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Vertical wavenumber spectra of three-dimensional winds revealed by radiosonde observations at midlatitude

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Abstract. By applying 12-year (1998–2009) radiosonde data over a midlatitude station, we studied the vertical wavenumber spectra of three-dimensional wind fluctuations. The horizontal wind spectra in the lower stratosphere coincide well with the well-known "universal spectra", with mean spectral slopes of -2.91 ± 0.09 and -2.99 ± 0.09 for the zonal and meridional wind spectra, respectively, while the mean slopes in the troposphere are -2.64 ± 0.07 and -2.70 ± 0.06, respectively, which are systematically less negative than the canonical slope of -3. In both the troposphere and lower stratosphere, the spectral amplitudes (slopes) of the horizontal wind spectra are larger (less negative) in winter, and they are larger (less negative) in the troposphere than in the lower stratosphere. Moreover, we present the first statistical results of vertical wind fluctuation spectra, which revealed a very shallow spectral structure, with mean slopes of -0.58 ± 0.06 and -0.23 ± 0.05 in the troposphere and lower stratosphere, respectively. Such a shallow vertical wind fluctuation spectrum is considerably robust. Different from the horizontal wind spectrum, the slopes of the vertical wind spectra in both the troposphere and lower stratosphere are less negative in summer. The height variation of vertical wind spectrum amplitude is also different from that of the horizontal wind spectrum, with a larger amplitude in the lower stratosphere. These evident differences between the horizontal and vertical wind spectra strongly suggest they should obey different spectral laws. Quantitative comparisons with various theoretical models show that no existing spectral theories can comprehensively explain the observed three-dimensional wind spectra, indicating that the spectral features of atmospheric fluctuations are far from fully understood.

Keywords. Meteorology and atmospheric dynamics (middle atmosphere dynamics; waves and tides)

1 Introduction

The earth atmosphere is always disturbed by gravity waves (GWs), which can exert significant impacts on local and even global atmospheric dynamical and thermal structures via various dissipation processes. The most novel feature of atmospheric gravity wave fluctuations is the so-called "universal" frequency and wavenumber spectra (VanZandt, 1982). Here the term "universal" refers to the spectral structure of wind and temperature fluctuations, especially the spectral slope in the high wavenumber region, and is nearly independent of height, season, and geographical location. Decades of observations by different ground-based and satellite-borne instruments (Dewan et al., 1984; Dewan and Good, 1986; Smith et al., 1987; Fritts and Chou, 1987; Fritts et al., 1988; Tsuda et al., 1989; Wilson et al., 1990; Senft et al., 1993; Eckermann, 1999) further confirmed the universality. These observations revealed that in the large wavenumber region, i.e., the spectral tail region, the spectrum has a nearly universal index of -3. Thus, the value of -3 is also regarded as the canonical spectral slope.

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Numerous efforts have been carried out to explain the observed universal vertical wavenumber spectra. The most widely accepted explanation is the linear instability spectrum theory, which assumes that the spectral intensity at a large wavenumber is restricted by the saturated GW amplitudes (Dewan and Good, 1986). By applying monochromatic GW theory, Smith et al. (1987) determined that the spectral intensity should be proportional to $\frac{N^2}{m^3}$, where m is the vertical wavenumber and N is the atmospheric buoyancy frequency. However, in the low atmosphere, especially in the troposphere, we could not expect the waves to be saturated due to small wave amplitudes there. Besides the saturation spectral theory, Weinstock (1990), Gardner (1994), and Zhu (1994) proposed the diffusive filtering theory, which suggests that the nonlinear interactions among waves acting diffusively upon the wave ensemble are the key physical mechanism in forming the m^{-3} spectral shape. Hines (1991, 1997) proposed the Doppler-spread theory, which predicts that nonlinear interactions due to the horizontal advection lead to the observed m^{-3} tail of the vertical wavenumber spectrum. Based on these GW spectral theories, many GW parameterizations were carried out and implemented in general circulation models to implicitly represent the GW force on background atmosphere. These different GW parameterizations are an important contributing factor with regard to differences among various general circulation models. In addition, the differences between the model results and observations are also non-ignorable. All these spectral theories are essentially based on anisotropic scaling and linear gravity wave dispersion relations. Schertzer and Lovejoy (1985a, b) proposed a 23/9D empirical model based on the classic isotropic turbulence theory. Schertzer and Lovejoy (1985a, b) suggested that the actual strongly anisotropic turbulence may occur dimensionally in the transition between three dimensions (3-D) and two dimensions (2-D), i.e., 23/9D. Under this assumption and by a scaling analysis, they derived the slopes of the zonal wind and vertical wind fields to be -2.4 and -0.6, respectively. All these differences among different theories imply that the essential mechanism accounting for the spectral structure has not been fully understood until now.

Owing to the very small value, it is challenging to directly measure vertical wind with a high resolution. Therefore, most previous observational studies of the vertical wavenumber spectrum were concentrated on horizontal wind and temperature, while there were only very few spectral studies on vertical wind fluctuation from radar (Larsen et al., 1986, 1987; Fritts and Chou, 1987; Yamamoto, et al., 1996) and lidar (Gardner et al., 1998) observations. It is clear that the vertical wind fluctuation spectra were overlooked, and there are not even any systematical statistical results about the vertical wind spectra so far. On the other hand, it should be noted that all aforementioned GW spectral theories are able to reproduce the universal m^{-3} for the horizontal wind and temperature fluctuations at high m, but they predict different spectral slopes for the vertical wind fluctuation spectra. For instance, the linear instability theory (Dewan and Good, 1986; Smith et al., 1987) pointed out that the vertical wind fluctuation wavenumber spectrum has the same spectral index of -3 as those in the horizontal wind fluctuations, while the diffusive filtering theory (Gardner, 1994) predicted that the vertical wind fluctuation should have a different spectral slope. Thus, the spectrum of the vertical wind fluctuation may be an important clue in discriminating which spectral theory is more reasonable.

Accordingly, several questions naturally arise. (1) Does the vertical wavenumber spectrum of the vertical wind have the same or similar structures as those of the horizontal wind and temperature? (2) Does the vertical wavenumber spectrum of the vertical wind also have a universal spectral structure? (3) Can the existing spectral theories account for the observed spectral structures of the vertical wind? Motivated by answering these questions, we present the first statistical analyses on the vertical wavenumber spectra of vertical wind fluctuations from radiosonde observations. The structure of this paper is as follows. In the following section, we introduce the data and the calculation method. Some examples are presented in Sect. 3. In Sect. 4, we present the statistical analyses. In the last section, we give the main conclusions and remark on these.

2 Data description and analysis approach

Twelve-year (1998–2009) routine high-resolution radiosonde data over a middle latitude station, Miramar NAS (naval air station; 32.87° N, 117.15° W), CA, are adopted in this study. These routine radiosondes are launched twice daily at 00:00 and 12:00 UT. In each sounding, pressure, temperature, relative humidity, horizontal winds, and balloon ascent rate are determined via GPS from the ground surface up to the balloon burst height, typically at 25–30 km altitude. For the purpose of statistical significance, in this study the upper limit of the data is set to be 25 km. Moreover, we choose 1 km as the lowest height to avoid some potential measurement uncertainties. The soundings are sampled at 6 s temporal interval; considering that the ascent rate of balloon is about $5\,\mathrm{m\,s^{-1}}$ in the troposphere and increases to about $8\,\mathrm{m\,s^{-1}}$ in the lower stratosphere, these temporal intervals correspond to an irregular height resolution. For convenience, all the data were interpolated onto a 50 m height grid.

In the radiosonde sounding, we cannot measure the vertical wind but Reeder et al. (1999) and Lane et al. (2000) proposed that the perturbation component of the ascent rate can be treated as the vertical velocity fluctuation. By applying this assumption, Zhang et al. (2012, 2013, 2014) directly calculated complete GW parameters and compared them with the results from other methods. The good consistency con-

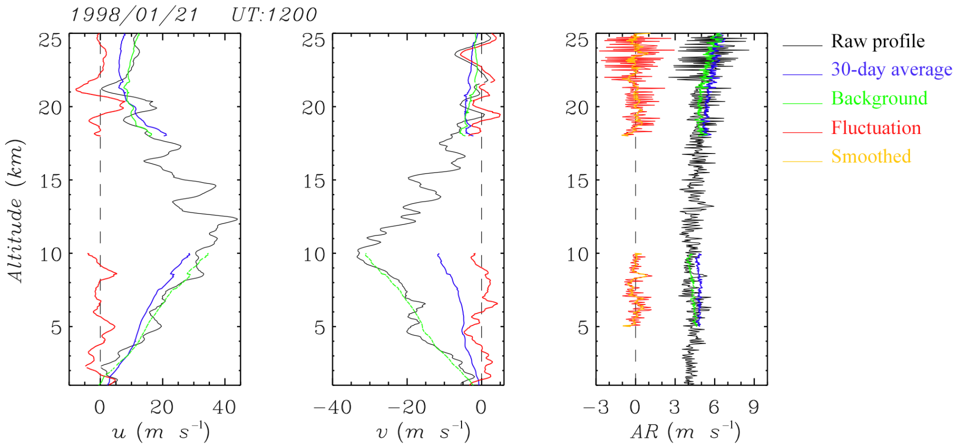

Figure 1. An example of the raw profiles and the extracted fluctuations on 21 January 1998. The left, middle, and right panels represent, respectively, the zonal and meridional winds and ascent rate profiles. The black curves denote the raw horizontal winds and ascent rate profiles. The blue and green curves denote, respectively, the 30-day averaged and total background components, i.e., the 30-day averaged components plus second-order polynomial fitted components. Red curves are the extracted wind fluctuations. The orange curves in the right panel are the smoothed vertical wind fluctuations.

firmed the reasonability of this assumption. Thus, in this study we take the ascent rate perturbations of balloons as the vertical wind fluctuations. Note that Wang et al. (2008) pointed out that the strong convection in the lower troposphere below 5 km may bring some uncertainty in the estimation of vertical wind; thus, for horizontal wind spectra we analyzed the data in the height range of 1–25 km, while for the vertical wind spectra, we only analyze data at heights of 5–25 km.

Some spectral theories have pointed out that the wavenumber spectral intensity depends on the background atmosphere's buoyancy frequency N (Smith et al., 1987). Considering the distinct buoyancy frequency values in the troposphere and lower stratosphere and the sharp transition of N around the tropopause, our wavenumber spectrum analyses are carried out individually in two separate height regions. One is the tropospheric region, from 1 up to 10 km (5 up to 10 km) for horizontal (vertical) wind fluctuation; the other one is the lower-stratospheric region, from 18 up to 25 km.

For extracting the wind fluctuation, firstly, for a specific vertical ascent or horizontal wind raw profile, an average over a 30-day window centered at the specified profile is removed from the raw profile. Secondly, a second-order polynomial was fitted individually to the residual profile in both the tropospheric and lower-stratospheric regions and was also removed. Then the residual perturbations were regarded as the wind fluctuations. It should be noted that although the radiosonde measurement of horizontal wind is believed to have a very high accuracy, the extracted vertical wind fluctuation from the ascent rate may have some bias due to the turbulence and pendulum motion of the balloon. Therefore, a smoothing process for the vertical fluctuation is necessary. We apply an unweighted five-point sliding average (corresponding to a smoothing window of 200 m) to the vertical wind fluctuation profiles before doing the Fourier transform.

Figure 1 presents an example of the raw profiles and the extracted fluctuations on 21 January 1998 to demonstrate the extraction of wind fluctuations. The black curves denote the raw data. For horizontal winds, the zonal wind exhibits a strong jet with a maximum value larger than 40 m s^{-1} at around 12 km. Since the day-to-day variation of horizontal winds is significant, the 30-day averaged (blue curves in Fig. 1) horizontal winds depart apparently to the raw profiles. The green curves in Fig. 1 denote the background components, i.e., 30-day averaged components plus second-order polynomial fitted components. Red curves in Fig. 1 represent the remainder components, i.e., the wind fluctuations. A similar process was performed on the ascent rate to extract vertical wind fluctuation. The ascent rate is around 5 m s^{-1} in the troposphere and increases to more than 6 m s^{-1} in the lower stratosphere. It can be observed that the ascent rate is not as smooth as in the horizontal winds due to the pendulum effects and self-induced motion of the balloon. The smoothing of the vertical wind fluctuations is illustrated by the orange curves in Fig. 1.

Subsequently, we perform a Fourier transform on the wind fluctuation profiles to calculate the vertical wavenumber spectra. The spectra are then characterized by the two quantities slope (S) and spectral amplitude (A). The slope is linearly fitted in the region of wavenumbers lower than 5 cycles km^{-1}, corresponding to a vertical scale of 200 m, which is equal to the smoothing window for the vertical wind fluctuations. The spectral amplitude is defined as the maximum spectral intensity. To improve statistical confidence, monthly averaged slopes and spectral amplitudes are investigated. The monthly averaged spectral amplitudes are calculated as un-

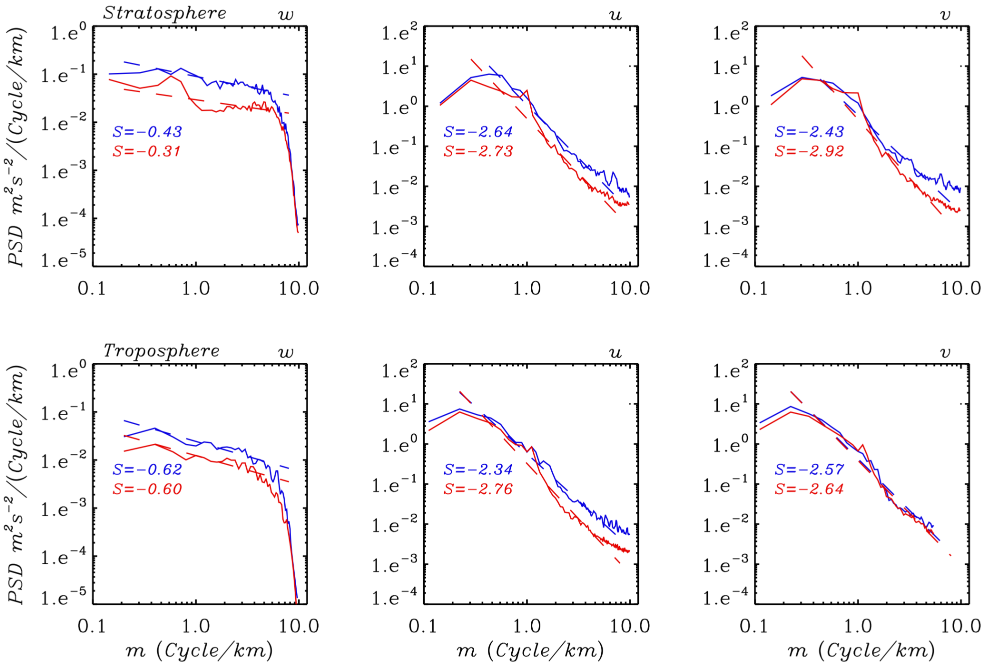

Figure 2. Monthly averaged vertical wavenumber spectra of the vertical (left panel), zonal (middle panel), and meridional (right panel) wind fluctuations in logarithm coordinates at Miramar NAS (32.87° N, 117.15° W), CA. The bottom and upper panels denote, respectively, the spectra in the tropospheric and lower-stratospheric regions. The spectra in a winter month (January 1998) and a summer month (July 2003) are plotted in blue and red curves, respectively. The dashed lines represent the linear fitting, and the corresponding fitted slopes are also marked.

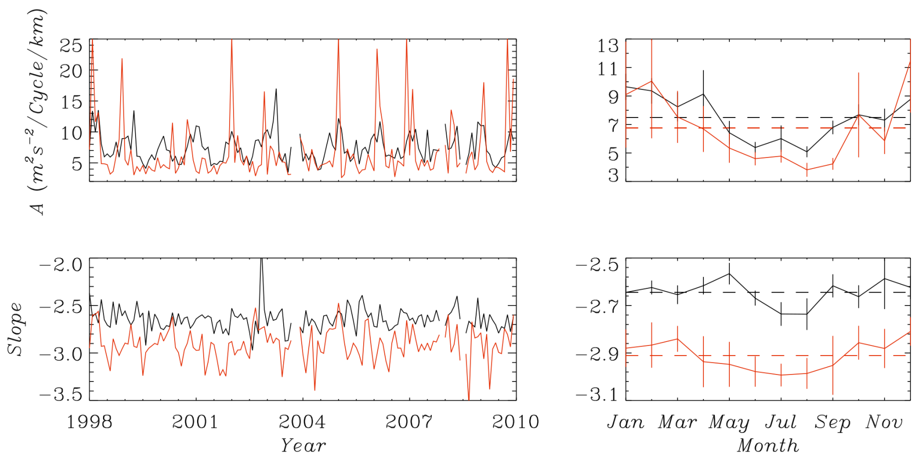

Figure 3. Monthly averaged (left panel) spectral amplitudes (upper panel) and slopes (bottom panel) of the zonal wind fluctuation. The gaps mean no measurements. The annual variations of these spectral parameters are illustrated in the right panel. The black and red curves denote, respectively, the spectral parameters in the troposphere and lower stratosphere. The horizontal and vertical lines in the right panel represent the mean values and standard deviations, respectively.

weighted averages over the amplitudes of the spectra of individual profiles within a calendar month. For the calculation of the monthly mean spectra and their slope, normalized individual spectra are averaged within 1 month. This ensures that all spectra contribute equally to the shape of the mean spectrum (Allen and Vincent, 1995).

3 Example

As an example, we show the monthly averaged vertical wavenumber spectra for three-dimensional wind fluctuations in logarithm coordinates in both the tropospheric and lower-stratospheric regions in Fig. 2. For comparing the spectra in different seasons, we plot the spectra for arbitrarily chosen

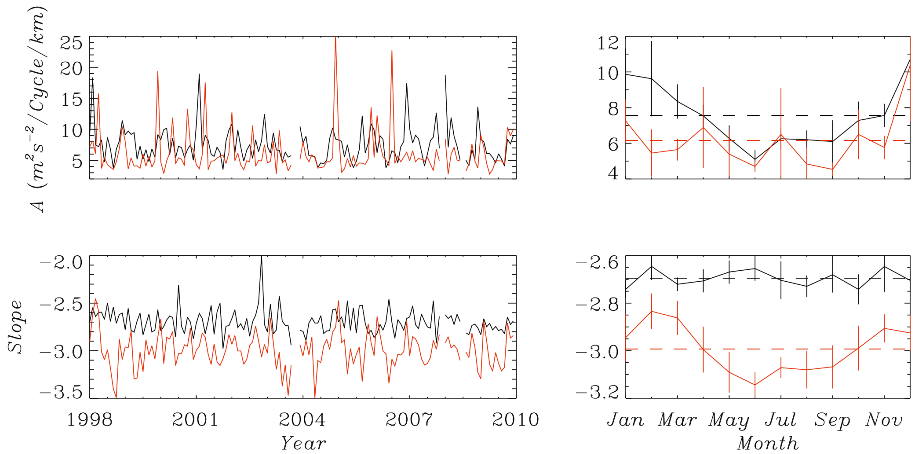

Figure 4. Same as for Fig. 3, but for meridional wind fluctuation.

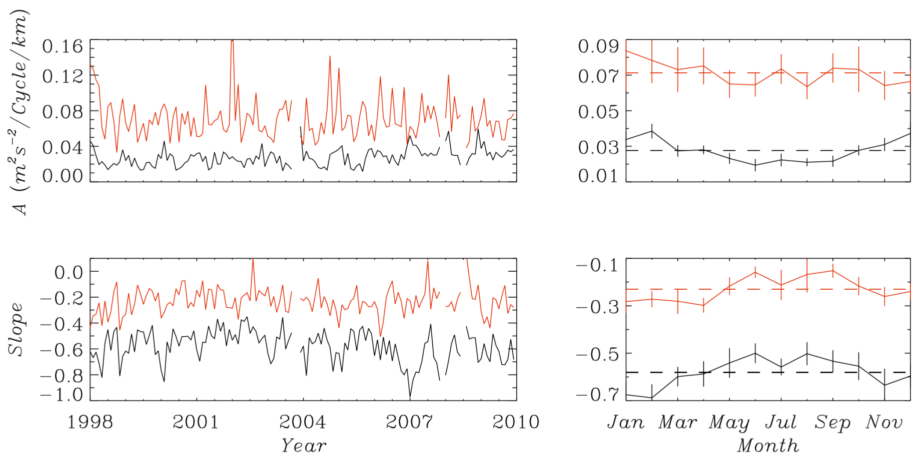

Figure 5. Same as for Fig. 3, but for vertical wind fluctuation.

winter and summer months in blue and red colors, respectively. Figure 2 illustrates that in both winter and summer, the spectral structures of zonal and meridional winds in the lower stratosphere coincide well with the previously reported "universal spectrum". In the spectral tail region, their slopes vary from −2.92 to −2.64, which are close to but less negative than the canonical value of −3. However, in the troposphere, the slopes are obviously less than −3 and vary from −2.64 to −2.34. Notably, the vertical wind fluctuation spectra have a distinct structure from those of the horizontal wind fluctuations. In the spectral region with a wavenumber smaller than 5 cycles km^{-1}, corresponding to the smoothing window width (200 m), the power spectra are much shallower than those of the horizontal wind fluctuations, with small negative slopes varying in a small range of −0.62 to −0.31. In the wavenumber region larger than 5 cycles km^{-1}, the spectral density drops dramatically with wavenumber, which results from the artificial smoothing process rather than having geophysical causes.

4 Statistical results

In the following, we present statistical characteristics of the spectral parameters, including the spectral amplitude (A) and spectral slope. Figures 3, 4, and 5 illustrate the temporal variation of the monthly averaged spectral parameters for zonal, meridional, and vertical wind fluctuations, respectively. The annual variations of the spectral parameters shown in these figures are computed by averaging the results over the same months within the whole observation period.

4.1 Horizontal wind fluctuation spectra

The spectral amplitude of the zonal wind fluctuation in the troposphere varies in the range of 3–17 $\mathrm{m^2\,s^{-2}}$ $\mathrm{cycle^{-1}\,km^{-1}}$, with a mean value of $7.5 \pm 1.2\,\mathrm{m^2\,s^{-2}\,cycle^{-1}\,km^{-1}}$. In the lower stratosphere, the spectral amplitude is more variable, ranging from 2 to larger than 25 $\mathrm{m^2\,s^{-2}}$ $\mathrm{cycle^{-1}\,km^{-1}}$, with a mean value of $6.8 \pm 2.5\,\mathrm{m^2\,s^{-2}}$ $\mathrm{cycle^{-1}\,km^{-1}}$. These spectral amplitudes agree well with previous spectra observations (Zhang et al., 2006). The spectral amplitudes in the troposphere and lower stratosphere exhibit similar seasonal variations, with larger values in winter and smaller values in summer, which is consistent with the seasonal variation of gravity wave energy in the lower atmosphere (Zhang et al., 2010, 2013). In the midlatitude region, strong tropospheric jet-induced shear instability and/or imbalance flow are important GW excitation sources. A stronger jet in winter tends to excite stronger GWs. Except in two winter months (December and February), the spectral amplitude in the troposphere is larger than in the lower stratosphere, which might be the result of tropospheric wind filtering of upward-propagating GWs.

The mean slopes for the zonal wind spectra shown in Fig. 3 are -2.64 ± 0.07 and -2.91 ± 0.09 in the troposphere and lower stratosphere, respectively, indicating that the spectra in the lower stratosphere are steeper. The slopes in the troposphere have only slight seasonal variation but they exhibit evident seasonal variation in the lower stratosphere, with more negative values (steeper spectra) in summer. Seasonal variations of both spectral magnitudes and slopes are consistent with the radiosonde observations in China (Zhang et al., 2006). Generally, the spectral slopes of the zonal wind in the troposphere are systematically larger than -3.0, which is different from the spectral models based on linear gravity wave theory but rather close to the 23/9D anisotropic model (Schertzer and Lovejoy, 1985a, b; Lovejoy and Schertzer, 2013) prediction. In the lower stratosphere they are rather close to the previously discussed universal spectra slope -3.

Both GW source characteristics and propagation effects, such as the effects of background wind, could impact wavenumber spectral structures. In the troposphere, which is believed to be the main source region of GWs, the GWs could not be expected to be saturated and the spectra should be controlled mainly by the source characteristics. Previous observations (Zhang et al., 2012) indicated that the vertical wavelengths of GWs in the troposphere have no obvious seasonal variation. Thus, the spectral slopes of the horizontal wind fields only exhibit a slight seasonality, with mean values evidently less negative than the canonical slope of -3.

The seasonality of the spectral slopes in the lower stratosphere may come from the background wind. At the presented middle latitude station, the strongly seasonally dependent tropospheric jet stream (with maximum magnitude in winter) will exert significant Doppler shifting effects on the upward-propagating GWs. GWs propagating upward from troposphere into the lower stratosphere will pass through the jet stream region, where the zonal wind increases and decreases dramatically with height below and above the jet, respectively. When a GW propagates along the wind that increases with height, due to the Doppler shift induced by the jet, the intrinsic frequency will decreases with height until it is close to the Coriolis frequency f, and then the wave encounters its critical layer and the wave energy will be absorbed by the background wind. In contrast, when a GW propagates upward and against the zonal wind that increases with height, the wave-intrinsic frequency will increase until approaching the buoyancy frequency, and then the wave will be evanescent or even reflected. It is noted that the horizontal winds are more sensitive to horizontal advection activity, which usually has a low frequency and is more easily absorbed by the critical layer. Therefore, in winter months, resulting from the selective filtering of the strong jet, most GWs propagate against the zonal wind above the jet (Zhang et al., 2012). As a result of zonal wind decreasing with height, the intrinsic frequencies of GWs propagating against the zonal wind decrease with height. According to the GW theory, the vertical wavelength will decrease with the decrease in the intrinsic frequency, thus leading to the energy transportation from large vertical scale to smaller scale, in turn yielding a shallower spectrum in winter in the lower stratosphere. Radiosonde observations of gravity waves at the same station (Zhang et al., 2012) also indicated that GWs in the lower stratosphere have shorter vertical wavelengths compared with those in the troposphere. As a result, the horizontal wind spectral slopes have evident seasonal variations, with less negative values in winter.

The spectral features of the meridional wind, including the spectral amplitudes and slopes as well as their seasonal variations, are rather close to those of the zonal wind, suggesting the universality of the horizontal wind fluctuation spectra. The mean spectral amplitudes (slopes) are 7.6 ± 1.4 (-2.70 ± 0.06) and $6.2 \pm 1.8\,\mathrm{m^2\,s^{-2}\,cycle^{-1}\,km^{-1}}$ (-2.99 ± 0.09) in the troposphere and lower stratosphere, respectively. It is suggested that all spectral models based on linear gravity wave theories (Dewan and Good, 1986; Smith et al., 1987; Weinstock, 1990; Gardner, 1994; Hines, 1991, 1997) can predict the horizontal wind spectra in the lower stratosphere well, but the 23/9D anisotropic model (Schertzer and Lovejoy, 1985a, b; Lovejoy and Schertzer, 2013) has an advantage in describing the spectra in the troposphere.

4.2 Vertical wind fluctuation spectra

We have examined all monthly averaged vertical wavenumber spectra of the vertical wind fluctuations and found that all these spectra have a relatively shallow structure compared with those of horizontal wind fluctuations. This indicates that it is a robust and universal spectral structure, which has never been statistically studied before.

The spectral amplitudes of the vertical wind fluctuations are very weak. In the troposphere, the amplitude varies from 0.02 to $0.05\,\mathrm{m^2\,s^{-2}\,cycle^{-1}\,km^{-1}}$, while a much larger spectral amplitude can be observed in the lower stratosphere, with the largest amplitude greater than $0.2\,\mathrm{m^2\,s^{-2}\,cycle^{-1}\,km^{-1}}$ occurring in January 2002. In the troposphere, the spectral amplitudes show a regular annual cycle, with a larger (smaller) value in winter (summer), which is consistent with the gravity wave energy seasonal variation at middle latitude (Zhang et al., 2012). These spectral amplitudes are comparable to those radar observations in the lower atmosphere (Larsen et al., 1986, 1987; Fritts and Chou, 1987) but obviously smaller than the lidar observations in the mesopause region (Gardner et al., 1998). The seasonality of the spectral amplitude in the lower stratosphere is somewhat complex, with a prominent peak occurring in January. The mean amplitude in the lower stratosphere is $0.07 \pm 0.01\,\mathrm{m^2\,s^{-2}\,cycle^{-1}\,km^{-1}}$, which is 2 times larger than the value in the troposphere ($0.03 \pm 0.005\,\mathrm{m^2\,s^{-2}\,cycle^{-1}\,km^{-1}}$). Such a height difference is the opposite of that in the horizontal wind spectra. This departure might be explained by the fact that the horizontal wind fluctuations usually have low frequency (Geller and Gong, 2010; Zhang et al., 2012). When these low-frequency horizontal wind fluctuations propagate upward from the troposphere into the lower stratosphere, they are more easily absorbed by the tropospheric jet-induced critical layer, leading to the weaker horizontal wind spectral amplitudes in the stratosphere compared with those in the troposphere, whereas the vertical wind fluctuations usually have a relatively high frequency (Geller and Gong, 2010; Zhang et al., 2012). Radar observations (Larsen et al., 1986) also revealed that the dominant contribution to the high-frequency disturbance is the vertical velocity component. These high-frequency vertical wind fluctuations are relatively difficult to be absorbed by the tropospheric jet. As a result, we cannot observe the decrease in vertical wind spectral amplitude in the lower stratosphere.

The most notable feature of the vertical wind fluctuation spectra is the spectral slope. In the troposphere, the slopes vary in a small range, from -1.0 to -0.4, indicating the universality of the slopes. The slopes in the lower stratosphere are generally larger than -0.4 and occasionally even reach small positive values. The mean slope in the troposphere is -0.58 ± 0.06, while a less negative value of -0.23 ± 0.05 is seen in the lower stratosphere, indicating that the spectra in the lower stratosphere are shallower. Such a height variation is also in contrast with those of the horizontal wind fluctuation spectra. The slopes in both regions have evident and almost the same seasonal variations, with more negative values in winter. This seasonality is weak and contrasts with that of horizontal wind spectral slopes. This can be partly explained by the seasonally dependent background wind effects. As mentioned above, the vertical wind is associated with the high-frequency convection activity, which is not as easily Doppler-shifted by the strong tropospheric jet in winter to a low frequency and a shorter vertical wavelength but is more easily shifted to a higher frequency and longer vertical wavelength. Thus, we can find that the vertical wind spectra are steeper and accompanied by a larger amplitude in winter.

In our study the vertical wind fluctuation is derived from the balloon ascent rate, which is not solely dependent on the vertical velocity. Other factors, e.g., heating, may also contribute to the variation of the ascent rate and thus may affect the derivation of vertical wind fluctuation. To examine this effect, we sort the statistical results of the vertical wind spectra into two groups according to the local launching times of the balloons, which are 16:00 and 04:00 LT. Figures 6 and 7 illustrate the comparison of the spectral parameters of the vertical winds at different local times in the troposphere and stratosphere, respectively. We can observe from these figures that in the troposphere, the month-to-month variation and the annual variation of the vertical wind spectral parameters at different local times are very similar, and the mean values over the whole observation period at two different local times are also close to each other. In the lower stratosphere, although the annual variation of the spectral magnitudes at different local times exhibits some differences, their mean values and the variation of spectral slopes show considerable consistence at different local times. These consistencies indicate that the vertical wind derivation and the presented vertical wind spectra are reliable.

In addition it is noteworthy that as shown by Lane et al. (2000), the balloon ascent rate containing mainly higher-frequency waves is affected by the tilted trajectories of the balloon. Thus, due to the horizontal drift of balloons, the derived vertical wavenumber spectrum of vertical wind may be contaminated by the horizontal wavenumber spectrum, which is known to be quite shallow. Since at the present midlatitude station (Zhang et al., 2012), a strong tropospheric jet (greater than $40\,\mathrm{m\,s^{-1}}$) occurs in winter, while in summer the zonal wind is considerably weak (less than $10\,\mathrm{m\ s^{-1}}$), we can estimate the influence of the balloon horizontal drift on the vertical wavenumber spectra by comparing the spectra in different seasons. Our results clearly demonstrate that the vertical wavenumber spectra of the vertical wind have the same seasonal variation in the troposphere and lower stratosphere: with smaller slopes in winter and less negative slopes in summer when the background wind is rather weak, which strongly suggests that the shallow spectral structure is the geophysical essence of the vertical wind wavenumber spectrum rather than contamination by the horizontal motion of the balloon. Moreover, the zonal wind in the lower-stratospheric region is less than $10\,\mathrm{m\,s^{-1}}$, which is usually much smaller than that in the tropospheric region (Zhang et al., 2012). Interestingly, we can observe the vertical wind spectra in the lower-stratospheric region are evidently shallower than those in the tropospheric region, which further confirms that the shallow vertical wavenumber spectra of the vertical wind are a realistic geographical phenomenon rather than contamination by the horizontal wavenumber spectra.

The slopes of the vertical wind fluctuation spectra are obviously different from the well-known canonical spectral slope -3 as in the horizontal wind fluctuation spectra. Radar observations (Larsen et al., 1986, 1987; Fritts and Chou, 1987; Yamamoto, et al., 1996) also revealed shallow vertical wind fluctuation spectra. For instance, by analyzing hundreds of wind profiler measurements, Larsen et al. (1986) suggested that the line-of-sight wavenumber spectra of the radical velocity are rather shallower, with slopes between -1.5 and -1. Larsen et al. (1987), Fritts and Chou (1987),

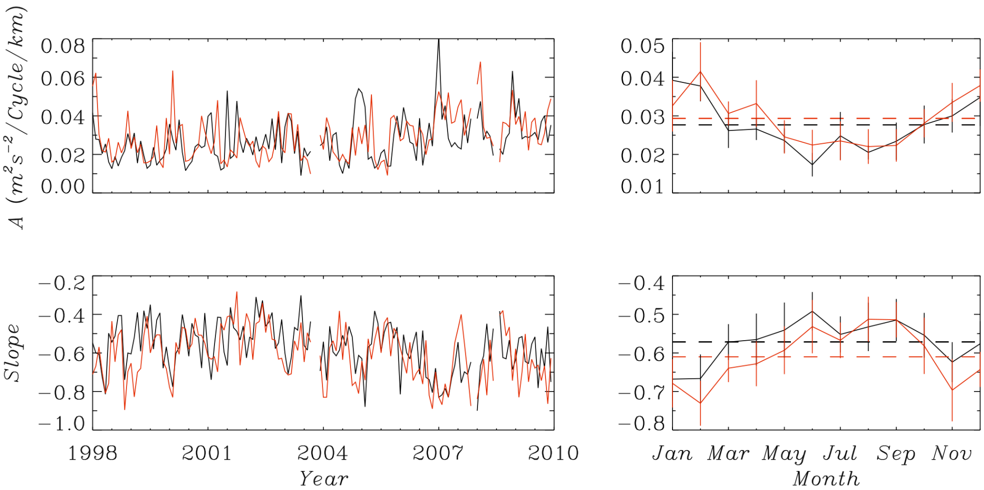

Figure 6. Monthly averaged (left panels) spectral magnitudes (upper panels) and slopes (bottom panels) of the vertical wind fluctuation at different local times in the troposphere. The gaps mean no measurements. The annual variations of these spectral parameters are illustrated in the right panels. The black and red curves denote, respectively, the spectral parameters at 00:00 UT (16:00 LT) and 12:00 UT (04:00 LT). The horizontal and vertical lines in the right panels represent the mean values and standard deviations, respectively.

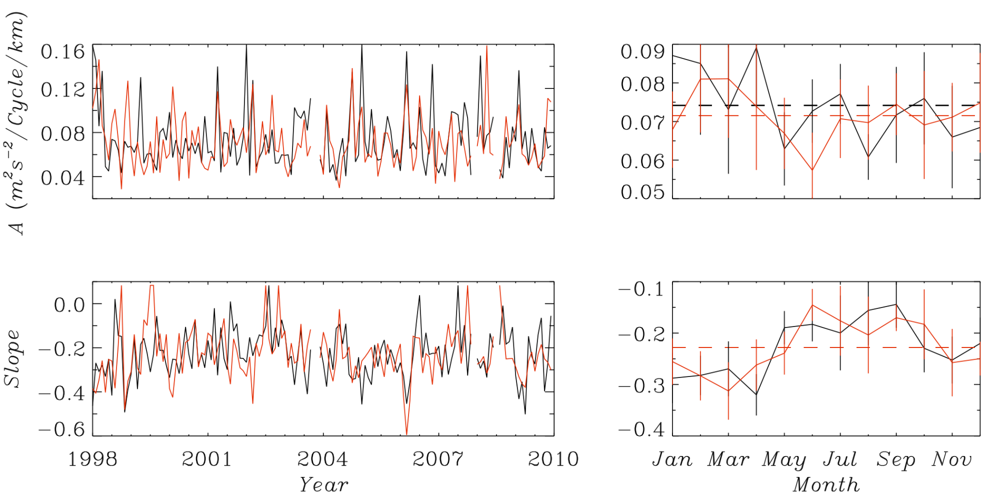

Figure 7. Same as for Fig. 6, but in the lower stratosphere.

and Yamamoto et al. (1996) revealed that even though vertical wind measurements from VHF radar might be contaminated by the horizontal wind, the slopes of vertical wavenumber spectra of the vertical wind are much less negative than -3. Additionally, the shallow vertical wind spectrum with a slope of -1.08 has also been revealed by an 8-night lidar observation (Gardner et al., 1998). This departure could be explained as follows: (1) the vertical wind measurements from VHF radar may inevitably be contaminated by horizontal wind; (2) the radiosonde has a higher height resolution, and the previously reported lidar observations (Gardner, et al., 1998) only had 8 nights of data; (3) the Doppler effect, which influences the ground-based radar and lidar measurements, causing steeper spectra.

These obvious departures between the slopes and their height and seasonal variations of vertical and horizontal wind spectra strongly suggest that the vertical wind fluctuation should obey different spectral laws. However, the linear instability theory (Dewan and Good, 1986) predicted that the spectral slope of the vertical wind had the same value of -3 as those of the horizontal wind. The diffusive filtering theory (Gardner, 1994) points out that the spectral slope of the vertical wind is different from that of the horizontal wind, but the slope of the vertical wind spectrum is predicted to be 1 instead of a small negative value as revealed by our observations. Furthermore, Gardner et al. (1998) considered the effects of the Doppler shift and filtering caused by critical level interactions and suggested theoretically that the slope of vertical wind spectrum might be a small negative value, which will depend on the frequency of the gravity waves, vertical profiles of the background zonal wind and buoyancy frequency, and so on. It is likely that the model proposed by Gardner et al. (1998) could reproduce a slope around -1, whereas our results indicate that the slopes for the vertical wind in the troposphere are generally smaller than -0.6 in magnitude, and in the lower stratosphere these slopes are even smaller in magnitude. Although the 23/9D anisotropic model (Schertzer and Lovejoy, 1985a, b; Lovejoy and Schertzer, 2013) can predict the slope in the troposphere well, its prediction evidently departs from the slope in the lower stratosphere. Hence, no existing theories could fully explain the presented slopes of the vertical wind fluctuation.

5 Conclusions and remarks

By applying 12-year (1998–2009) radiosonde observations over a middle latitude station, Miramar NAS (32.87° N, 117.15° W), CA, we statistically study the vertical wavenumber spectra of three-dimensional wind fluctuations. Notably, this is the first time statistical results for the vertical wind spectra have been processed.

For the vertical wavenumber spectra of the horizontal wind fluctuations, in the troposphere, the slopes of the horizontal wind fluctuations are systematically larger than the universal value of -3, with a mean slope of -2.64, which departs from those spectral models based on gravity wave theories (Dewan and Good, 1986; Smith et al., 1987; Weinstock, 1990; Gardner, 1994; Hines, 1991, 1997) but is close to the 23/9D anisotropic model (Schertzer and Lovejoy, 1985a, b; Lovejoy and Schertzer, 2013) prediction. The spectra in the lower stratosphere are steeper, with a mean slope of -2.91, which is in good agreement with the universal value of -3 and spectral models based on gravity wave theories. Compared with in the troposphere, the spectra in the lower stratosphere are much weaker. Both the spectral amplitudes and slopes in the troposphere and lower stratosphere have similar seasonal variations. The spectra in winter are stronger and shallower. The shallower spectra in winter may come from the Doppler shifting by the strong tropospheric jet in winter, which can lead to the energy transfer of the horizontal wind field from large scale to small scale. The meridional wind spectral characteristics are generally consistent with those of the zonal wind spectra.

The vertical wavenumber spectral structure of the vertical wind fluctuation is rather different from those of the horizontal wind fluctuations. The vertical wind fluctuation spectrum is much weaker than those of horizontal wind fluctuation spectra. Compared with the troposphere, the vertical wind fluctuation in the lower stratosphere is obviously stronger. Such a height variation is in contrast with that of the horizontal wind fluctuation spectra. The most notable feature of the vertical wind fluctuation spectra is that they are much shallower than the well-known universal spectra in the horizontal wind spectra. The spectral slopes of the vertical wind spectra in the lower stratosphere are around -0.23; steeper spectra with a mean slope of -0.58 are observed in the troposphere. These evidently depart from the canonical value of -3 in the horizontal wind fluctuations. Both the spectral magnitudes and slopes in the troposphere and lower stratosphere have consistent seasonal variations, with stronger and steeper spectra in winter. The steeper spectra in winter may also be explained by strong tropospheric jet-induced Doppler shifting in winter, which can lead to the energy transfer of the vertical wind field, which has a relatively high frequency, from small scale to larger scale. The height and seasonal variations of the vertical wind spectra are completely different from those of the horizontal wind spectra.

The present study reveals a new vertical wavenumber spectral structure of the vertical wind fluctuations, which is inconsistent with that of the horizontal wind fluctuations. The inconsistence between the spectra of the vertical wind and horizontal wind fluctuations may be attributed to the following factors. (1) The vertical wind itself may obey a different spectral law, and/or (2) compared with the horizontal wind, the vertical wind is more sensitive to high-frequency motion, which seems to imply that the frequency–wavenumber spectra of GW field should be nonseparable (Gardner et al., 1998).

Although all gravity-wave-associated spectral models can reproduce the observed slopes in the horizontal wind spectra in the lower stratosphere well, they all display a systematical discrepancy from our horizontal wind spectra observations in the troposphere. Furthermore, all these models fail to exhibit the observed vertical wind spectra in both the troposphere and lower stratosphere. The 23/9D anisotropic model can explain both the horizontal and vertical wind spectral structures in the troposphere well; however, the model prediction departs from the observation in the lower stratosphere for both the horizontal and vertical wind spectra.

Finally, from our study we answer the questions listed in the introduction: (1) the vertical wavenumber spectrum of the vertical wind has a rather different structure from that in the horizontal wind; (2) the vertical wavenumber spectrum of the vertical wind does have a nearly universal spectral structure; (3) no existing spectral theories can comprehensively explain the observed three-dimensional wind spectra. Hence, the spectral features of atmospheric fluctuations are far from fully understood. More observations with a high temporal and spatial resolution and large temporal and spatial coverage as well as more theoretical efforts are needed.

6 Data availability

The radiosonde data can be downloaded from the Stratospheric Processes and Their Role in Climate Data Center (ftp://ftp.ncdc.noaa.gov/pub/data/ua/rrs-data/bufr).

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