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Altitude dependence of middle and low-latitude daytime thermospheric disturbance winds measured by WINDII

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[1] Thermospheric neutral winds exhibit strong altitudinal and latitudinal variation during geomagnetically quiet and active times. We use daytime middle and low-latitude neutral winds measured by the Wind Imaging Interferometer (WINDII) instrument on board the Upper Atmosphere Research Satellite (UARS) over the 90–275 km height range to study the altitude and season dependent climatology of disturbance winds (i.e., with quiet time patterns removed) in magnetic coordinates. The daytime perturbations winds are generally equatorward and westward and decrease toward the magnetic equator. Both the zonal and meridional components decrease sharply below 120 km and are essentially insignificant below 100 km. The seasonal dependence of the disturbance winds is strongest in the early morning sector. The zonal disturbance winds are predominantly westward with largest magnitudes near 150 km at late afternoon upper midlatitudes (magnetic latitudes 50°–60°) and near 120 km in the early morning winter between 25° and 45°. At upper midlatitudes in the early morning sector, the zonal perturbation winds are eastward above ~120 km in the summer hemisphere and between ~120 and 180 km in the winter hemisphere. Eastward perturbations are also observed near the magnetic equator with largest values near noon at 250 km. The meridional perturbation winds are mostly equatorward and nearly height-independent above ~150 km, with the largest magnitudes occurring in the early morning winter hemisphere. In the late afternoon sector the upper midlatitude meridional disturbance winds are poleward, and there is a small winter-to-summer flow at lower latitudes. In general, the storm-time dependence of the disturbance winds is quite complex. The perturbation winds begin to develop after 0–9 hours and tend to saturate 12–24 hours after storm onset. The development does not change much with altitude, and the seasonal dependence is generally important only at storm times greater than 6–12 hours. INDEX TERMS: 3369 Meteorology and Atmospheric Dynamics: Thermospheric dynamics (0358); 3309 Meteorology and Atmospheric Dynamics: Climatology (1620); 2427 Ionosphere: Ionosphere/ atmosphere interactions (0335); 0358 Atmospheric Composition and Structure: Thermosphere—energy deposition; KEYWORDS: thermosphere, wind, geomagnetic storms, disturbance


1. Introduction

[2] Thermospheric neutral winds exhibit substantial spatial and temporal variations over a wide range of scales, which produce large corresponding changes in the ionospheric plasma density, composition, temperature, and electrodynamics. The thermospheric wind response to geomagnetic storms is generally characterized by enhanced and expanded wind convection in the polar regions (owing to enhanced ion convection) and divergent flows originating at high latitudes (owing to Joule heating-induced expansion). Numerous ground-based optical and radio observations as well as in situ and remote sensing satellite measurements have been used to determine the basic characteristics of middle and low-latitude $F$ region thermospheric winds during geomagnetically quiet and disturbed conditions and have also been used for the development of empirical neutral wind models and for detailed testing of results from general circulation models [e.g., Hedin et al., 1991; Hagan, 1993; Burns et al., 1995; Fuller-Rowell et al., 1994, 1997; Titheridge, 1995; Fesen et al., 1995, Fesen, 1997; Dyson et al., 1997; Emery et al., 1999; Buonsanto and Witasse, 1999; Emmert et al., 2001; Fejer et al., 2002; Zhang and Shepherd, 2002].

[3] The measurement of the motion of the neutral gas is considerably more difficult in the lower thermosphere than
at higher altitudes. As a result, until recently only a very limited database has been used to study the morphology and altitudinal variation of the lower thermospheric winds at middle and low latitudes, particularly during and after geomagnetically disturbed times. Daytime midlatitude neutral winds at altitudes below 150–180 km have been derived from incoherent scatter radar (ISR) measurements at Millstone Hill (43°N, 72°W, apex latitude 54°N) and Saint Santin (45°N, 2°E, apex latitude 40°N) [e.g., Mazaudier and Bernard, 1985; Mazaudier et al., 1987; Salah et al., 1996; Salah and Goncharenko, 2001]. This technique requires the knowledge of the ion-neutral collision frequency and therefore of the neutral composition, which is not well characterized during geomagnetically disturbed times [e.g., Salah and Goncharenko, 2001]. These radar studies have not determined the climatology of the lower thermospheric winds during disturbed times, but they have shown that intense geomagnetic activity (Kp > 6) is required for significant perturbations of lower thermospheric winds at midlatitudes. Several theoretical studies have examined the global response of lower thermospheric winds to geomagnetic activity [e.g., Roble, 1992; Fuller-Rowell et al., 1997; Fesen et al., 1991, 1993], but these model results remain largely untested.

[4] Extensive thermospheric wind measurements over the 90–300 km altitudinal range were made by the Wind Imaging Interferometer (WINDII) instrument on the Upper Atmosphere Research Satellite (UARS) from November 1991 through August 1996. These measurements have been used mostly for extensive studies of the dynamics of the quiet time lower thermosphere [e.g., McLandress et al., 1996; Wang et al., 1997] but recently have also been used for thermospheric disturbance wind studies. Zhang and Shepherd [2000] used WINDII measurements during a large storm to investigate the occurrence of large disturbance winds down to altitudes of about 120 km. Fejer et al. [2000] presented average latitudinal profiles of upper thermospheric perturbation winds (i.e., relative to their quiet time values) derived from WINDII data. Emmert et al. [2001] used middle and low-latitude WINDII measurements averaged over the 225–275 km height range to determine the global climatology and to obtain empirical models for the F region disturbance winds associated with enhanced geomagnetic activity. This later study showed noticeable discrepancies between these satellite data and results from both empirical and general circulation models. Recently, Zhang and Shepherd [2002] examined global scale lower thermospheric winds and O1S emission rates measured by WINDII during geomagnetic storms in April 1993 and February 1994. They showed that at altitudes above 130 km, the winds follow the two-cell ion convection patterns at polar latitudes, while at lower altitudes they change from complex patterns at geomagnetic latitudes above ~50° to tidal structures at lower latitudes.

[5] In this work, we use the entire database of WINDII daytime wind measurements from 90 to 275 km to determine, for the first time, the height dependent climatology of middle and low-latitude thermospheric perturbation winds during disturbed conditions. We use magnetic coordinates because the forcing by both Joule heating and ion drag are oriented in magnetic rather than geographic coordinates. In the following sections, we first summarize the basic characteristics of the WINDII measurements and our experimental methodology. Then, we describe the local time, height, and latitude dependent climatology of the perturbation winds at magnetic latitudes below 60°. Finally, we briefly study their storm-time dependence following a sudden increase in geomagnetic activity.

2. Data and Methodology

[6] The UARS mission was described by Reber et al. [1993]. This satellite is in a nearly circular orbit at an altitude of 585 km, with a 57° inclination and a period of 96 min. In this study, we use measurements from WINDII, a limb-scanning Michelson interferometer [Shepherd et al., 1993] that probes the thermosphere from 42°S to 42°N continuously and up to 72° in alternating hemispheres every 36 days. The wind data are derived from measurements of the 557 nm O1S (green line) and 630 nm O1D (red line) airglow emissions. Daytime green line data cover a height range of 90–300 km, and red line data cover 200–300 km. The temporal resolution is ~30 s (2° at the equator); the height resolution is 3 km below 120 km and 5 km at higher altitudes. The uncertainties in the wind measurements are ~40 m/s and 20 m/s for the upper thermospheric green and red line data, respectively; near 125 km, the green line wind measurements have an uncertainty of ~10 m/s.

[7] The WINDII data used in this study were produced by the V4.98 processing software and consist of ~250,000 daytime height profiles of the horizontal wind velocity in the 90–275 km altitudinal range; these data were obtained on ~600 days from November 1991 through August 1996. The database also includes ~135,000 nighttime profiles, but these are mostly restricted to 90–110 km, with a much smaller number of red line observations at 200–300 km; these data are not extensive enough to perform the analyses described in this paper. The average decimetric solar flux index (F10.7) during the period covered by the database was ~115. The largest (smallest) number of measurements was obtained during December (June) solstice months. High solar activity data, gathered during the early part of the mission, is mostly from December solstice and equinox periods [Emmert et al., 2001].

[8] Our data analysis initially consists of subtracting out the local time, latitude, and season dependent average quiet time (Kp < 3, for an average Kp ~ 1.7) winds for each 3–5 km height level. This technique and other experimental procedures used for obtaining the perturbation winds and for minimizing experimental bias (dependent on the orbital configuration and measurement geometry) were described by Fejer et al. [2000] and Emmert et al. [2001]. Figure 1 shows a sample of the longitudinally averaged quiet time zonal and meridional winds at three height levels and for three seasons: December solstice (November–February), June solstice (May–August) and equinox (March, April, September, October). The contours were obtained by first averaging the data along the satellite orbit (as described by Fejer et al. [2000]) and then smoothing these averages using cubic splines in local time and Legendre functions in latitude. Notice the change of the local time dependence from diurnal variation in the upper thermosphere to a more semidiurnal character in the lower thermosphere, which is consistent with tidal theory [e.g., Forbes and Garrett, 1979].
and with winds derived from radar observations [e.g., Goncharenko and Salah, 1998; Buonsanto and Witasse, 1999].

The values shown in the 250 km case of Figure 1 contain uncertainties on the order of ±30 m/s associated with the orbital configuration of the satellite [Emmert et al., 2001]. The relative local time and latitude dependences are more accurate, and the systematic biases become insignificant below 150 km. Furthermore, because we subtract out the orbit-dependent quiet time averages to obtain the per-

Figure 1. Quiet time (Kp < 3, average Kp ≈ 1.7) average wind patterns derived from WINDII measurements, as a function of local time and geographic latitude. The contour interval is 20 m/s.
turbation winds, the biases are largely removed from the residual winds even in the upper thermosphere.

In order to account for the more complex seasonal dependence of the lower thermospheric winds [e.g., McLandress et al., 1996], we calculated the quiet time reference averages in 2-month seasonal bins rather than the 4-month bins shown in Figure 1. Also, in contrast to our earlier studies, we analyzed the disturbance winds as a function of geomagnetic instead of geographic latitude; this gives a slightly better picture of the disturbance winds, particularly at middle and higher latitudes [e.g., Zhang and Shepherd, 2002]. Because the quiet time reference averages are calculated in geographic (orbital) coordinates, it was necessary to apply a small correction to each residual to account for sampling differences between the geographic latitude dependent and geomagnetic latitude dependent quiet time winds. This correction was obtained by averaging the quiet time perturbation winds as a function of geomagnetic latitude. The adjustment is only important at upper midlatitudes, where it is less than 20 m/s. Note that although we use geomagnetic latitude for analysis, the perturbation wind meridional and zonal components refer to geographic north and east. Finally, our results correspond to longitudinally averaged conditions; longitudinal effects will be discussed in a future study.

3. Results

3.1. General Characteristics of the Height-Dependent Average Disturbance Winds

Figure 2 presents the height dependent climatology of the seasonally averaged zonal and meridional thermospheric disturbance winds measured by WINDII in three local time sectors, for an average \( \Delta Kp \approx 3.7 \) over the average quiet time value (\( Kp \approx 1.7 \)). The curves were obtained by first averaging the perturbation winds at each height level in 10° geomagnetic latitude bins and then smoothing the resulting values with a running average. The error bars correspond to the standard errors of the means obtained by dividing the standard deviations by the square root of the number of independent days (rather than the number of observations).

Figure 2 shows that in general, the zonal and meridional perturbation winds have largest amplitudes at the higher latitudes. Both components decrease sharply below 120 km and become insignificant below 100 km. In the early morning period the upper midlatitude zonal disturbance winds are eastward above 100 km, with largest magnitudes of \( \sim 40 \) m/s near 140 km. Later in the morning, starting at the upper latitudes, the eastward winds turn westward. The westward perturbation winds have largest magnitudes in the afternoon sector near 150 km with values of about 100 m/s and decrease sharply between 140 and 120 km. At geomagnetic latitudes between \( \sim 45° \) and \( 25° \), the zonal disturbance winds are westward at all local times, have largest magnitudes in the early morning sector (peaking near 120 km) and do not change much with height above 140 km. Near the geomagnetic equator, the early morning zonal disturbance winds are eastward above \( \sim 220 \) km, become increasingly westward with decreasing altitude down to \( \sim 120 \) km, and decrease sharply between \( \sim 120 \) and 100 km. In the late morning sector the disturbance winds are eastward at latitudes below \( 25° \), with magnitudes increasing with height above \( \sim 120 \) km. The afternoon zonal perturbation winds are eastward near the magnetic equator and westward at higher latitudes, but below \( 35° \) they have only small magnitudes.

The meridional perturbation winds are generally equatorward at all daytime hours (except in the late afternoon at upper midlatitudes, as described below) and their magnitudes increase with increasing latitude. They are largely height independent above \( \sim 150 \) km and decrease sharply below 120–140 km. In the afternoon sector the midlatitude equatorward perturbations have largest magnitudes at \( \sim 140 \) km and decrease sharply toward the equator.

3.2. Climatology of the Height-Dependent Disturbance Winds

Emmert et al. [2001] showed that in geographic coordinates, the WINDII disturbance winds at \( \sim 250 \) km are largely independent of season except for a winter-to-summer flow in the afternoon sector. For this study we developed an empirical model of the season and \( Kp \) dependence of the perturbation winds as a function of local time, altitude, and geomagnetic latitude. We first sorted the data into three \( Kp \) bins (4 \( \leq Kp \leq 5 \), 5 \( \leq Kp \leq 6.3 \), \( Kp \geq 6 \)) and into summer (defined here as April–September in the Northern Hemisphere and October–March in the Southern Hemisphere) and winter conditions. These broad seasonal bins maximize the statistical significance of our results. We then calculated average disturbance winds in 3-hour local time bins (at 1.5-hour intervals) and 10° geomagnetic latitude bins (at 5° intervals) for each height level. The latitude dependence was treated continuously from the summer hemisphere, across the equator, and into the winter hemisphere. Finally, we performed a weighted least squares fit (with the weights determined by the errors of the mean) of the binned averages to three-dimensional nonperiodic cubic splines in local time, latitude, and altitude. We used node spacings of 6 hours for the local time functions, 15°–30° for the latitude functions (nodes at 0°, 30°, 45°, and 60°), and 20–90 km for the altitude functions (nodes at 90, 110, 130, 180, and 270 km).

Figures 3 and 4 present the model zonal and meridional disturbance winds as a function of latitude and height at three local times and three levels of geomagnetic activity. Results corresponding to summer (winter) are shown on the left (right) side of each plot. We should note that the WINDII database contains about twice as many observations during December solstice than June solstice [Emmert et al., 2001]. Therefore the winter (summer) results shown in Figures 3 and 4 come predominantly from the Northern (Southern) Hemisphere. However, we have verified that the seasonal differences reported below are also observed when the Northern and Southern Hemispheres are analyzed separately.

Figure 3 indicates that the seasonal dependence of the zonal disturbances is most pronounced in the early morning sector at high geomagnetic activity levels. In this local time sector the upper midlatitude eastward winds above 180 km are noticeably larger in the summer hemisphere than in the winter hemisphere, where strong eastward winds are largely restricted to altitudes between 120 and 180 km. The early morning midlatitude winds change from eastward to westward at geomagnetic latitudes of \( \sim 50° \) in the summer hemisphere for an average \( Kp \) of...
Figure 2. Seasonally averaged disturbance winds as a function of height for a disturbed condition of $K_p \geq 4.7$. The top three panels show the zonal component for three different local time sectors, and the bottom three panels show the meridional component. Each panel contains results from several $10^5$ geomagnetic latitude bins (centered on the indicated latitudes). Error bars denote the estimated uncertainty of the means.
about 6.4. In the winter hemisphere, eastward disturbances occur below 60° only between 120 and 180 km, where they again extend down to 50°. The midlatitude westward perturbations decrease in magnitude toward the geomagnetic equator. These westward disturbances are stronger in the lower thermosphere than in the upper thermosphere (particularly during winter), and extend below 110 km, which is also illustrated in the top panel of Figure 2.

By noon, the upper midlatitude zonal winds shift from eastward to westward, but these have magnitudes larger than 20 m/s only above ~50°. They are small for latitudes from 50° to 20°, where they become eastward, with magnitudes increasing toward the equator and with increasing altitude. The equatorial eastward winds increase sharply with $K_p$ and are less evident when averaged over a large range of altitudes, as done by Emmert et al. [2001]. The noon zonal perturbation winds are very small below 120 km at all latitudes. In the late afternoon the upper midlatitude westward disturbances become much stronger and extend more equatorward (to 30° latitude for an average $K_p = 6.4$), particularly in the winter hemisphere.

The disturbance meridional winds shown in Figure 4 are mostly equatorward, extend down to ~120 km and do not change much with altitude above ~150 km. They have largest magnitudes in the early morning sector in the winter hemisphere where they reach values in excess of 120 m/s at 50° for $K_p \approx 6.4$. In general, the equatorward disturbances are latitudinally more widespread in the winter hemisphere, and in the afternoon there is a small winter-to-summer flow across the geomagnetic equator (as indicated by the discontinuity in the colors of the right-hand panels of Figure 4). In the late afternoon, there are strong poleward flows at latitudes above 50°–55°.

Figures 5 and 6 show the local time dependence of the zonal and meridional disturbance winds at four latitudinal sectors, for an average $\Delta K_p \approx 3.7$. Notice the different vertical scales of the right-most panel of each figure. The data with the error bars were obtained by averaging the results from summer and winter conditions in 3-hour overlapping bins at 1.5-hour intervals; the error bars are the
estimated standard errors of the mean. The smooth curves show corresponding results from our empirical model. The perturbation wind patterns shown in Figures 5 and 6 for an altitude of 250 km are similar to the results in geographical coordinates presented by Emmert et al. [2001], except at upper midlatitudes where the longitudinally averaged zonal (meridional) disturbance winds averaged in geomagnetic coordinates are weaker (stronger).

Figure 5 shows that the early morning upper midlatitude eastward perturbations (which are most prominent in the summer) change to westward around 0900–1000 LT at 250 km and at 1100–1200 LT at 130 km. The disturbance winds subsequently become increasingly westward with increasing local time, reaching values on the order of 150 m/s by dusk. At midlatitudes during winter, the perturbations are westward at all local times, are weakest near noon, and are strongest in the early morning sector, reaching values of 75 m/s at a height of 117–130 km. These westward perturbations are observed at heights as low as 108 km. The midlatitude disturbances during summer are weaker than in winter, and they have a weak dependence on local time. Near the geomagnetic equator, the disturbance winds are predominantly eastward with largest magnitudes of 20–30 m/s. Figure 6 shows that the local time dependence of the meridional disturbance winds is largely the same for summer and winter conditions, except at low and midlatitudes in the afternoon sector, where the equatorward perturbations are stronger during winter. Also, the early morning perturbations near 40° in the lower thermosphere are stronger during summer. At upper midlatitudes, the equatorward disturbances shift to poleward around 1400–1500 LT in the summer, and at 1600–1700 LT in the winter.

Figure 7 presents the horizontal vector disturbance winds derived from our empirical model as a function of local time and latitude, for two $K_p$ conditions and four height levels. The upper (lower) half of each panel corresponds to winter (summer) conditions. Note that the scaling is different for the two $K_p$ levels. The patterns for the 250 km, $K_p = 4.4$ case are similar to those of Figure 8 in Emmert et al. [2001], except that the upper midlatitude winds are more equatorward and less westward in the present study, as previously discussed, owing to the use of geomagnetic latitude. Figure 7 shows that the disturbance wind patterns are very similar for the two activity levels, other than an increase in magnitude. However, in the morning sector at summer upper midlatitudes, there is a rotation of the disturbance wind vector toward the east-
ward direction as the $Kp$ level increases. Also, the afternoon perturbation flows are more pronounced at the higher activity level. At a height of 108 km the disturbance winds are very weak, except for westward flows in the morning sector at midlatitudes (especially during winter), and late afternoon poleward flows at summer upper midlatitudes.

3.3. Storm-Time Dependent Wind Responses

[22] In this section we will examine the storm-time dependent response of the disturbance winds following a sudden increase of geomagnetic activity. In particular, we will show that most of the climatological features develop relatively quickly following storm onset and that the climatological large-scale circulation changes described in the preceding section generally develop over the same timescales in the upper and lower thermosphere. We will not discuss in much detail the recovery of the winds following disturbed periods because this recovery is strongly dependent on the initial disturbed conditions (e.g., length of storm), making it difficult to obtain statistically significant results, particularly for the meridional wind component.

[23] We concentrate on the development of the disturbance wind patterns following the idealized storm shown in Figure 8, which corresponds to a change in average $Kp$ index from 1.7 to ~5.5. The following figures show the variations of the disturbance winds as a function of storm time, obtained using a superposed epoch analysis [e.g., Burns and Killeen, 1992; Fejer et al., 2002]. Figure 8 approximately represents the average $Kp$ development of the storms we included in our analysis. We only considered storms for which the $Kp$ index reached at least 5.0, and we defined the onset of a storm as the time when $Kp = 3.7$, which occurs ~3 hours after the steady increase of the index above its initial idealized stationary quiet time value.

[24] Figures 9 and 10 show the average zonal and meridional disturbance winds as a function of storm time, for three local time sectors, four latitude bins, and three height levels. Notice the different vertical scales of the right-most panels of each figure. The symbols with the error bars were obtained by averaging the data in 5-hour storm time bins at 3-hour intervals (we used larger bins for the later storm times). Because of the coarse resolution of the $Kp$ index, our results depict only large-scale changes in global wind circulation and not transient events such as gravity wave surges.

[25] Figure 9 shows that in general, significant westward perturbations begin to develop within 0–9 hours following
storm onset and the saturation of these disturbance winds occurs 12–24 hours after onset. There are no obvious differences between the upper and lower thermospheric initial response times or saturation times. In the early morning sector the disturbances appear after ~6 hours at 55° and 40°, 9 hours at 25°, and 12–21 hours at 5°. The storm development is similar in summer and winter at low latitudes, but at midlatitudes the zonal wind perturbations after ~12 hours become eastward during summer. This is probably a manifestation of the equatorward expansion of the region of eastward winds associated with the high-latitude ion convection (see Figure 3). In this region the winter perturbation winds become increasingly westward and reach magnitudes of 80–120 m/s 18 hours after storm onset (for a steady state $Kp = 5.5$). In the late morning sector the midlatitude summer and winter equatorward winds have comparable magnitudes and saturate ~15–18 hours after the storm onset; the low latitude winds are very small. In the late afternoon sector the summer and winter upper midlatitude equatorward perturbation winds are first nearly identical, but the summer disturbance winds change to poleward at ~9 hours, when the winter winds are in steady state. In this local time sector the largest equatorward winds are observed at magnetic latitudes of ~45° during winter (see Figure 3).

Figure 6. Same as Figure 5 for the meridional component.

[26] Figure 10 shows that the development of the early morning equatorward disturbance winds is similar to that of the zonal winds, with no consistent differences in the timescales of the upper and lower thermospheric responses. The summer and winter midlatitude equatorward winds are nearly identical for the first 9 hours of increased geomagnetic activity, but at later storm times the summer winds begin to abate, and the winter winds continue to increase and reach much larger magnitudes, especially at the upper latitudes and altitudes. At low and equatorial latitudes, the summer and winter perturbations are nearly identical for storm times smaller than ~12 hours, when the summer disturbance winds saturate and the winter perturbations rapidly decrease. In the late morning sector the midlatitude summer and winter equatorward winds have comparable magnitudes and saturate ~15–18 hours after the storm onset; the low latitude winds are very small. In the late afternoon sector the summer and winter upper midlatitude equatorward perturbation winds are first nearly identical, but the summer disturbance winds change to poleward at ~9 hours, when the winter winds are in steady state. In this local time sector the largest equatorward winds are observed at magnetic latitudes of ~45° during winter (see Figure 3).

[27] As mentioned earlier, the recovery of the storm time winds is considerably more complex, since it depends on the initial conditions. Our preliminary results indicate that in the early morning sector the westward perturbations winds develop after 12 hours and they quickly saturate. In this local time sector the equatorial zonal perturbation winds are very small at all altitudes.
Figure 7. Vector disturbance winds from empirical WINDII model, as a function of local time and geomagnetic latitude for two activity levels and four height levels. Summer conditions are shown on the bottom half of each panel, and winter conditions are shown on the top. Note the different scale factor for the two activity levels.
perturbation winds at midlatitudes recover more quickly, within \(~3 \text{ hours}\). Here again, the lower thermospheric time constant is somewhat larger.

4. Discussion

4.1. Comparison With Previous Results

[28] The variation of the daytime WINDII upper thermospheric disturbance winds as a function of geographic latitude was discussed in our previous study [Emmert et al., 2001]. The upper thermospheric results described here, which utilized geomagnetic latitude, are very similar to those of our earlier work, with the following exceptions. First, at upper midlatitudes, the zonal (meridional) disturbance winds are weaker (stronger) in geomagnetic coordinates. Second, the eastward perturbation winds near the equator are much more pronounced than in our earlier paper. This difference is mostly due to the stronger levels of activity used in the present study, coupled with a sharp increase in this feature with increasing \(Kp\). The use of geomagnetic coordinates and not averaging over height also caused small increases in this feature. Finally, the seasonal dependence of the disturbance winds in the early morning sector is much more pronounced as a result of the use of magnetic coordinates. In our previous work we also pointed out that the thermospheric perturbation winds do not change much with altitude above \(~150 \text{ km}\). As illustrated in Figure 3, the largest altitude variation of the disturbance winds above 150 km occurs at upper midlatitudes in the early morning sector during winter for strongly disturbed conditions. However, it is in the lower thermosphere, below 150 km, that the altitudinal variations are strongest.

[29] Daytime thermospheric winds below \(~150 \text{ km}\) have been studied mostly using ISR measurements from Millstone Hill and Saint-Santin. The climatology of quiet time Millstone Hill radar winds at altitudes between 90 and 130 km shows dominant semidiurnal structure at all seasons [e.g., Goncharenko and Salah, 1998] and is consistent with results from WINDII [Zhang et al., 2002]. This tidal oscillation is generally maintained during disturbed times, but the amplitude may increase or decrease depending on the duration and time of the storm event [Salah and Goncharenko, 2001]. Case studies using Millstone Hill radar measurements also indicate that geomagnetic disturbances with \(Kp > 6\) are required for noticeable lower thermospheric wind responses using this radar technique. Intense storms of short duration affect mostly the zonal component without significant changes in the meridional component, but strong long-lasting storms result in large (in excess of \(~60 \text{ m/s}\)) zonal and meridional perturbation winds over the duration of the storm [Salah and Goncharenko, 2001]. The Millstone Hill meridional disturbance winds reverse from equatorward to poleward below \(~120 \text{ km}\), which is consistent with Saint-Santin radar observations [Reddy, 1974; Mazaudier et al., 1987]. No observable perturbation wind effects were detected from radar measurements below 100 km.

[30] The WINDII lower thermospheric disturbance wind patterns, with larger equatorward enhancements in the upper thermosphere and stronger response of the zonal winds at lower heights is consistent with the radar results. The statistical approach using the large satellite database allows us to infer the morphology of the disturbances for smaller levels of geomagnetic activity (down to \(Kp = 4.5\)) by largely removing the complex day-to-day variability. It is also important to remember that during geomagnetically disturbed times the uncertainty of the ion-neutral collision frequency increases the statistical uncertainty of the lower thermospheric radar-derived winds [Salah et al., 1996]. Radar studies have not examined the seasonal dependence of the disturbance winds. Our results clearly indicate larger climatological eastward or smaller westward perturbation winds in the midlatitude summer hemisphere and complex seasonal effects in the meridional component.

[31] The strong zonal disturbance winds observed by WINDII at upper midlatitudes, with a shift from eastward to westward perturbations around 0900–1200 LT, are consistent with acceleration by enhanced and expanded ionospheric convection in the polar regions [e.g., Thayer, 1990; Burns et al., 1995; Müller-Wodarg et al., 2001; Zhang and Shepherd, 2002]. Notice that Millstone Hill (54° magnetic latitude) lies near the magnetic latitude where the early morning WINDII disturbance winds shift from eastward to westward with increasing latitude, which could help to account for the different zonal wind responses of the storms studied by Salah and Goncharenko [2001].

[32] The westward disturbances at midlatitudes in the morning sector are consistent with the westward wind enhancements described by Emery et al. [1999] and Blanc and Richmond [1980]. However, the fact that these westward perturbations are strongest in the lower thermosphere is in apparent disagreement with these two numerical studies, although the authors did not specifically examine the postdawn sector and were not able to compare with lower thermospheric observations. These strong midlatitude westward perturbations are probably associated with the transport of angular momentum from higher latitudes.

[33] The WINDII midlatitude meridional disturbance winds show evidence of small poleward flows below 120 km in the morning sector, especially for strong activity levels and prolonged disturbances, which is consistent with Millstone Hill and Saint-Santin radar observations [Reddy, 1974; Mazaudier et al., 1987; Salah et al., 1996] and theoretical model predictions [Blanc and Richmond 1980]. The poleward winds below 120 km are possibly indicative of the return flow of the stronger higher-latitude equatorward motions.

[34] The most striking characteristic of the low-latitude disturbance winds is the large eastward perturbation of the
The occurrence of equatorial eastward neutral wind and ion drift perturbations during geomagnetically active times was predicted by the ionospheric disturbance dynamo model of Blanc and Richmond [1980]. Equatorial and low-latitude $F$ region ion drift perturbations derived from incoherent scatter radar observations [Fejer, 1997] are in good agreement with the predictions from this theory. The WINDII results indicate that the eastward disturbance dynamo ion drift is confined to the region near the equator, which maximizes at the magnetic equator. The occurrence of equatorial eastward neutral winds and ion drift perturbations during geomagnetically active times was predicted by the ionospheric disturbance dynamo model of Blanc and Richmond [1980]. Equatorial

![Figure 9. WINDII average zonal disturbance winds as a function of storm time during the development phase, corresponding to the storm conditions shown in Figure 8. The solid curves were calculated by taking a running average of the raw values. Error bars denote the estimated uncertainty of the mean. Note the different vertical scale of the right-most panels.](image)
magnetic equator, probably owing to the sharp decrease of the corresponding eastward neutral wind perturbations below $\sim 200$ km with increasing magnetic latitude.

4.2. Storm-Time Dependence

[35] Several experimental and numerical modeling studies have examined the time-dependent response of middle and low-latitude upper thermospheric winds to high-latitude forcing during disturbed conditions [e.g., Roble et al., 1987; Crowley et al., 1989; Burns et al., 1995; Fuller-Rowell et al., 1994, 1996; Reddy and Mayr, 1998]. The middle and low-latitude thermospheric perturbation winds are initially due to storm-generated fast-traveling atmospheric waves (phase velocities of $\sim 600$ m/s), which propagate to lower

\[\text{Figure 10. Same as Figure 9 for the meridional component.}\]
latitudes as surges and penetrate the opposite hemisphere, carrying with them changes in the global circulation. In the second phase the zonal perturbation winds develop via the Coriolis interaction, which limits the meridional dynamical response. The wind perturbations maximize in the nighttime sector and at the geographic longitude of the magnetic pole [e.g., Fuller-Rowell et al., 1996].

[36] Only a few studies have examined the storm-time dependence of the lower thermosphere winds. Mazaudier et al. [1987] reported a time lag of 3–4 hours between the sudden increase of the AE index and the occurrence of lower thermospheric winds over Saint-Santin. They also showed that the Saint-Santin radar observations of storm-time electric fields and neutral winds during two disturbed periods were generally consistent with the predictions of the two-dimensional model developed by Blanc and Richmond [1980]. Millstone Hill observations indicate a prompt response of the ion drift velocity and a delayed response in the neutral wind and that in the lower thermosphere the zonal wind component is the most sensitive component to geomagnetic disturbances [Salah et al., 1996; Salah and Goncharenko, 2001]. During the large geomagnetic storm of September 1998, the maximum response of the lower thermospheric daytime horizontal winds occurred ~13 hours after the onset of the storm impulse [Salah and Goncharenko, 2001].

[37] Our longitudinally and seasonally averaged data are not well suited for studying the transient response of the middle and low-latitude thermospheric winds to the fast-traveling atmospheric surges, but they provide an insight to the more slowly developing changes in the atmospheric circulation due to enhanced high-latitude convection and energy deposition. However, the initial response times (0–9 hours) of the WINDII disturbance winds are generally consistent with the results of previous experimental and modeling studies of upper thermospheric storms. The mid-latitude zonal wind results of Blanc and Richmond [1980] and Emery et al. [1999] suggested a 1-hour delay between the upper and lower (125 km) thermospheric response, which is consistent with the WINDII result of similar response times in the upper and lower thermosphere; note that the time resolution of the WINDII results is at best 3 hours, so a 1-hour delay is possible.

[38] The saturation times (12–24 hours) of the WINDII daytime perturbation winds do not have clear altitudinal dependence, and the midlatitude zonal wind saturation times appear to be consistent with the lower thermospheric predictions of the disturbance dynamo model of Blanc and Richmond [1980]. However, the theoretical study indicated shorter saturation times (4–6 hours) in the upper thermosphere. The results of Emery et al. [1999] suggested that the maximum westward response for lower latitudes at 125 km occurs after 24 hours, whereas the low latitude westward disturbances from WINDII are very strong within 24 hours. It should be noted that accurate saturation times are difficult to determine in both data and model results, since the response frequently continues to increase at more gradual rates after the initial curtailment. Also, although gravity wave propagation is predicted to be slightly slower in the lower thermosphere [e.g., Balthazar and Moffett, 1997], Fuller-Rowell [1995] pointed out that in the lower thermosphere, nonlinear advection provides an additional mechanism for transporting the wind field.

[39] The WINDII data indicate that for the first 6–12 hours following the increase of geomagnetic activity the disturbance winds are largely season independent, except for the late afternoon upper midlatitude zonal component. Numerical simulations of 12-hour storms by Fuller-Rowell et al. [1996] also indicate similarities (at most heights) between surges in summer and winter and in the Northern and Southern Hemispheres, whereas the strongest effects are predicted to occur during equinox. At later storm times, the WINDII data show large seasonal effects on the disturbance winds at magnetic latitudes higher than ~35°, with the winter perturbations being generally more equatorward/westward. The stronger overall westward perturbations during the early morning winter are consistent with the 0000 LT numerical simulations of Emery et al. [1999]. At lower latitudes the seasonal effects are weaker and mostly restricted to the early morning meridional perturbation winds.

5. Summary

[40] We have used daytime middle and low-latitude thermospheric neutral winds measured by WINDII to study the latitude, altitude, and season dependent climatology and storm time response of disturbance winds over the height range of 90–275 km. These perturbation winds are generally equatorward and westward and have magnitudes that increase with magnetic latitude. The meridional perturbation winds have largest magnitudes in the morning sector, at upper midlatitudes, and at altitudes above ~180 km. The zonal disturbance winds are mostly westward with largest magnitudes near 150 km in the upper midlatitude afternoon sector. Both components decrease sharply below 120 km and are negligible below 100 km. The seasonal dependence is strongest at midlatitudes and in the early morning hours, where the meridional perturbation winds are mostly equatorward with nearly height-independent magnitudes above ~150 km, and the height variation of the zonal winds is season dependent. The storm-time dependence of the disturbance winds is complex, varying with latitude and season, but does not change much with altitude. The perturbation winds begin to develop after ~0–9 hours and generally tend to saturate 12–24 hours after storm onset. The initial development of the disturbances is largely independent of season. The recovery of the thermospheric winds following geomagneticquieting is even more complex, since it also varies with the duration of the storm. These WINDII results are basically consistent with and extend the results of earlier studies of thermospheric winds during geomagnetically disturbed periods.

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