Large wind shears and stabilities in the mesopause region observed by Na wind-temperature lidar at midlatitude

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[1] Unexpected large horizontal winds and wind shears in the lower thermosphere have been observed by rocket soundings and lidars for decades. From 4 years of the Colorado State University sodium wind-temperature lidar data set (2002–2005; total of ~1600 nocturnal h), we observed an altitude distribution of high wind velocity and wind shears between 80 and 105 km, similar to the results of chemical release experiments. Our lidar data show conclusively that when the observed wind shears are plotted as a function of the squared Brunt-Vaisala frequency, $N^2$, they are below the value corresponding to the Richardson number of 1/4, which is a necessary condition of the onset of dynamic instability. This suggests that large wind shears can be sustained in the region of high static stability, for example, in the lower thermosphere, where large wind shears are often observed by rocket sounding. The full-diurnal-cycle lidar data enable the extraction of tidal wave components with periods of 24, 12, 8, and 6 h, therefore allowing us to reveal the strong correlation of 60% between large wind shears (>50 m s$^{-1}$ km$^{-1}$) and tidal waves. The lidar-measured seasonal variation in $N^2$ and tidal amplitudes in the mesosphere and lower thermosphere are found to be consistent with the difference in altitude distribution of strong wind shears between winter and summer.


1. Introduction

[2] Large winds and wind shears in the mesosphere and lower thermosphere (MLT) region not only play an essential role in the dynamics by causing Kelvin–Helmholtz instability, they also determine the critical levels for gravity wave propagation. Large wind shears may drive electrodynamics and plasma irregularities in the ionosphere as well. For example, the Lorentz force associated with the opposite zonal wind in a strong wind shear has been considered as a dominant mechanism for the formation of sporadic E layers [Whitehead, 1970; Mathews, 1998; Collins et al., 2002]. The same force is also associated with quasi-periodic radar echoes [Larsen, 2000]. Furthermore, large winds and wind shears drive the equatorial electrojet [Hysell et al., 2002]. Enhanced winds and strong wind shears have been observed at 100–110 km over a wide range of latitudes, longitudes, seasons, and local times by rocket sounding via chemical release experiments for the past five decades [Larsen, 2002]. Large wind shears can often exceed 70 m s$^{-1}$ km$^{-1}$ and sometimes reach 100 m s$^{-1}$ km$^{-1}$. These large winds and wind shears have also been verified by other techniques. For example, VHF radar is able to measure the wind velocity between 93 and 110 km by tracking nonspecular meteor echoes: large winds and large wind shears above 95 km have been reported [Oppenheim et al., 2009].

[3] Stronger winds at 100–110 km can be reproduced by the thermosphere–ionosphere–mesosphere–electrodynamics general circulation model (TIME-GCM) with a spatial resolution finer than standard resolution [Larsen and Fesen, 2009]. This implies that the wind-driving waves are well represented in TIME-GCM, with better vertical resolution. However, the strong wind shear observed by Larsen [2002] is still unachievable in any first-principle or empirical models. By comparing the vertical profiles of wind shear and squared Brunt-Vaisala (buoyancy) frequency in the MLT, Liu [2007] argued that large shears could be closely associated with the larger static stability above the mesopause, that is, in the lower thermosphere, where the temperature is cold and the temperature gradient is positive and large.

[4] There are a few dynamical processes proposed to drive the formation of the large wind shears. For example, the large wind shear could be due to the superposition of gravity waves (GWs) and the mean state and/or atmospheric tides [Liu, 2007]. Zhao et al. [2003] and Li et al. [2005a, 2005b] reported case studies of the GW-tide interaction resulting in large shears and dynamic instabilities. These studies hinted that tidal waves set up the environment for GW breaking and induce instability and large wind shear. With the help of full-diurnal-cycle lidar observation and 35 cases of nocturnal large wind shears, we can define tidal-period perturbations and establish a conclusive relation between tides and large wind shears. This is reported in section 3.1. In the E-region
ionosphere, magnetized Rossby waves are also proposed to more effectively drive wind and wind shear compared to neutral dynamics [Kaladze et al., 2007; M. Larsen, personal communication].

[5] The wind-temperature sodium lidar is capable of measuring temperature and horizontal wind fields simultaneously and continuously with a high temporal and vertical resolution. This facilitates investigation of the neutral dynamics associated with large wind shears. For example, both an unusually large wind shear \(100 \text{ m s}^{-1} \text{ km}^{-1}\) and a large temperature gradient \(100 \text{ K} \text{ km}^{-1}\) were observed by the sodium lidar at high latitudes, between \(~85\) and \(95 \text{ km}\) [Fritts et al., 2004; She et al., 2006]. Seven years of wind profile data from the Doppler sodium lidars in New Mexico \(35.0^\circ\text{N}, 106.5^\circ\text{W}\) [Liu et al., 2002] and Hawaii \(20.7^\circ\text{N}, 156.3^\circ\text{W}\) [Franke et al., 2005] has been compared to the results of chemical release experiments. A good agreement is reached below \(105 \text{ km}\) [Larsen and Fesen, 2009]. Because the sodium density decreases dramatically above \(100 \text{ km}\), Na lidars can seldom measure horizontal winds above \(~105 \text{ km}\).

[6] As of 2009 the Colorado State University (CSU) sodium lidar \(41^\circ\text{N}, 105^\circ\text{W}\) had been monitoring MLT dynamics for 18 years [She et al., 2009]. Between 2002 and 2005, more than 1600 h of nocturnal mesopause region temperature and zonal and meridional wind data was collected. Because analysis of stabilities and large wind shears requires high temporal and vertical resolutions [Zhao et al., 2003; Sherman and She, 2006], we use only the nocturnal lidar data for characterization of stabilities and wind shears. The available daytime data have a lower signal-to-noise ratio, so they are not suitable for stability and shear study. However, the 24 h continuous lidar observation at 1 h intervals were utilized to determine the mean state and tides in this study.

[7] In section 2 we present lidar data analysis of wind shears and the squared Brunt-Vaisala frequency. In section 3 we report the general agreement between the CSU lidar wind and the rocket sounding data at \(80\)–\(105 \text{ km}\). This is followed by a discussion of tidal-period oscillation as a possible large wind shear driver in the midlatitude MLT. Since large wind shears are constrained by atmospheric dynamic instability, sustainable large wind shear exists in the region of high static stability. The seasonal and altitude dependence of the large wind shear is addressed. A brief conclusion is given in section 4.

2. Lidar Data Analysis

[8] In May 2002 the CSU sodium lidar upgraded to 24 h continuous monitoring of the middle-atmosphere temperature and horizontal wind with a high temporal and vertical resolution on a campaign basis [She et al., 2003]. Between 2002 and 2005 the lidar was operated in a two-beam arrangement (eastward and northward, \(30^\circ\) off zenith) to simultaneously measure zonal and meridional winds along with the neutral temperature [Yuan et al., 2008a].

[9] To study the large wind and wind shear the 2 min photon count profiles are first integrated over 15 min and vertically smoothed using a Hanning window of 2 km full-width half-maximum, reported at 0.5 km intervals. Figures 1a and 1b show the superposition of zonal and meridional wind

Figure 1. Composite profiles of 15 min averaged (a) zonal wind, (b) meridional wind, and (c) total wind shear magnitude, along with (d) hourly averaged wind shear magnitude. (c, d) The boundary of the largest wind shear at each height is plotted by the bold line. Wind profiles were taken by the Colorado State University (CSU) Na lidar in 2002–2005.
profiles in 2002–2005 at 15 min and 2 km resolution. The large winds measured by the University of Illinois Na lidar in New Mexico and Hawaii and those by measured chemical release experiments in various geographical locations (see Figure 2 of Larsen and Fesen [2009]) have comparable magnitudes. It is recognized that GWs, tides, and planetary waves are superposed on the mean state, from which the local atmospheric stability is determined [Li et al., 2005a]. Although the 2 km vertical resolution and 15 min temporal resolution of the lidar measurement can only cover part of the GW spectrum with periods longer than 30 min, this temporal resolution is adequate for extracting the dynamics of tidal and planetary waves. The photon noise in most cases prevents further increase in time resolution. To extract the tidal-period components during nights with large wind shears, we include hourly mean wind and temperature profiles from both day and night in 2002–2005, as described by Yuan et al. [2008a, 2008b].

[10] The averaged measurement uncertainties due to photon noise for nocturnal temperature and horizontal wind between 85 and 100 km are 3 K and 4.5 m s\(^{-1}\), respectively, for 15 min and 2 km resolutions. At the peak of the Na layer these uncertainties are smaller, typically 2 K and 3.5 m s\(^{-1}\). At the edge of the sodium layer the measurement uncertainty increases dramatically because of the decreasing sodium atom density. In this study we set a maximum allowed uncertainty of 5 K and 10 m s\(^{-1}\) for temperature and horizontal wind, following Sherman and She [2006]. Data with errors higher than this threshold are discarded. The gradient of temperature and horizontal wind is calculated with a vertical interval of 2 km. At the peak of the sodium layer the averaged measurement errors for the temperature gradient and wind shear are 3 K km\(^{-1}\) and 5 m s\(^{-1}\) km\(^{-1}\), respectively. The uncertainty of temperature gradient and wind shear below 85 km and above 100 km are typically larger than 8 K km\(^{-1}\) and 10 m s\(^{-1}\) km\(^{-1}\) because the Na density is low. For the 1 h– and 2 km-resolution lidar data the average measurement errors for the nighttime temperature and horizontal wind are 1 K and 1.5 m s\(^{-1}\), respectively.

[11] With the simultaneous temperature \(T\), zonal wind \(U\), and meridional wind \(V\) measured by the CSU sodium lidar, we can deduce the quantities indicating local stability, that

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**Figure 2.** Lidar data on the night of 14 May 2003: (a) zonal wind (m s\(^{-1}\)), (b) meridional wind (m s\(^{-1}\)), and (c) wind shear (m s\(^{-1}\) km\(^{-1}\)). In each row, raw data are on the left, and reconstructed mean state plus tides is on the right.
is, the Brunt-Vaisala frequency and Richardson number. The static stability is characterized by the square of the Brunt-Vaisala frequency, $N^2$:

$$N^2 = \frac{g}{T} \left( \frac{dT}{dz} + \Gamma_d \right),$$  \hspace{1cm} (1)

where $\Gamma_d = g/c_p$ is the dry adiabatic lapse rate (~9.5 K km$^{-1}$ in the mesopause region), $g$ is the gravitational acceleration (~9.5 m s$^{-2}$), and $c_p$ is the atmospheric specific heat at constant pressure (1004 J K$^{-1}$ kg$^{-1}$). When the atmospheric lapse rate $\Gamma = -dT/dz$ is greater than the adiabatic lapse rate $\Gamma_d$, that is, $N^2$ is negative, the atmosphere becomes convectively unstable. On the basis of measurement uncertainties of temperature and its gradients, the typical error of $N^2$ of 15 min resolution is $1.5 \times 10^{-4}$ s$^{-2}$. The large wind shear of $S = \sqrt{(dU/dz)^2 + (dV/dz)^2}$ can overcome the stability of a stratified atmosphere and induce dynamic instability. The dimensionless Richardson number $R_i$ is used to characterize the dynamic stability:

$$R_i = \frac{N^2}{(dU/dz)^2 + (dV/dz)^2} = \frac{N^2}{S^2}.\hspace{1cm} (2)$$

On the basis of the linear theory, dynamical instability may occur and Kelvin-Helmholtz billows can form when $R_i$ is less than 1/4.

[12] It is worth mentioning that the Richardson number criterion is a necessary but not sufficient condition for dynamic instability [Dutton, 1986]. This Richardson number criterion may not rigorously apply in a number of scenarios, for example, for slantwise (the wind shear is tilted from zenith) instabilities [Hines, 1971; Sonmor and Klaassen, 1997, and references therein] or when the molecular viscosity is important [Liu, 2007]. Because this dynamic instability threshold is derived from the linear theory, the criterion does not apply to nonlinear flows.

### 3. Observation of Large Wind Shears

[13] While the typical tracing time for chemical release experiments is 10 min or less [Larsen et al., 2003], Na Doppler lidars can change the time resolution by varying the period over which measurement profiles are integrated. Along with simultaneous temperature measurement, this allows us to investigate the stability associated with large wind shears. Figures 1c and 1d show the superposition of wind shear profiles in 2002–2005 observed by the CSU Na lidar with the same vertical resolution (2 km) but different time resolutions (15 min vs. 1 h). The envelopes of the largest wind shears at each altitude are plotted by bold lines. Profiles in Figure 1c are calculated from the nocturnal lidar data at 15 min intervals. Table 1 reports the occurrence probability of large wind shears (~40 m s$^{-1}$ km$^{-1}$) observed in Figure 1c at different heights based on data with a resolution of 15 min and 2 km. The occurrence probability for the same large shear magnitude will increase with higher temporal and vertical resolution. As shown, the Na lidar observes more large wind shears (>40 m s$^{-1}$ km$^{-1}$) above 95 km. The fact that the maximum shear magnitude increases with altitude and reaches the peak value at ~100 km is consistent with the vertical profile of the maximum wind shear in the rocket sounding data [Larsen, 2002]. The amplitude of the Na lidar-measured wind shear is, however, lower than that of the chemical release experiment results. Note that the rocket chemical release experiment and Na lidar are indeed very different techniques. Na lidar measurement is a Eulerian measurement; that is, the lidar measures along the line of sight of laser beams in space. The chemical release measurement is a Lagrangian measurement; that is, the wind velocity is determined by following the tracer with time. The chemical release wind is almost instantaneous at a given height, while the lidar wind is averaged over time. Therefore, chemical release experiments enable observation of the strongest wind shears. The uncertainty of lidar measurement is large at the edges of the sodium layer; for chemical release experiments the larger uncertainties occur in the region where the wind shear is large [Larsen et al., 2003]. Although a good agreement between lidar and chemical release measurements has been demonstrated in recent campaigns [e.g., Larsen et al., 2003], it is difficult to trace all of the measurement uncertainties in the rocket sounding wind profiles included by Larsen [2002].

[14] Figure 1d shows the composite shear profiles averaged at 1 h and 2 km resolution. Comparing the results to those with a time resolution of 15 min (Figure 1c), we see a similar altitude dependence but a slightly smaller (<10 m s$^{-1}$ km$^{-1}$) difference in the shear amplitude. This shows that GW perturbations with periods of between 30 min and 2 h contribute to the large wind shears but are not dominant. The periods referred to in this paper are apparent periods relative to the ground. Larsen and Fesen [2009] stated that because the enhanced horizontal wind at 100–110 km can be reproduced using the TIME-GCM with a higher vertical resolution, the enhanced wind is likely driven by large-scale oscillations, that is, tides and planetary waves. If we assume that both enhanced horizontal winds and strong wind shears at 100–110 km are caused by the same mechanism, then large-scale waves like tides could be a significant driver. As presented in the next section, indeed, the very large wind shears we observed were dominantly driven by tidal perturbations.

#### 3.1. Large Wind Shears and Tidal Waves

[15] We employ two data sets for this study. One includes only nighttime data during full-diurnal-cycle observation at 15 min and 2 km resolutions in 2002–2005. Another data set of full-diurnal-cycle observations data at 1 h and 2 km resolutions is used to retrieve the tidal-period oscillations. The tidal amplitude $A$ and phase $\theta$ at each altitude $z$ is determined by linearly fitting the temperature $T$ and hori-
Zonal wind $U$, $V$ to a constant plus the sum of oscillations with 24, 12, 8, and 6 h tidal periods [She, 2004].

$$T, U, V(z, t) = T, U, V(z) + \sum_j A_j T, U, V(z) \cos \left( \frac{2\pi j}{24} (t - \theta_j T, U, V) \right) + R_T, U, V(z, t).$$  \hspace{1cm} (3)

The residue term $R$ represents variations with periods other than tides and shorter than 1 day. As an example, the case study based on data from 14 May 2003 is presented in Figure 2, showing zonal wind, meridional wind, and wind shear. The lidar measurement is shown at the left; the reconstruction of the mean-state plus tidal-period perturbations deduced from the full-diurnal-cycle observations over 24 h is shown at the right. Judging from visual comparison between the two columns in Figure 2, the observed large wind shear can be attributed mainly to the semidiurnal tidal perturbation. There are discrepancies between them at certain heights and times, for example, 92 km at 0600 UT, because internal and inertia GWs are not included in the reconstructed contour. As shown in Figure 2c (left) there is a downward progression band of large wind shear ($40–50$ m s$^{-1}$ km$^{-1}$) lasting almost all night. This is consistent with the wind shears and unstable regions described by Zhao et al. [2003] and Li et al. [2005a]. However, the nighttime-only data set was unable to resolve all tidal-period perturbations. The wind shear, ~$40$ m s$^{-1}$ km$^{-1}$, extracted from the reconstructed superposition of mean-state and tidal perturbations (Figure 2c, right) more or less reproduces the structure of the wind shear in the raw data. The meridional wind (Figure 2b, right) shows an unusually rapid transition between ~85 and 95 km.

Figure 3 provides the corresponding tidal components in the meridional wind deduced from the 24 h continuous data set on 14 May 2003. The tidal amplitudes in the meridional wind at 90 km are reported in Table 2. We can see that in Figure 3b, the phase of the semidiurnal tide (dashed green line) transits rapidly between 85 and 94 km. Thus the vertical wavelength of the semidiurnal tide is approximately 13.5 km. This transient wavelength is much shorter than the typical vertical wavelength of 45–100 km for monthly mean semidiurnal tides [Yuan et al., 2008b]. Table 2 shows that the amplitude of semidiurnal tide (46 m/s) in meridional wind dominates other components in magnitude. As a result, the rapid tidal wind phase transition along with the magnitude enhancement of the semidiurnal tide contributes significantly to this nightlong large wind shear. Moreover, in Figure 2c, following the large shear band around 90 km, there are maxima and minima in shear magnitude. This is due to the constructive and destructive interference among diurnal and semidiurnal tides and other tidal components. This work with tidal analysis confirms previous lidar studies showing that the large wind

![Figure 3](image-url)

**Figure 3.** Tidal components in the meridional wind on 14 May 2003: (a) amplitude (m s$^{-1}$); (b) phase (hours local time).

![Figure 4](image-url)

**Figure 4.** Tidal removed wind shear on 14 May 2003.

<table>
<thead>
<tr>
<th>Period</th>
<th>Zonal wind</th>
<th>Meridional wind</th>
<th>Temperature</th>
</tr>
</thead>
<tbody>
<tr>
<td>24 h</td>
<td>14</td>
<td>18</td>
<td>5</td>
</tr>
<tr>
<td>12 h</td>
<td>23</td>
<td>46</td>
<td>6</td>
</tr>
<tr>
<td>8 h</td>
<td>15</td>
<td>14</td>
<td>6</td>
</tr>
<tr>
<td>6 h</td>
<td>7</td>
<td>6</td>
<td>3</td>
</tr>
</tbody>
</table>

**Table 2.** Characteristics of Tidal Components in Temperature, Zonal, and Meridional Wind at 90 km on 14 May 2003

![Table 2](image-url)
shears follow the downward tidal phase progression in the mesopause region [Zhao et al., 2003; Li et al., 2005a, 2005b].

Figure 1 shows that GWs with periods of between 30 min and 2 h contribute to the shears but are not dominant. Although high-frequency GWs with periods shorter than 1 h presumably carry the most momentum flux [Fritts and Vincent, 1987], a number of GWs with periods of between 5 and 18 h were still found in the study by Hu et al. [2002], using the Na lidar data and hodograph method. To illustrate that GWs with periods longer than 30 min were not responsible for the formation of large shears on the night of 14 May 2003, we remove all of the tidal-period oscillations from the raw data, that is, the left column in Figure 2 minus the right column. The residue wind shear is plotted in Figure 4. The resulting shears are much smaller than the raw data and bear no resemblance to the wind shear structure shown in Figure 2c.

Although we show only one example to illustrate the correlation between large wind shear and tides, more than 60% of the large wind shears (>50 m s⁻¹ km⁻¹) observed on 35 nights are driven directly by tidal-period perturbations at 85–100 km in 4 years. Similar conclusions about these nights can be reached by repeating the same analysis as presented for 14 May 2003 (not shown). These perturbations also have shortened vertical wavelengths and enhanced amplitudes of tides, which suggest that tidal variability is statistically one significant driver for large wind shears. The interaction between tides and transient planetary waves or GWs could be
Figure 7. Brunt–Vaisala frequency squared versus wind shear in winter with a 15 min integration time and 2 km resolution at the altitude of (a) 90–95 km, (b) 95–100 km, and (c) 100–105 km.

Figure 8. Brunt–Vaisala frequency squared versus wind shear in summer with a 1 h integration time and 2 km vertical resolution at the altitude of (a) 90–95 km and (b) 95–100 km.
responsible for the significant tidal variability [Liu and Hagan, 1998; Liu et al., 2000; Hagan and Roble, 2001; She et al., 2004; Li et al., 2005a; Ortland and Alexander, 2006; Liu et al., 2007, 2008]. About 40% of case studies on large wind shear show patterns with little coherence throughout the night, without clear tidal progression. Moreover, Figure 5a shows the local time dependence of the occurrence of large wind shears (>40 m s$^{-1}$ km$^{-1}$) at 95–100 km in winter. The tidal-period variation is clear. The large wind shear is likely to be observed around 1900–2100 local solar time (LST), while it is most unlikely after midnight. Figure 5b gives the local time distribution of large zonal and meridional wind shear (>30 m s$^{-1}$ km$^{-1}$ for each component). The peak of large zonal wind shears occurs at 0500–0600 LST and about 1900–2000 LST; and large meridional wind shears likely occur around 0300–0400 LST. The phase of the semidiurnal tide in zonal wind at 95–100 km in winter is at about 0600 and 1800 LST; the phase of the

![Figure 9](image)

Figure 9. Brunt–Vaisala frequency squared versus wind shear in winter with a 1 h integration time and 2 km resolution at the altitude of (a) 90–95 km, (b) 95–100 km, and (c) 100–105 km.

Table 3a. Occurrence Probability (%) of Dynamical Instability for Different Values of Wind Shear and 15 min Resolution

<table>
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<tr>
<th>Shear (m s$^{-1}$ km$^{-1}$)</th>
<th>90–95 km</th>
<th>95–100 km</th>
<th>90–95 km</th>
<th>95–100 km</th>
<th>100–105 km</th>
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<td>1.2</td>
<td>2.7</td>
<td>2.3</td>
</tr>
<tr>
<td>10—20</td>
<td>0.8</td>
<td>1.1</td>
<td>2.6</td>
<td>4.0</td>
<td>3.6</td>
</tr>
<tr>
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<td>36.8</td>
<td>24.7</td>
<td>24.0</td>
<td>16.4</td>
<td>8.6</td>
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Table 3b. Occurrence Probability of Dynamical Instability for Different Values of Wind Shear and 1 h Resolution

<table>
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<th>Shear (m s⁻¹ km⁻¹)</th>
<th>Summer 90–95 km</th>
<th>Summer 95–100 km</th>
<th>Winter 90–95 km</th>
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<td>0</td>
<td>0</td>
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<td>0</td>
</tr>
<tr>
<td>10—20</td>
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<td>0.4</td>
<td>0.3</td>
</tr>
<tr>
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<td>1.5</td>
</tr>
<tr>
<td>30—40</td>
<td>16.1</td>
<td>3.6</td>
<td>17.6</td>
<td>21.3</td>
<td>4.5</td>
</tr>
</tbody>
</table>

3.2. Correlation Between Large Wind Shear and Brunt-Vaisala Frequency

[19] Besides the driving mechanism of the large wind shear, a question of equal importance is how these large wind shears, \( S \), are sustained in the lower thermosphere. A proposed mechanism is that a large \( S \) can be maintained in the region of large \( N^2 \) without breaking or too much damping [Liu, 2007]. Using equation (2), when \( N^2 \) is large, \( S \) can reach a large value before the local atmosphere becomes dynamically unstable, thus limiting further growth of the wind shear. The maximum sustainable magnitude of wind shear \( S_{\text{max}} \) is then \( 2N \), based on the necessary (but not sufficient) condition for dynamic instability with \( Ri < 1/4 = N^2/S_{\text{max}} \) [Liu, 2007]. For a given value of \( N \) representing the local atmospheric stability, an \( S \) value larger than \( S_{\text{max}} \) may induce dynamic instability and the onset of turbulence. Then the wind shear will be reduced by turbulence mixing until the local stability is recovered. As a result, large wind shears can exist longer in a region with a higher static stability, which depends on the temperature lapse rate. With a statistically large lidar data set at midlatitude, next we show the limitation of \( S \) by \( N^2 \) in different seasons.

[20] In Figures 6 and 7 we show simultaneously observed wind shear \( S \) and \( N^2 \) from nocturnal lidar data at 15 min and 2 km resolution, taken in 2002–2005 at different altitude ranges for summer (June–July) and winter (November–February), respectively. Table 3a lists the corresponding occurrence probabilities of dynamical instability (above the solid blue line) for different wind shears in Figures 6 and 7. The lidar data in summer covers less altitude range than the data in winter because of the lower sodium density in summer. The solid blue line in each Figures 6 and 7 indicate the maximum allowed wind shear as a function of \( N^2 \), that is, \( S_{\text{max}} = 2\sqrt{N^2} \). Above this curve, \( Ri < 0.25 \) and the atmosphere could be dynamically unstable. When the flow becomes unstable and results in the onset of turbulence, the wind shear amplitude decreases rapidly [Kim et al., 2009]. There are few data points above the dashed blue line in either summer or winter. Such occasional cases are to be expected due to the facts that \( Ri < 0.25 \) is only a necessary, and not a sufficient, condition for dynamic instability and that it takes a finite time for turbulence mixing to restore atmospheric stability [Fritts et al., 1996]. Figures 8 and 9 show similar scattered data points of \( S \) versus \( N^2 \) at 1 h and 2 km resolution in summer and winter. And Table 3b lists the corresponding occurrence probabilities of dynamical instability (above the solid blue line) for different wind shears in Figures 8 and 9. The statistics for shears of 40–50 m s⁻¹ km⁻¹ is not significant because there are few samples at 1 h resolution. With a longer integration time (correspondingly, a longer time for large wind shears to relax with turbulence), hardly any observed wind shear exceeds the maximum allowed limit, \( 2N \). Meanwhile, the upper bound of the large wind shears in 1 h is slightly smaller than that in 15 min, which is consistent with Figures 1c and 1d, since wind shears induced by internal GWs with periods of between 30 min and 2 h are excluded. In general, the observed wind shear and temperature gradient must satisfy the condition of static and dynamic stability.

[21] Now we discuss the possible seasonal variation in altitudes where large wind shears exist. As reported in Table 3, in summer, when shears become larger than 30 m s⁻¹ km⁻¹, the chance of dynamical instability rises dramatically. In contrast, in winter at high altitudes (i.e., 100–105 km), the correlation between large wind shear and greater probability of occurrence of dynamical instability is weaker. To explain the seasonal variation of large wind shears, Figure 10 shows the maximum allowed wind shear deduced from mean \( N^2 \) profiles in summer and winter. Following equation (1), the maximum \( N^2 \) exists where the temperature gradient \( dT/dz \) is steep and the temperature \( T \) is low. Only the mean temperatures right above the mesopause meet this requirement in the
MLT region. Because the mesopause is at \( \sim 86 \text{ km} \) in summer and at \( >100 \text{ km} \) in winter at midlatitude [She and von Zahn, 1998; Yuan et al., 2008a], the mean \( S_{\text{max}} = 2 \sqrt{N^2} \) has its maximum at 88 km in summer and at \( >103 \text{ km} \) in winter, as shown in Figure 10. At 85–90 km, the maximum allowed wind shear in summer is \( \sim 10 \text{ m s}^{-1} \text{ km}^{-1} \) greater than that in winter. Because the seasonally integrated \( N^2 \) contains very small measurement uncertainty, this \( 10 \text{ m s}^{-1} \text{ km}^{-1} \) difference cannot be ignored.

[22] To put the seasonal variation of the maximum allowed wind shear, \( S_{\text{max}} = 2N \), in perspective and to investigate further the difference between \( S_{\text{max}} \) and observed wind shears, we plot the superposition of the actual observed wind shears in winter and summer using the 1 h– and 2 km–resolution nocturnal data in Figures 11a and 11b. The mean and the mean plus standard deviation of wind shear as a function of altitude are plotted in Figure 11c. Between summer and winter the mean and standard deviation show little difference. The actual wind shear envelope in winter increases upward, from \( \sim 30 \text{ m s}^{-1} \text{ km}^{-1} \) at 90 km to \( \sim 50 \text{ m s}^{-1} \text{ km}^{-1} \) at 105 km, while the actual shear envelope between 87 and 98 km is \( \sim 40 \text{ m s}^{-1} \text{ km}^{-1} \) in summer. The altitude dependence of the maximum wind shears in Figure 11 is in qualitative agreement with the allowed wind shears deduced from the mean \( N^2 \) profiles in Figure 10, except that the observed maximum wind shears in summer (with values typically \( <40 \text{ m s}^{-1} \text{ km}^{-1} \)) appears to be \( \sim 10 \text{ m s}^{-1} \text{ km}^{-1} \) less than the maximum allowed value. To explain this discrepancy, note that the large wind shear at midlatitude can be driven by one mechanism, such as tides, while its maximum allowed wind shear can be limited by another mechanism, that is, the dynamic stability. Therefore, the observed magnitude of large wind shears should be determined by both mechanisms. We discussed in section 3.1 that tides can contribute significantly to the large wind shears in the MLT. The semidiurnal tidal amplitude in zonal and meridional wind is plotted in Figure 12. Above 90 km the tidal amplitude grows much more rapidly as a function of height during winter than in summer. The vertical wavelength of the semidiurnal tide is also shorter (resulting in larger wind shears) in winter than in summer [Yuan et al., 2008b].

**Figure 11.** Composite of the total wind shear magnitude profiles in 2002–2005 with a 1 h integration time in (a) winter and (b) summer. (c) Mean magnitude and mean plus standard deviation of winter (blue) and summer (red) composite wind shear magnitudes are plotted.
suggests that the large wind shears caused by the semidiurnal tide will grow in magnitude as the altitude increases and can approach its maximum allowed wind shear at a height above 105 km. In summer, however, the growth rate of semidiurnal tidal amplitude and its induced wind shear would be much slower, leading to an observed maximum wind shear above summer mesopause near 90 km much lower than the maximum allowed windshear, 2N.

[23] Since the Na lidar does not have enough signal to measure winds above 105 km, we cannot study the rapidly decreasing large wind shears and N^2 in the thermosphere above 110 km presented in chemical release experiments [Larsen, 2002]. Although the semidiurnal tidal amplitude above the turbopause (~110 km) no longer increases owing to high molecular diffusion in this region [Forbes, 1982], the large wind shear still requires a stable background. Upward from the mesopause, N^2 and the maximum allowed wind shear decrease rapidly because the temperature increases rapidly with altitude owing to absorption of highly energetic solar radiation. Using TIME-GCM simulation, Liu [2007] reports profiles of the maximum allowed wind shear at 80–140 km similar to those reported by Larsen [2002].

4. Conclusion

[24] The CSU Na lidar data set between 2002 and 2005 contains 1600 h of nocturnal observations. Since nighttime observation has a sufficient signal-to-noise ratio, we have analyzed temperature, zonal, and meridional wind profiles at 2 km and 15 min resolutions. With the data acquired in a full diurnal cycle, we also compile profiles of the hourly mean; these data are used to determine the mean-state and tidal-period oscillations. By comparing wind shear profiles with different integration times (15 min and 1 h) and the same vertical resolution, in the 35 cases we investigated, about 60% of the large wind shear formation is driven by long-period waves, for example, local tidal-period perturbations. A case study is presented to illustrate the strong correlation between large wind shears and tides.

[25] The CSU lidar data have demonstrated that the value of the squared Brunt-Vaisala frequency, N^2, confines the maximum magnitude of wind shear. Those wind shears with magnitudes greater than 2N become dynamically unstable and, thus, are prevented from growing. Owing to the corresponding relationship between temperature gradient and N^2, the altitudes with the highest wind and wind shear exist at a few kilometers above the mesopause. Combining the seasonal variations of tidal wave amplitude and mean temperature profiles, we can provide a qualitative explanation of the altitude distribution of large wind shears during summer and winter in the MLT.

[26] The seasonal variation in large wind shear and the convective stability lead to the question of “wave turbopause,” the altitude where the temperature fluctuation maximizes [Offermann et al., 2006]. The wave turbopause height shows clear annual variation at high latitudes: ~90 km in summer and >100 km in winter. We note that this seasonal variation is consistent with the seasonal variation in mesopause altitudes and, thus, the variation of the maximum Brunt-Vaisala frequency. It is therefore likely that both the wave turbopause and the large wind shear are both determined by the atmospheric static stability in the MLT.

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References


Hysell, D. L., J. L. Chau, and C. G. Fesen (2002), Effects of large horizon-