Altitude variations of the horizontal thermospheric winds during geomagnetic storms

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It is generally assumed that viscosity smooths out vertical gradients of the horizontal thermospheric winds in the upper thermosphere, and thus observations of neutral winds at one height can be used at other altitudes in this region. In this paper we present neutral wind simulations of the May 1997 geomagnetic storm using the Coupled Magnetosphere-Ionosphere-Thermosphere (CMIT) model. The model results show that during quiet periods, the assumption of a shearless vertical profile of the horizontal winds is generally valid in low and middle latitudes, although vertical shears do occur in wind profiles in the upper thermosphere in some locations at higher latitudes. During disturbed periods, large variations in the vertical profiles of the upper thermospheric winds are seen globally in the model simulations. A diagnostic analysis of the forcing processes in the neutral momentum equations shows that (1) during quiet time, there are shearing forces, most noticeably the pressure gradient and ion drag, in the upper thermosphere that result in a net momentum forcing that changes with height; this induces altitude variations in the wind profiles at high latitudes and sometime even at middle latitudes. (2) During storm time, momentum advection, which is relatively weak during the quiet time, becomes a dominant force globally. Pressure gradient forces are also significantly enhanced. Ion drag, on the other hand, can be enhanced or suppressed, depending on the location of positive or negative effects. All these forces exhibit significant altitude variations that lead to a net force that is greatly enhanced and has large vertical shears. This produces globally enhanced neutral winds that vary with height. (3) Viscosity is less important than other forcing processes during both the quiet and storm periods and thus cannot prevent shears from occurring in the vertical profiles of the horizontal winds. Viscosity has an insignificant effect on vertical shears that change with height linearly. It, however, does restrict vertical shears that vary nonlinearly with height. The effectiveness of the viscosity in restricting such shears depends on its magnitude. In our simulations, viscosity is weaker than other forcing processes and thus is a relatively slow process, so it will take a few hours for viscosity to reduce such shears.


1. Introduction

The notion of a lack of vertical variations in the horizontal neutral winds has a long history. Rishbeth [1972] used the Jacchia [1965] neutral atmospheric model and a simple ionospheric model to estimate thermospheric winds and their vertical profiles at middle latitudes under geomagnetically quiet conditions. Nonlinear advection terms in the neutral momentum equations were neglected in the analysis to simplify the calculation. They concluded that vertical gradients of the horizontal neutral winds in the upper thermosphere are smoothed out completely by molecular viscosity because it increases with altitude as neutral densities decrease. This result has since been used widely in interpreting observations and constructing models [e.g., Bailey and Sellek, 1990; Titheridge, 1991; Crickmore, 1993; Davis et al., 1995]. They all assumed that there were no altitude variations in the horizontal neutral winds in the upper atmosphere, under both quiet and active conditions. Hedin et al. [1988, 1991] also used this assumption to construct a widely used empirical horizontal wind model (HWM) which gives uniform vertical profiles of the horizontal winds in the upper thermosphere.

Killeen et al. [1982] showed that thermospheric meridional winds observed by the Dynamics Explorer 2 (DE2) FPI varied very little with altitude under geomagnetically quiet conditions. Wharton et al. [1984] also reported that the seasonal-averaged zonal winds measured in situ by the DE2 WATS instrument increased slightly in magnitude
with increasing altitude but preserved their directions in low and middle latitudes. The DE2 wind measurements used in these studies, however, were obtained in a very limited altitude range, and the results were averaged over many satellite orbits. Thus any vertical variations in the horizontal winds were likely to be almost completely smoothed out. Fejer et al. [2000] and Emmert et al. [2001, 2002, 2004] studied the climatological characteristics of the neutral wind observed by the UARS-Wind Imaging Interferometer (WINDII) instrument by binning the data set for different Kp cases and seasons. Their results, which were also averaged over a number of vertical profiles, showed that the horizontal winds did not change significantly with height in a statistical sense.

However, spatial and temporal changes in the thermospheric winds are expected to be of short duration, so averaging removes any vertical shears that might exist. In fact, there are reasons to believe that vertical gradients of the horizontal winds can occur in the thermosphere, since momentum and energy inputs from the magnetosphere at high latitudes during geomagnetic storms are both dynamical and localized. It is known that significant amount of energy is deposited at high latitudes during storms by means of neutral Joule heating [e.g., Killeen, 1987]. This Joule heating increases neutral temperatures and changes neutral densities at all altitudes above about 120 km. The most significant impact of Joule heating, however, occurs in the upper thermosphere [Wang et al., 2004; Thayer and Semeter, 2004]. This produces horizontal pressure gradients that increase with altitude. These pressure gradients are greater than those predicted by the empirical Jacchia model which tends to give smoothed distributions of neutral density and temperature. In addition, the neutrals are also accelerated at high latitudes through ion-neutral collisions or ion drag [Killeen and Roble, 1984]. These pressure gradient and ion drag forcings drive high-speed neutral winds which can then be transferred to middle and lower latitudes by momentum advection and the Coriolis force. Because of the altitude distributions of neutral and electron densities, ion-neutral coupling maximizes in a specific altitude range [Richmond and Matsushita, 1975; Kwak and Richmond, 2007] that may introduce altitude variations to the neutral winds.

In fact, observations have indicated that vertical gradients of the horizontal winds exist in the upper thermosphere. Vertical shears in horizontal wind profiles were observed in the upper thermosphere at lower latitudes by Burnside et al. [1983, 1991] and by Schlesier and Buonsanto [1999a, 1999b] at middle latitudes. Gravity waves are also known to produce neutral wind shears in the upper thermosphere [Richmond and Matsushita, 1975; Millward et al., 1993]. Schlesier and Buonsanto [1999a, 1999b] also had to introduce vertical shears in the horizontal winds in their model to simulate the observed ionospheric electron density profiles at Millstone Hill. Zhang and Shepherd [2002], using UARS-WINDII data, showed that large vertical shears of the zonal winds occurred during a major storm in the local morning hours in the thermosphere. Although their wind profiles were used only up to 200 km, there were, however, no indications that the observed large vertical gradients were decreasing with altitude at this top height. These results suggest that there are greater vertical variations in the horizontal winds on a global scale than previous theoretical analysis have predicted. Thus there is a need to systematically address the discrepancies seen between the theoretical analysis and observations, and to detail the physical processes involved in defining the vertical profiles of the horizontal winds under both quiet and disturbed conditions.

In this paper, we investigate storm-time variations of thermospheric wind profiles that were simulated by the Coupled Magnetosphere-Ionosphere-Thermosphere (CMIT) model. We also examine storm-time changes of various forcing processes, including the viscous force, to study their response to high-latitude energy and momentum inputs, and their relative contributions to the storm-time perturbations seen in the modeled neutral wind profiles. The paper begins with a brief description of the CMIT model in Section 2, which is followed by presentations of simulation results in Section 3 and discussions of the results in Section 4. Summaries of main findings of this work are given in Section 5.

2. The Coupled Magnetosphere-Ionosphere-Thermosphere Model

The CMIT model consists of two codes: the Lyon-Fedder-Mobarry (LFM) global magnetospheric MHD code and the Thermosphere-Ionosphere Nested Grid (TING) model. The LFM code solves ideal MHD equations in the magnetosphere using the Partial Interface Method on a distorted spherical mesh and Yee type grid [Lyon et al., 2004]. The TING model solves time-dependent, three-dimensional, coupled equations of momentum, energy, and mass continuity for the neutrals, and O′ transport on both global coarse and local fine grids [Roble et al., 1988; Wang et al., 1999]. In the CMIT model, these two codes are coupled by exchanging parameters across their interfaces through an M-I coupler module. These parameters include field-aligned currents, auroral precipitation from the LFM code, and neutral wind induced currents and ionospheric conductance from the TING model. A detailed description of the CMIT model and its coupling procedure is given by Wilberger et al. [2004] and Wang et al. [2004].

In this study, a diagnostic package for the TING model has also been applied to study the physical processes that produce altitude profiles of the horizontal winds seen in the model simulations during geomagnetic active periods. This diagnostic package performs term-by-term analyses of the neutral momentum, energy, composition and ionospheric density and energy equations solved in the TING model [Killeen and Roble, 1984]. It thus enables us to determine the physical and chemical processes that are responsible for establishing the thermospheric and ionospheric structures that are seen in the model and for introducing perturbations to these structures. Recently, this diagnostic package has been used to study the solar cycle dependence of the response of the thermosphere to geomagnetic storms [Burns et al., 2004], and ionospheric electron temperature variations during storms [Wang et al., 2006].

The thermospheric wind is driven by a number of processes. These processes include: viscosity, the Coriolis force, ion drag, momentum advection and the pressure
gradient force. The diagnostic package reads in model outputs and calculates all the terms that represent these forcing processes in the neutral momentum equation [Killeen and Roble, 1984, 1986]. The temporal and spatial variations of these terms are then analyzed to show their relative importance in determining neutral wind vertical profiles as well as their changes during geomagnetic storms.

3. Model Results

[10] Figure 1a gives the solar wind and interplanetary magnetic field (IMF) data measured by the WIND satellite for the 14–17 May 1997 event. These data were used to drive the CMIT model. Around 0200 UT on 15 May, solar wind speed jumped abruptly from ~300 km/s to ~500 km/s, while the strength of the IMF increased also suddenly from about 10 nT to 25 nT over a very short period of time. The IMF B_z component turned southward at 0600 UT, and remained mostly southward for about 12 hours till ~1800 UT when it turned to northward. There was a very short period of time around 0900 UT when B_z was northward. This solar wind condition caused a major geomagnetic storm on 15 May, with a maximum Kp value of 7. In addition, the $F_{107}$ solar radio flux was around 74 (10^{-22} W/m^2/Hz) during this event, corresponding to solar minimum conditions.
The CMIT response to this event is shown in Figure 1b. Significant amounts of energy were deposited in the high-latitude upper atmosphere in the form of Joule heating and particle precipitation. A rapid increase in cross-polar cap potential and Joule heating occurred at about 0200 UT on 15 May, which was associated with large increases in the solar wind speed and the magnitude of IMF $B_y$. A more significant enhancement in cross-polar cap potential, Joule heating and particle precipitation hemispheric power occurred immediately after the IMF $B_z$ component turned southward at about 0600 UT. The magnitudes of all of these parameters dropped significantly during the short period when $B_z$ was northward around 0900 UT. Large values of polar cap potential, Joule heating and hemispheric power were seen between 1200 and 1800 UT. They then decreased gradually toward their prestorm values as $B_z$ turned slowly from southward to northward. There were some periods of high polar cap potential and Joule heating on 16 May, but they were not as significant as those that occurred on 15 May.

Figure 1 (right panels) also illustrates Northern Hemisphere high-latitude ion convection patterns at two times: 1200 UT on 14 May and 1500 UT on 15 May, corresponding to quiet and active conditions, respectively. It is evident that during the storm period, the ion convection pattern was significantly enhanced, and expended to lower latitudes. This enhancement in ion drifts produced strong ion drag and Joule heating, leading to a neutral wind circulation that was enhanced globally. It is worth noting here that the cross-polar cap potential calculated by CMIT appears to be too high, although the storm simulated here was a major one: IMF $B_z$ was $-25$ nT and Kp was 7. The global magnetospheric MHD code tends to give high cross-polar cap potentials that lead the CMIT model to overestimate ion drift velocities and Joule heating [Wang et al., 2004; Merkin et al., 2007]. However, we do not think this will affect our results since we are more interested in the changes of the winds and the relative importance of each forcing that drives the winds, instead of their absolute values. We are going to look into the issue of changes of neutral wind profiles with levels of geomagnetic activity and other geophysical conditions in future studies.

### 3.1. Wind Profiles

Figure 2 illustrates quiet time (1200 UT, 14 May) vertical profiles of the horizontal winds at local noon (right panels) and midnight (left panels) in the Northern Hemis-
sphere. Zonal winds (positive eastward) are shown in the upper panels, whereas meridional winds (positive northward) are shown in the lower panels. This UT corresponded to a period when the IMF \(B_z\) and \(B_y\) components were both slightly negative; but there were no significant variations in solar wind speed or density (Figure 1a). The energy deposition into the high-latitude upper atmosphere simulated by the CMIT model was also insignificant: the magnitudes of polar cap potential, Joule heating and hemispheric power were all very small (Figure 1b). Because of this weak magnetospheric driver at high latitudes, middle and low-latitude neutral winds at this UT were mostly determined by the internal dynamical processes of the thermosphere and ionosphere system, and the wind speed was relatively small.

[14] Middle and low-latitude zonal winds were westward at midnight between 100 km and 200 km and eastward above about 200 km (Figure 2, upper left panel). At high latitudes between 60°N and 80°N, zonal winds were westward at all heights. The strongest zonal wind occurred at about 65°N. In the local noon sector (Figure 2, upper right panel), zonal winds were predominantly westward with two regions of strong winds: one occurred at high latitudes between 60°N and 80°N, and another at latitudes equatorward of 30°N. Largest vertical gradients of the winds occurred between 150 km and 200 km in both sectors. Above 250 km, zonal winds showed insignificant vertical variations. Winds at high latitudes were, in general, stronger than those at middle and lower latitudes, due mainly to energy and momentum inputs from the magnetosphere, although the magnetospheric driver was relatively weak during this period.

[15] Meridional winds (Figure 2, lower panels) were much stronger at high latitudes than they were at the middle and low latitudes for both the noon and midnight sectors. The meridional winds in the noon sector were predominantly poleward at high and middle latitudes. In the midnight sector, winds were equatorward at all latitudes above ~150 km. Thus winds blew from the dayside across the pole into the night side middle latitudes as expected for quiet conditions [e.g., Mayr et al., 1978; Hedin et al., 1988].

[16] Low-latitude meridional winds at local midnight did not vary in the vertical above 200 km, whereas they varied considerably at middle and high latitudes in the same altitude range. At local noon, meridional winds and their associated vertical shears were relatively weak at middle and low latitudes. Strong meridional winds occurred in the very high latitude regions (>80°N) in the midnight and noon sectors; vertical gradients of the winds were also large in these regions. Figure 2 thus shows that even during quiet periods latitudinal and altitude variations can exist in the horizontal winds at high latitudes and sometimes at middle latitudes in the upper thermosphere.

[17] These vertical variations of the horizontal winds can be more clearly seen in Figure 3, which gives altitude profiles of zonal and meridional winds for three latitudes: 22.5°N (solid lines), 42.5°N (dash-dotted lines) and 62.5°N (dashed lines). In the midnight sector, zonal winds at 22.5°N and 42.5°N were westward in the lower thermosphere and had a peak at ~140 km (Figure 3, upper left panel). Between 150 km and 200 km they became eastward. The zonal wind at 62.5°N, however, was westward at all altitudes. It is evident, however, that zonal winds at all latitudes were asymptotic vertically above about 250 km, and thus no perceptible wind shears existed above this height.

[18] The zonal winds in the noon sector (Figure 3, upper right panel) had different altitude variations from those in the midnight sector. The winds in this sector were purely westward at all three latitudes, and more variations were seen in the lower thermosphere. Above 200 km, the winds appeared to increase with altitude and there were weak, but discernible, vertical shears in the wind profiles.

[19] Meridional neutral winds were almost entirely equatorward in the midnight sector (Figure 3, lower left panel). At low latitudes these winds did not vary greatly with altitude at heights above 250 km. At middle and high latitudes (42.5°N and 62.5°N), however, meridional winds continued to increase with altitude, showing large vertical gradients even above 250 km. In the noon sector, meridional winds were mostly poleward. Peaks of wind speed occurred between 160 km and 200 km for all three latitudes. It is evident that meridional winds at local noon also varied with altitude in the upper thermosphere (>200 km), but their vertical variations were not as significant as those at midnight.

[20] It is worth mentioning here that although vertical shears exist in the upper thermospheric horizontal winds, the absolute vertical gradients of these winds and their speeds are, nevertheless, not large at middle and low latitudes, when compared with those during the storm time. Therefore during quiet times upper thermospheric winds do not change significantly with altitude above 250 km over most of the low and middle latitudes, so the basic conclusions of Rishbeth [1972] and Killeen et al. [1982] are valid. At high latitudes, however, significant vertical gradients can exist even under relatively quiet conditions (compare Figure 2). The assumption of no vertical shears in the horizontal winds in the upper thermosphere breaks down in this region.

[21] Figure 4 gives horizontal winds during a geomagnetic storm period (1500 UT on 15 May). IMF \(B_z\) was almost continuously southward for about 9 hours (it was briefly northward at about 0900 UT). A significant amount of energy was deposited into the thermosphere during this period (Figure 1). As a result, the thermosphere was heated up and expanded vertically. This is evident in Figure 4, which illustrates that, in the upper thermosphere, the altitudes (right hand side vertical axis in the plots) of each pressure level were much higher than those during the quiet time case (Figure 2). In addition, the neutral winds were significantly enhanced globally, and associated with large vertical variations. For instance, in the midnight sector (Figure 4, upper left panel) westward winds had a peak around 300 km at high latitudes (>60°N); this peak occurred at lower heights at lower latitudes. Multiple peaks of zonal winds were evident at middle latitudes between 30°N and 50°N. At lower latitudes, an enhancement in westward winds happened at all altitudes above 200 km.

[22] The most striking feature at noon was the strong westward jet that occurred at about 200 km and extended from high latitudes to almost the equator (Figure 4, upper right panel). The maximum wind speed of this jet was close to 600 m/s (compare Figure 5). Above 300 km, the winds were westward and had relatively low speeds at low
latitudes. Large vertical shears occurred between 200 km and 300 km, where the strong westward jet decreased with altitude.

[23] Meridional winds were predominantly equatorward at midnight (Figure 4, lower left panel). They had a peak around 250 km at latitudes equatorward of 40°N. This peak increased with altitude at higher latitudes between 40°N and 70°N. A weak equatorward jet occurred from middle to low latitudes at about 200 km, otherwise, the local noon meridional winds were primarily poleward with two high-speed zones: one occurred between 60°N and 70°N above 300 km with a speed that increased with altitude; and another occurred poleward of 80°N. Therefore there were strong altitude and horizontal variations in both meridional and zonal winds during storm periods. These variations extended from high latitudes to low latitudes in the upper thermosphere above about 150 km, and had significantly different vertical structures at different latitudes.

[24] Figure 5 shows vertical profiles of the horizontal winds in the same format as Figure 3, but for the storm period at 1500 UT on 15 May. Note that lower panels have different scales for wind speeds. Compared to Figure 3, storm-time zonal winds (upper panels) were significantly enhanced at all latitudes and exhibited greater altitude variations. They were westward above about 150 km in both the noon and midnight sectors at all three latitudes, whereas zonal winds were eastward at 22.5°N and 42.5°N during quiet times (Figure 2 and 4). The vertical profiles of zonal winds were also significantly different between quiet and storm times. During quiet times, above about 200 km, zonal winds did not change significantly with altitude for both time sectors. During storm times, however, a large wind peak occurred at about 300 km in the midnight sector at both 42.5°N and 62.5°N, and there was also a prominent peak near 180 km at noon at all three latitudes. Large altitude variations occurred both above and below these peaks. This means that strong wind shears in the vertical direction existed for almost the entire thermosphere at this time.

[25] Midnight meridional winds were also much stronger than those that occurred in the quiet time case (Figure 5, lower left panel). The storm-time wind had strong latitudinal and altitude variations. A wind peak occurred at 200 km at 22.5°N and the wind speed decreased with altitude. Above about 400 km the wind did not vary much with altitude; it approached an almost constant value of about 250 m/s. At 42.5°N the wind peak happened at a much higher altitude (~350 km). The wind then decreased at higher altitudes, so large shears occurred in the wind profiles. At 62.5°N there were no peaks in the wind profiles:

\[\text{Figure 3.} \quad \text{Neutral winds profiles at three latitudes (22.5°N, 42.5°N, and 62.5°N) during quiet time at 1200 UT on 14 May 1997. The left-hand side panels are winds at local midnight (0000 LT), and the right-hand side panels are winds at local noon (1200 LT). Top panels are zonal winds, and lower panels are meridional winds. In these plots, the left-hand side vertical axis is pressure levels, while the right-hand side shows heights.}\]
the wind speed increased monotonically with height. Large vertical shears thus existed throughout the entire thermosphere.

There was an equatorward wind peak just below 200 km at noon at 22.5° N and 42.5° N (Figure 5, lower right panel). The winds at these two latitudes turned poleward over a very short vertical distance. At 62.5° N, however, the wind was purely poleward. A minima of the wind speed occurred at the same height as the peaks did at the other two latitudes. Above this region the wind speed increased monotonically with height; no constant speed was approached. The storm-time winds at local noon thus changed their direction compared with the quiet time winds, enhanced their strength significantly, and exhibited quite different vertical structures which included large vertical shears.

The CMIT simulations shown in Figures 2–5 indicate that variations exist in the vertical profiles of the horizontal winds not only in different locations, but also at different heights and times. These variations become significantly larger during geomagnetically active periods. They are caused by changes in momentum forcing. In the next section, we will investigate how these forcings vary with geophysical conditions, and their relative importance in producing the wind structures seen in model simulations.

### 3.2. Diagnostic Analysis: Quiet Times

Figure 6 illustrates momentum forcing during the quiet period (1200 UT, 14 May) at 22.5° N for local midnight and noon, respectively. This forcing includes: horizontal momentum advection, the Coriolis force, ion drag, the pressure gradient force and viscosity. Vertical momentum advection is neglected since its contributions to the overall momentum balance are relatively insignificant.

Pressure gradients and the Coriolis force dominated the zonal force balance at midnight (Figure 6, upper left panel). Ion drag became important above 250 km in this plot, since significant ion densities existed only in the F2 region at night. Viscosity and horizontal advection were of the same magnitude, which was less than 25% of the values of either the pressure gradient or the Coriolis force. The net acceleration (thick solid line, sum of all 5 forcings) was small. It had an eastward peak around 150 km and a broad westward peak above 200 km. Compared with the wind profile (solid line in Figure 3, upper left panel), this net acceleration reduced the westward wind peak at 150 km and the eastward flow above 200 km. It is worth noting that the
total change in wind speed, however, was also related to the length of time that this net acceleration was applied to a parcel of air: the accelerations shown here were only a snapshot of the instantaneous forcing at the time.

The same analysis was also applied to meridional accelerations at midnight (Figure 6, lower left panel). Again, horizontal advection and ion drag were weak and thus contributed little to the overall momentum balance. The Coriolis acceleration was poleward in the lower thermosphere below about 150 km and became equatorward above that height. The change in the Coriolis acceleration direction was caused by the altitude variations of the zonal winds (Figure 3). This acceleration was also balanced primarily by the pressure gradient force, although there were some contributions from viscosity. Thus in a crude approximation upper thermospheric winds at this latitude and local time can be regarded as geostrophic. It is interesting to note that since there were fewer ions in the nighttime ionosphere the ion drag force was relatively weak in both the zonal and meridional directions. The net meridional acceleration also had altitude variations, but it was of insignificant magnitude. Thus large changes in the meridional wind speed were not expected at midnight.

Zonal momentum forcing at local noon was very different from the forcing at midnight (Figure 6, upper right panel). Pressure gradients were balanced by ion drag; there were negligible contributions from other forces above 200 km. The net acceleration was weak, thus the temporal variations in the winds were very small. Ion drag became significant in this region, since F2 region ion densities were large in daytime. It is also interesting to see that horizontal advection was close to zero indicating that there were almost no horizontal variations in wind velocities (compare Figure 2, upper panels). In the meridional direction, however, strong pressure gradient forcing caused a large imbalance in the total forcing. The net acceleration that resulted from this imbalance was equatorward and increased with altitude in the upper thermosphere. This would induce changes in the wind profile (Solid line in Figure 3, lower right panel).

At 42.5°N and midnight (Figure 7, left panels), zonal acceleration was dominated by the Coriolis and pressure gradient forces. The net zonal acceleration was westward and increased with altitude. In the meridional direction the pressure gradient force was significantly stronger than other forces above 150 km, leading to a net acceleration that increased almost linearly with altitude. In this case, the equatorward meridional wind that was seen in Figure 3 (dash-dotted line in the low left panel) would speed up and vertical shears of the horizontal winds would become larger. Viscosity, which increased with altitude, became very important in the overall forcing of meridional winds above 150 km. This was caused by the large vertical shears that already existed in the horizontal winds: viscosity acted to reduce these shears. However, since viscosity was much smaller than the pressure gradient force, the net acceleration increased with height, enhancing the existing vertical shears instead of diminishing them. At local noon (Figure 7, right panels).
panels), ion drag became an important force in both the zonal and meridional directions. It had a magnitude that was comparable to or even greater than those of the pressure gradient and Coriolis forces. Momentum advection and viscosity were insignificant compared to other forcing. The net accelerations in both the zonal and meridional directions were large, changing with height, and opposite to the wind direction, so it acted to reduce wind speeds.

High-latitude meridional acceleration at midnight was dominated by the Coriolis and pressure gradient forcing, with some contributions from ion drag and viscosity above 250 km (Figure 8, lower left panel). Pressure gradient forcing was equatorward, indicating that heating in the polar region drove meridional winds toward lower latitudes. There were also contributions to the pressure gradient forcing from the day-to-night temperature differences. Pressure gradient acceleration increased monotonically with altitude and was significantly larger than other forces above 200 km. All other forces acted in the opposite direction to the pressure gradient forcing. These forces were almost balanced which resulted in a net acceleration that had a complicated vertical profile of small magnitude. Thus the wind profile did not change significantly with time at this location and time.

The midnight zonal net acceleration was westward. It increased rapidly with altitude above 150 km (Figure 8, upper left panel) since almost all of the major forces were in the same direction. Thus no force balance was achieved. This was probably the cause of westward zonal wind seen in the upper left panel of Figure 3. Stronger vertical shears also occurred in the wind profile due to this rapid increase in the net acceleration with height.

The Coriolis force became dominant in the thermosphere in the noon sector (Figure 8, right panels). It acted poleward in the meridional direction and eastward in the zonal direction. Horizontal advection became significant above 150 km. The pressure gradient force was small zonally, but was very important in the meridional direction where it acted with momentum advection to balance the Coriolis forcing. Ion drag was large in the zonal direction and less important in the meridional direction. In addition, the magnitude of the viscous acceleration was less than those of the other forcing terms, although it did increase with altitude as neutral density decreased exponentially with height. The net acceleration in the noon sector was small in both the meridional and zonal directions. Thus temporal changes in the wind speed were very small.

The ion drag at 62.5° and midnight in the upper thermosphere was significantly larger than it was at middle and lower latitudes (Figure 8, left panels). This increase in ion drag at this local time was produced by two factors: large ion drift velocities and enhanced ion densities [Killeen and Roble, 1984]. Figure 1 shows that at this UT the cross-polar cap potential was about 30 KV. This electric field
drove the ions to move faster at high latitudes than they did in middle and low latitudes. F region ion densities were also enhanced by the transportation of daytime plasma across the polar cap into the night auroral region and by particle precipitation.

3.3. Diagnostic Analysis: Storm Times

[37] Figures 9–11 illustrates storm-time momentum forcing at 1500 UT on 15 May. Momentum accelerations at this UT were significantly larger than those during the quiet period at almost all latitudes and local times. At 22.5°N and midnight, horizontal advection was the dominant forcing in the zonal direction above ~150 km (Figure 9, upper left panel). In fact, the altitude profile of the net momentum acceleration followed closely that of the zonal acceleration by momentum advection. This indicates that significant horizontal gradients existed in the winds at this time (compare Figures 4 and 5). It is also interesting to observe that the acceleration by the pressure gradient force maximized at two heights: one was eastward and occurred at about 300 km; the other was westward and occurred at ~200 km. Zonal pressure gradients thus changed direction from eastward below 350 km to westward above that height. This was quite different from the quiet time horizontal pressure gradient profiles which either increased or were constant with altitude. Zonal accelerations by ion drag and viscosity were negligible at this local time. The net zonal acceleration was eastward and increased with height above 250 km.

[38] The equatorward excursion of the midnight meridional winds between 200 km and 300 km at 22.5°N (Figures 4 and 5, lower left panels) was probably caused by the strong horizontal advection at the same height (Figure 9, lower left panel). The pressure gradient acted to reduce the forcing by momentum advection. These two were the dominant processes here. Viscosity contributed to the overall momentum forcing, albeit weakly, and it varied significantly with height, with a peak that occurred around 350 km. This vertical variation of the viscosity corresponded to the meridional wind profiles (Figure 5, lower left panel). The net meridional acceleration was evidently large and equatorward. It increased with altitude above 200 km. Thus the equatorward meridional winds (Figure 5, lower left panel) sped up. The net meridional accelerations at midnight were large and increased with height; thus winds at this latitude changed significantly with time and exhibited large vertical variations.

[39] Zonal forcings at local noon were smaller than those at midnight (Figure 9, upper right panel) at the same latitude. Ion drag maximized around 200 km and was the dominant force at that height. Both the Coriolis force and horizontal advection had westward peaks at about 180 km. Coriolis forcing decreased with height and became less significant in the upper thermosphere, whereas horizontal advection turned eastward and maximized at about 250 km,
and continued to be one of the dominant forcing terms in the upper thermosphere. Viscosity and pressure gradient forcing were of the same magnitude but acted in different directions. They both increased with altitude, and contributed to the overall momentum forcing above 350 km. The net acceleration above 200 km was eastward, and acted to decrease the westward winds that were seen in Figure 5. In the meridional direction at noon (Figure 9, lower right panel), the two major forcing processes were the Coriolis and pressure gradient forces; both acted in the poleward direction. They sped up the northward meridional wind (Figure 5, lower right panel). Nevertheless, the net meridional acceleration at noon was smaller than it was at midnight above 200 km; thus changes in neutral wind velocities were also smaller than those that occurred at midnight.

Figure 10 gives storm-time forcing at 42.5°N. Note that a different scale is used in the lower left panel as large accelerations occurred in the meridional direction at midnight (Figure 10, left panels). At this local time and in both the meridional and zonal directions, ion drag was negligible compared with the other forces. Both the pressure gradient force and horizontal advection increased rapidly with height. Viscosity also increased with height above 350 km, but its magnitude was much smaller than those of the pressure gradient force and horizontal advection.

Coriolis acceleration was the dominant process between 150 km and 300 km in the zonal direction, speeding up the westward wind in that altitude range (Figure 10, upper left panel). Above 300 km, the net zonal acceleration resulted from a combination of four processes: the Coriolis force, horizontal advection, the pressure gradient force and viscosity. This net acceleration was westward and had a peak at an altitude of about 400 km. This peak occurred because the large increase of the eastward pressure gradient force could not offset the combined effect of westward acceleration by momentum advection and the Coriolis force.

Momentum forcing in the meridional direction was totally dominated by two processes with opposite directions above 200 km (Figure 10, lower left panel): a poleward pressure gradient force and equatorward horizontal advection. There were contributions to the net acceleration from the Coriolis force and viscosity, but they were minor. The net momentum acceleration peaked at about 350 km. It thus reinforced the equatorward meridional wind that also peaked at about 350 km (Figure 5, lower left panel).

The forcing terms showed significant altitude variations at noon (Figure 10, right panels). Ion drag was the dominant process in the zonal direction. Multiple peaks were seen in horizontal advection, the pressure gradient and Coriolis accelerations. These three processes had roughly the same magnitude, and all were smaller than the acceleration by ion drag above 200 km. The net acceleration was, nevertheless, comparable to that at midnight and increased with altitude.

Below 200 km horizontal advection and the Coriolis acceleration were the major processes acting in the merid-
ional direction at noon (Figure 10, lower right panel). Above 300 km, the contributions from the pressure gradient and ion drag accelerations became noticeable, but their values were still only about half of those of horizontal advection and the Coriolis acceleration. There were no large changes in acceleration with altitude above 350 km for any of the processes, so the net acceleration was close to zero at these altitudes. Hence changes in the meridional winds with time were small at these heights. The equatorward peak at 150 km and poleward peak at 200 km of the net acceleration, however, reduced the corresponding wind peaks seen in Figure 5.

[45] At high latitude (62.5° N), storm-time forcing was also significantly larger than the forcing that occurred during quiet time (Figure 11). Coriolis acceleration was the dominant process at midnight in the zonal direction throughout the upper thermosphere above 150 km (Figure 11, upper left panel). This force was balanced by eastward ion drag between 180 km and 250 km, and by eastward horizontal advection and the pressure gradient force above about 300 km. The effect of viscosity became important only above 350 km. However, the magnitude of the net acceleration of these processes was not large since they tended to cancel each other out. Thus the zonal wind profile shown in Figure 5 did not change significantly with time.

[46] Contributions from ion drag and viscosity were negligible in the meridional direction at midnight (Figure 11, lower left panel). The major drivers of the winds were the equatorward pressure gradient acceleration and poleward horizontal advection with some contributions from Coriolis acceleration. The net acceleration was very weak, and thus did not cause significant changes in the wind profile seen in Figure 5. The weak ion drag in this plot, at this particular time and location, occurred because of the low ionospheric conductance there. In regions of high conductance, nighttime ion drag can be very significant.

[47] Complicated vertical structures were seen at local noon in all zonal momentum accelerations. Ion drag, horizontal advection, and the pressure gradient and Coriolis accelerations were the major processes that drove the zonal neutral wind in this location. The contribution of viscosity was negligible. Each process peaked at a different altitude. This produced a net acceleration that had a very complicated vertical structure, and consequently there was a very complicated zonal wind profile (Figure 5, upper right panel).

[48] The pressure gradient acceleration in the meridional direction at noon also had a vertical profile that was equatorward in the lower thermosphere and poleward in the upper thermosphere (Figure 11, lower right panel). It became the dominant force above 350 km. Ion drag, horizontal advection, and the Coriolis and pressure gradient forces were all important between 150 km and 300 km. Above 300 km the pressure gradient acceleration increased.
with height. As the pressure gradient force was not balanced by other forces there, this increasing acceleration with height created poleward winds that also increased with altitude. Ion drag, horizontal advection and the Coriolis force decreased with altitude, thus contributing less to the overall momentum forcing than they did in the lower thermosphere. Viscosity increased with altitude and became very large in the highest parts of the thermosphere (above 450 km). It acted to reduce the poleward acceleration that was driven by the pressure gradient force. Nevertheless, the viscous acceleration was still much smaller than the pressure gradient acceleration, thus the meridional wind would continue to increase its poleward speed with time and altitude.

4. Discussion

Figures 2 and 3 showed that quiet time, low- and middle latitude horizontal winds in the upper thermosphere varied little in the vertical most of the time. Wind profiles were very close to those which Rishbeth [1972] obtained using a steady state thermosphere model and a simplified ionospheric model at middle latitudes. There were, however, cases in which significant departures from these analytical vertical profiles occurred. An example of such departures happened at local midnight when meridional winds at 42.5°N had large vertical shears in the upper thermosphere (Figure 3, lower left panel). Both the vertical distribution of horizontal winds obtained from Rishbeth’s analysis and data averaged over long periods of time [Killeen et al., 1982; Wharton et al., 1984; Fejer et al., 2000; Emmert et al., 2001, 2002, 2004] are, therefore, generally descriptive of the behavior of quiet time, steady state, low- and middle latitude winds in a statistical sense. At high latitudes, horizontal winds sometimes can have significant vertical variations even under relatively weak magnetospheric driving conditions in the upper thermosphere. Thus these steady state, statistical wind profiles must be applied very carefully when constructing models or interpreting observations of winds that may be dynamic and change with altitude.

The quiet time vertical shears in the horizontal winds occur due mainly to the height variations in the pressure gradient and ion drag forces. The ion drag force is determined by two factors: electron densities and velocity differences between the ions and the neutrals. The ion drag variations seen in this simulation were caused mainly by the changes of electron densities with height: maximum ion drag happened at the F2 peak height. The vertical variation of the pressure gradient force is related to the horizontal and vertical distribution of thermospheric temperatures. The Rishbeth [1972] analysis used the Jacchia model which is smooth both temporally and spatially, and thus may underestimate the thermospheric pressure gradient force and its contribution to the distribution of the neutral winds in both the vertical and horizontal directions. In fact, our simulations indicate that large altitude variations in temperatures,
and thus in pressure gradient forces can occur even in quiet times. These variations can cause vertical shears in the horizontal winds at some locations (see Figures 7 and 8, lower left panels). In addition, since neutral temperatures, pressure gradients and electron densities vary significantly with solar cycle, season, and geomagnetic activity, the vertical profiles of the horizontal winds will change with these geophysical conditions. We will fully address this issue in future studies.

Viscosity is not only related to the neutral densities, whose vertical profiles give viscous drag coefficients that increase with height, but it is also related to the second-order vertical derivatives of the winds [Rishbeth, 1972]. Thus viscosity is important only at altitudes where there are height changes in the vertical gradients of the horizontal winds, that is, $\partial^2 U_n / \partial h^2 \neq 0$, where $U_n$ and $h$ are the neutral wind speed (zonal or meridional) and height, respectively. Our simulations show that in the upper thermosphere $\partial^2 U_n / \partial h^2$ is small in most cases; this causes the magnitudes of the viscous acceleration to be less than those of the other forces, although these magnitudes do increase with altitude at most times as neutral density decreases exponentially with height. It is also evident that in our simulations that $\partial U_n / \partial h \neq 0$ can occur even during quiet times at some locations, such as in the meridional winds at local midnight at 42.5°N and 62.5°N. This gradient is maintained by a net vertical shearing forcing that is produced by vertical variations of ion drag and the pressure gradient force. Since the weak viscous forcing cannot balance these vertical variations, the net acceleration by these forces also changes with height, and so do the vertical profiles of the horizontal winds. Diagnostic analysis shown in Figures 6–8 also suggests that the “uniform” vertical profiles of the horizontal winds seen in Figures 2 and 3 are most likely related to a lack of altitude variation in their driving forces, rather than to the smoothing effect of the viscous diffusion.

It is also worth noting that the relatively weak viscosity simulated in this study, compared with other forcing, does not mean that the vertical profiles of the horizontal winds do not change with height, rather it indicates that the vertical gradients of the wind profiles are nearly constants, thus the second-order vertical derivatives of the winds are close to zero. This suggests that the viscous diffusion tends to remove any nonlinear deviations of the wind profiles. In other words, viscosity often forces the winds to change linearly with height. This is clearly shown, for instance, in the cases of midnight meridional winds at 62.5°N (Figure 5, dashed line in the lower left panel) and midday zonal winds at 42.5°N (Figure 5, dash-dotted line in the upper right panel) where the horizontal winds increased almost linearly with height, the viscosity, nevertheless, was very small in both cases (compare Figure 11, lower left panel, and Figure 10, upper right panel).

It is very difficult to determine how long it takes for viscosity to remove existing vertical shears in our simula-

![Figure 11. Same as Figure 9, but for 62.5°N.](image-url)
tions, since other forces are constantly changing with time. However, a very crude estimation can be made based on wind speeds and the magnitude of the viscous forcing. For instance, at 62.5°N and 500 km, the midnight meridional wind was 750 m/s; to reduce this wind speed to the value of that at 200 km, which was 550 m/s (Figure 5, lower left panel), required about an hour by viscous acceleration acting alone. Here, we assumed that the viscous acceleration did not change with time and had a constant value of 0.05 m/s² (Figure 11, lower left panel), and all other forces suddenly became zero.

[54] Of course the real situation is much more complicated than this simple estimation. In fact, viscosity decreases as shears decrease, increasing the recovery time considerably. Other forces acting on the neutral air parcel can also either enhance or reduce the shears. Nevertheless, we expect that these vertical shears can exist for from as less as a few minutes to as long as a few hours.

[55] Storm-time, vertical profiles of the neutral winds were significantly different from those at quiet time. The most noticeable differences were the large vertical and horizontal gradients that occurred globally (Figures 4 and 5). Storm-time neutral winds are not in a steady state. They are time varying since energy and momentum inputs from the magnetosphere at high latitudes are dynamic and the response of the thermosphere and ionosphere to these inputs is also nonlinear and dynamic (Figure 1). As we have discussed in Section 3.3, storm-time viscosity is not an important force for most of the time, instead it is often smaller than the sum of the other forcing processes, and acts in the opposite direction. Viscosity is a relatively weak force that requires time to smooth out nonlinear vertical variations in the horizontal winds. The constantly changing conditions during storms do not allow sufficient time for this smoothing to occur, so there are often strong vertical shears in the upper thermospheric horizontal winds during these events. At higher altitudes above the top boundary of the model, viscosity becomes very large so winds tend to be constant. In fact, the wind profiles and forcing terms shown in this paper illustrate just that. There are significant shears at most heights, but, close to the top of the model, viscosity becomes large and the wind profiles tend to be constant even for storm conditions.

[56] The altitude variations in momentum forcing happen in several ways during storms. Variations in ion drag occur when there are changes in electron densities, and in ion and neutral velocities. Nighttime ion drag is weaker in the high and middle latitudes due to negative storm effects (decreases in electron density at the F2 peak) [e.g., Prölls, 1980]. These effects tend to occur near midnight and rotate toward early morning hours [Prölls, 1981]. The lower electron densities in this region during storms reduce the ion drag coefficient and hence the ion drag. This becomes particularly important as these regions of negative storm effects rotate into the dayside. In the noon sector ion drag is significantly enhanced at all latitudes due mainly to the enhanced electron densities (positive storm effects) [e.g., Burns et al., 2007], and also at high latitudes due to the enhanced ion drift velocities. Ion drag has distinct variations with altitude that are related to the altitude distribution of ionospheric electron densities, since in the model ion drift velocities are driven by ExB drifts at all heights, and thus do not have altitude variations at F region. The temporal variations of ion drag are induced by the changes in the high-latitude convection pattern and precipitation that vary with solar wind driving conditions and internal dynamic processes in the magnetosphere.

[57] The pressure gradient force contributes significantly to the establishment of neutral wind profiles during both quiet and storm times. Storm-time pressure gradients are enhanced globally and have large vertical variations (Figures 9–11). These variations are caused by changes in the thermospheric heating during storms. One major source of such changes is the enhancements of high-latitude Joule heating. Joule heating occurs where large differences between ion and neutral velocities occur and has the most significant impact on the neutral temperatures in the upper thermosphere during storms [Thayer and Semeter, 2004; Wang et al., 2004; G. Jee et al., Continual initialization of the TING model with GAIM electron densities: Ionospheric effects on the thermosphere, submitted to Journal of Geophysical Research, 2007]. Since both the high-latitude convection pattern and particle precipitation can change rapidly during storms, the high-latitude conductivities can also change rapidly during these events both vertically and horizontally. Joule heating responds to these changes and is also very variable in both the horizontal and vertical directions. Thus large vertical and horizontal variations can occur in neutral temperatures and pressure gradients during storms. Changes in Joule heating and in ion drag also change global wind patterns, this pattern has large variations both horizontally and vertically. Convergent and divergent flows associated with these changing winds induce compressional heating and cooling globally [Burns et al., 1992]. This, in turn, causes further variations in the neutral temperatures and consequently the pressure gradient forces and the neutral winds.

[58] Momentum advection is not a major force during quiet time at the middle and low latitudes (Figures 6–8). However, at high latitudes and sometime at middle latitudes (Figure 7, lower left panel), advection can contribute significantly to the overall momentum forcing. These large accelerations due to momentum advection result from large horizontal wind gradients (Figure 2). During the storm, momentum advection is enhanced globally and becomes one of the major forces for the horizontal neutral winds. It is also interesting to note that, unlike the quiet time case when the momentum advection force almost always increases with height, storm-time momentum advection tends to have peaks at certain altitudes. This is caused by the horizontal neutral winds varying both vertically and horizontally (Figure 4).

[59] At middle and low latitudes, nighttime momentum advection, and the pressure gradient and Coriolis forcing are the three dominant forces during the storm (Figures 8 and 9, left panels). The net acceleration is mostly in the direction of momentum advection. At 22.5°N, the net acceleration actually follow the vertical shape of momentum advection, indicating that changes in horizontal winds at low latitudes are induced by high-speed winds advecting from high and middle latitudes to low latitudes. At middle latitudes the pressure gradient forcing that is produced by high-latitude Joule heating and compressional heating/cooling by wind circulation acts against momentum advection.
The Coriolis force sometimes also changes with altitude during quiet times. This vertical change is a reflection of vertical variations of the winds. It is interesting to note here that at night ion drag becomes less important because electron densities are low; thus thermospheric momentum balance is sometimes achieved just between pressure gradient and Coriolis forcing at low and middle latitudes, resulting in geostrophic winds. Daytime momentum balance is more complicated due to the involvement of ion drag and thus, in general, daytime winds are not geostrophic.

5. Conclusions

In this paper, we presented CMIT simulations of upper thermospheric winds during a specific storm event under solar minimum, equinox conditions. Significant vertical shears were seen in the horizontal neutral wind profiles during the storm. These vertical shears occurred globally. Quiet time neutral wind profiles were less variable, but there were still noticeable vertical gradients in the wind profiles in the upper thermosphere. Our results suggest that the assumption of a shearless wind profile in the upper thermosphere can be applied at low and middle latitudes during quiet times. However, this assumption is less applicable at high latitudes for both nondisturbed and disturbed conditions. At middle and low latitudes storm-time winds also show significant vertical and horizontal variations, thus they cannot be regarded as uniform vertically in the upper thermosphere.

The vertical variations of the winds at high latitudes were caused by changes in the net momentum forcing with altitude. In response to the enhanced and rapidly varying energy and momentum inputs from the magnetosphere, high-latitude neutral temperatures, electron densities and ion velocities had significant temporal and spatial variations. These variations introduced changes in major forcing processes, such as the pressure gradient force and ion drag, enhanced the neutral wind speeds and produced vertical shears in wind profiles.

There were significant altitude and horizontal variations in storm-time neutral winds at middle and low latitudes. These variations were caused by changes in the pressure gradient that were associated with neutral temperature changes, variations in ion drag that were induced by storm-time electron density variations, such as positive and negative storm effects, and by significantly enhanced momentum advection from higher latitudes. In fact, momentum advection, which was relatively weak in the quiet time case, became a dominant force that drove the middle and low-latitude neutral winds during the storm. The changes in neutral temperatures at middle and low latitudes were due to the greater compressional/expansion heating/cooling that was associated with enhanced global neutral wind circulation and heat conduction from high latitudes.

Our simulation also illustrates that under both storm and quiet conditions viscosity was not a major forcing mechanism for the neutral winds at most heights. Thus when the net forces change with height in the upper thermosphere, they cause vertical shears of the horizontal winds in this region. Viscosity is a slow process due to its relatively small magnitude, so that other, faster processes can set up significant shears in the upper thermosphere in a relatively short period of time, particularly during storms.

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