Dependence of the high-latitude lower thermospheric momentum forcing on the interplanetary magnetic field

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We analyze the forces acting on the high-latitude lower thermospheric wind system below 170 km for Southern Hemisphere summer conditions, as a function of the interplanetary magnetic field (IMF) direction, on the basis of numerical simulations. The pattern and magnitude of the forces and their relative contributions to the wind system vary strongly with respect to the direction of the IMF. At higher altitudes, above 130 km, for negative B_y, strong anticyclonic winds are accelerated primarily by rotational Pedersen ion drag and are maintained by an approximate balance among the divergent/convergent Coriolis, horizontal advection, and relatively weak pressure-gradient accelerations. For positive B_y, the pressure-gradient acceleration is increased, while the inertial forces are extended. For negative B_z, in comparison with negative and positive B_y, the winds and forces extend to lower latitudes. The patterns of the accelerations for positive B_z are similar to those for negative B_z, but the magnitudes tend to be significantly smaller. At lower altitudes, below 120 km, the horizontal advection acceleration is less important but still contributes significantly to the maintenance of the neutral circulation in the polar cap region for positive B_y. The difference of winds and forces above 130 km for negative and positive B_y, with respect to winds and forces for zero IMF, show a simple structure with a strong anticyclonic or cyclonic vortex near the pole, respectively, centered differently for the two B_y directions. The difference of winds and forces for negative and positive B_z are more complex than those for negative and positive B_y and extend to lower latitudes. Below 120 km, the difference of winds and forces for negative and positive B_z are much stronger near the pole than for negative and positive B_y, indicating that the IMF B_y component tends to dominate effects on the neutral winds in the polar cap at low thermospheric altitudes. For all IMF conditions, at higher altitudes, the rotational ion-drag acceleration makes the dominant contribution to the neutral velocity tendency. This feature is most pronounced when the IMF B_z is negative.


1. Introduction

[2] Analyses of the dynamics of winds in the high-latitude thermosphere based on theoretical and numerical models have led to a better understanding of the forcing mechanism involved in establishing the high-latitude thermospheric neutral circulation [e.g., Mayr and Harris, 1978; Roble et al., 1982; Mikkelsen and Larsen, 1983; Larsen and Mikkelsen, 1983, 1987; Fuller-Rowell and Rees, 1984; Killeen and Roble, 1984; Walterscheid and Boucher, 1984; Walterscheid et al., 1985; Walterscheid and Schubert, 1986; Gundlach et al., 1988; Walterscheid and Brinkman, 2003].

[3] Recently, Kwak and Richmond [2007] (hereafter called K&R) analyzed the dynamics of winds for negative interplanetary magnetic field (IMF) B_z in the high-latitude lower thermosphere, where there is a strong vertical variation of the characteristics of horizontal forcing by ionospheric electric currents, with the aid of the National Center for Atmospheric Research Thermosphere-Ionosphere Electrodynamics General Circulation Model (NCAR TIE-GCM) [Richmond et al., 1992]. It was shown that a transition of the forcing patterns and their relative contributions to the high-latitude lower thermospheric wind system occurs around 123 km under various conditions, weak or strong IMF, summer or winter. Winds around and above 123 km are horizontally accelerated mainly by rotational Pedersen ion-drag force, while other horizontal forces (pressure-gradient, Coriolis, and horizontal momentum advection), though often larger than ion drag, are mainly divergent/convergent and largely balance through gradient-flow adjustment. Gradient-flow balance is defined as this balance among the Coriolis, centrifugal, and pressure-gradient forces [Walterscheid and Brinkman, 2003]. Below 123 km, winds are accelerated by the rotational component of the ion-drag force, while the horizontal pressure-gradient force...
and the divergent/convergent components of the Coriolis and Hall ion-drag forces largely balance through modified geostrophy.

[4] It is known that high-latitude ionospheric plasma convection and current depend on the direction of the IMF [e.g., Heppner, 1972; Heppner and Maynard, 1987; Foster et al., 1986; Ruohoniemi and Greenwald, 1996; Weimer, 1995, 2001]. Thus the high-latitude thermospheric wind, influenced by interaction with the ionospheric plasma, also depends on the IMF direction [e.g., McCormac and Smith, 1984; McCormac et al., 1985, 1991; Killeen et al., 1985, 1995; Meriwether and Shih, 1987; Thayer et al., 1987; Rees and Fuller-Rowell, 1989, 1990; Sica et al., 1989; Hernandez et al., 1991; Niciejewski et al., 1992, 1994; Won, 1994; Richmond et al., 2003; Deng and Ridley, 2006]. From such results, one can expect that the momentum forcing on the high-latitude lower thermospheric wind system should vary with respect to the direction of the IMF.

[5] Analysis of the IMF dependence of the forcing has not previously been carried out and is the subject of the present study. In this paper, we extend the study by K&R [2007] by examining the forces that are mainly responsible for maintaining the summertime high-latitude lower thermospheric wind system over the Southern Hemisphere as a function of the IMF orientation.

2. TIE-GCM

[6] The TIE-GCM used in this study is an extension of the NCAR TIGCM described by Roble et al. [1988] and self-consistently calculates electrodynamical interactions in the coupled thermospheric/ionosphere system. The model was described in more detail in K&R [2007]. The external inputs required by the model are the solar extreme ultraviolet and ultraviolet fluxes, the auroral particle precipitation, the ionospheric convection, and the upward-propagating tides from the middle atmosphere.

[7] We analyze steady state forcing for different IMF directions, for seasonal and solar conditions representative of 23 January 1993. This period is characteristic of conditions for Upper Atmospheric Research Satellite (UARS)-Wind Imaging Interferometer (WINDII) wind observations analyzed by Richmond et al. [2003]. On this date, the daily F10.7 index was recorded as $102.7 \times 10^{-22}$ W/m$^2$/Hz. The total hemispheric power (HP) was used to specify the high-latitude auroral particle precipitation pattern. The $Kp$ index was used to represent the HP (in gigawatts) as $HP = -2.78 + 9.33 \times Kp$; this formula is from the work of Maeda et al. [1989]. The mean $Kp$ index on 23 January 1993 was 1, although the actual time-varying values for this date were used. The empirical electric potential model by Weimer [2001] was adopted to specify the pattern of the ionospheric convection. The prescriptions of the upward-propagating diurnal and semidiurnal tides are taken from the Global Scale Wave Model [Hagan and Forbes, 2002].

[8] To investigate the response of the model fields to varying IMF, five TIE-GCM simulations for IMF ($B_y$, $B_z$) values of $(-3.2, 0.0)$, $(+3.2, 0.0)$, $(0.0, -2.0)$, $(0.0, +2.0)$, and $(0.0, 0.0)$ nT are made. The magnitudes of the nonzero reference values of $B_y$ and $B_z$ is that, 3.2 nT and 2.0 nT, are their respective root-mean square values for the temporally smoothed data set analyzed by Richmond et al. [2003].

3. Analysis Procedure

[9] The main horizontal acceleration terms acting on a parcel of gas in the upper atmosphere are pressure-gradient, Coriolis, ion-drag (Pedersen plus Hall components), viscous, nonlinear horizontal momentum advection, and vertical momentum advection accelerations. The individual forcing terms are obtained by using the momentum equations with outputs of the model, where vertical and horizontal finite differencing is performed to calculate forces at a given altitude for each universal time. The forcing terms calculated are defined by K&R [2007].

[10] We first bin each force term at each altitude in magnetic coordinates for different orientations of the IMF; we analyze the forces in Quasi-Dipole (QD) coordinates [QD-latitude and magnetic local time (MLT)] [Richmond, 1995]. The QD-latitude/MLT region poleward of 47.5°S is divided into 145 subregions of approximately equal size, each having 5° width in latitude except one at the pole, and a variable, latitude-dependent, width in MLT. The number of MLT subregions at a given latitude decreases from 32 at 50°S to 4 at 85°S, plus one at the pole. Each subregion has bins every half-scale height in altitude. We then carry out an averaging for 24 UT hours of each forcing term in each bin for each of the different IMF conditions. The resultant force patterns can be then mapped out over magnetic latitude and MLT.

4. Results

4.1. Wind Patterns With the IMF Orientation

[11] In this section, the high-latitude lower thermospheric wind patterns for different IMF orientations are shown, and our model results are compared with measurements by the WINDII on the UARS as analyzed by Richmond et al. [2003].

[12] Figures 1a and 1b shows the average neutral wind circulations around 140 and at 111 km over the southern hemisphere for IMF ($B_y$, $B_z$) values of $(-3.2, 0.0)$, $(+3.2, 0.0)$, $(0.0, -2.0)$, and $(0.0, +2.0)$ nT from WINDII data and the TIE-GCM runs, respectively. To compare with other IMF conditions, we include the wind patterns and forces for negative $B_z$ shown in K&R [2007]. As for all other figures in this paper, these projections are as if one were looking up on the thermosphere from below. As shown in Figure 1, the wind patterns in the high-latitude lower thermosphere depend on the orientation of the IMF. Around 140-km altitude, the TIE-GCM wind patterns agree with the WINDII observations as follows, although the latter have a data gap around MLT midnight. The winds for negative $B_y$ show a strong anticyclonic (clockwise) vortex on the dusk side with enhanced antisunward flow inside the polar cap and a return sunward flow at lower latitudes in the vicinity of the auroral oval. The morning cell virtually disappears in the neutral wind field, and there is an equatorward flow in the early morning hours. The winds for positive $B_y$ around 140 km have a two-cell pattern. The winds poleward of $-70^\circ$ on the dusk side tend to show weaker anticyclonic motion than when $B_y$ is negative, while the cyclonic (anticlockwise) circulation on the dawn side is more fully
developed than for negative $B_y$ and the equatorward flow in the post-midnight sector is enhanced. For negative $B_z$, the two-cell wind pattern is also evident, with a larger and stronger anticyclonic vortex on the dusk side and with relatively stronger equatorward flow in the early morning sector than for other IMF conditions. The winds poleward of $70^\circ$ around 140 km tend to be the weakest when $B_z$ is positive. Around 140-km altitude, the modeled maximum wind speeds for negative and positive $B_y$ and negative and positive $B_z$ are 268, 250, 271, and 158 m s$^{-1}$, respectively. At 111-km altitude, the modeled wind patterns show agreement with WINDII observations as follows. Although the global-scale background winds associated with atmospheric tides are present at all latitudes for all patterns, for negative $B_y$, weak winds as a part of the anticyclonic dusk cell appear in the polar cap, and for positive $B_y$, the cyclonic flow as a part of the dawn cell is dominant in the polar cap. When $B_z$ is negative, a weak cyclonic flow still appears in the late morning sector. At 111-km altitude, the modeled maximum wind speeds for negative and positive $B_y$ and negative and positive $B_z$ are 67, 107, 74, and 46 m s$^{-1}$, respectively.

Figure 1. The neutral winds over the Southern Hemisphere for IMF ($B_y$, $B_z$) values of (left to right) $(-3.2, 0.0)$, $(+3.2, 0.0)$, $(0.0, -2.0)$, and $(0.0, +2.0)$ nT from (a) WINDII data and (b) the TIE-GCM run, respectively. These vectors are plotted against a fixed length of arrow of 200 m s$^{-1}$. 
It is found that there is a rapid transition in the wind patterns and magnitudes between about 120 and 130 km. As discussed by K&R [2007], at higher altitudes, ion drag efficiently generates winds having a similar pattern with the ion drift. At lower altitudes, where the ion-neutral collision frequency is comparable to or larger than the ion gyrofrequency, the ion drift is strongly influenced by collisions with the neutral atmosphere.

The TIE-GCM wind patterns show disagreements with the WINDII observations in the following aspects. In all four patterns around 140 km, the equatorward turning of the wind in the late evening region for latitudes $-50^\circ$ to $-65^\circ$ that is seen in the WINDII observations is not seen in the modeled wind patterns, and the modeled wind magnitudes are usually weaker than the WINDII observations. The cyclonic dawn cell, weak or absent in the WINDII observations, is stronger in the TIE-GCM wind patterns except for negative $B_y$. At 111-km altitude, the modeled wind magnitudes, especially in the post-noon sector, are weaker than the observations. The cyclonic or anticyclonic flows of the modeled results in the polar cap tend to stand out more clearly against the background winds than can be seen for the observations. The differences between the observed and modeled winds may be associated with inaccuracy in estimating the model ion-drag forcing, inaccuracy in the tidal boundary conditions for the TIE-GCM, or possibly with numerical smoothing features of the model.

In general, the model wind patterns are very consistent with the WINDII observations. Since the behavior of high-latitude winds in the lower thermosphere varies strongly with the IMF direction, one can expect that the momentum forcing on the wind system should also vary strongly with the IMF direction.

4.2. Analysis of Momentum Forcing Dependence on the IMF Orientation

In order to address the dependence of the forcing on the IMF direction, we analyze forces in the high-latitude lower thermosphere over the southern hemisphere for the different IMF orientations.

Figure 2 illustrates the dominant horizontal acceleration terms acting on the neutrals at 142 km for the TIE-GCM winds shown in Figure 1b, (top to bottom) the ion-drag, pressure-gradient, Coriolis, and horizontal advection accelerations. These vectors are plotted against a fixed vector length of 0.02 m s$^{-2}$ so that the relative importance of these individual forces can be readily seen. The vertical advection and viscous accelerations are not shown here because of their relatively small values in comparison with other momentum forcing terms.

When $B_y$ is negative (the first column in Figure 2), the ion-drag acceleration maximizes in the vicinity of the auroral oval on the early morning side in the sunward direction. This is expected since the strong sunward ion drift shown in Figure 3 is located in this region and the ion density is considerable owing to auroral precipitation. There is also strong anticyclonic ion-drag acceleration on the dayside at the inner boundary of the auroral oval. Inside the polar cap, the neutrals and ions are moving in nearly the same direction and the ion-drag acceleration is not large. Nevertheless, the neutrals in the polar cap are accelerated in the antisunward direction by the ions, in the same direction as the day-to-night solar-driven neutral circulation, reinforcing the antisunward flow of the neutrals. The divergent pressure-gradient acceleration occurs at latitudes $-75^\circ$ to $-80^\circ$ around noon and in the early morning hours. The maximum pressure-gradient acceleration is much smaller than the maximum ion-drag, Coriolis, and horizontal advection accelerations. There is the convergent Coriolis acceleration vectors on the dusk side of the polar cap because of the strong anticyclonic vortex flow pattern for negative $B_y$, shown in Figure 1b. The Coriolis force maximizes in the dawn sector within the polar cap where neutrals have their maximum flow speed. The horizontal advection acceleration is directed equatorward from the dusk sector of the polar cap. This divergent horizontal advection acceleration on the dusk side opposes the convergent Coriolis acceleration in the same region and reinforces the divergent pressure-gradient acceleration. It is evident that for negative $B_y$ at 142-km strong dusk anticyclonic winds accelerated primarily by the ion drag (mainly the approximately rotational Pedersen ion drag shown in Figure 4) involve an approximate balance between the Coriolis and horizontal advection accelerations, although the balance is modulated by the relatively weak pressure-gradient acceleration. The balance among these accelerations constitutes gradient-wind balance [Mayr and Harris, 1978; Larsen and Mikkelsen, 1983; Walterscheid and Boucher, 1984; Walterscheid and Brinkman, 2003; K&R, 2007].

From the second column in Figure 2, one can see that not only the patterns but also the magnitudes of the forces for positive $B_y$ are different from those for negative $B_y$. Strong eastward ion-drag acceleration occurs at latitudes $-80^\circ$ to $-85^\circ$ on the dayside. This is in the opposite direction to that for negative $B_y$, and is produced by strong ion drift at those latitudes in the morning (Figure 3). The ion-drag acceleration in the nighttime polar cap is smaller than when $B_y$ is negative. The pressure-gradient acceleration is, in general, more enhanced than when $B_y$ is negative. The maximum pressure-gradient acceleration is greater than that for negative $B_y$ by more than a factor of 3. Divergent and convergent regions of the pressure-gradient acceleration are located on the postnoon sector and the dawn side of the polar cap/auroral zone boundary, respectively. There are strong convergence and divergence of the Coriolis force vectors on the dusk and dawn sides of the polar cap/auroral zone boundary, respectively, because of the two-cell flow pattern of the neutrals for positive $B_y$. The horizontal advection acceleration when $B_y$ is positive has two strong divergent regions, a dusk side divergence that opposes the convergent Coriolis force at the same place and a divergent horizontal advection acceleration on the dawn side that reinforces the divergent Coriolis force there. It is evident that, for positive $B_y$ at 142 km, the two-cell neutral flows which are accelerated by ion drag involve an approximate balance between the Coriolis and horizontal advection accelerations on the dusk side and an approximate balance of the pressure-gradient, Coriolis, and horizontal advection accelerations on the dawn side.

For negative $B_y$ (the third column in Figure 2), in comparison with negative and positive $B_y$, it is obvious that the forces extend to lower latitudes. As discussed by K&R [2007], it is evident that for negative $B_y$ at 142 km, the
strong anticyclonic cell accelerated by the effects of ion drag on the dusk side, as well as the equatorward flow in the early morning hours, involve an approximate balance between the horizontal advection force and the Coriolis force, although the relatively weak pressure-gradient force contributes to the balance. On the other hand, the dawn cyclonic vortex involves an approximate balance among a relatively strong pressure-gradient force and relatively weak horizontal advection and Coriolis forces. Analysis of forces for the positive $B_z$ (the last column in Figure 2) shows that the patterns of the accelerations are similar to those for negative $B_z$, but the magnitudes tend to be significantly smaller. The reason for this is that the $E \times B$ drift (Figure 3) for the positive $B_z$ has a similar pattern but smaller magnitude than for negative $B_z$.

Figure 2. Vectors of the dominant acceleration terms acting on the neutral winds at 142 km over the Southern Hemisphere for IMF ($B_y$, $B_z$) values of ($-3.2$, $0.0$), ($+3.2$, $0.0$), ($0.0$, $-2.0$), and ($0.0$, $+2.0$) nT: (top to bottom) the ion-drag, pressure-gradient, Coriolis, and horizontal advection accelerations. These vectors are plotted against a fixed length of arrow of 0.02 m s$^{-2}$.
142 km, and, in particular, the horizontal advection acceleration tends to be much weaker. For all IMF conditions except when $B_y$ is positive, the main forces are the pressure-gradient, Coriolis, and ion-drag forces. The neutral flow is accelerated by the rotational component of the ion drag and involves an approximate balance of the pressure-gradient, Coriolis, and divergent/convergent ion-drag forces (mainly Hall ion-drag, as shown in Figure 6) in the polar cap. That is, the neutral flows are sustained by modified geostrophic balance [Larsen and Walterscheid, 1995]. On the other hand, for positive $B_y$, in comparison with negative $B_y$, the horizontal advection acceleration is still significant around the polar cap. Thus, the convergent pressure-gradient, divergent Coriolis, and divergent horizontal advection accelerations all contribute to the maintenance of the cyclonic neutral flow in the polar cap region at 111 km.

Figure 7 shows vectors of the sum of all the acceleration terms, equal to the velocity tendency (i.e., the partial derivative of the velocity with respect to time), at 142 and 111 km. For all four IMF conditions, the sum of accelerations at 142 km is predominantly rotational and has a pattern with notable similarities to the ion-drag acceleration poleward −60°, indicating that ion-drag acceleration tends to be the largest contributor to the time rate of change of the neutral wind velocity even though it is generally smaller than the other accelerations. This feature is most pronounced when the IMF $B_z$ is negative, as discussed by K&R [2007]. At 111-km altitude, although the sum of accelerations has a pattern with similarities to the ion-drag acceleration (mainly Hall ion drag) in the polar cap, it no longer resembles the ion-drag acceleration at lower latitudes.

Ion–drag Acceleration at 142 km

Figure 4. Vectors of the Pedersen and Hall ion-drag accelerations at 142 km.
4.3. Effect of the IMF on the Momentum Forcing

[23] By considering the difference of winds and forces, obtained by subtracting values of a simulation run with zero IMF from those with nonzero IMF, we can emphasize the IMF dependency in order to examine more closely how the winds and forces are influenced by the orientation of the IMF.

[24] The difference of neutral winds at 142 and 111 km and the difference of thermospheric $E \times B$ drift over Southern Hemisphere high latitudes for the four IMF ($B_y$, $B_z$) values are shown in Figure 8. The difference of winds for negative $B_y$ show a high-latitude anticyclonic vortex with a stronger speed on the dawn side. The apparent center of the vortex lies off the pole toward early morning sector at 142 km but lies closer to the pole at 111 km. The difference winds for positive $B_y$ show a high-latitude cyclonic vortex and equatorward flow in the midnight and post-midnight sectors. At 111 km, there is a roughly symmetric vortex over the polar cap. The difference neutral winds for negative $B_z$ have, at 142 km, a two-cell structure with evening anticyclonic and late-morning cyclonic vortices and extend to subauroral latitudes. There is also an equatorward flow in the early morning hours. The pattern of the cross-polar winds at $-70^\circ$ magnetic latitude is rotated toward later

Figure 5. Same as Figure 2 except at 111 km.
MLTs at 111 km. The negative-$B_z$ difference winds are stronger than the positive-$B_z$ difference winds, indicating that negative $B_z$ has a stronger effect on the winds than does positive $B_z$.

Figure 6. Same as Figure 4 except at 111 km.

[25] Figure 9 shows the difference accelerations corresponding the difference neutral winds at 142 km shown in Figure 8a. (Note that, because the horizontal advection acceleration is a nonlinear function of the wind, it cannot be

**Ion–drag Acceleration at 111 km**

- $B_y = -3.2$ nT
- $B_y = 3.2$ nT
- $B_z = -2.0$ nT
- $B_z = 2.0$ nT

Figure 7. Vectors of the sum of various accelerations acting on the neutral winds at 142 and 111 km altitudes over the southern summer hemisphere for IMF ($B_y, B_z$) values of (left to right) ($-3.2, 0.0$), ($+3.2, 0.0$), ($0.0, -2.0$), and ($0.0, +2.0$) nT.
When $B_y$ is negative, there is a strong anticyclonic vortex pattern of difference ion-drag acceleration at high latitudes, with maximum acceleration on the dayside around $-80^\circ$. There is also a sunward difference ion-drag acceleration in the early morning hours in the vicinity of the auroral oval. The difference ion-drag acceleration reflects well the pattern of the difference $E \times B$ drift shown in Figure 8b. Regions of strongly and weakly divergent difference pressure-gradient accelerations are located on the dawn side and in the postnoon sector. There is a convergence of the difference Coriolis acceleration vectors directed to the center of the polar cap because of the anticyclonic one-cell vortex shown in Figure 8a. This convergent difference Coriolis acceleration is offset by the divergent difference pressure-gradient and horizontal advection accelerations at roughly the same place. The difference horizontal advection acceleration has a strongly divergent pattern centered on the dusk side of the polar cap. Larger difference horizontal advection acceleration occurs on the dawn side inside the polar cap to reinforce the divergent difference pressure-gradient acceleration there. The horizontal advection acceleration also has a rotational component that counteracts part of the rotational difference ion drag, allowing a steady state mean zonal wind.

Differences not only in the signs but also in the patterns of the difference accelerations are apparent between negative and positive $B_y$, most clearly in the ion-drag and horizontal advection accelerations. The difference ion-drag acceleration disappears on the nightside polar cap and in the early morning sector for positive $B_y$, in contrast to that for negative $B_y$. There is a dusk convergence and a dawn divergence with a strong dawn-to-dusk difference horizontal advection acceleration across the polar cap for positive $B_y$, while there is a stronger divergence of the difference horizontal advection acceleration centered on the dusk side of the polar cap for negative $B_y$.

Unlike the negative- and positive-$B_y$ difference accelerations that have a simple structure over the polar cap and auroral oval, the negative-$B_z$ difference accelerations...
Difference Accelerations at 142 km

Figure 9. Vectors of the difference acceleration terms at 142 km over the Southern Hemisphere for IMF $(B_y, B_z)$ value of $(-3.2, 0.0)$, $(+3.2, 0.0)$, $(0.0, -2.0)$, and $(0.0, +2.0)$ nT: (top to bottom) the difference ion-drag, pressure-gradient, Coriolis, and horizontal advection accelerations. These vectors are plotted against a fixed length of arrow of 0.02 m s\(^{-2}\).

tions extend to subauroral latitudes. As discussed by K&R [2007], the pattern of ion-drag acceleration shows a two-cell rotational structure at 142 km. The maximum difference ion-drag accelerations occur in the vicinity of the auroral oval in the early morning hours, with a sunward direction. The difference pressure-gradient acceleration has divergent regions in the early morning and postnoon sectors in the vicinity of the auroral oval and convergent regions on the dusk and dawn sides of the polar cap. There are a convergent difference Coriolis acceleration on the dusk side and a divergent acceleration on the dawn side in the vicinity of the auroral oval because of the dusk anticyclonic and dawn cyclonic difference neutral flow shown in Figure 8a. The difference horizontal advection acceleration shows a strong divergence on the dusk side that reinforces the difference pressure-gradient acceleration and counteracts the difference Coriolis acceleration there. There is also a weak divergence of the difference horizontal advection accelerations.
tion on the dawn side that counteracts the difference pressure gradient and reinforces the difference Coriolis acceleration.

The difference acceleration terms at 142 km for positive $B_z$ are generally opposite in direction to those for negative $B_z$. In general, the magnitudes for positive $B_z$ are weaker than for negative $B_z$. The divergent difference pressure-gradient acceleration on the dawn side is offset by the convergent difference Coriolis and horizontal advection accelerations at roughly the same location. The divergent difference pressure-gradient and Coriolis accelerations on the dusk side offset the convergent difference horizontal advection acceleration.

Figure 10 illustrates vectors of the difference acceleration terms 111 km over the Southern Hemisphere. At 111 km, a nearly convergent pattern dominates the overall difference ion-drag acceleration for negative $B_y$. The difference pressure-gradient acceleration shows a divergent pattern radiating outward from the center of the polar cap. The difference Coriolis acceleration vectors direct to the center of the polar cap. The difference horizontal advection acceleration is small compared with other forcing terms. For positive $B_y$, the contribution of the difference horizontal advection acceleration is comparatively greater in comparison with that for negative $B_y$. The strengths of the difference pressure-gradient and Coriolis accelerations for
positive $B_y$ are stronger than for negative $B_y$. For negative $B_z$, the magnitudes of the difference pressure-gradient and Coriolis accelerations over the polar cap are much smaller than for the negative and positive $B_y$ cases. The divergence horizontal advection acceleration is relatively small. There are dusk convergence and dawn divergence patterns of the difference ion-drag acceleration that counteract the difference pressure-gradient acceleration and reinforce the difference Coriolis force. For positive $B_z$, the convergent difference Coriolis and ion-drag accelerations on the dawn side are offset by the divergent difference pressure-gradient acceleration there. The divergent difference Coriolis and ion-drag accelerations on the dusk side are offset by the convergent difference pressure-gradient acceleration at roughly the same place.

5. Summary and Conclusions

[30] To understand the forcing processes that control the high-latitude lower thermospheric dynamics below 170 km with respect to the direction of the IMF, we quantify the forces acting on the wind system as a function of the IMF orientation. We analyze steady state forcing for different IMF directions, for summer conditions in the Southern Hemisphere, on the basis of simulations with the NCAR TIE-GCM.

[31] The primary forces and the approximate balance of divergent/convergent accelerations for all IMF orientations are generally consistent with what was found for negative $B_z$ by K&R [2007]. A transition of the forcing patterns and their relative contributions to the high-latitude lower thermospheric wind system occurs around 120–130 km. At higher altitudes (above 130 km), the high-latitude lower thermospheric winds accelerated primarily by ion drag (mainly rotational Pedersen ion drag) involve an approximate balance among the Coriolis, horizontal advection, and pressure-gradient accelerations through the gradient-wind adjustment mechanism [Mikkelsen et al., 1981a, 1981b; Killeen and Roble, 1984; Killeen and Roble, 1986; Walterscheid and Brinkman, 2003]. At lower altitudes (below 120 km), the horizontal advection force tends to be much weaker, and the neutral flows accelerated by the rotational component of ion drag are maintained by the pressure-gradient and Coriolis forces and the largely divergent/convergent Hall ion-drag forces through the modified geostrophic balance process [Larsen and Walterscheid, 1995].

[32] However, the pattern and magnitude of the forces and their relative contributions to the high-latitude lower thermospheric wind system vary strongly with respect to the direction of the IMF. At higher altitudes, for negative $B_y$, strong anticyclonic winds accelerated primarily by the approximately rotational Pedersen ion drag are maintained by an approximate balance among the divergent/convergent Coriolis, horizontal advection, and relatively weak pressure-gradient accelerations. For positive $B_y$, the pressure-gradient acceleration is increased, while the inertial terms (Coriolis and horizontal advection accelerations) are reduced. The neutral circulations are accelerated by ion drag and are maintained mainly by an approximate balance between the Coriolis and horizontal advection accelerations on the dusk side and by an approximate balance among the pressure-gradient, Coriolis, and horizontal advection accelerations on the dawn side. For negative $B_z$, in comparison with negative and positive $B_y$, it is obvious that the forces extend to lower latitudes. The patterns of the accelerations for positive $B_z$ are similar to those for negative $B_y$, but the magnitudes tend to be significantly smaller. At the lower altitude of 111 km, especially for positive $B_z$, the horizontal advection acceleration is still significant around the polar cap. Thus, the convergent pressure-gradient, divergent Coriolis, and divergent horizontal advection accelerations mainly contribute to the maintenance of the cyclonic neutral flow in the polar cap region.

[33] The difference winds and accelerations, which are obtained by subtracting values for zero IMF from those for nonzero IMF, emphasize the differences with respect to the direction of IMF. At higher altitudes, while the difference winds and accelerations for negative and positive $B_z$ show a simple structure with a strong vortex near the pole, those for negative and positive $B_y$ extend to lower latitudes. Moreover, between negative and positive $B_y$, differences not only in the signs but also in the patterns of the difference accelerations are apparent, most clearly in the horizontal advection and ion-drag accelerations. The difference winds and accelerations for negative $B_y$ are stronger than those for positive $B_y$. At lower altitudes (111 km), the difference winds and accelerations for negative and positive $B_y$ are much stronger near the pole than for negative and positive $B_z$. Thus, although negative IMF $B_y$ has a stronger effect on the ionospheric convection, in general, than does negative or positive IMF $B_z$, the IMF $B_y$ component tends to dominate effects on the neutral winds in the polar cap at low thermospheric altitudes.

[34] As discussed by K&R [2007] for negative $B_z$, the nonlinear horizontal advection term is also a major force at higher altitudes for other IMF directions, although its pattern and magnitude vary with the direction of the IMF. At 111 km, for positive $B_y$, the horizontal advection force still contributes significantly to the balance of forces.

[35] For all IMF conditions, as discussed by K&R [2007] for negative $B_z$, the wind tendency (time rate of change) is predominantly rotational at higher altitudes and tends to resemble ion-drag acceleration poleward $-60^\circ$, indicating that ion-drag acceleration makes the dominant contribution to the neutral velocity tendency. This feature is most pronounced when the IMF $B_z$ is negative.

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References


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