 and 3 August. In the UTLS region, horizontal bands of smaller and larger $\epsilon$ were evident, although no clear change in the magnitude of $\sigma_r$ at the tropopause height was noticed.

Figure 12 shows $\epsilon$ in the UTLS calculated from the individual profiles during the campaign. A layered structure of $\epsilon$ is apparent because of its dependence upon $N$. Small-scale, high-frequency changes were regularly noticed with $\epsilon$ in the UTLS region, which has previously been observed in MU radar RASS data in the lower troposphere (Furumoto and Tsuda 2001; Alexander et al. 2007). Large $\epsilon$ around 0000 LT 4 August at 16.5 km were observed, but otherwise $\epsilon$ were generally lower below the tropopause than above it.

The height-relative median profiles of $\sigma_r$ and $\epsilon$ during the campaign are shown in Figs. 13a,b. Medians are used instead of the means in order to minimize effects of any outlying data. Standard errors are shown in each panel. As shown in Fig. 13a, $\sigma_r \sim 0.5$ m s$^{-1}$ below 10 km, with a small increase at 7 km related to the peak of
the convective activity. The maximum $\sigma_t \sim 0.7 \text{ m s}^{-1}$ occurred in the statically unstable upper troposphere between 13 and 16 km. The $\sigma_t$ decreased slightly from 16 to 18 km but was a relatively constant $0.6 \text{ m s}^{-1}$ in the lower stratosphere. Figure 13b shows that $\epsilon$ was minimum at 10–12 km, where it was about $2 \times 10^{-4} \text{ m}^2 \text{ s}^{-3}$, before increasing to be $\sim 10^{-3} \text{ m}^2 \text{ s}^{-3}$ at 20 km.

In the lower and middle troposphere, $\sigma_t$ had a local maximum at 7 km. The values of $\epsilon$ in the UTLS calculated here were larger than the climatological values at the MU radar of $\sim 10^{-4} \text{ m}^2 \text{ s}^{-3}$ obtained using radiosondes by Fukao et al. (1994). Nastrom and Eaton (1997) reported climatological summertime UTLS $\epsilon$ at the White Sands, New Mexico, radar of $\sim 5 \times 10^{-4} \text{ m}^2 \text{ s}^{-3}$ and a similar value at Vandenburg, California (Nastrom and Eaton 2005).

These campaign turbulence parameter height profiles do not reveal the full details across the tropopause because the tropopause height is not constant. Instead, the campaign medians are replotted relative to the RASS WMO tropopause height in order to examine their change across the tropopause. The WMO tropopause is used instead of the cold-point tropopause because the former was previously found to better separate the tropospheric and stratospheric air masses when considering turbulent mixing (Alexander and Tsuda 2007). As shown in Fig. 13c, there was only a slight decrease in $\sigma_t$ across the tropopause boundary. In the 2.5 km below the tropopause, $\sigma_t = (0.65 \pm 0.02) \text{ m s}^{-1}$, while in the 2.5 km above it, $\sigma_t = (0.6 \pm 0.02) \text{ m s}^{-1}$.

From Fig. 13d, it can be observed that the median $\epsilon$ went from $(0.5 \pm 0.1) \times 10^{-3} \text{ m}^2 \text{ s}^{-3}$ in the upper troposphere to $(0.7 \pm 0.1) \times 10^{-3} \text{ m}^2 \text{ s}^{-3}$ in the lower stratosphere. Uncertainties describe the standard errors in the medians. A local maximum in $\epsilon$ was noticed within 1 km above the tropopause. Seasonal estimates of $\epsilon$ obtained using $N$ from radiosonde data relative to the tropopause were also made with radars in the United States (Nastrom and Eaton 1997, 2005), where it was shown that most of the tropopause-relative structure of $\epsilon$ was due to the tropopause-relative structure of $N$. This was also observed with the MU radar RASS dataset, as can be seen by comparing the increase of $N^2$ in the lower stratosphere (Fig. 3f) with that of $\epsilon$ (Fig. 13d).

4. Discussion

The dominant wave period observed in the $T_v$, $u$, and $v$ spectra during the campaign period was 24 h. This oscillation fits the known characteristics of the diurnal tide because the phases exhibited standing wave behavior in the midtroposphere and downward phase front progression in the stratosphere. The amplitudes of the wind components of the diurnal tide were largest in the middle troposphere, where the large amplitudes were confined between two regions of enhanced stability. These regions showed a clear descending motion during the experiment. The maximum phases of all components in this region occurred in the afternoon at a time
when convection and gravity wave activity in $w$ was observed. The absence of phase shifting in this region suggests standing wave behavior. Above 15 km, downward phase (upward energy) propagation became evident, both in the diurnal tide and higher-frequency components. This coincided directly with the height of the first layer of increased stability in the UTLS region and was therefore well below the mean tropopause height.

The convectively generated vertical velocity gravity wave components had amplitudes of less than 0.1 ms$^{-1}$ above 15 km and did not appear to influence the WMO tropopause height. A low-stability region occurred between 11 and 14 km during the entire campaign, which acted to dissipate most of the upward-propagating gravity waves. The convection changed the background stability in the lower and middle troposphere while it was present. Wavelike motions were visible in the $N^2$ data but on none of the days was the convection strong enough to destroy the long-lasting descending layers of enhanced stability at 5 and 8 km. Indeed, during each night these layers restrengthened as the convection ceased.

The WMO and cold-point tropopause heights varied throughout the experiment and only coincided when the tropopause height was constant around 0000 LT 4 August and briefly when there was a tropopause discontinuity at 1200 LT 5 August and 0000 LT 6 August. The tropopause discontinuities on 5 and 6 August corresponded to downward phase-shifting (upward energy propagating) gravity waves, with large positive temperature perturbations.

In the UTLS region, high-frequency fluctuations in $N^2$ occurred along with layers of enhanced stability, which persisted over several days. These layers were also observed in the $\epsilon$ data. The results of studying $\sigma$, and $\epsilon$ relative to the tropopause as opposed to height above ground level indicated that averaging data in altitude coordinates can wash away sharp transitions in these parameters at the tropopause. Climatological studies of midlatitude $\sigma$, and $\epsilon$ using standard 12-h radiosondes data to obtain $N^2$ also showed that measurements averaged in altitude underestimated the tropopause transitions (Nastrom and Eaton 1997, 2005). Birner (2006) showed that climatological $N^2$ profiles in southern Germany exhibited a strong inversion layer of a few kilometers depth immediately above the tropopause. This was not apparent in the RASS data, perhaps because the RASS $T_v$ were only obtainable a couple of kilometers above the tropopause, and also because the summertime tropopause above Shigaraki was much higher than in Germany.

5. Conclusions

The high-resolution MU radar RASS $T_v$ dataset allowed a detailed analysis of $N^2$ and high-frequency changes in both the WMO and cold-point tropopause heights. A campaign-averaged difference in altitude between the two tropopauses showed the cold-point tropopause to be $(1.0 \pm 0.6)$ km higher. The tropopause height varied on time scales shorter than the 6-hourly radiosondes could detect, although 12-h-period downward progressions of the tropopause height were also observed.

Convective gravity waves were generated each afternoon but the vertical velocity perturbations did not reach the UTLS region. However, 12-h-period upward energy-propagating gravity waves observed in the $T_v$ data corresponded with the height discontinuities of the tropopause. The temperature and horizontal wind amplitude and phase structures of the diurnal tide were observed. The $u$ and $v$ components showed large amplitudes between the 5-km stable layer and the start of the 11-km unstable region, with a 12-h phase shift in all components at the 5-km stable layer boundary. The kinetic energy dissipation rate $\epsilon$ was studied in the UTLS region. Relative to the RASS-derived WMO tropopause, $\epsilon$ increased from $(0.5 \pm 0.1) \times 10^{-3}$ m$^2$ s$^{-3}$ in the upper troposphere to $(0.7 \pm 0.1) \times 10^{-3}$ m$^2$ s$^{-3}$ in the lower stratosphere.

This unique MU radar RASS dataset provided the opportunity to examine the thermodynamics of the UTLS region with a time resolution not previously possible. The high-resolution variability in the tropopause height and its relation to gravity wave activity could be studied, as well as the UTLS diurnal tide and turbulence parameters.

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