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ABSTRACT

By using the vertical temperature profiles obtained by the radio occultation measurements on the European Space Agency (ESA)’s Venus Express, the vertical wavenumber spectra of small-scale temperature fluctuations that are thought to be manifestations of gravity waves are studied. Wavenumber spectra covering wavelengths of 1.4–7.5 km were obtained for two altitude regions (65–80 and 75–90 km) and seven latitude bands. The spectra show a power-law dependence on the high-wavenumber side with the logarithmic spectral slope ranging from \(-3\) to \(-4\), which is similar to the features seen in Earth’s and Martian atmospheres. The power-law portion of the spectrum tends to follow the semiempirical spectrum of saturated gravity waves, suggesting that the gravity waves are dissipated by saturation as well as radiative damping. The spectral power is larger at 75–90 km than at 65–80 km at low wavenumbers, suggesting amplitude growth with height of unsaturated waves. It was also found that the wave amplitude is larger at higher latitudes and that the amplitude is maximized in the northern high latitudes. On the assumption that gravity waves are saturated in the Venusian atmosphere, the turbulent diffusion coefficient was estimated. The diffusion coefficient in the Venusian atmosphere is larger than those in Earth’s atmosphere because of the longer characteristic vertical wavelength of the saturated waves.

1. Introduction

Internal gravity waves are vertically propagating, small-scale waves for which buoyancy is the main restoring force. Gravity waves are known to have a variety of sources (topography, convection, fronts, jet stream, etc.) and are thought to be one of the most essential elements driving the atmospheric circulation in Earth’s stratosphere and mesosphere (e.g., Fritts and Alexander 2003). Among various dissipation processes of gravity waves, saturation is considered to be of particular importance in Earth’s atmosphere. The amplitude growth with height of gravity waves leads to wave breaking via convective and/or shear instability, and then the amplitude is limited by turbulent diffusion of wave energy; this process is called saturation (e.g., Lindzen 1981; Fritts and Alexander 2003). Smith et al. (1987) and Tsuda et al. (1991) derived a semiempirical formula for the vertical wavenumber spectrum of saturated gravity waves. In terms of the temperature perturbation $T'$, it is written as

\[ T' = T_0 + \frac{1}{2} \xi \eta \cdot \nabla \xi \eta \cdot \nabla T' \]

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where \( k_z \) is the vertical wavenumber with the unit of cycles per meter, \( T_0 \) is the background temperature, \( N \) is the Brunt–Väisälä frequency, and \( g \) is the gravitational acceleration. Another expression, which adopts a conventional unit for \( k_z \), is given in Ando et al. (2012).

Observations using radars, radiosondes, and GPS radio occultation measurements showed that the spectra of gravity waves in Earth’s atmosphere are roughly consistent with (1) (i.e., a logarithmic spectral slope of \(-3\) for wavelengths shorter than 2–5 km) and show flattening at longer wavelengths (e.g., Fritts et al. 1988; Tsuda et al. 1989, 1991; Tsuda and Hocke 2002). While retaining such spectral features indicative of saturation, the amplitude varies over space and time. For example, the spectral density is usually larger in the equatorial region than in higher latitudes, and this equatorial enhancement is attributed to strong cumulus convection (e.g., Tsuda and Hocke 2002; Sato et al. 2003).

Although vertical wavenumber spectra provide fundamental information about the sources and dissipation processes of gravity waves, studies of the vertical wavenumber spectrum are limited in other planetary atmospheres. Recently, Ando et al. (2012) obtained vertical wavenumber spectra of gravity waves in the Martian atmosphere using radio occultation data taken by Mars Global Surveyor. They found that the spectral shape is qualitatively similar to that of Earth’s gravity waves: the spectra are roughly consistent with (1) for wavelengths shorter than 3–8 km. This indicates that saturation plays an important role, although radiative damping should also be important in the Martian atmosphere, which is composed of CO\(_2\). Spectral densities tend to be maximized in the equatorial region in any seasons, suggesting a contribution of convectively generated waves.

Wavenumber spectra have never been obtained for Venus, which is another terrestrial planet with a thick atmosphere. To reproduce the realistic atmospheric dynamics in Venusian general circulation models (GCMs), the momentum transport by gravity waves and the eddy diffusion associated with gravity wave breaking should be included similarly to Earth’s GCMs. At the present stage, however, the lack of information about the gravity wave spectrum prevents realistic parameterizations of such gravity wave effects (e.g., Lebonnois et al. 2010).

Many observations have been made for Venusian gravity waves. Small-scale periodic structures attributable to gravity waves have been observed in temperature profiles obtained by entry probes (Seiff et al. 1980; Counselman et al. 1980) and radio occultation measurements (Kliore and Patel 1980; Kolosov et al. 1980; Yakovlev et al. 1991; Hinson and Jenkins 1995; Tellmann et al. 2012). The Vega balloon detected vertical wind oscillations over large topographic rises, suggesting propagation of topographically generated gravity waves (Sagdeev et al. 1986; Young et al. 1987; 1994). Recently, the European Space Agency (ESA)’s Venus Express observed gravity wave–like features by a variety of methods. Peralta et al. (2008) identified gravity waves at cloud heights using images taken by the Visible and Infrared Thermal Imaging Spectrometer (VIRTIS) onboard Venus Express. They found that the observed locations of the waves are not correlated with topographic features or the local time, implying that convection in the cloud layer generates the waves. Garcia et al. (2009) detected gravity waves using images of nonlocal thermodynamic equilibrium (non-LTE) CO\(_2\) emission taken by VIRTIS and suggested that the polar vortex might be the source of gravity waves based on the geographical distribution of the waves and the orientation of wave fronts. Markiewicz et al. (2007) found, using cloud images taken by the Venus Monitoring Camera (VMC), that wave trains, located perpendicularly to the direction of cloud streaks, are ubiquitously present at the low-latitude side of the bright polar band. Tellmann et al. (2012) studied small-scale temperature perturbations detected by radio occultation measurements and found that gravity wave activity increases with latitude. They also showed that the greatest wave activity occurs in the vicinity of Ishtar Terra, the highest topographical feature on Venus, and that gravity wave activity in the low latitudes is enhanced near noon.

The present study obtains vertical wavenumber spectra of gravity waves in the Venusian atmosphere for the first time and compares them with those in Earth’s and Martian atmospheres. The spectra are calculated from temperature profiles obtained by the Venus Radio Science experiment (VeRa) performed during the Venus Express mission (Häusler et al. 2006; Pätzold et al. 2007; Tellmann et al. 2009, 2012). Radio occultation has an advantage of providing high-vertical-resolution and high-precision temperature profiles, which are essential to studies of gravity waves. We also estimate the turbulent diffusion coefficients in the Venusian atmosphere based on the wave saturation theory. Section 2 introduces the procedure for obtaining vertical wavenumber spectra, and section 3 gives the result. Section 4 compares the wave characteristics among the Venusian, Martian, and Earth’s atmospheres. Section 5 estimates turbulent diffusion coefficients. Section 6 gives the summary.

### 2. Dataset and analysis procedure

Venus Express was launched in November 2005 and inserted into a polar orbit with the period of 24 h. The
radio occultation experiments are conducted using radio waves with wavelengths of 3.6 (X band) and 13 cm (S band) transmitted by the spacecraft and received on Earth. From the frequency variation of the received signal during each occultation, the angle of the ray bending due to the atmosphere of the planet is obtained as a function of the impact parameter, and then the vertical density profile is obtained by Abel inversion. The vertical profile of the atmospheric pressure is obtained from the density profile under the assumption of hydrostatic equilibrium. Finally, the atmospheric temperature profile is retrieved from the density and pressure profiles by using the ideal gas law. A broad latitudinal coverage from the equatorial region to the polar region is achieved, with a dense sampling in the high latitudes because of the polar orbit of the spacecraft. Details of the measurements are given in Häusler et al. (2006), Pätzold et al. (2007), and Tellmann et al. (2009, 2012).

In this study, we analyzed X-band open-loop data obtained by 286 occultations during the period from July 2006 to June 2010. Unlike the closed-loop receiver, which obtains the frequency and amplitude in real time by means of phase-lock loop technology, the open-loop receiver digitizes the received signal after down conversion to \( \sim 100 \text{kHz} \), and the data are analyzed offline by a phase-unwrapping technique (Imamura et al. 2011). An advantage of using open-loop data is its high-frequency sampling; thanks to the wider spectral coverage, one can examine the influence of the Fresnel diameter as shown below.

The method of analyzing temperature profiles is similar to Ando et al. (2012). A cubic function is fitted to each profile (Fig. 1), and the fitted function is regarded as \( T_0 \). The subtraction of \( T_0 \) from the original temperature profile gives \( T' \), which is considered as a manifestation of gravity waves (Fig. 2). The result given in the next section is virtually unchanged by the change of the order of fitting to 1 or 2, suggesting that the uncertainty in the determination of the background does not have a significant influence on the result. The minimum vertical wavelength resolved in this study is \( \sim 1.4 \text{ km} \) because the diameter of the first Fresnel zone is \( \sqrt{\lambda L} \approx 0.7 \text{ km} \), where \( L \approx 12000 \text{ km} \) is the maximum distance between the transmitter and the point of the ray’s closest approach to Venus during the measurements and \( \lambda = 3.6 \text{ cm} \) is the wavelength of the radio wave (e.g., Häusler et al. 2006; Tellmann et al. 2012).

Temperature profiles in the altitude regions 65–80 and 75–90 km are analyzed separately with a 1024-point fast Fourier transform (FFT) including a Welch window. The gravest Fourier components are excluded from the analysis because of the possible influence of the finite Fourier length, and thus the maximum wavelength analyzed is \( \sim 7.5 \text{ km} \). Figure 3 shows typical examples of the spectra obtained in the northern and southern high-latitude occultations corresponding to small and large \( L \) cases, respectively. The vertical wavenumber is defined by \( k_z = 1/\lambda_z \) in the rest of this paper, where \( \lambda_z \) is the vertical wavelength. A near power-law dependence which is consistent with (1) is seen at low wavenumbers.
and a deviation from the power law is observed at high wavenumbers. The change of the slope occurs around 2.0 km^{-1} (wavelength ≈ 0.5 km) for the northern high-latitude case and 1.0 km^{-1} (wavelength ≈ 1.0 km) for the southern high-latitude case. This is attributed to blurring by the finite vertical resolution, because $L = 5000$ km for the northern high latitudes and $L = 12000$ km for the southern high latitude mean their respective Fresnel diameters are ~0.4 and ~0.7 km. The change of the spectral slope at this wavenumber suggests that the spectrum for wavelengths longer than twice the Fresnel diameter is free from the influence of the vertical resolution; herein the analysis is limited to wavelengths longer than 1.4 km. Obtained spectra are classified into seven latitude bands (90°S–75°S, 75°S–45°S, 45°S–15°S, 15°S–15°N, 15°S–45°N, 45°–75°N, and 75°–90°N) and then are averaged separately. The number of profiles used in the analysis and the background Brunt–Väisälä frequencies are summarized in Table 1.

It should be noted that atmospheric waves other than gravity waves can also contribute to the spectra. The wavenumber-1 Kelvin wave seen at the cloud top is considered to have a vertical wavelength of ~7 km (Del Genio and Rossow 1990), although the wave is considered to be significantly damped above clouds (e.g., Imamura 2006). Thermal tides have a vertical wavelength of ~30 km according to the infrared radiometer measurement in the Pioneer Venus mission (Schofield and Taylor 1983). The wavenumber-1 Rossby wave observed at the cloud top will not influence the spectra since the wave is considered to have a vertical wavelength of ~30 km (e.g., Del Genio and Rossow 1990).

### 3. Result

#### a. General features of the wavenumber spectra

Figure 4 shows the obtained vertical wavenumber spectra divided by $N^4$. The semiempirical spectrum of saturated gravity waves given by (1) (dotted) is also plotted in Fig. 4. The accuracy of each spectrum is evaluated based on the scatter of the spectral densities before averaging; the half-width of the 95% confidence interval is given by 2 times the standard deviation divided by the square root of the number of data and is typically 8% for the number of data of 100 and 12%–25% for the number of 10. Based on this estimate, the spectral features described below are considered statistically significant. The contribution of the measurement noise is evaluated on the assumption that the temperature measurement error is random with a magnitude of ~0.1 K (200 K)^{-1} (Häusler et al. 2006) as

$$
\epsilon = \frac{(\Delta T/\overline{T_0})^2}{\Delta k_z} N^4 \sim 1 \text{ km s}^{-4},
$$

### Table 1. Number of data values and the mean Brunt–Väisälä frequencies for seven latitude bands and two altitude regions. The mean Brunt–Väisälä frequency is obtained by averaging the Brunt–Väisälä frequencies over the altitude region and all the profiles in each latitude bin.

<table>
<thead>
<tr>
<th>Altitude range</th>
<th>Latitude range</th>
<th>Number of data values</th>
<th>Brunt–Väisälä frequency (rad s^{-1})</th>
</tr>
</thead>
<tbody>
<tr>
<td>75–90 km</td>
<td>75°–90°N</td>
<td>122</td>
<td>0.018</td>
</tr>
<tr>
<td></td>
<td>45°–75°N</td>
<td>24</td>
<td>0.018</td>
</tr>
<tr>
<td></td>
<td>15°–45°N</td>
<td>9</td>
<td>0.019</td>
</tr>
<tr>
<td></td>
<td>15°S–15°N</td>
<td>31</td>
<td>0.020</td>
</tr>
<tr>
<td></td>
<td>15°–45°S</td>
<td>49</td>
<td>0.020</td>
</tr>
<tr>
<td></td>
<td>45°–75°S</td>
<td>31</td>
<td>0.018</td>
</tr>
<tr>
<td></td>
<td>75°–90°S</td>
<td>20</td>
<td>0.019</td>
</tr>
<tr>
<td>65–80 km</td>
<td>75°–90°N</td>
<td>119</td>
<td>0.018</td>
</tr>
<tr>
<td></td>
<td>45°–75°N</td>
<td>27</td>
<td>0.019</td>
</tr>
<tr>
<td></td>
<td>15°–45°N</td>
<td>11</td>
<td>0.017</td>
</tr>
<tr>
<td></td>
<td>15°S–15°N</td>
<td>31</td>
<td>0.017</td>
</tr>
<tr>
<td></td>
<td>15°–45°S</td>
<td>47</td>
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<tr>
<td></td>
<td>75°–90°S</td>
<td>20</td>
<td>0.018</td>
</tr>
</tbody>
</table>

![Figure 3](image-url)  
**Fig. 3.** Examples of the vertical wavenumber spectra in the altitude range of 65–80 km. Averages for the latitude bands of 90°–75°S (red) and 75°–90°N (blue) are plotted. Semiempirical saturation curve given by (1) (dotted) is also plotted.
where $\Delta k_z$ is the observation bandwidth that is taken to be equal to the maximum wavenumber of $\sim 0.7 \, \text{km}^{-1}$. This noise level is considered to be low enough.

We see a general tendency that the spectral density decreases with the wavenumber similarly to those in Earth’s and Martian atmospheres. The logarithmic spectral slope is close to $-3$ at high wavenumbers in all spectra and becomes flatter at low wavenumbers. The inflection wavenumber is $k_z \approx 0.2 \, \text{km}^{-1}$ (wavelength $\approx 5 \, \text{km}$) in the altitude range of 65–80 km, while it is not clearly seen in 75–90 km. The observed spectral densities are closer to the saturation curve in the high

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**Fig. 4.** Vertical wavenumber spectra of the normalized temperature fluctuation divided by $N^4$ within the altitude range of 65–80 (red) and 75–90 km (blue) for seven latitude bins. Semiempirical saturation curve given by (1) (dotted) is also plotted for comparison.
latitudes than in the low latitudes. More specifically, for the altitude region 65–80 km, the spectra observed in 75°–90°N, 45°–75°N, and 90°–75°S almost coincide with the saturation curve at \( k_z = 0.3–0.7 \), and 0.2–0.7 km\(^{-1} \), respectively. In other latitudinal bins the spectral densities do not reach the saturation value. For other hand, at high wavenumbers of height due to the density decrease with height. On the heights (wavelengths less than 4 km) the spectral densities are almost invariant in this height region, suggesting that the waves are subject to saturation and other dissipative processes.

At low wavenumbers of \( k_z < 0.3 \) km\(^{-1} \) (wavelength >3 km) the spectral densities are higher at the higher level with the amplitude roughly doubled per two scale heights (~10 km), suggesting amplitude growth with height due to the density decrease with height. On the other hand, at high wavenumbers of \( k_z > 0.3 \) km\(^{-1} \) (wavelength < 3 km), the spectral densities are almost invariant in this height region, suggesting that the waves are subject to saturation and other dissipative processes.

\[ h = 2\sqrt{2d(R + z)}, \]  
where \( R \) is the radius of the planet and \( z \) is the altitude of the tangential point along the ray path. In the Venus Express radio occultation experiments, substituting \( d = 700 \) m, \( R = 6052 \) km, and \( z = 70 \) km into (3) yields \( h \approx 200 \) km.

Figure 6 shows the ratio of the apparent amplitude to the true one for a 200-km-width running average applied to sinusoidal waves with horizontal wavelengths shorter than 4 km, which have been observed by radio occultation. The apparent saturated spectra obtained in this study seem to contradict those arguments. However, considering the wavelength dependence of the radiative damping rate, longer-vertical-wavelength waves might not be strongly dissipated. The amplitude growth with height can be written locally as

\[ T' \propto \exp(\beta z), \]
where the growth rate $\beta$ is given by (e.g., Imamura and Ogawa 1995)

$$\beta = \frac{1}{2H} \frac{1}{2N} \frac{2\pi \lambda_z}{\tau},$$  \hspace{1cm} (5)$$

where $H$ is the scale height, $\lambda_z$ is the horizontal wavelength, and $\tau$ is the radiative relaxation time. Positive $\beta$ means that the amplitude grows with altitude overcoming radiative damping, while negative $\beta$ means that the amplitude decreases with height. Here we estimate $\beta$ as a function of the altitude and the horizontal wave-number for $\lambda_z = 5$ and $2.5$ km. The background atmosphere is assumed to have $H = 5$ km and $N = 0.02$ rad s$^{-1}$. The radiative relaxation time (Fig. 7) is estimated from the model result for $\lambda_z = 7$ km given by Crisp (1989) by assuming that the damping time is proportional to the vertical wavelength. The calculated $\beta$ shown in Fig. 8 is mostly positive for $\lambda_z = 5$ km in the 60–90-km altitude range, while it is negative for $\lambda_z = 2.5$ km. This implies that the amplitude of a gravity wave can increase with altitude and reach the saturation amplitude depending on the vertical wavelength.

4. Comparison with Earth and Mars

We compare the spectra with those in the Martian atmosphere (Ando et al. 2012) as well as those in Earth’s atmosphere. The Martian atmosphere is mostly composed of CO$_2$ like the Venusian atmosphere, and its density near the surface is one order of magnitude lower than the density around the cloud top of the Venusian atmosphere. It is noteworthy that the three planetary atmospheres exhibit similar spectral features that a roughly power-law dependence with an index of around $-3$ is seen on the high-wavenumber side and that this portion of the spectrum is close to the semiempirical saturation model. This suggests that common fluid dynamical processes, including saturation, create the gravity wave spectrum in these atmospheres. The inflection wavenumbers in the Venusian and Martian atmospheres seem to be lower than that in Earth’s stratosphere of 0.2–0.4 km$^{-1}$ (Tsuda and Hocke 2002).

![Fig. 5. Typical vertical profiles of the atmospheric stability in the Venusian atmosphere obtained by Venus Express radio occultation.](image1)

![Fig. 6. Ratio of the apparent amplitude of the wave to the true one for a 200-km-width running average applied to sinusoidal waves with horizontal wavelengths of 200–800 km.](image2)
This implies that the wave amplitude before being suppressed by saturation is larger in the Venusian and Martian atmospheres than in Earth’s middle atmosphere and that the influence of saturation extends to lower wavenumbers in the former than in the latter.

In the short-wavelength range, where the logarithmic spectral slope is near $-3$, the height variation of the spectral density is small in both the Venusian and Martian atmospheres. At longer wavelengths the spectral density tends to increase with altitude in the Venusian atmosphere, while it is almost constant with altitude in the Martian atmosphere. This difference might be explained in terms of radiative damping, which is stronger in the Martian atmosphere than in the Venusian atmosphere because radiative damping is more efficient in optically thinner atmospheres provided that the atmospheric composition is the same (Imamura and Ogawa 1995).

In Earth’s and Martian atmospheres the spectral density is usually maximized in the equatorial region where strong insolation drives intense convection (e.g., Tsuda and Hocke 2002; Creasey et al. 2006). In the Venusian atmosphere, on the other hand, the spectral density is maximized in the high latitudes. If the primary source of Venusian gravity waves is cloud-level convection, then the unusual latitudinal dependence is attributed to the long radiative relaxation time of the Venusian atmosphere, which allows convection to be driven by infrared flux from the thermally homogenized lower atmosphere (Imamura et al. 2014).

5. Turbulent diffusion coefficients

As shown in Fig. 4, the spectra in the low-altitude range in the Venusian high latitudes show a tendency of...
The turbulent diffusion coefficient in the Venusian atmosphere is $D = 2.7–31 \text{ m}^2\text{s}^{-1}$ for the altitude range of 65–80 km. Combined with the value of $4 \text{ m}^2\text{s}^{-1}$ around 60 km estimated from the scintillation of the received signal intensity by Woo and Ishimaru (1981), the diffusion coefficient seems to increase with altitude. Zhang et al. (2012) assumed in their one-dimensional photochemical model an altitude-dependent diffusion coefficient increasing monotonically from $4 \text{ m}^2\text{s}^{-1}$ at 60-km altitude to 100 m$^2$s$^{-1}$ at 100 km; such height dependence is roughly consistent with our estimate.

Lindzen (1981) proposed a parameterization of the turbulent diffusivity induced by a steady, monochromatic, saturated gravity wave on the assumption that the eddy diffusion of heat and momentum prevents the amplitude growth with height and keeps the amplitude at the saturation level. Holton and Zhu (1984) extended this formula to gravity waves damped by the combination of saturation and radiative damping. The diffusion coefficient is given by

$$D = \frac{\lambda^2}{16\pi^3} \left( \frac{N}{\lambda_x} \frac{\lambda^2}{\lambda_x} - \frac{6\pi \partial \sigma}{\partial z} \frac{\lambda_z}{\lambda_x} - \frac{2\pi}{\tau} \right), \quad (6)$$

where $\pi$ is the zonal velocity and $z$ is the altitude. The values of $N$ and $H$ are obtained from radio occultation data, and $\tau$ is obtained from the estimate by Crisp (1989) assuming that radiative relaxation time varies with vertical wavelength linearly. The dominant wavelength $\lambda_z$ is taken to be the inverse of the inflection wavenumber. The possible range of $\lambda_z$ is given by two constraints: the lower limit is determined by considering the effect of the horizontal smoothing in radio occultation measurements as described in section 3e, which is $\sim 300 \text{ km}$, and the upper limit is the wavelength above which radiative damping inhibits amplitude growth with height; that is, $\beta = 0$ (see Fig. 8). The adopted parameter values and the estimated diffusion coefficients are summarized in Table 2. The vertical shear of the background wind is ignored in this study because the background wind profile in the Venusian mesosphere is not well constrained. The possible effect of the vertical shear is mentioned later.

Recent observational results (e.g., Piccialli et al. 2012) show that the vertical shear of the zonal wind is nearly absent around the cloud-top level and that the negative shear becomes larger with altitude reaching approximately $-2 \text{ m} \text{s}^{-1} \text{ km}^{-1}$ in the middle latitudes. Then the turbulent diffusion coefficient can increase by up to 3 times when we consider the contribution of the vertical shear in (6). This is also true for Earth’s case.

Given the estimates above, the characteristic time of vertical diffusion in the Venusian atmosphere is estimated to be $H^2/D = 8.2–95 \text{ days}$ at altitudes of 65–80 km. The time scale of the meridional circulation of $\sim 90 \text{ days}$ around the cloud top evaluated by a diagnosis
of the temperature distribution (Imamura 1997) is within the range of the diffusion time. Thus eddy diffusion seems to make nonnegligible contributions on transport processes in the Venusian atmosphere. Also in Earth’s stratosphere the turbulent diffusion and the meridional circulation equally contribute; the time scale of the former is estimated to be \( H^2 / D = 0.26–2.6 \) years when Fukao et al.’s (1994) diffusion coefficient is adopted, while that for the latter is \( \sim 4 \) years (e.g., Fritts and Alexander 2003).

It should be noted that the Lindzen’s (1981) parameterization of turbulent diffusion is based on a highly simplified situation. Achatz (2007) and Fritts et al. (2009a,b) studied the temporal development of turbulence generated by a breaking of gravity wave and the attenuation of the wave by using nonlinear models. They suggested that wave breaking occurs even when the condition of convective or dynamical instability is not satisfied and that, once a wave begins to break, turbulence rapidly develops and reduces the amplitude of the primary wave to \( \sim 1/3 \) of the original value. This implies that a wave does not propagate over a long vertical range, keeping its amplitude near saturation level; rather, a wave packet rapidly decays once its amplitude reaches a threshold that is lower than the saturation level, and strong turbulence occurs during a short time interval after the wave breaking. Based on these results, the wave field observed in this study is considered as a superposition of waves that are about to break, waves undergoing breaking, and waves significantly attenuated by turbulence. When a certain amplitude level is considered for waves that are about to break, the resultant turbulence in real atmospheres might be stronger than the Lindzen’s (1981) formula.

6. Summary

By using the temperature profiles obtained by the radio occultation measurement on the Venus Express mission, we obtained vertical wavenumber spectra of small-scale temperature perturbations that are thought to be associated with gravity waves. The obtained spectra generally show remarkable similarities to those in Earth’s atmosphere. The spectral density decreases with the wavenumber with a characteristic slope of around \(-3\) at wavelengths shorter than about 5 km, and the spectrum slightly flattens near the long-wavelength end in the high latitudes. The power-law portion of the spectrum is close to the semiempirical spectrum of saturated gravity waves. The remarkable similarity of the gravity wave spectrum among Venusian, Martian, and Earth’s atmospheres suggests that common dissipation processes, including saturation, are at work in these atmospheres.

Temperatures in the two altitude ranges of 65–80 and 75–90 km are analyzed separately, and vertical wavenumber spectra covering wavelengths of 1.4–7.5 km are obtained for each altitude region. The power is larger at 75–90 km than at 65–80 km for wavelengths larger than 3 km, suggesting amplitude growth with height. In the Martian atmosphere, contrary to the case of Venus, the power does not vary noticeably with altitude (Ando et al. 2012). This might be due to weaker radiative damping in the Venusian atmosphere than in the Martian atmosphere.

The gravity wave amplitude is enhanced in the high latitudes in the Venusian atmosphere, while it is maximized in the equatorial region in Earth’s and the Martian atmospheres. This is attributed to the difference of the latitudinal distribution of the convective activity generating gravity waves. Recent studies suggest that the cloud-level convection in the Venusian atmosphere is more intense at higher latitudes (Tellmann et al. 2012; Imamura et al. 2014). On the other hand, it is thought that strong insolation drives more intense convection in the lower latitudes in the Martian atmosphere as suggested in Creasey et al. (2006). Convection in equatorial latitudes is thought to be an important source of gravity waves also in Earth’s atmosphere (e.g., Tsuda and Hocke 2002), although surface topography and stratospheric jet streams are also major sources (Wu and Waters 1996).

The vertical eddy diffusion coefficient in the Venusian atmosphere was estimated based on the wave saturation theory for the first time. The result shows that the diffusion coefficient in the Venusian atmosphere is generally larger than those in Earth’s atmosphere. The characteristic time of vertical diffusion is comparable to that of advective exchange in the Venusian atmosphere, implying that vertical diffusion plays a role as important as meridional circulation in the transport of energy, momentum, and various atmospheric constituents.

We should note that the process determining the wavenumber spectrum is still under debate. For example, Weinstock (1985), Hines (1991), and Dewan (1997) suggested that the frequently observed power-law spectrum might be created by interaction among waves. Scale-dependent radiative damping can also contribute to the power-law dependence (Zhu 1994). Comparing the statistical characteristics of gravity waves on various planets whose atmospheric environments are different from each other might be helpful in constraining the mechanism to create the spectral form.
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