Sources of low-latitude ionospheric $E \times B$ drifts and their variability
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[1] The complete mechanism of how upward propagating tropospheric tides connect to the upper atmosphere is not yet fully understood. One proposed mechanism is via ionospheric wind dynamo. However, other sources can potentially alter the vertical $E \times B$ drift: gravity and plasma pressure gradient driven current, the geomagnetic main field, and longitudinal variation in the conductivities. In this study we examine the contribution from E- and F-region neutral winds to the vertical drift shows that the geomagnetic main field do not change the longitudinal variation of the vertical drift significantly. Modifying the geomagnetic main field will change the vertical drift at 5–6 LT, 18–19 LT and 23–24 LT at almost all longitudes. In general the influence of the geomagnetic main field on the vertical drift is larger, with respect to the maximum difference, at 18–19 LT and 23–24 LT, equal at 5–6 LT, and smaller at 14–15 LT than the influence due to nonmigrating tidal components in the neutral winds. Examination of the contribution from E- and F-region neutral winds to the vertical drift shows that their importance depends on the local time and the solar activity. This implies that the vertical drift has to be analyzed at specific local times to examine the relation between the wave number in the vertical drift and in the neutral winds.


1. Introduction

[2] The general behavior of the average ionospheric electric field and the corresponding upward $E \times B$ drift at low latitude during the day is well understood [e.g., Richmond et al., 1976; Heelis, 2004; Kelley, 2009]. During the day the ions in the E-region are dragged along by the neutral winds and an eastward electric field is set up to enforce current continuity. During the nighttime the vertical $E \times B$ drift is in general downward due to the westward electric field. The neutral wind at night is still important [e.g., Rishbeth, 1981; Crain et al., 1993; Fesen et al., 2000; Jin et al., 2008]. However, on a daily basis the electric field is very variable, and understanding the variability is a challenge since many factors play a role, e.g., geospace conditions, preconditioning, wave and tidal variability, seasons, local time and location [e.g., Fejer, 1997; Fuller-Rowell et al., 1997; Kelley, 2009].

[3] In recent years with more space based observations it has been possible to study the longitudinal variation of the ionosphere. IMAGE-FUV observation showed that the Equatorial Ionospheric Anomaly (ELA) in the early evening is enhanced at certain longitudes [Sagawa et al., 2005; England et al., 2006a]. Immel et al. [2006] made the connection that the 4-peak longitudinal structure in the IMAGE-FUV is related to the longitudinal variation in the upward propagating non-migrating tides which originate in the tropics. England et al. [2006b] showed that the daytime electric field in the E-region correlates well with the Global Scale Wave Model (GSWM) which has the effects of non-migrating tides included [Hagan and Forbes, 2002, 2003].

[4] Many studies examined the longitudinal variation of ionospheric and thermospheric quantities; to mention only a few: the ionospheric nightglow brightness from IMAGE-FUV [Sagawa et al., 2005], the equatorial electrojet based on Oersted was studied by Jadhav et al. [2002], a few years later the equatorial electrojet (EEJ) based on CHAMP was studied by Lühr et al. [2008] and with TIE-GCM by Dombia et al. [2007], vertical $E \times B$ drift from ROCSAT-1 [Kid et al., 2007], and from DMSP by Hartman and Heelis [2007], longitudinal and height structure of the F-region electron density based on FORMOSAT-3/COSMIC by Lin et al. [2007] and from CHAMP by Lühr et al. [2007], the F-region current by Park et al. [2010], and the longitudinal and seasonal variation

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of the $S_q$ current system by Pedatella et al. [2011]. In the lower thermosphere nitric oxide from the SNOE satellite showed nonmigrating tidal signatures [Oberheide and Forbes, 2008]. In the thermosphere CHAMP could observe signature of longitudinal variations in the zonal winds [Häusler and Lith, 2009] and in the neutral density [Liu et al., 2009].

[5] Different coupling mechanism have been proposed to explain the F-region longitudinal variations. Modeling studies help to understand possible mechanisms. The study from England et al. [2010] using SAMI2 (SAMII is another Model of the Ionosphere) and comparing with results from the Thermosphere Ionosphere Mesosphere Electrodynamics General Circulation Model (TIME-GCM) concluded that changes in the thermospheric $O/N_2$ ratio, meridional winds at F-region heights, and vertical drift around sunset are all possible mechanism to explain longitudinal variation in the nighttime ionosphere. Liu et al. [2009] examined the equatorial neutral mass density at 400 km altitude and found a longitudinal wave number-4 pattern which is not in phase with the Equatorial Ionosphere Anomaly (EIA), and therefore direct penetration of non-migrating tides into the upper thermosphere and modulation of the density via thermal budget influence by NO was proposed. Häusler and Lith [2009] examined multiple years of CHAMP accelerometer data to deduce zonal winds at 400 km altitude, and specified the seasonal variation of the wave number-4 pattern. Due to the large data set they could show that a nonmigrating diurnal tide with wave number 3 (DE3) that penetrates into the upper thermosphere is the primary cause of the observed zonal wind pattern. One of the most accepted coupling mechanism is the electrodynamic one explained below, although other mechanisms have to play a role, especially for the neutral thermosphere as indicated above.

[6] The electrodynamic coupling is driven largely by the E-region wind dynamo which produces an upward $E \times B$ drift during the day. Throughout the literature papers examined and noted different sources which influence the upward drift, and in the following we describe only a few relevant to our study. Hartman and Heeles [2007] studied longitudinal variations of the equatorial drift measured by DMSP at 9:30 local time (LT). They attributed part of the longitudinal variation in vertical drift in the topside ionosphere to the F-region meridional winds which drive zonal currents of different magnitude and direction at each foot point of the flux tubes. Fang et al. [2009] used ROCSAT ion drifts to drive the Global Ionosphere Plumassphere numerical model and showed that the model driven by the ROCSAT observations can reproduce the ionospheric observations. Fesen et al. [2000] used the Thermosphere Ionosphere Electrodynamics General Circulation Model (TIEGCM) to study the effect of the semidiurnal migrating component on the day time upward $E \times B$ drift, and found that the semidiurnal migrating tidal component largely determines the magnitude and phase of the daytime upward $E \times B$ drift. Fang et al. [2008b] used the TIEGCM to examine the influence of neutral winds in different altitude regions on the upward $E \times B$ drift and the magnetic perturbations at the ground, and found that the altitude variation of the neutral winds at low latitudes can change the ground magnetic perturbations a few degrees off the equator. Vichare and Richmond [2005] suggested among other things that the longitudinal variation of the declination will introduce longitudinal dependence in the vertical drift. Park et al. [2010] examined vertical current strength in the F-region based on CHAMP magnetic field observations and pointed out that wave number-4 longitudinal variation could also originate from a wave number-4 pattern in the F-region Pedersen conductivity. Eccles [2004] examined and discussed the influence of gravity driven current due to the $m g \times B_e(qB_e^e)$ drift of $O^+$, where $m$ is the ion mass, $g$ is the gravitational constant, $B_e$ is the geomagnetic main field with magnitude $B$, and $q$ the charge, and pressure gradient current due to the $-\nabla P \times B_e(qB_e^e)$ drift of electrons and $O^+$, where $\nabla P$ is the total plasma pressure gradient, on the $E \times B$ drift. He showed that the gravity and plasma pressure gradient driven current will change the upward $E \times B$ drift up to 15 m/s at dawn and dusk. The gravity driven current was observed by the CHAMP satellite [Maus and Lith, 2006]. The magnitude of the gravity driven current varies with the $O^+$ density. In the morning and early evening when strong east–west gradients in the $O^+$ exist due to the build up and decay of the ionosphere, respectively, polarization electric fields are set up to maintain current continuity. During the day gradients in the $O^+$ densities produce negligible electric fields, since the current continuity is maintained through currents connecting to the highly conducting E-region. The effect of the pressure gradient term is much smaller than the gravity driven current since this current closes naturally along pressure contours. Maruyama et al. [2005] examined the effect of penetration electric field and disturbance dynamo on the equatorial electric field for the March 2001 storm. In the early stage of the storm the penetration electric field is dominant during the daytime. However, at night both effects are comparable in magnitude. They found that both mechanisms are strong enough to restructure the low latitude ionosphere.

[7] Although there have been many studies about the upward $E \times B$ drift at low latitude and many possible sources identified to contribute to the upward $E \times B$ drift no study is known to have examined the upward $E \times B$ drift in a systematic way. The present paper does a first step to quantify the contribution to the vertical drift from the different sources. This will help to better understand what causes changes in the vertical drift and therefore the possible influence on the coupling of the E- and F-region.

[8] In the following we try to distinguish between the magnitude and the longitudinal variation of the average low-latitude upward $E \times B$ drift whenever it is possible. We think that the distinction is important because an increasing number of studies focus on the tidal signature in the F-region [e.g., England et al., 2006a; Forbes et al., 2009; Häusler et al., 2010; Lin et al., 2007; Oberheide et al., 2009; Park et al., 2010; Huang et al., 2010; Liu et al., 2009] and an electrodynamic connection [e.g., Sagawa et al., 2005; Immel et al., 2006; England et al., 2010; Hagan et al., 2007; Jin et al., 2008, 2011; Akmaev, 2011]. The longitudinal variation of the upward $E \times B$ drift is one mechanism connecting the E-region to the longitudinal variation of the F-region [e.g., Immel et al., 2006; Kil et al., 2007; Jin et al., 2008; England et al., 2010]. The magnitude of low latitude upward $E \times B$ drift can contribute to the onset of E-region irregularities at certain local times [e.g., Kelley et al., 1981; de la Beaujardière et al., 2009; Huba et al., 2010].

[9] The paper is organized as follows. In section 2 we introduce the numerical experiment using output from the TIME-GCM. In section 3 we discuss the magnitude and
diurnal variation of the upward $E \times B$ drift due to the neutral winds, penetration electric field, gravity and gradient pressure currents. The longitudinal variation of the upward $E \times B$ drift is examined in section 4 for selected local time bins. In section 5 the neutral wind contributions with respect to migrating and nonmigrating tides, including secondary waves, as well as the contributions due to neutral winds at different altitudes and directions, are studied. Changes in the longitudinal variation of upward $E \times B$ drift with solar cycle are discussed in section 6. The effects of the conductivities and the geomagnetic main field variation on the longitudinal variation of the upward $E \times B$ drift are shown in section 7. The main conclusions are summarized in section 8.

2. Model Simulation

[10] In this study we use a NCAR TIME-GCM run which was described by Hagan et al. [2007]. We summarize the most important points. The TIME-GCM is a three dimensional, time-dependent general circulation model. It simulates the dynamics, temperatures, composition and electrodynamics of the middle and upper atmosphere as well as the ionosphere from first principles. The TIME-GCM is described by Roble [1995] and Roble and Ridley [1996]. Richmond [1995] describes the electrodynamics used in TIME-GCM. The model uses a realistic geomagnetic main field model, the International Geomagnetic Reference Field (IGRF), with a modified apex coordinate system [Richmond, 1995]. At the lower boundary around 30 km tidal perturbations from the Global Wave Scale Model (GSWM02) [Hagan and Forbes, 2002, 2003] are specified. The migrating and nonmigrating, diurnal and semi-diurnal tidal components from westward propagating wave number 6 to eastward propagating wave number 6 are included in GSWM02. The perturbations of geopotential height and neutral temperature are prescribed while in this TIME-GCM version the neutral winds are calculated self-consistently at the lower boundary. The calculated neutral wind perturbations at the lower boundary agree well with the neutral wind perturbation from GSWM02. The upper boundary of TIME-GCM is approximately at 450–600 km depending on the solar activity. The $O^+$ flux and electron heat flux are prescribed at the upper boundary.

[11] For this study we use March equinox conditions with a medium solar radio flux number $F_{10.7} = 150$ sfu (1sfu = $10^{-22}$W/m$^2$/Hz). We run the TIME-GCM with a spatial resolution of 2.5° by 2.5° in latitude and longitude and 4 grid points per scale height for geomagnetic quiescent conditions with a cross polar cap potential drop of 30 kV. The hemispheric power of total precipitating electrons is 8 GW. We use the high latitude convection model from Heelis et al. [1982]. The model was run long enough to reach a diurnally reproducible state for day of year 80 (March 21), so that no transient features are present. In the following we use the ionosphere and thermosphere state from this representative day.

[12] For the numerical experiments in this study we solve the standard electrodynamo equation, which in three dimensions is given by

$$\nabla \cdot \left[ \sigma_p (\nabla \Phi)_z + \sigma_b \mathbf{b}_0 \times \nabla \Phi + \sigma_J (\nabla \Phi)_b \right] = \nabla \cdot \left[ \sigma_p \mathbf{u}_n \times \mathbf{B}_0 + \sigma_b \mathbf{b}_0 \times (\mathbf{u}_n \times \mathbf{B}_0) + \mathbf{J}_f \right]$$

(1)

with the Hall conductivity $\sigma_H$, the Pedersen conductivity $\sigma_p$, and the parallel conductivity $\sigma_J$. $\mathbf{b}_0$ is the unit vector in the direction of the geomagnetic main field. The neutral winds are given by $\mathbf{u}_n$, the ionospheric current due to gravity and plasma pressure gradient are $\mathbf{J}_f$. We assume that geomagnetic field lines are equipotential, setting the parallel electric field to zero. For solving for the electric potential we reduce equation (1) to two dimensions by integration along geomagnetic field lines using a modified apex coordinate system. We use the neutral winds $\mathbf{u}_n$, the conductivities $\sigma_H$ and $\sigma_p$, and the ionospheric current $\mathbf{J}_f$ from the representative day. We calculate the electric field and upward $E \times B$ drift every 10 min. We want to emphasize that the ionosphere, thermosphere and electrodynamics results from the representative day are self-consistent, but in the following analysis there is no feedback from the electrodynamics to the ionosphere or neutral atmosphere. This makes it possible to isolate the influence of the neutral winds, ionospheric currents, geomagnetic main field and conductivities on the upward $E \times B$ drift.

[13] The longitudinal variation during the daytime is dominated by the eastward propagating diurnal tide with wave number 3 (DE3) (the eastward propagating semidiurnal tide with wave number 2 (SE2) is much smaller in TIME-GCM). Hagan et al. [2007] showed the magnitude of the DE3 zonal wind amplitude for this TIME-GCM simulation which is about 50 m/s around 105 km. Compared with the few observations available it was shown that TIME-GCM DE3 amplitude is probably too large. Oberheide et al. [2011] used Sounding the Atmosphere using Broadband Emission Radiometry (SABER) neutral temperature and TIMED Doppler Interferometer (TIDI) neutral wind measurements to deduce the DE3 components of neutral temperature and neutral winds in the E-region. They found a 6–12 m/s DE3 zonal neutral wind amplitude at 103 km. Häusler et al. [2010] compared zonal neutral wind derived from CHAMP accelerometer data with zonal neutral winds from TIME-GCM simulations at CHAMP altitude of 400 km. They found that for March the TIME-GCM (with $F_{10.7} = 75$ sfu) zonal neutral wind amplitude of the DE3 is 3 times larger than the ones derived from CHAMP observations (average $F_{10.7} = 128$ sfu). However, they also showed that with increasing solar flux the amplitude of DE3 in the F-region is decreasing which can explain part of the difference. For our study we reduced the DE3 amplitude of neutral winds and geopotential height of the self-consistent simulation to 1/3 the original model values to get better agreement with published results of the DE3 component. The conductivities are not reduced in a similar way, and the longitudinal variation of the conductivities might be too large. However, we show in section 7 that the longitudinal variation of the conductivities (except the geomagnetic main field variation) has a minor influence on the longitudinal variation of the upward $E \times B$ drift.

3. Magnitude of the Upward $E \times B$ Drift

[14] Before we study the details of the vertical $E \times B$ drift as simulated by TIME-GCM the general behavior of the vertical drift is briefly put into perspective. Fejer et al. [2008] binned vertical drift from ROCSAT-1 satellite for $K_p < 3$ and adjusted them to a solar radio flux of 150 sfu for
equinox conditions (March–April, September–October). The longitudinal variation over local time is shown in the middle plot of their Figure 4 [Fejer et al., 2008] which is reproduced in the present paper in Figure 1 (right). The day time peak in TIME-GCM in Figure 1 (left) is around 13–14 LT while ROCSAT-1 data as well as the Scherliess-Fejer empirical model [Scherliess and Fejer, 1999] shown for 12 UT in Figure 2 have the day time peak between 10–11 LT. The differences might be caused by inaccurately modeling the tides in TIME-GCM. The magnitude of the day time vertical drift in TIME-GCM is in general smaller than the one derived from ROCSAT-1 observations and the Scherliess-Fejer empirical model. The local time of the prereversal enhancement (PRE) from TIME-GCM, ROCSAT-1 observations and Scherliess-Fejer empirical model agrees within half an hour. The magnitude of the PRE in the TIME-GCM model is smaller than seen by ROCSAT-1 and the Scherliess-Fejer empirical model. In the 4–6 LT sector the vertical $\mathbf{E} \times \mathbf{B}$ from TIME-GCM as well as based on ROCSAT-1 data shows strong longitudinally varying downward drift. The longitudinal variation of the downward drift of TIME-GCM and ROCSAT-1 is similar; however, TIME-GCM has a sharper local time gradient between 5 and 6 local time than the one shown in the ROCSAT-1 data. Additional TIME-GCM simulations indicate that the local time gradient of the vertical drift in the 5–6 local time sector is sensitive toward the ratio between day- and nighttime ionization, which depends on the nighttime ionization rates, the E-region ionization rates, and the prescribed $O^+$ flux at the upper boundary. Modeling a realistic nighttime ionosphere is still part of ongoing research. The simplified prescribed upper boundary $O^+$ flux is upward during the daytime and downward at night, and does not change with latitude or geophysical conditions. The E-region conductivities in the model are lower than the ones based on the IRI and MSIS empirical models. This effect will be discussed later in this section.

[15] Overall TIME-GCM can reproduce the general local time and longitudinal variation of the observed vertical $\mathbf{E} \times \mathbf{B}$ drift. However, the exact longitudinal location of day time enhancements is shifted in some sectors. Hagan et al. [2007] already pointed out that the TIME-GCM simulation of the F-region electron density peaks are shifted in longitude with respect to IMAGE-FUV observations [Immel et al., 2006]. One source of the longitudinal variation are nonmigrating tides which are excited in the troposphere due to latent heat release. Hagan et al. [2007] speculated that one possible reason can be the difference between the climatological GSWM02 tidal forcing and the real tropospheric tidal forcing.

Figure 1. (left) Upward $\mathbf{E} \times \mathbf{B}$ drift [m/s] at magnetic equator as simulated by TIME-GCM for geomagnetic quiescent March equinox conditions with solar flux F10.7 = 150. (right) Equatorial vertical drifts for the average solar flux conditions ($Sa = 150$) during the ROCSAT-1 mission for equinox conditions (March, April, September, October) (reproduced from Fejer et al. [2008, Figure 4b]).

Figure 2. Upward $\mathbf{E} \times \mathbf{B}$ drift [m/s] at geomagnetic equator at 12 UT due to neutral winds, gravity and gradient pressure current, penetration electric field (solid/green lines); due to neutral wind (long dashed/blue lines); gravity and gradient pressure current (short dashed/orange lines), penetration electric field (dash-dotted/red lines), compared to Scherliess/Fejer empirical model [Scherliess and Fejer, 1999] (thin purple double dotted line).
above approximately 140 km (short dashed/orange lines). Neutral wind in the E-region below approximately 140 km (long dashed/blue lines); due to neutral wind in the F-region above approximately 140 km (short dashed/orange lines).

3.1. Contributions to the Magnitude of the Vertical $E \times B$ Drift

[16] First, we examine the magnitude of the upward $E \times B$ drift due to different contributions i.e. neutral winds, gravity and plasma pressure gradient driven current, and penetration electric field at 12 UT in Figure 2. For calculating the upward drift due to neutral wind we set in equation (1) the ionospheric gravity and gradient pressure current $I$ to zero, and reduce the cross polar cap potential drop which defines the high latitude convection pattern to 0.1 kV, so that the penetration electric field is negligible. For the upward $E \times B$ drift due to penetration electric field we set the neutral wind $u_\parallel$ and the ionospheric current $J_\parallel$ to zero in equation (1), and solve for the electric potential. For the influence of the ionospheric current $J_\parallel$ on the upward $E \times B$ drift we set the neutral wind $u_\parallel$ to zero, and reduce the cross polar cap potential drop to 0.1 kV. The total upward $E \times B$ drift is shown by the solid green line. The penetration electric field (red/dot-dashed line) is small in our study since this simulation was done for geophysical quiet time. In the early morning between 4–6 LT and evening around 18–20 LT the gravity and plasma pressure gradient driven current is up to 12 m/s and $-6$ m/s, respectively. In the early morning and evening the ionosphere is building up and decaying, respectively, and since the conductivities are small polarization electric fields are set up [Eccles, 2004]. The magnitude of change is larger in Eccles [2004] by almost a factor of two. Eccles used a model which is based on flux tubes with an apex height of 2000 km. One possible reason for the difference in magnitude could be that the TIME-GCM upper boundary, which is approximately at 500–600 km, is low, and has a prescribed $O^+$ flux. Compared to the International Reference Ionosphere (IRI) empirical model the local noon $O^+$ densities integrated from 90 to 1000 km at the equator are approximately 50% larger than the integrated TIME-GCM $O^+$ densities through the TIME-GCM model domain (90 to 510 km). TIME-GCM underestimates the integrated $O^+$ densities which can account for the magnitude difference in gravity and gradient pressure driven current. In our simulation the upward $E \times B$ drift due to the gravity driven current is counterbalanced by the penetration electric field which depends strongly on the geophysical condition. At all local times the major contribution to the upward $E \times B$ drift is due to the neutral wind driven dynamo shown by the long dashed blue line.

3.2. E- and F-Region Neutral Wind Contributions to the Vertical $E \times B$ Drift

[17] The neutral wind at all altitude regions can potentially contribute to the upward $E \times B$ drift. However, the importance of the wind in different altitude regions also depends on the magnitude of Hall and Pedersen conductivity. In the following we examine if the winds in a particular altitude region are a major contributor to the upward $E \times B$ drift at low latitudes. We distinguished between E-dynamo winds, defined as neutral winds acting in the E-region below approximately 140 km (pressure level–3 in our simulation under medium solar flux conditions) where the Hall conductivity is dominant, and the F-dynamo winds acting above approximately 140 km, where the neutral wind is growing in strength, but the Hall conductivity is negligible compared to the Pedersen conductivity [see Fambitakoye et al., 1976; Fang et al., 2008b]. The importance of the neutral winds in the different altitude regions depends very much on the local time considered [Crain et al., 1993; Fesen et al., 2000; Heelis, 2004]. In Figure 3 we show the contribution due to the neutral winds (green solid lines, same as the blue long dashed line in Figure 2). The upward $E \times B$ drift due to neutral winds only in the E-Dynamo region is shown by the long dashed blue line, while the upward $E \times B$ drift due to neutral winds in the F-dynamo region is shown by the shorted dashed red line. One can see as expected that during the nighttime the winds in the F-dynamo region are more important than in the E-region where the conductivity is very small at nighttime. During the daytime neutral winds in the E-dynamo region are more important, but still the neutral wind in the F-dynamo region is also relevant in our simulation. However, we found that during the daytime it is the lower F-region, below approximately 200 km, where the neutral winds are most important. In the TIME-GCM simulation the Hall conductivities are approximately 2/3 of the conductivities calculated with IRI [Bilitza et al., 1993] and Mass Spectrometer Incoherent (MSIS) [Hedin, 1991] empirical model. Fang et al. [2008a] increased the soft X-rays fluxes (wavelengths between 8 and 70 Å) by a factor of 4.4 in their TIE-GCM simulation to get a more realistic E-region, and better agreement with ground magnetometer data. TIE-GCM and TIME-GCM are very similar with respect to the E-region ionospheric physics. We did not tune the TIME-GCM E-region ionosphere this way, since we are not comparing with observations. However, we keep in mind that the E-region is weak in the
model simulation, and the Hall to Pedersen conductance (field-line integrated conductivities) ratio is smaller. In general, this leads to a smaller E-dynamo region contribution.

4. Longitudinal Variation of the Upward $E \times B$ Drift

[18] In this section we focus on the longitudinal variation of the upward $E \times B$. We solve the electrodynamo equation (1) using the neutral wind, conductivities and ionospheric current from the representative March equinox day. We calculate the average of the upward $E \times B$ drift between $\pm 30^\circ$ geomagnetic latitude using 10 min model data and bin the values according to solar local time in one hour LT bins. We tested the sensitivity of the average upward $E \times B$ drift due to the latitude range. Using a smaller latitude range gives somewhat larger amplitudes but the general behavior is similar. Therefore we use the larger range to show that the results are not limited to a small latitudinal region. The local time bins we examine in the following are 5–6 LT, 13–14 LT, 18–19 LT, and 23–24 LT.

**Figure 4.** Average upward $E \times B$ drift [m/s] between $\pm 30$ deg magnetic latitude for (top left) 5–6 LT, (top right) 13–14 LT, (bottom left) 18–19 LT, and (bottom right) 23–24 LT: due to neutral wind in the E-region below approximately 140 km (blue/long dashed lines), neutral winds in the F-region above approximately 140 km (blue/dotted curves), due to gravity and plasma pressure gradient driven current (orange/dash-dotted lines), due to penetration electric field (red/dash-double dotted lines) and due to all the above contribution (black/solid lines).
Figure 4 shows the longitudinal variation of the upward $E \times B$ drift for the different local times with top left for the 5–6 LT bin, top right for the 13–14 LT bin, bottom left for the 18–19 LT bin, and bottom right for the 23–24 LT bin. Each of the four plots is organized in the same way with the average upward $E \times B$ drift shown over magnetic longitude. (Zero magnetic longitude is at approximately $-70^\circ$ geographic longitude.) The total upward $E \times B$ drift is shown by the solid black line.

In the 13–14 LT bin the regular longitudinal variation of the upward $E \times B$ drift is determined by the neutral winds. In the E- and lower F-region the nonmigrating tidal signature is still very strong, especially the DE3 components [Hagan et al., 2007; Oberheide et al., 2011] which leads to a wave-4 structure at a particular local time. We want to remind the reader that the TIME-GCM DE3 component was reduced to 1/3 of its value from the REMIND the reader that the TIME-GCM DE3 component was reduced to 1/3 of its value from the Hagan et al. study [2007] to better compare with observations from CHAMP [Häusler et al., 2010], TIDI, and SABER [Oberheide et al., 2011] (see section 2). In this simulation both the E-dynamo winds (blue long dashed) and the F-dynamo winds (blue dotted) contribute in almost equal parts to the longitudinal variation in the 13–14 LT bin. This result is in contrast to Fang et al. [2008b] who found using the related TIE-GCM with the lower boundary at 97 km that the day time drift is mainly driven by E-region winds below 123 km. There are some possible reasons for the different findings. Fang et al. [2008b] divided the altitude differently at 123 km versus 140 km in the present study. Below 123 km the Hall conductivity is larger than the Pedersen conductivity in their simulation. They forced the model at the lower boundary at 97 km with migrating tides from GSWM02, compared to using migrating and nonmigrating tides from GSWM02 at the lower boundary at 30 km in the present study, which, as the tides propagate upward, can result in the generation of secondary waves due to nonlinear interactions between different tidal components. Finally, Fang et al. [2008b] increased the E-region conductivity by increasing the soft X-ray fluxes to get better agreement with IRI electron densities. Because our E-region Hall conductivities are approximately 1/3 smaller than those based on IRI and MSIS, our Hall to Pedersen conductance (field-line integrated conductivities) ratio is smaller. This leads to a smaller contribution from the E-dynamo region compared to the F-dynamo contribution. The F-dynamo contribution (blue dotted) shows a structured wave-4 pattern as does the E-dynamo contribution. Further numerical experiments show that the F-dynamo contribution comes mainly from the lower F-dynamo region below 200 km. Ionospheric gravity and plasma pressure gradient current and penetration electric field do not contribute to the longitudinal variation in the 13–14 LT bin.

The 5–6 LT bin was chosen because the gravity driven current can be large in the morning, as shown by Eccles [2004]. Although the contribution from the gravity and plasma pressure gradient driven current is up to 8 m/s the longitudinal variation is not significant. The main contribution to the upward $E \times B$ drift comes from the neutral wind above 140 km (blue dotted) with some weak modulation from the neutral winds in the E-dynamo region (blue dashed). In the 18–19 LT bin the contribution from the gravity and plasma pressure driven current can be between $-6$ and $-9$ m/s. However, similar to the 5–6 LT bin, this ionospheric current does not contribute significantly to the longitudinal variation. Also for our geophysical conditions the penetration electric field is almost constant with respect to longitude in the 18–19 LT bin. As discussed before, the penetration electric field strongly depends on the geophysical condition. The neutral wind contributes the most to the upward $E \times B$ drift. However, different from the 5–6 LT bin for which the neutral wind in the F-region dominates the upward $E \times B$ drift, in the 18–19 LT bin the F-dynamo neutral winds lead to an upward $E \times B$ drift of approximately 30 m/s on average, but it is partly counteracted by the E-dynamo neutral wind effect which is around $-8$ m/s. The longitudinal variation of the upward $E \times B$ drift is mainly defined by the F-region neutral wind contribution.

In the 23–24 LT bin the neutral wind contribution to the upward $E \times B$ drift dominates the longitudinal variation. Ionospheric gravity driven current and penetration electric field have a minor influence. Different from the 18–19 LT bin for which the F-dynamo contribution was positive and the E-dynamo contribution negative, in the 23–24 LT bin E- and F-dynamo contributions lead to a downward $E \times B$ drift similar to the behavior in the 5–6 LT bin. However, in the 5–6 LT bin the F-dynamo contribution was dominant while in the 23–24 LT bin, although the F-dynamo has the much larger contribution to the downward drift, the longitudinal variation of the drift is still modulated by the E-dynamo neutral winds. In section 6 we will show that the importance of the E-dynamo wind in the 23–24 LT bin will increase with decreasing solar radio flux.

The importance of the neutral winds in the E- and F-region depends on the local time. This agrees with the general idea that at dawn, dusk, and during the nighttime due to the increased ratio of Pedersen to Hall conductances the F-region winds are more important [Crain et al., 1993; Heelis, 2004] than during the daytime. It was shown by Fang et al. [2008b] that the winds above 123 km dominate the upward $E \times B$ drift between dusk and dawn. Our study agrees qualitatively with it.

5. Contributions of Neutral Winds

In the following, we want to quantify the contribution of the migrating tidal components and the contributions due to the nonmigrating tidal components and the secondary waves due to nonlinear interactions. Migrating tidal components move westward with the apparent motion of the sun, and they have periods which are harmonics of the solar day. At a fixed local time migrating tidal components do not contribute to the longitudinal variation [Oberheide and Gusev, 2002]. Nonmigrating tidal components move faster or slower than the apparent motion of the sun. At a fixed local time they vary with longitude. The major migrating tidal components are the migrating diurnal and semidiurnal tides. The major nonmigrating component in the modeled dynamo region is the diurnal eastward propagating tide with zonal wave number 3 (DE3) [Hagan et al., 2007]. Other nonmigrating components are present but not as strong in the E-region [Oberheide et al., 2011; Hagan et al., 2009]. Due to the nonlinear interaction between different tidal components, secondary stationary planetary waves (SPW) can be generated, and they can be significant in the E-region, as shown by...
The difference in vertical components is shown by the thin lines, and the case with only migrating tidal components is shown by the thick lines, and the case with migrating and nonmigrating tidal components (thick lines) and with migrating components (thin lines).

Hagan et al. [2009]. Our labeling of “nonmigrating tidal case” refers to the tides specified at the lower boundary.

5.1. Longitudinal Variations Due to Migrating Tidal Components

[25] In the following we consider only the neutral wind contribution when calculating the upward $E \times B$ drift. We use our representative March equinox day which includes migrating and nonmigrating tidal forcing at the lower boundary. A spectrum of tidal components are present in the E- and F-region due to upward propagation and nonlinear interaction of the tides. For the migrating tidal component case we extracted the migrating tidal components from the neutral winds- namely, the diurnal, semidiurnal and tri-diurnal migrating components. We use the zonal mean wind condition of the representative March day as the background winds for the migrating only case since we do not have the self-consistent background winds for the migrating case. Therefore we neglect the changes in the zonal mean wind due to the different tidal forcing. With these reconstructed neutral winds we solve equation (1) for the electric field. We want to point out that this is different from a self-consistent run with only migrating tidal forcing at the lower boundary. However, in this study we are not focusing on the dynamics in the middle atmosphere which would include the nonlinear interaction between tidal components and with the zonal mean wind. Figure 5 shows the comparison for the different local times considered in section 4. The case we presented before which included migrating and nonmigrating tidal components is shown by the thick lines, and the case with only migrating tidal components is shown by the thin lines. The difference in vertical $E \times B$ drift between the two cases is due to nonmigrating tidal components and stationary planetary waves generated by nonlinear interaction. Probably the best known effect of nonmigrating tides is the distinct wave-4 structure during the day time (brown, short dashed, thick line) in 13–14 LT bin. The diurnal eastward propagating tide with zonal wave number 3 (DE3) is the major nonmigrating component in the model [Hagan et al., 2007] in the E-region. Migrating components do not vary with longitude at a fixed local time and therefore the regular wave number 4 pattern disappears (brown, short dashed, thin line). At the other local time bins the contribution of nonmigrating tides cannot be easily characterized. The small scale longitudinal variation is reduced, although the larger scale longitudinal variation remains the same, and is still significant. The largest absolute difference between the cases with and without nonmigrating tides is in the 13–14 LT bin (~7.4 m/s). The maximum change in the longitudinal variation between the two cases (absolute difference between maximum and minimum difference) is very similar at the different local times: at 13–14 LT bin (10.7 m/s), 5–6 LT bin (10.6 m/s), 18–19 LT bin (9.2 m/s), and 23–24 LT bin (8.6 m/s). In section 7 we will compare these changes with the changes due to the geomagnetic main field configuration.

[26] The longitudinal variation in the migrating tidal case is still significant and could be caused by the longitudinal variation of the conductivities and by the longitudinal variation of the geomagnetic main field. In the following we use results from this study to propose likely causes. In section 7 we show that the longitudinal variation of the conductivity has only a minor effect on the longitudinal variation of the upward $E \times B$ drift. But the geomagnetic main field has a large and very distinct influence on the longitudinal variation of the upward $E \times B$ drift. Therefore we suggest that the main contribution to the longitudinal variation of the upward $E \times B$ drift with migrating tides is mainly due to the interaction of the neutral winds with the longitudinal varying geomagnetic main field. In future studies we can quantify the contribution to the longitudinal variation due to the geomagnetic field strength, orientation or geometry.

5.2. Longitudinal Variation Due to Zonal and Meridional E- and F-Region Winds

[27] In Figure 4 we showed that depending on the local time, neutral winds in the E-dynamo region or in the F-dynamo region dominate the longitudinal variation of the upward $E \times B$ drift. In the following we examine the influence of the zonal and meridional wind components on the upward $E \times B$ drift. We focus on daytime (14–15 LT bin), and around midnight (23–24 LT bin). Previously we showed results from the 13–14 LT bin, which are very similar to the ones from the 14–15 LT bin. However, for solar minimum conditions, which are presented in section 6, the strongest wave-4 longitudinal signature can be found in the 14–15 LT bin, which is very similar to that for the solar medium case in the 14–15 LT bin. Therefore we decided to show the 14–15 LT bin to be able to better compare with the solar minimum case presented in section 6. We show the average upward $E \times B$ drift only due to neutral wind (migrating and nonmigrating tidal components included), but excluding effects from ionospheric gravity and pressure gradient driven current and penetration electric fields. Figure 6 shows on the left side the average upward $E \times B$ drift between ±30 deg magnetic latitude due to neutral wind for 5–6 LT (blue/long dashed), 13–14 LT (brown/dotted), 18–19 LT (red/dash-dotted), and 23–24 LT (green/solid) with migrating and nonmigrating tidal components (thick lines) and with migrating components (thin lines).

Figure 5. Average upward $E \times B$ drift [m/s] between ±30 deg magnetic latitude due to neutral wind for 5–6 LT (blue/long dashed), 13–14 LT (brown/dotted), 18–19 LT (red/dash-dotted), and 23–24 LT (green/solid) with migrating and nonmigrating tidal components (thick lines) and with migrating components (thin lines).
drift for the 14–15 LT bin, and on the right for the 23–24 LT bin. In the 14–15 LT bin the wave-4 structure comes from the geographically zonal neutral wind in the E-dynamo (thin blue dashed double dotted line) and in the F-dynamo (thin green dashed double dotted line) region. We want to remind the reader that the E-region Hall conductivities are underestimated in this TIME-GCM simulation (see section 3) which will lead to a weaker E-dynamo contribution compared to the F-dynamo contribution. Although the zonal wind has a distinct wave 4 pattern the contribution to the upward $E \times B$ drift from the geographically meridional wind is of the same magnitude, only with less longitudinal variation. In the 23–24 LT bin the longitudinal variation comes mainly from the F-dynamo neutral winds (solid green line); however, neither the zonal nor the meridional wind component is dominant (thin green lines). The contribution from the E-dynamo neutral winds (thin blue lines) are much smaller, and do not contribute significantly to the longitudinal variation of the upward $E \times B$ drift. We studied the longitudinal variation of the upward $E \times B$ drift caused by the zonal wind in the F-region for every local time (only two local time bins shown in this study). We could observe that the drift goes smoothly from being upward at dusk (e.g. 18–19 LT around 20 m/s) to a very small contribution around midnight (23–24 LT around −5 m/s), to negative/downward drift in the morning (5–6 LT around −10 to −15 m/s). Therefore in the 23–24 LT bin the zonal F-region winds contribution (thin green dashed double dotted line) to the upward drift is very small. To explain the behavior of the zonal wind in the F-region it would be necessary to study the height integrated Pedersen weighted winds which is not within the scope of the present study.

One implication of the varying importance of the neutral winds with direction and altitude at different local times is that it is not meaningful to analyze the upward $E \times B$ drift as thermospheric quantities are to derive period and zonal wave number. To examine the relation between the wave number in the upward $E \times B$ drift and the wave number in the neutral winds the upward $E \times B$ drift has to be analyzed at specific local times.

6. Variation With Solar Cycle

The present study was done for strong solar medium conditions with an $F_{10.7}$ solar radio flux number of 150 sfu. For solar minimum condition the ratio of F-region to E-region conductivities is much smaller [Takeda et al., 1986]; however, Häusler et al. [2010] showed that during solar minimum conditions the direct penetration of the tides into the F-region is stronger than under increased solar activity. Therefore the magnitude and longitudinal variation of the upward $E \times B$ drift will differ. In section 3 we found that the neutral wind is the major contributor to the magnitude of the upward $E \times B$ drift. Therefore we focus on the neutral wind contribution when we examine the average upward $E \times B$ drift at low latitudes under solar minimum conditions with $F_{10.7} = 75$ sfu. As for the solar medium model simulation we reduced the amplitude of the diurnal eastward propagating tide with wave number 3 (DE3) to 1/3 of its original value. We separated the neutral wind into two regions separated at pressure level $z = 3$, which corresponds approximately to 135 km for the solar minimum. In the E-dynamo region (approximately up to 135 km) the Hall conductivity exceeds the Pedersen conductivity, while in the F-dynamo region (approximately above 135 km) the Pedersen conductivity is dominant. Overall the absolute magnitude of the upward $E \times B$ drift is decreasing with decreasing solar radio flux. This was shown previously by empirical models [e.g., Scherliess and Fejer, 1999], observations [e.g., Fejer et al.,
2008], and numerical models [e.g., Takeda et al., 1986]. The change in magnitude of the upward $E \times B$ drift for a specific local time is mainly due to the decrease in F-region plasma density, and therefore decrease in Pedersen conductivity [Takeda and Araki, 1985]. The Hall conductivity does not change so significantly, decreasing by about 25% from solar medium to solar minimum at 14 LT at its peak altitudes, while the Pedersen conductivity decreases in the upper F-region by approximately 80%. This agrees qualitatively with findings from Takeda and Araki [1985] that between solar maximum and minimum the E-region conductivities vary only by roughly a factor of 1/2 while the F-region conductivities vary by up to a factor of 1/50 from solar maximum to minimum.

7. Contributions of the Conductivities and the Geomagnetic Main Field Model

7.1. Contributions of the Conductivities

[31] In the following we examine the effects of the longitudinal variation of the conductivities on the longitudinal variation of the upward $E \times B$ drift. [32] Park et al. [2010] examined F-region currents derived from CHAMP observations. They discuss that one possibility to induce longitudinal variation in the current is the longitudinal variation of the Pedersen conductivities. However, no coordinated observations are available to deduce the longitudinal variation of the conductivities. Liu et al. [2009]
used CHAMP observations to examine the neutral and ion density longitudinal variations. They applied a Fourier transformation to extract the dominant wave number 4 component of the equatorial ionization anomaly maximum and the equatorial mass density anomaly maximum. They showed that in the crest region (within ±10°–20° magnetic latitude) the equatorial mass anomaly longitudinal maxima are 30° east of the equatorial ionization anomaly maxima. However, in the trough region (within ±5° magnetic latitude) the equatorial mass density anomaly longitudinal maxima are only 10° east of the equatorial ionization anomaly maxima. The F-region Pedersen conductivity depends on the product of ion and neutral density. Since variations in neutral and ion densities are not always in phase this could indicate that the longitudinal variation in the conductivity is small.

[33] For the following numerical experiment we recalculate the conductivities using the ion and neutral density composition, and the ion, electron and neutral temperature for our representative March equinox day. We organize the ion and neutral density composition and the temperatures with respect to longitude and local time over geographic latitude to quantify the longitudinal variation for each quantity. We remove the longitudinal variation for each field, i.e., ion and neutral composition, and ion, electron and neutral temperature. With these longitudinal invariant fields we calculate the conductivities which we use in equation (1) to get the electric field. Note that the conductivities still vary with longitude since we did not remove the longitudinal variation of the geomagnetic main field. We removed the longitudinal variation at all latitudes including at high latitude from the aurora conductivities, which is not realistic. So we would expect that the result shows even larger differences than using unmodified conductivities in the aurora region. In this numerical test we include neutral wind effects together with effects from ionospheric gravity and pressure-gradient currents and penetration electric fields. Figure 8 shows the average upward $E \times B$ drift for the four different local time bins. The maximum absolute difference is in the 23–24 LT bin with 5.8 m/s, followed by the 5–6 LT bin with 3.9 m/s, then 18–19 LT bin with −3.14 m/s, and in the 13–14 LT bin with −2.9 m/s. It can be seen that there is only a minor influence on the longitudinal variation of the upward $E \times B$ drift due to calculating the conductivities with longitudinal invariant ion and neutral density composition, and ion, electron and neutral temperatures. The changes in the upward $E \times B$ drift are much smaller than due to the inclusion of nonmigrating tidal components examined in section 5.

7.2. Contributions of the Geomagnetic Main Field Model

[34] Next we study the effect of the geomagnetic main field on the upward $E \times B$ drift. Changes in the vertical drift due to the geomagnetic main field have already been studied. Hartman and Heeles (2007) examined the longitudinal variation of the topside ionosphere at 9:30 LT using DMSP ion drift data. They found a correlation between vertical drift changes and season as well as declination angle. They attributed the correlation to meridional winds and conductivity differences at the foot point of the magnetic field line in the two hemispheres with respect to the terminator. Vichare and Richmond [2005] used a global circulation model the Magnetosphere-Thermosphere-Ionosphere Electrodynamics-GCM (MTIEGCM) to study the longitudinal variation of the vertical drift in the evening sector at the geomagnetic equator. They found that the model simulation agreed with observation with respect to the increased strength in the vertical drift in the American-African sector, where the geomagnetic field is weaker and more distorted. They examined possible causes for the longitudinal variation of the vertical drift by examining the conductivities, zonal neutral wind, geomagnetic declination, but found no clear correlation. The following numerical experiment quantifies the effect of the geomagnetic field on the longitudinal variation of the upward $E \times B$ drift based on the TIME-GCM without any feedback to the ionosphere and thermosphere. We use the neutral winds and conductivities from the representative March equinox day for solar medium conditions. In the following we consider only the neutral wind effect when calculating the electric fields. We vary the geomagnetic main field $B_0$ in equation (1) from the full IGRF approximation to a tilted dipole field. Note that the conductivities remain unchanged and they are calculated using the full IGRF magnetic field. The tilted dipole magnetic equator has a simple cosine variation with respect to the geographic equator, while the full IGRF is most distorted in the 0°–90° geomagnetic longitude region. The difference between the two geomagnetic equator approximation is up to 15° in latitude in this region. However, the two geomagnetic field approximations differ at all latitudes and longitudes in the geomagnetic field strength and direction. Another changes is the mapping of the neutral wind and the
conductivities from the geographic coordinate system to the geomagnetic coordinate system (dipole or full IGRF with apex coordinates). Figure 9 shows the average upward $\mathbf{E} \times \mathbf{B}$ drift over geomagnetic longitude for the four local time bins.

The absolute differences are largest at 18–19 LT with a minimum difference of $-9.1$ m/s, followed by 23–24 LT with a maximum difference of 7.6 m/s. The magnitude of the changes (difference between maximum and minimum of difference for a certain local time bin) may be an indicator for the change in shape of the longitudinal variation. The maximum-minimum difference is largest for 18–19 LT (15 m/s), then 23–24 LT (12 m/s), followed by 5–6 LT (9.8 m/s), and 13–14 LT (4.9 m/s). From Figure 9 it can be seen that the shape of the curve does not change for 13–14 LT bin, which also has the smallest change in maximum-minimum difference. At 18–19 LT the longitudinal variation has a peak around $-10^\circ$ geomagnetic longitude using the dipole approximation (red/dashed dotted thin line) which is not there using the full IGRF approximation (thick red/dashed dotted line). In the 23–24 LT bin the shape of the longitudinal variation is changed mainly in the $-45^\circ$ to $180^\circ$ geomagnetic longitude sector. The changes due to a different geomagnetic main field approximations are larger than the changes due to the inclusion of the nonmigrating tides with the exception of the 13–14 LT bin for which the nonmigrating tides have a larger influence on the upward $\mathbf{E} \times \mathbf{B}$ drift than the geomagnetic main field model. In an addition numerical experiment we used the tilted dipole approximation also for the conductivity calculation. The upward $\mathbf{E} \times \mathbf{B}$ drift differences between using the full IGRF (Figure 9, thick lines) and a tilted dipole approximation in equation (1) (Figure 9, thin lines) are larger than the differences between using a tilted dipole approximation for both the conductivity calculation and the electrodynamo equation versus only for the equation (1). Only in the 5–6 LT bin is the change in the upward $\mathbf{E} \times \mathbf{B}$ drift due to different geomagnetic main field models for the conductivity calculation larger than the change in the upward $\mathbf{E} \times \mathbf{B}$ drift due to the different geomagnetic main field models in the electrodynamo equation alone.

8. Summary

This numerical study was done for March equinox conditions for which the ionospheric and thermospheric response in both hemispheres is more or less symmetric. Since the conductivities and the neutral wind system is changing with season this study can only be used as guidance for the contributions to the upward $\mathbf{E} \times \mathbf{B}$ drift at other months.

In the following we summarize the main findings:

1. The longitudinal variation of the ion and neutral density composition, and ion, electron and neutral temperatures in the conductivity calculation does not have a significant contribution to the longitudinal variation of the upward $\mathbf{E} \times \mathbf{B}$ drift.

2. The second smallest effect adding to the longitudinal variation of the upward $\mathbf{E} \times \mathbf{B}$ drift at low latitudes is the gravity and plasma pressure gradient driven current. Although the magnitude of the upward $\mathbf{E} \times \mathbf{B}$ drift due to these currents can be up to $8$ m/s at 5–6 LT and $-9$ m/s at 18–19 LT, the longitudinal variation is small. This is not a significant source of the longitudinal variations in upward $\mathbf{E} \times \mathbf{B}$ drift, contrary to the speculation by Hartman and Heelis [2007] that it may explain longitudinal variation in the ionospheric topside equatorial vertical drift. However, the change in magnitude in the average low-latitude upward $\mathbf{E} \times \mathbf{B}$ drift due to gravity driven current can be important for the initiation of plasma irregularities [e.g., Kelley et al., 1981; de la Beaujardière et al., 2009; Huba et al., 2010].

3. A change in the geomagnetic main field configuration from full IGRF to a tilted dipole approximation can have a major influence on the longitudinal variation of upward $\mathbf{E} \times \mathbf{B}$ drift at 18–19 LT (15 m/s), 23–24 LT (12 m/s), and 5–6 LT (9.8 m/s). Note that the conductivities were not recalculated, and used the full IGRF approximation. The result gives some guidance about what longitudinal variation differences to expect when ionospheric and thermospheric quantities are exchanged between numerical models with different geomagnetic main field approximations. Since the longitudinal variation of upward $\mathbf{E} \times \mathbf{B}$ drift will change due to the change in the geomagnetic main field the F-region response due to the electrodynomic coupling will be different. Changing the geomagnetic field approximation additionally in the calculation of the conductivity had only a small effect, and was only in the 5–6 LT bin important.

4. The neutral winds is the main contributor to the upward $\mathbf{E} \times \mathbf{B}$ magnitude and longitudinal variation. However, depending on the local time neutral winds in different directions and altitude regions are important. During the daytime in the 13–14 LT bin the neutral winds below and above 140 km contribute approximately in equal parts to the upward $\mathbf{E} \times \mathbf{B}$ drift, with the caveat that the Hall conductivities in our model simulation are too small compared to...
empirical models. The longitudinal variation is driven by zonal neutral winds at all altitudes, while the meridional neutral wind contributes to the magnitude of the upward $E \times B$ drift. In the early morning in the 5–6 LT bin the neutral wind above approximately 140 km has the main contribution to the magnitude and longitudinal variation of the upward $E \times B$ drift. The longitudinal variation is still slightly modulated by the neutral winds below 140 km. In the early evening in the 18–19 LT bin winds below and above 140 km contribute to the magnitude and longitudinal variation. However, the neutral wind above 140 km causes more small scale longitudinal variation in the upward $E \times B$ drift. At night in the 23–24 LT bin the main contribution comes from the neutral wind above 140 km but the longitudinal variation is modulated by the neutral winds below 140 km. There is no dominant wind direction during that time, although the meridional wind above 140 km contributes more to the magnitude while the zonal wind above 140 km contributes more to the longitudinal variation. One implication of the time local dependence of the neutral wind with respect to altitude and direction is that a space-time analysis is as it is done for the neutral thermosphere to extract zonal wave number and period is not meaningful for the upward $E \times B$ drift. For the upward $E \times B$ drift the analysis has to be restricted to the wave number at certain local times.

5. Nonmigrating tides contribute significantly to the longitudinal variation of the upward $E \times B$ drift during the daytime in the 13–14 LT bin. In the 5–6 LT bin, 18–19 LT bin, and 23–24 LT bin there is a strong longitudinal variation even without nonmigrating tides, and the nonmigrating tides do not contribute as much to the longitudinal variation as during the daytime.

6. The importance of the neutral winds with respect to altitude for the upward $E \times B$ drift depends on the solar cycle. In general for solar minimum conditions the neutral wind above 135 km diminishes in importance for the wind-dynamo. The longitudinal variation as well as the magnitude of the upward $E \times B$ drift in the 14–15 LT bin, and in the 23–24 LT bin is mainly driven by the neutral winds below 135 km. Especially at nighttime the neutral winds above 135 km have an insignificant contribution to the longitudinal variation of upward $E \times B$ drift. However, at 5–6 LT bin and 18–19 LT bin (not shown in this study) the neutral wind above 135 km still contributed to the longitudinal variation although not as much as for solar medium conditions.

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