The Role of Thermospheric Neutral Winds in the Mid-latitude Ionospheric Evening Anomalies

Levan Lomidze

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THE ROLE OF THERMOSPHERIC NEUTRAL WINDS IN THE MID-LATITUDE IONOSPHERIC EVENING ANOMALIES

by

Levan Lomidze

A dissertation submitted in partial fulfillment of the requirements for the degree of

DOCTOR OF PHILOSOPHY

in

Physics

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UTAH STATE UNIVERSITY
Logan, Utah
2015
ABSTRACT

The Role of Thermospheric Neutral Winds in the Mid-latitude Ionospheric Evening Anomalies

by

Levan Lomidze, Doctor of Philosophy
Utah State University, 2015

Major Professor: Ludger Scherliess
Department: Physics

One of the intriguing features of the F-region ionosphere are anomalous evening enhancements of the electron density over certain mid-latitude sites. The most prominent example of this enhancement is the Weddell Sea Anomaly. Although the evening anomalies have been known for several decades, their generation mechanisms are still under debate and their accurate modeling remains a challenge. In this dissertation, the role of thermospheric neutral winds in the generation of these anomalies is investigated.

Thermospheric winds play an important role in the dynamics of the F-region ionosphere, and, as it will be shown, in the generation of the evening anomalies. However, to date, their reliable estimation remains a challenge. To mitigate this shortcoming, data assimilation models were employed. First, seasonal global maps of F-region peak parameters (NmF2 and hmF2) from COSMIC radio occultation measurements were assimilated into the Global Assimilation of Ionospheric Measurements Full Physics (GAIM-FP) model. The model estimates magnetic meridional winds at low and mid-latitudes. GAIM-FP estimated winds were shown to be in good agreement with independent ground-based wind observations. Next, in order to address the role of neutral wind components in the generation of anomalies, a separate, 3-D physics-based Thermosphere Wind Assimilation Model (TWAM) was developed. TWAM is based on an implicit Kalman filter technique, and combines GAIM-FP magnetic meridional wind data with the equation of motion of the
neutral gas to provide the climatology of the thermospheric wind components. The neutral
wind components estimated by TWAM were also found to be in close quantitative agreement
with independent ground-based wind observations, and were shown to accurately reproduce
NmF2 and hmF2 over the anomalies.

To understand the physical mechanism behind the anomalies, the plasma production,
loss, and transport processes were analyzed. It was found that, due to the action of the
equatorward wind, the evening density maximum forms at altitudes where the recombin-
ation rate is relatively small. It was revealed that at this time and altitude, plasma loss due
to transport also weakens. As a consequence, the relative role of solar production increases
over the net loss process and the electron density enhancement occurs.
The Role of Thermospheric Neutral Winds in the Mid-latitude Ionospheric Evening Anomalies

Levan Lomidze

The Earth’s ionosphere, which is the ionized part of the upper atmosphere, plays a very important role in ground- and satellite-based radio communication and navigation. During the last half century, a considerable experimental, theoretical, and modeling effort has been directed towards understanding the physical processes that affect the variation of the ionization in the ionosphere at various spatial and temporal scales. Even though tremendous progress has been made in many directions of ionospheric research, many questions remain to be answered.

One of the intriguing features in the ionosphere is an anomalous evening enhancement that occurs over certain mid-latitude locations, where the ionization may become more than twice as large during the evening/night hours than during noon. The phenomenon takes place during summer and is most prominent over a region west of the Antarctic Peninsula where it is known as the Weddell Sea Anomaly. The generation mechanisms of this anomaly are still being debated and its accurate modeling remains a challenge.

In this dissertation, the physical mechanisms behind this evening anomaly are investigated with an emphasis on the role of winds in its generation. For this, data assimilation techniques were used for the first time to estimate the global wind velocities. These winds were then used together with a physics-based model of the ionosphere to establish the relative importance of plasma production, recombination, and transport processes in the generation of the evening anomaly.
Dedicated to my wife, Tamari Chitadze, and daughters, Nia and Martha.
I would like to express my deepest gratitude to my advisor, Dr. Ludger Scherliess, for his exceptional advising, encouragement, and continuous support. The experience I have gained while working with him will be invaluable for my future career. I am grateful to the members of my dissertation committee, Dr. Robert Schunk, Dr. Michael Taylor, Dr. Bela Fejer, and Dr. Charles Swenson, for all of their time and assistance.

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Levan Lomidze
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CHAPTER 1
INTRODUCTION

1.1. Overview and Motivation

The Earth’s ionosphere, which is the ionized part of the upper atmosphere, plays a very important role in ground- and satellite-based radio communication and navigation. The ionospheric plasma is a dispersive medium and when electromagnetic waves interact with it, their phase, amplitude, and direction of propagation are modified. These modifications depend on the ionospheric electron density distribution, and correspondingly, during the last half century a considerable experimental, theoretical, and modeling effort has been directed toward understanding the physical processes that affect the variation of the ionospheric plasma density at various spatial and temporal scales. Even though tremendous progress has been made in many directions of ionospheric research, many aspects of its spatio-temporal variations still require further investigation.

In the beginning of active studies of the ionosphere, an observed ionospheric phenomenon that could not be explained by the Chapman theory of solar photoionization of the neutral atmosphere [Chapman, 1931] was often called anomalous. Later it was realized that at altitudes (heights of F region) where these anomalies where observed, the plasma was no longer in photochemical equilibrium, and that transport processes due to diffusion, neutral winds, and electromagnetic drifts played an important role [Hanson and Moffett, 1966; Rishbeth, 1967], and deviations from the simple Chapman theory could be expected.

Nevertheless, one type of ionospheric anomaly, which was first noticed in the late 1950’s [Bellchambers and Piggott, 1958], is still not fully understood and still remains an active topic of research [Burns et al., 2011; Ren et al., 2012; Chen et al., 2013; Slominska et al., 2014]. The phenomena takes place west of the Antarctic Peninsula and because of its proximity to the Weddell Sea, it is called the Weddell Sea Anomaly (WSA). It is characterized by larger evening/night values of the F-region electron density compared to corresponding noontime values, is most prominent during local summer, and completely disappears during
winter. Currently, the generation mechanism of the WSA is being debated and its modeling remains a challenge. The WSA was either reported to be entirely missing from the output of ionospheric models, or the quantitative comparison of the modeled electron density parameters to their observations over the WSA region seems to be problematic. The WSA is the most extreme manifestation of mid-latitude ionospheric evening anomalies. Similar density enhancements during evening hours have also been observed in the northern hemisphere during local summer [Eyfrig, 1963; Evans, 1965; Kohl et al., 1968; Papagiannis and Mullaney, 1968; Lin et al., 2010; Burns et al., 2011; Slominska et al., 2014], but with smaller amplitudes.

Among the many suggested explanations of the evening anomalies, the earlier proposed mechanism of solar photoionization superimposed on an upward drift of plasma due to the neutral winds [Eccles et al., 1971; Dudeney and Piggott, 1978] still dominates. Indeed, the ionosphere is embedded in the neutral atmosphere and Earth’s magnetic field. At F-region heights, thermospheric neutral winds influence plasma transport by changing the field-aligned diffusion of ions, and hence, play an important role in the spatial distribution and dynamics of the ionospheric plasma. Because of this, neutral winds represent one of the key inputs for ionospheric models, and the accuracy of these models strongly depends on the accuracy of the used wind values. On the other hand, the ionosphere-thermosphere is a coupled system. The ionosphere exerts a drag force on the motion of the neutral atmosphere due to ion-neutral collisions, and therefore knowledge of the ionospheric electron densities is needed to realistically characterize the neutral dynamics in the models.

The recent advent of various measurement techniques for the ionosphere, and especially satellite-based observations (among them GPS radio occultation), provides an excellent database for studying the ionosphere globally and for specifying the ionospheric electron density for various applications. On the contrary, the progress in routine and accurate measurements of thermospheric neutral winds on a global scale has been relatively slow. The spatial and temporal sparseness of wind data, for example, affects the performance of empirical wind models, which often fail to reproduce even the observed climatology. Physics-
based general circulation models (GCMs), which generally couple models of different parts of the Earth’s space environment (ionosphere, thermosphere, plasmasphere, magnetosphere, mesosphere), can also provide wind estimations. However, results from these models can suffer from error propagation from model to model, and are sensitive to various input parameters and prescribed external forcing.

A relatively new approach in ionospheric specification and modeling is the incorporation of ionospheric measurements into ionospheric physics-based data assimilation models to compensate for the lack of knowledge in the various ionospheric drivers [Schunk et al., 2004]. Among several data assimilation techniques, the Kalman filter [Kalman, 1960; Evensen, 2009] is a widely used and promising method. One of these models that employs these techniques (Global Assimilation of Ionospheric Measurements Full Physics (GAIM-FP)) was developed at Utah State University [Scherliess et al., 2004, 2009] as part of the GAIM project [Schunk et al., 2004]. GAIM-FP can assimilate different types of ionospheric data and can provide information about the physical drivers of the ionosphere, including neutral densities, equatorial electric fields, and neutral winds in the direction of the magnetic meridian.

This dissertation presents work that employs data assimilation techniques to estimate thermospheric neutral winds globally (at low and mid-latitudes) and uses the obtained winds to model and understand the physical mechanism behind the evening anomalies. Chapter 1 presents a general introduction to the Earth’s thermosphere, thermospheric neutral winds and the Earth’s ionosphere. It also describes the main principles of various ionospheric models and provides a brief description of the ionospheric GPS radio occultation technique, used to obtain the ionospheric data needed for this study.

The main work of this dissertation is presented in Chapters 2-4. The ultimate objective of this work is to understand the physical mechanism behind the observed evening anomalies in the mid-latitude ionosphere. Specifically, the focus is on the study of the role of neutral winds in their generation. Ideally, direct observations of the neutral wind would be used for the study. Unfortunately, wind observations are extremely sparse over these regions and
consequently, wind estimates need to be obtained through other means. As a second choice, an empirical wind model (like HWMs) could be used. However, these models, when used in our ionospheric model, fail to satisfactorily reproduce the observed variation of ionospheric F-region peak density (NmF2) and corresponding height (hmF2), as seen in the observations by the Constellation Observing System for Meteorology Ionosphere and Climate/Formosa Satellite 3 (COSMIC). Therefore, to obtain neutral winds and their individual geographic components needed for this study, we use the existing wind estimation techniques developed for GAIM-FP together with a new data assimilation model for thermospheric neutral winds, which was specifically developed for this investigation. The estimated wind components obtained from this model are then used to investigate the role of the wind in the mid-latitude ionospheric evening anomalies. These three parts of the work are described in Chapters 2-4, which can be viewed as individual studies, but also represent necessary steps to achieve our main goal. A schematic view of the work described in this dissertation is illustrated in Figure 1.1 where the three parts correspond to the individual Chapters.

Chapter 2 presents the results of the magnetic meridional wind climatology, obtained from GAIM-FP by assimilating COSMIC seasonal NmF2 and hmF2 maps, for geomagnetically quiet, and solar minimum conditions. It demonstrates the ability of the data assimilation model to obtain accurate wind values and to significantly improve the agreement between the measured and observed F-region electron density parameters (NmF2 and hmF2) simultaneously on a global scale. The wind results are compared with corresponding ground-based measurements and empirical model values. Finally, the sensitivity of the obtained winds to various uncertain parameters is studied.

Chapter 3 describes a novel technique to estimate the 3-D geographic zonal, meridional, and vertical wind components in 3-D from the magnetic meridional wind data. It presents the recently developed physics-based Thermospheric Wind Assimilation Model (TWAM), which uses an implicit Kalman filter technique together with the equation of motion of the neutral gas in the thermosphere. The assimilated data is the magnetic meridional wind from GAIM-FP obtained in Chapter 2. The climatology of the quiet-time zonal and
Figure 1.1. Schematic view of the research described in this dissertation. Parts 1-3 are discussed in Chapters 2-4, respectively.
meridional winds from TWAM is presented and a comparison with corresponding empirical and ground-based wind measurements is performed.

In Chapter 4 the TWAM winds of Chapter 3 are used in the Ionosphere-Plasmasphere Model (IPM) to model the ionospheric evening anomalies. The effect of the neutral wind on the anomalies is studied and the physical mechanism behind their generation is investigated. An analysis of the plasma motion due to diffusion and winds, and plasma transport, production, and loss processes is performed to understand their relative roles in the evening density enhancements.

Finally, Chapter 5 summarizes our main results, discusses the limitations of the performed study, and provides recommendation for future work.

1.2. The Earth’s Thermosphere

The Thermosphere is a region of the Earth’s upper atmosphere between about 90 - 500 km altitude range. This is the region where most of the solar ultraviolet (UV) and X-ray radiation is absorbed, and consequently, it is characterized by high temperatures. Besides radiation, various processes contribute to the thermal balance in the lower thermosphere, including dissipation of atmospheric tides and gravity waves, and absorption of energetic charged particles. At the lower thermosphere, a large positive gradient of temperature exists due to downward heat conduction. At higher altitudes, the temperature may exceed 1000°K, but eventually becomes constant with altitude because of a decrease in the absorption of UV radiation and the large heat conductivity at higher altitudes. The thermospheric temperature also undergoes considerable diurnal, seasonal, geographical, and solar cycle variations. As a representative example, the left panel of Figure 1.2 illustrates the altitude profile of daytime neutral temperature in the thermosphere at a typical mid-latitude location for solar cycle minimum and maximum conditions. The region above the thermosphere is called exosphere. In the exosphere, the collisions between molecules become so infrequent that they move in ballistic orbits and the atmosphere can no longer be considered as a fluid.

Below the thermosphere the neutral atmosphere is composed of about 78% of N\textsubscript{2} and 21% of O\textsubscript{2}, and the content of the rest of the gases (mainly Ar and CO\textsubscript{2}) is about 1%. This
mixture holds well from ground up to about 100 km because of rapid mixing of gases due to atmospheric turbulence. Above 100 km, photodissociation of molecules (mostly of O\textsubscript{2}) becomes important and diffusive separation starts to dominate over turbulent mixing. As a result, atmospheric species become gravitationally separated and assume their independent height distributions. The right panel of Figure 1.2 shows the altitude profiles of different atmospheric species in the daytime mid-latitude thermosphere for medium solar activity conditions. The heavier molecules (N\textsubscript{2} and O\textsubscript{2}) dominate at low altitudes, atomic oxygen becomes dominant above 200 km, and even higher up lighter helium and hydrogen start to gradually dominate. The density of the atmospheric species decreases exponentially with height, and the rate at which the density is decreasing is higher for heavier molecules.
The expression describing the altitude distribution of the density for a given atmospheric species is derived as follows. The neutral atmospheric constituents in the thermosphere can be characterized by their own pressure and density. If we take a volume element of air and assume static equilibrium conditions, then gravity is balanced by the net pressure forces. Then, if partial pressure and mass density of a given species, $j$, are $p_j$ and $\rho_j$, the hydrostatic equation (also called barometric equation) takes the form

$$\frac{dp_j}{dz} = -\rho_j g = -n_j m_j g,$$  

(1.1)

where $n_j = \rho_j / m_j$ is number density and $g$ is the acceleration due to gravity at height $z$. To a good approximation, air behaves like an ideal gas; and therefore, satisfies the equation of state

$$p_j = n_j kT,$$  

(1.2)

where $k$ is the Boltzmann’s constant and $T$ is the temperature. Dividing Eq. (1.1) by Eq. (1.2) gives

$$\frac{1}{p_j} \frac{dp_j}{dz} = -\frac{m_j g}{kT} = -\frac{1}{H_j},$$  

(1.3)

and by using (1.2) in (1.3) it can be shown

$$\frac{1}{n_j} \frac{dn_j}{dz} = -\frac{1}{H_j} - \frac{1}{T} \frac{dT}{dz}.$$  

(1.4)

Here, the quantity $H_j = kT / m_j g$ is called the scale height of constituent $j$. If the temperature profile is known, the pressure and density at any height can be found by integrating (1.3) and (1.4), respectively. If isothermal conditions apply, (1.3) and (1.4) yields

$$\frac{p_j}{p_{j,0}} = \frac{n_j}{n_{j,0}} = exp \left[ -\frac{z - z_0}{H_j} \right],$$  

(1.5)

where $z_0$ is some reference height at which $p_j = p_{j,0}$ and $n_j = n_{j,0}$. Expression (1.5) shows that pressure and density of a given gas fall exponentially with altitude.

For more details about the Earth’s thermosphere, the reader is referred to Ratcliffe
1.3. Thermospheric Neutral Winds

During geomagnetically quiet-times, horizontally uneven heating of the thermosphere by solar extreme UV (EUV) radiation is a primary source of pressure gradient forces, which drive large scale atmospheric dynamics. The neutral winds primarily blow from the hottest part of the thermosphere (day side) towards the coldest part (night side). Since the Earth is rotating, the flow of neutral gas is subject to Coriolis forces that affect it. In addition, the so-called ion drag force, which is due to collision between neutral and charged particles, and atmospheric viscosity also affect the motion of air. The effects of a variety of high-latitude processes on atmospheric circulation become more important during geomagnetic disturbances.

The thermospheric wind velocity $u$ satisfies the following equation of motion [Rishbeth and Garriot, 1969]:

$$\frac{\partial u}{\partial t} + (u \cdot \nabla) u = -2\Omega \times u + \frac{1}{\rho} \nabla p + \frac{\mu}{\rho} \nabla^2 u - \nu_{ni} (u - u_i),$$  \hspace{1cm} (1.6)

where $\Omega$ is the Earth’s angular velocity, $\mu$ is dynamic (molecular) viscosity, $\nu_{ni}$ is neutral-ion collision frequency, and $u_i$ is the velocity of ions. In Eq. (1.6) the flow is assumed to be incompressible and $\mu$ to be constant. For quiet conditions, the vertical wind velocity is small and can be ignored in the Eq. (1.6). With increasing altitude the kinematic viscosity $\mu/\rho$ increases, smoothing out height variation of the wind. The typical magnitude of thermospheric wind velocities at low and mid-latitudes is about 100-300 $\text{ms}^{-1}$. During daytime, the flow is generally poleward and westward, and during night it becomes equatorward and eastward [Rishbeth, 1972; Titheridge, 1995; Emmert et al., 2003]. This is well demonstrated in Figure 1.3 where the typical quiet-time global distribution of the thermospheric neutral temperature together with the wind vectors is shown. The flow pattern is mainly from hot to cold regions and from the summer to the winter hemisphere. Note the phase difference of
Figure 1.3. Global distribution of neutral temperature and wind at 300 km for quiet-time and low solar activity conditions. The values correspond to December solstice conditions and are given for UT=12:00. The temperature is calculated from the MSISE-90 and the wind vectors from the HWM93 [Hedin et al., 1996] empirical models.

a few hours between the subsolar point and the diurnal temperature maximum, which lags behind for a few hours (for the temperature-density phase anomaly see Mayr et al. [1973]).

The study of thermospheric winds is important for understanding the global variation of thermospheric composition. The winds also play an important role in the variation of ionospheric plasma. In addition, they create dynamo electric fields and can influence plasma instabilities [Kelley, 2009]. Nevertheless, routine observations of thermospheric winds on a global scale are limited. Ground-based direct observations can be carried out using interferometers by measuring the Doppler shift in the airglow emission lines [Hays and Roble, 1971]. However, these observations can provide information only during cloudless nights, and the number of places that perform such measurements on the globe are rather limited [Drob et al., 2008]. Interferometric observations of winds can also be carried out from satellites [Shepherd et al., 1993]. In addition, wind velocities can also be obtained from
space-based measurements using spectrometers [Spencer et al., 1982] and accelerometers [Liu et al., 2006]. Yet, accuracy, continuity, and local time coverage are often the issues with wind measurements obtained from satellites.

Alternatively, information about thermospheric neutral winds can be inferred from ionospheric measurements. For example, the incoherent scatter radar (ISR) can measure the plasma velocity along the geomagnetic field. By calculating the diffusion part of this velocity, which involves its estimation using empirical neutral atmospheric models, one can estimate the horizontal wind along the magnetic meridian [e.g., Salah and Holt, 1974]. This method is generally applicable for mid-latitudes and is limited to a handful of locations where ISR data are available [Hedin et al., 1991; Drob et al., 2008]. Another method of obtaining winds is the use of variations in the measured ionospheric F2-layer peak parameters (density and height) and combining them with a so-called servo model [Rishbeth et al., 1978]. This method also gives the magnetic meridional wind, and is valid only for mid-latitudes. In general, errors in the wind estimation from ionospheric data are relatively large compared to direct wind observations, and are the result of numerous assumptions and uncertainties in the model parameters [Buonsanto et al., 1997; Buonsanto and Witasse, 1999].

There are also two types of models that calculate the global thermospheric wind field. First is the series of empirical wind models (HWMs) [Hedin et al., 1991, 1996; Drob et al., 2008], which incorporate the majority of available ground- and space-based wind data, and use spatial and temporal interpolation functional expressions to interpolate between the data gaps. Such models generally provide the average climatology of neutral winds and are less reliable over locations where limited or no data are available. The second type of models that calculate thermospheric winds are physics-based models (such as CTIM [Fuller-Rowell and Rees, 1980], TIEGCM [Richmond et al., 1992], TGCM [Ma and Schunk, 1995], WAM [Akmaev et al., 2008]). These models internally couple several models of various parts of the Earth’s atmosphere and calculate self-consistent thermospheric winds. However, in this approach, error propagation from model to model can lead to unreliable quantitative results [Schunk et al., 2002]. The model results may also be sensitive to errors in the imposed
boundary conditions and to uncertainties in the external driving forces.

1.4. The Earth’s Ionosphere

The Earth’s ionosphere is the partially ionized part of the Earth’s upper atmosphere. It starts at about 60 km above the surface and extends beyond 1000 km in altitude. The primary source of ionization is solar EUV and X-ray radiation; however, particle precipitations from the magnetosphere, secondary photoelectrons, and cosmic rays also play a role. Depending on the altitude, the ionospheric plasma consists of various ionic species. Heavier ions are present at lower altitudes and a gradual transition to lighter ions occurs as height increases. Though there are no sharp boundaries, the ionosphere is commonly divided into several regions in altitude, which differ by their characteristics.

The D region, which only exists during daytime, spans from 60 - 90 km and has lowest plasma density \(10^3 \text{ cm}^{-3}\) (due to plasma quasineutrality, the concentration of negative and positive charges are approximately equal throughout the ionosphere). There are both, positive and negative ions present in the D region. Between about 90 - 150 km is the E-region ionosphere with electron densities of up to \(10^5 \text{ cm}^{-3}\). The major ions in the E region are \(\text{O}_2^+\) and \(\text{NO}^+\). The D and E regions are dominated by photochemistry. During night the D region disappears and the E-region electron density drops down to D-region levels. The part of the ionosphere between about 150 - 500 km is called the F region, which is often further divided into F1 (150 - 200 km) and F2 regions (200 - 500 km). At F-region heights plasma transport processes start to become important. The dominant ion species in the F region is \(\text{O}^+\). The peak plasma densities in the ionosphere typically occur in the F2 region and are of the order of \(10^6 \text{ cm}^{-3}\). Around the peak height (around 300 km altitude), chemical loss processes are slow and the density survives at night. The uppermost part of the ionosphere is the topside ionosphere with a transition from \(\text{O}^+\) to \(\text{He}^+\) and \(\text{H}^+\) with altitude in this domain. Figure 1.4 shows typical altitude profiles of the density of various ion species for the daytime mid-latitude ionosphere during medium solar activity conditions.

A photoionization of neutral air constituents such as \(\text{N}_2\), \(\text{O}_2\), and \(\text{O}\) is the primary source of ionization in the ionosphere, and is accomplished by solar radiation in the X-ray
Figure 1.4. Ion composition of daytime mid-latitude ionosphere at medium solar activity conditions. Densities are computed from the empirical IRI2012 model [Bilitza et al., 2014]. $N_e$ denotes electron density, which at a given altitude is equal to the sum of the ion densities.

(0.1 - 10 nm) and EUV (10 - 120 nm) range. Let’s assume the incoming solar radiation intensity (photon flux) at wavelength, $\lambda$, and altitude, $z$, is $I(\lambda, z)$. Let $\sigma(\lambda)$ be the absorption cross section for ionizing neutral species and $\eta$ the ionizing efficiency. Then the production rate (also called Chapman production function) of an ion-electron pair from neutral species, $j$, will be

$$P_j(\lambda, z) = \eta\sigma(\lambda)n_j(z)I(\lambda, z),$$  \hspace{1cm} (1.7)

where $n_j(z)$ is the neutral gas density given by Eq. (1.5). When the radiation passes a distance, $ds$, in the atmosphere, its intensity will be reduced by $dI$ and

$$dI = -n_j(z)\sigma(\lambda)I(\lambda, z)ds = n_j(z)\sigma(\lambda)I(\lambda, z)dz/\cos(\chi),$$  \hspace{1cm} (1.8)
where $\chi$ is the solar zenith angle and $dz$ is the altitude change corresponding to $ds$. By integrating Eq. (1.8) from $z$ to infinity we obtain

$$I(\lambda, z, \chi) = I_\infty(\lambda) \exp[-\tau(\lambda, z, \chi)], \quad (1.9)$$

where

$$\tau(\lambda, z, \chi) = \int_z^\infty n_j(z') \sigma(\lambda) dz' / \cos(\chi). \quad (1.10)$$

To calculate the production rate for a given ion species, one should consider that there is a certain spectrum of radiation that is effective to ionize. If $\lambda_j$ is the ionization threshold wavelength for species $j$, the total ion production rate for species $j$ when considering (1.9) in (1.7) is

$$P_j(z, \chi) = n_j(z) \int_{\lambda_j}^{\lambda_j} I_\infty(\lambda') \exp[-\tau(\lambda', z, \chi)] \eta \sigma(\lambda') d\lambda'. \quad (1.11)$$

The total ionization rate of all species will then be the sum of $P_j$ over all species (for more details on ionization rates see Rishbeth and Garriot [1969]; Banks and Kockarts [1973b]; Schunk and Nagy [2009]).

The ionization in the ionosphere is lost in a series of charge exchange, dissociative recombination, and radiative recombination reactions. The major chemical reactions in the F region, where $O^+$ is the most abundant ion, are given in Table 1.1. In the lower F-region ionosphere, the plasma is, to a good approximation, in chemical photoequilibrium conditions, which means the production of atomic oxygen ions equals its loss:

$$P(O^+) = L(O^+). \quad (1.12)$$

$P(O^+)$ can be calculated from (1.11), which shows it is directly proportional to the atomic oxygen density. For the loss by using Table 1.1 one obtains

$$L(O^+) = [O^+][k_1[N_2] + k_2[O_2]], \quad (1.13)$$
Table 1.1. Major chemical reactions and reaction rates in the F region (from Schunk and Walker [1973]). $T_i$ and $T_e$ are ion and electron temperatures, respectively.

<table>
<thead>
<tr>
<th>Reaction</th>
<th>Rate (cm$^3$/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$O^+ + N_2 \rightarrow NO^+ + N$</td>
<td>$k_1 = 1.2 \times 10^{-12}(300/T_i)$</td>
</tr>
<tr>
<td>$O^+ + O_2 \rightarrow O_2^+ + O$</td>
<td>$k_2 = 2 \times 10^{-11}(300/T_i)^{0.5}$</td>
</tr>
<tr>
<td>$NO^+ + e \rightarrow N + O$</td>
<td>$\alpha_1 = 4.6 \times 10^{-7}(300/T_e)^{1.2}$</td>
</tr>
<tr>
<td>$O_2^+ + e \rightarrow O + O$</td>
<td>$\alpha_2 = 2.2 \times 10^{-7}(300/T_e)^{0.5}$</td>
</tr>
</tbody>
</table>

where the square bracket indicates the number density. Eqs. (1.12) and (1.13) can be used to calculate the equilibrium density of $O^+$ ions.

In the upper F region, above the F2-region peak, the ionosphere is not in photochemical equilibrium. At this altitude diffusive processes start to dominate and plasma transport becomes important. In fact, the maximum electron density ($NmF2$) occurs at the altitude ($hmF2$) where chemical and diffusion processes are of equal importance.

The ionospheric F-region plasma is well described by ambipolar diffusion approximation. Next, we derive the ion velocity expression in this approximation. The motion of charged particles in the ionosphere is influenced by the presence of the geomagnetic field. At the heights of the F region, the ion and electron collision frequencies are much smaller than their corresponding gyro frequencies; hence, the plasma is magnetized, or in other words, it is constrained to move along the geomagnetic field lines by the diffusion. Note, if external electric fields are present, the plasma, as a whole, can drift across the magnetic field. Gravity causes charge separation of electrons and ions, inducing a polarization electrostatic field that keeps ions and electrons together. The approximation in which the plasma moves as a single gas is called the diffusion approximation.

The set of continuity, momentum, and energy equations for plasma species $j$ ($j$ can denote electrons, $e$, or any ions, $i$) in their approximate form are [Schunk and Nagy, 2009]
\[ \frac{\partial n_j}{\partial t} + \nabla \cdot (n \mathbf{v}_j) = P_j - L_j \]  
(1.14)

\[ m_j n_j \left[ \frac{\partial \mathbf{u}_j}{\partial t} + (\mathbf{u}_j \cdot \nabla) \mathbf{u}_j \right] + \nabla p_j - n_j m_j \mathbf{g} - q_j n_j (\mathbf{E} + \mathbf{u}_j \times \mathbf{B}) = n_j m_j \nu_{jn} (\mathbf{u}_n - \mathbf{u}_j) \]  
(1.15)

\[ \frac{\partial p_j}{\partial t} + (\mathbf{u}_j \cdot \nabla)p_j + \frac{5}{3} p_j (\nabla \cdot \mathbf{u}_j) = \frac{\delta E_j}{\delta t}, \]  
(1.16)

where \( q_j \) is charge of species \( j \), \( n \) denotes parameters for neutral species, \( \nu_{jn} \) is collision frequency between species \( j \) and \( n \), \( \mathbf{E} \) and \( \mathbf{B} \) are electric and magnetic fields, respectively, \( p_j = n_j kT_j \) is the partial pressure and \( \delta E_j/\delta t \) is the energy change rate of species \( j \).

Note, if \( T_j \) is given, then Eq. (1.16) is no longer needed for the system closure. In an ambipolar diffusion approximation, the flow is steady and subsonic, thus inertial terms in the momentum equation (1.15) can be neglected. If we sum up the momentum equations of ions and electrons along the magnetic field lines and consider \( \mathbf{u}_i = \mathbf{u}_e, \quad n_i = n_e, \quad n_i m_i \nu_{ie} = n_e m_e \nu_{ei} \) (the latter is required from the conservation of momentum during collisions), we obtain

\[ \frac{\partial}{\partial s} (p_i + p_e) - n_i (m_i + m_e) g_s = n_i (m_i \nu_{in} + m_e \nu_{en}) (u_{n,s} - u_{i,s}), \]  
(1.17)

where subscript \( s \) denotes the projection along the geomagnetic field. Finally, if we assume collisions of ions with neutrals are more important than those with electrons (\( m_i \nu_{in} \gg m_e \nu_{en} \)) and also consider the fact \( m_i \gg m_e \), the solution for the plasma field-aligned velocity yields

\[ u_{i,s} = u_{n,s} - D_a \left( \frac{1}{n_i} \frac{\partial n_i}{\partial s} + \frac{1}{T_p} \frac{\partial T_p}{\partial s} - \frac{\sin I}{H_p} \right). \]  
(1.18)

Here, \( T_p = (T_i + T_e)/2 \) is the plasma temperature, \( H_p = 2kT_p/m_i g \) is the plasma scale height, \( D_a = 2kT_p/m_i \nu_{in} \) is the ambipolar diffusion coefficient, and \( I \) is the magnetic field inclination from the horizontal direction (positive in the northern hemisphere). Expression (1.18) shows that the diffusion velocity depends on the plasma density, temperature and their gradients, as well as on the neutral wind velocity. In addition, the collision frequency,
which determines the diffusion coefficient, also plays an important role. The collision frequencies of $O^+$ ions with the major neutral species of the F region are given in Table 1.2. Eq. (1.18) also shows the ion field-aligned velocity due to the neutral winds is equal to the wind velocity projection along the geomagnetic field; and thus, the horizontal winds are more effective in modifying plasma vertical motion over mid-latitudes.

The electron density in the F-region ionosphere undergoes variability at various temporal and spatial scales. It displays variations with local time, day of year, season, latitude, longitude, and solar and geomagnetic activity. These variations are the result of changes in various energy inputs in the coupled atmosphere-ionosphere system and accompanied changes in neutral winds, composition, electric fields and temperatures.

The geomagnetic field geometry plays a very important role in various ionospheric processes. At mid-latitudes, the geomagnetic field lines are closed and they connect the two hemispheres through the plasmasphere. At the topside ionosphere the accidentally resonant charge exchange reaction between charged and neutral atomic oxygen and hydrogen ($H^+ + O \rightarrow H + O^+$) proceeds rapidly because the first ionization potentials of these atoms are almost equal. During daytime the upward flux of $H^+$ from the topside ionosphere is refilling the plasmasphere. At night the flux becomes downward and provides an ionization source, helping to maintain the nighttime F-region ionosphere.

An interesting phenomenon takes place at low latitudes. Near the magnetic equator where the geomagnetic field lines become horizontal, the daytime eastward electric field produces an upward $E \times B$ drift. As a consequence, plasma is lifted up and then, due to the action of gravity, diffuses down along the magnetic field lines away from the equator. This is the so-called plasma fountain effect and the resultant density distribution shows a trough at the equator and crests at $15^\circ$ - $20^\circ$ north and south of it (known as the equatorial ionospheric anomaly). The cross-equatorial winds, which generally blow from summer to winter hemisphere, can alter the ionization at the anomaly crests and create an electron density asymmetry about the magnetic equator.

At high latitudes, the geomagnetic field lines are open and connect the ionosphere to the
Table 1.2. Collision frequencies for $\text{O}^+$ ions with neutral species in the F-region ionosphere (from Schunk and Walker [1973]). Here $T_r = (T_i + T_n)/2$. Densities are in $\text{cm}^{-3}$.

<table>
<thead>
<tr>
<th>Collision frequencies ($\text{s}^{-1}$)</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>$\nu(\text{O}^+, \text{O}) = 3.42 \times 10^{-11} [\text{O}] T_r^{0.5} \left[ 1.08 - 0.139 \log_{10} T_r + 4.51 \times 10^{-3} (\log_{10} T_r)^2 \right]$</td>
<td></td>
</tr>
<tr>
<td>$\nu(\text{O}^+, \text{N}_2) = 6.82 \times 10^{-10} [\text{N}_2]$</td>
<td></td>
</tr>
<tr>
<td>$\nu(\text{O}^+, \text{O}_2) = 6.66 \times 10^{-10} [\text{O}_2]$</td>
<td></td>
</tr>
</tbody>
</table>

magnetosphere. Various energetic particles from the sun and magnetosphere precipitate to ionospheric heights and provide an additional source of ionization. In addition, the electric fields from the magnetosphere map down into the high-latitude ionosphere and strongly influence the plasma distribution and dynamics.

The ionospheric behavior is very complex during geomagnetic storms. As described, for example, by Buonsanto [1999], geomagnetic storms, which start with solar wind-magnetosphere interaction, produce high-latitude electric fields, which can penetrate to low latitudes. At the same time, the enhanced energy input causes substantial heating of the ionized and neutral gases and the resulting uneven expansion of the thermosphere produces pressure gradients, which drive strong neutral winds. The disturbed thermospheric circulation alters the neutral composition, and also produces polarization electric fields by a dynamo effect. These electric fields in turn affect the neutrals and the ionospheric plasma.

The various fundamental ionospheric processes and phenomena are described in detail in many textbooks, among them in Rishbeth and Garriott [1969]; Banks and Kockarts [1973a,b]; Brekke [1997]; Hunsucker and Hargreaves [2003]; Kelley [2009]; Schunk and Nagy [2009].

1.5. Ionospheric Models

Ionospheric modeling has always been an integral part of ionospheric research, and consequently, various types of ionospheric models have been developed over the years. They compute a variety of plasma parameters and may be used for ionospheric specification, forecasting, or for better understanding of the various physical processes. There are three main types of ionospheric models: empirical, physics-based, and data assimilation models.
Empirical ionospheric models are based on ionospheric data that are obtained by a variety of measurement techniques at various times and conditions. To construct such models, the data are binned by time of day, month, season, altitude, latitude, longitude, and solar and geomagnetic activity, and various analytical expressions and/or orthogonal functions are used to fit them. This allows to obtain an estimation of ionospheric parameters beyond measured times or spatial coordinates. One well known and widely used empirical ionospheric model is the International Reference Ionosphere (IRI) [Bilitza et al., 1990, 2011, 2014]. It includes data from a worldwide network of ionosondes, incoherent scatter radars, topside sounders and in-situ data from satellites and rockets. Given the approach used in the model development, IRI describes monthly averages of parameters (i.e., climatology). Similar to many empirical models, the accuracy of IRI depends on the spatio-temporal coverage and quality of the included data; and therefore, the model is expected to be less reliable over oceans, for example.

Theoretical first-principle models of the ionosphere numerically solve for the set of continuity, momentum and energy equations for electrons and ions (Eq. 1.14 - 1.16); and therefore, they determine self-consistent plasma parameters in the ionosphere. There are two types of physics-based models, stand-alone and coupled ionospheric models. Stand-alone models, such as the Ionosphere-Plasmasphere Model (IPM) [Schunk et al., 2004; Scherliess et al., 2004] or SAMI2 [Huba et al., 2000]), solve equations for charged particles only, and thus require inputs of the neutral composition and temperature, neutral winds, and electric fields. These parameters (ionospheric drivers) can be provided by empirical models like the MSISE-90 [Hedin, 1991], HWM93 [Hedin et al., 1996] and Scherliess-Fejer vertical drift model [Scherliess and Fejer, 1999], respectively. In order to simplify numerical schemes, the equations are often solved along the geomagnetic field lines (1-D problem) and then the 3-D picture is obtained by following a large number of field lines and mapping the solution on a spherical grid. The accuracy of the model output is sensitive, though at various degree, to the provided empirical inputs. Even though the stand-alone models solve
Another type of physics-based models are coupled models that numerically combine different parts of the near-Earth space environment. For example the thermosphere-ionosphere-electrodynamics general circulation model (TIEGCM) [Richmond et al., 1992; Qian et al., 2014], is a global model of the thermosphere-ionosphere system which self-consistently solves for ionospheric and thermospheric parameters, as well as dynamo electric fields. There are certain parameters, however, that are still needed to be prescribed such as momentum and energy transfer from the lower boundary, topside plasmaspheric flux and high-latitude forcing, and they are provided by separate models (some could be empirical). Even though the self-consistent description of the coupled systems has certain advantages, the error propagation from model to model can lead to unreliable quantitative results [Schunk et al., 2002].

The described common limitations of physics-based models, could be further supplemented by missing physics and uncertain model parameters that impact the output, e.g., see Schunk et al. [2012]. Regardless, the physics-based models, either stand alone or coupled, are very important to gain physical insight into many important ionospheric phenomena.

One way to overcome the limitations of physics-based models and take advantage of available ionospheric data is to combine the two in an optimal fashion. The process of objectively combining the information from observation and model is called data assimilation. Data assimilation fills the observational gaps, constrains the model with observations and helps to estimate unobserved quantities [Lahoz et al., 2010]. Historically data assimilation has been used in meteorology and oceanography, however, its application has also gained prominence in ionospheric modeling as well [Schunk et al., 2002, 2004; Pi et al., 2003]. There are many different types of data assimilation techniques; however, one of the most promising data assimilation techniques for ionospheric application is the Kalman filter (KF). It is a recursive, weighted, least squares estimator that, together with assimilating
different types of data, takes into account the uncertainties in both the model and data to produce an optimal estimation of the electron density distribution at a given time based on all priori information. The classical KF is commonly used for linear problems; however, for nonlinear systems and in cases when the number of estimated variables are large, an ensemble Kalman filter (EnKF) approach is a powerful method [Evensen, 2009]. The Global Assimilation of Ionospheric Measurements Full Physics (GAIM-FP) model [Scherliess et al., 2009], developed at Utah State University, is based on the EnKF approach. The model uses the physics-based IPM model and can assimilate a variety of ground- and satellite-based ionospheric data. In addition of producing an improved description of the electron density field, which is generally close to the observations, GAIM-FP can provide estimations of the drivers including neutral composition, low-latitude electric fields and magnetic meridional winds that are needed for the better model-data agreement. Hence, GAIM-FP provides valuable information about critical ionospheric drivers, which are very important for accurate ionospheric modeling but otherwise not always well known.

1.6. GPS Radio Occultation

Radio occultation (RO) is a method of studying planetary atmospheres or ionospheres by radio waves [Fjeldbo and Eshleman, 1968; Kursinski et al., 1997; Melbourne, 2004]. It involves a transmitter, a receiver and an atmosphere/ionosphere that needs to be probed. When the emitted signal passes through the atmosphere/ionosphere, which is a refractive medium, it experiences changes in phase, amplitude, velocity, and direction of propagation before arriving at the receiver. The measured phase and Doppler shifts of the radio wave can be related to the refractive index. The same refractive index can be theoretically calculated as a function of various atmospheric and ionospheric parameters, and finally information about these parameters can be obtained.

One possible scenario of accomplishing a radio occultation experiment to study Earth’s atmosphere is to use the signal transmitted by a setting or rising Global Positioning System (GPS) satellite. The signal passes through the atmosphere/ionosphere and is received by Low-Earth orbit (LEO) satellites [Kursinski et al., 1997; Hajj and Romans, 1999; Schreiner
et al., 1999]. The typical radio occultation geometry is illustrated in Figure 1.5. For the neutral atmosphere the refractive index is independent of the wave frequency and depends on the atmospheric temperature, pressure, water vapor, and liquid water contents resulting in a bending of the signal as it traverses the atmosphere. The bending becomes more important below \( \sim 60 \) km, and by separating the ionospheric contribution to the total bending, information about the neutral atmosphere can be obtained. Specifically, as time evolves, the RO geometry changes (Figure 1.5) and profiles of bending angle are generated. This makes it possible to retrieve the height distribution of neutral atmospheric pressure, density, and temperature, and water vapor pressure.

Earth’s ionosphere is a dispersive medium (refractive index varies with frequency), and its index of refraction depends on the plasma density. Even though it is possible to obtain the electron density height distribution in the ionosphere from the bending angle measurements, at GPS frequencies \( (f_1=1.57542 \text{ GHz}, \ f_2=1.22760 \text{ GHz}) \) the bending of the radio signal and ray separation can be neglected if the aim is to monitor the F2-region ionosphere. Therefore, it is more common to assume a straight-line propagation of the GPS signal in the ionosphere and take advantage of its dispersive nature [Schreiner et al., 1999; Lei et al., 2007]. In this case the measured parameter is the so-called slant total electron content (sTEC), which is inverted to obtain the electron density. The phase delay due to the neutral atmosphere is negligible and it is assumed that signals with two different GPS frequencies travel exactly the same path to the receiver (LEO). This RO geometry is schematically given in Figure 1.6. The index of refraction \( (n) \) for the plasma, subject to a magnetic field \( (B) \), is given by the Appleton-Hartree equation [Hunsucker and Hargreaves, 2003]

\[
n^2 = 1 - \frac{X}{1 - iZ - \frac{Y_L^2}{2(1-X-iZ)} + \left( \frac{Y_T^2}{4(1-X-iZ)^2} + Y_L^2 \right)^{1/2}},
\]

where \( i = \sqrt{-1}, \ X = \omega_{pe}^2/\omega^2, \ Y = \Omega/\omega, \ Y_L = Y \cos \theta, \ Y_T = Y \sin \theta, \ Z = \nu_e/\omega, \ \omega_{pe} = \sqrt{N_e e^2/\epsilon_0 m_e} \) is the electron plasma frequency, \( \Omega = eB/m_e \) is the electron gyrofrequency, and \( \nu_e \) is the electron collision frequency. \( \theta \) indicates the angle between the direction of propagation of the electromagnetic wave and the magnetic field, \( \omega \) is an angular frequency.
of the wave, and $+$ and $-$ correspond to ordinary and extraordinary waves, respectively. At UHF frequencies (0.3-3 GHz) the effects of collisions and magnetic field can be neglected, which yields

$$n \approx \sqrt{1 - \frac{\omega^2_{pe}}{\omega^2}} \approx 1 - \frac{\omega^2_{pe}}{2\omega^2} = 1 - \frac{40.03}{f^2} N_e.$$ \hspace{1cm} (1.20)

Here, $N_e$ is in cm$^{-3}$ and $f$ is in Hz.

According to the definition of sTEC, it is an integral of the electron density along the signal path. From Figure 1.6 and by considering Eq. (1.20) one can write sTEC as a function of the tangent altitude

$$sTEC(a) = \int N_e ds = \frac{f^2}{40.3} \int (1 - n) ds = \frac{f^2}{40.3} \left( L_{GL} - \int nds \right) = -\frac{S(a)f^2}{40.3},$$ \hspace{1cm} (1.21)

where $L_{GL}$ is the distance between the GPS and LEO satellites and quantity $S(a) = \int nds$ –
Figure 1.6. The GPS RO geometry for the straight line propagation of signal. Point O indicates the center of the Earth.

$L_{GL}$ is the so-called excess phase path, which can be determined from precise knowledge of the satellite orbits and readings from transmitter and receiver clocks. Since it is assumed that the GPS transmitted signals travel the same path, sTEC should be the same if it is calculated from $S_1$, $S_2$ or from $S_1 - S_2$:

$$sTEC(a) = -\frac{S_1 f_1^2}{40.3} = -\frac{S_2 f_2^2}{40.3} = \frac{(S_1 - S_2) f_1^2 f_2^2}{40.3(f_1^2 - f_2^2)}.$$  \hfill (1.22)

The advantage of using the combination of the two frequencies in (1.22) to find sTEC is that both orbit and clock errors are automatically eliminated in the difference [Schreiner et al., 1999]. In order to express sTEC through the electron density, which is to be made a function of distance from the Earth, one has to assume spherical symmetry. In this case

$$sTEC(a) = \int_s N_e(r) ds = \left[ \int_a^{r_{LEO}} + \int_a^{r_{GPS}} \right] \frac{N_e(r)r}{\sqrt{r^2 - a^2}} dr.$$  \hfill (1.23)

By calculating sTEC for all tangent altitudes, $a$, one can eliminate the part of sTEC that is the contribution from the integration from LEO to GPS (for details see Schreiner et al. [1999]). For the remaining calibrated $\tilde{sTEC}$ one has

$$\tilde{sTEC}(a) = 2 \int_a^{r_{LEO}} \frac{N_e(r)r}{\sqrt{r^2 - a^2}} dr.$$  \hfill (1.24)
Note, the form of the integral will be similar if one assumes the contribution of electron density from LEO to GPS is negligible for sTEC. Finally, one can apply the Abel transform to (1.24) to obtain the electron density altitude distribution

\[ N_e(r) = -\frac{1}{\pi} \int_{r}^{r_{\text{LEO}}} \frac{d(sT\dot{E})(a))}{da} \frac{da}{\sqrt{a^2 - r^2}}. \] (1.25)

The most significant source of errors in the electron density while using the described method is the assumption of spherical symmetry [Yue et al., 2010]. The spherical symmetry assumption is not satisfied at low latitudes where the electron density has strong horizontal gradients. Similarly, larger errors are expected at low altitudes where the downward error propagation from the F region also contributes to the net error. In addition, a possible large bending of the signal in the E region, which is thought to be associated with sharp density gradients (as in sporadic E) [Hajj and Romans, 1999], could make the assumption of straight line propagation of the signal no longer valid. The assumption of spherical symmetry may also be less valid even at mid-latitudes during geomagnetic storms, which is known to create various density gradients at different latitudes. In general, the expected accuracy for quiet-time F2-regions peak parameters are around 15-20% for NmF2 and 2-5% for hmF2 [Schreiner et al., 1999; Yue et al., 2010; Krankowski et al., 2011].

References


Richmond, A. D., E. C. Ridley, and R. G. Roble (1992), A thermosphere/ionosphere general


CHAPTER 2
MAGNETIC MERIDIONAL WINDS IN THE THERMOSPHERE
OBTAINED FROM GLOBAL ASSIMILATION OF
IONOSPHERIC MEASUREMENTS
(GAIM) MODEL

Abstract
Thermospheric neutral winds play an important part in the dynamics of ionospheric plasma at various temporal and spatial scales, and represent one of the key inputs for ionospheric physics-based models. Yet, available wind data is scarce and generally lacks the global coverage and continuity. To mitigate this shortcoming, a data assimilation model is used to estimate neutral winds in the low- and mid-latitude thermosphere simultaneously with the ionospheric electron density and the low-latitude electromagnetic drifts. Seasonal global maps of NmF2 and hmF2 generated from COSMIC GPS radio occultation (RO) measurements for geomagnetically quiet and low solar flux conditions are assimilated into the Utah State University Global Assimilation of Ionospheric Measurements Full Physics (GAIM-FP) model. The model uses a physics-based Ionosphere Plasmasphere Model (IPM), employs an ensemble Kalman Filter technique, and significantly improves the agreement between the modeled and measured NmF2 and hmF2 globally compared to the IPM model. In the present work, global quiet-time magnetic meridional winds in the thermosphere are derived for December and June solstices and March equinox. The morphology of the derived winds is analyzed and to validate the results, the estimated winds are compared with corresponding ground-based measurements and with wind values from the empirical horizontal wind model. The GAIM-FP estimated winds are shown to agree well with the wind observations. Furthermore, the sensitivity of the derived wind velocity to uncertain model parameters, including the O⁺-O collision cross section and the thermospheric neutral composition, is investigated. The analysis shows that uncertainties in these parameters are more important for nighttime winds over the mid-latitudes. The effect of the number of
radio occultations and the effect of uncertainties in the error estimates of the assimilated data on the estimated wind velocities are also studied. A random exclusion of 1/3 of the used data (∼50,000 RO measurements) to generate the global maps of NmF2 and hmF2, or a 30% increase of their errors is found to have only a small effect on the obtained wind velocity. The Results of this work indicate that thermospheric wind estimation from the ionospheric data assimilation model is a valuable tool for wind specification over regions where limited or no wind measurements exist.

2.1. Introduction

Thermospheric neutral winds play an important role in the dynamics of the low- and mid-latitude ionosphere [Rishbeth, 1972; Titheridge, 1995; Emmert et al., 2006; Makela et al., 2013]. At F-region heights, where the ion gyrofrequency is much larger than the ion-to-neutral collision frequency, mid-latitude neutral winds are effective to move ionospheric plasma along the tilted geomagnetic field lines. As a result, winds modify the field-aligned ionospheric plasma diffusion which, in turn, changes the altitude of the maximum electron density (hmF2) where chemical loss and diffusion processes approximately balance each other. Furthermore, the different recombination and ionization rates at the new hmF2 leads to a change in the maximum electron density (NmF2) [Rishbeth, 1972; Titheridge, 1995]. At low-latitudes, cross-equatorial winds [Meriwether et al., 2011] alter the ionization at the equatorial ionospheric anomaly crests and contribute to the global electron density asymmetry about the magnetic equator [Abub-Robb and Windle, 1969; Rishbeth, 1972; Anderson and Roble, 1981; Balan et al., 1995]. In addition, thermospheric winds influence the F-region electron density distribution indirectly by generating dynamo electric fields [Rishbeth, 1971; Heelis, 2004].

Since thermospheric neutral winds have a critical effect on the temporal and spatial variation of ionospheric plasma, they constitute a very important input for stand-alone ionospheric models, and uncertainties in their representation affect the accuracy of ionospheric modeling and specification [Miller et al., 1989; Richards, 1991; Titheridge, 1995; Schunk et al., 2004a]. Besides the fact the ionosphere-thermosphere is a strongly coupled
system and the thermospheric winds are important for ionospheric studies, the knowledge of neutral winds is also important for a better understanding of the general circulation of the thermosphere itself [Fuller-Rowell, 1998].

The thermospheric neutral winds can be calculated from physics-based global circulation models; however, their accuracy is affected by many uncertainties in the model, including ionospheric drag, which is not always well specified [Meriwether et al., 2013]. As regards to empirical wind models, the most widely used ones are Horizontal Wind Models, HWM93 [Hedin et al., 1991] and HWM07 [Drob et al., 2008], which are based on many years of wind measurements by various ground- and satellite-based instruments. Even though the performance of these models to specify the wind climatology are expected to be generally good over stations from where the wind data were incorporated into the model, studies show they may fall short of reproducing the observed climatology over places where no or few observations were included [Titheridge, 1995; Makela et al., 2012]. Therefore, it is vital to continue efforts toward advancing the existing wind models and to develop new methods of global wind estimation. This, in turn, will improve the accuracy of ionospheric models.

Valuable information about the thermospheric neutral wind can be obtained through direct observations; but, unfortunately, currently these observations are sparse both in space and in time. Ground-based thermospheric wind measurements, which are based on interferometric methods (such as Fabry-Perot Interferometer (FPI)) [Burnside et al., 1981; Meriwether et al., 2008], are mainly limited to nighttime and cloudless conditions, and, in contrast to ionospheric observations, are restricted to a handful of locations [Hedin et al., 1991; Emmert et al., 2006; Drob et al., 2008]. Space-based wind measurements can be performed with interferometers and spectrometers onboard satellites, and cross-track wind estimations are also possible from satellite-based accelerometers [Hedin et al., 1988; Liu et al., 2006; Doornbos et al., 2010]. However, current space-based missions lack the ability of routine wind measurements over a given location for all local times.

Alternatively, ionospheric observations can be used to deduce neutral winds along the magnetic meridian (or magnetic meridional winds) for any local time. For example, the
magnetic meridional wind can be derived from field-aligned plasma velocities obtained from Incoherent Scatter Radar (ISR) measurements [Salah and Holt, 1974; Oliver and Salah, 1988; Oliver, 1990]. An analysis of combined FPI and ISR observations by Burnside et al. [1983] indicates that the agreement between ISR-derived and FPI-measured magnetic meridional winds are expected to be within 35 ms$^{-1}$ over a typical mid-latitude site. However, considering the overall statistical uncertainty of the wind derivation from the ISR together with uncertainties in neutral atmospheric parameters and the ion-neutral collision cross section used in the estimation, the estimated overall wind accuracy during quiet and moderate geomagnetic conditions is thought to be up to 70 ms$^{-1}$ [Oliver and Salah, 1988]. An obvious limitation of this method is that it is also restricted to the very few locations of current ISRs.

Another widely used technique to obtain neutral winds in the thermosphere is the well-known servo method, which uses the variation of the NmF2 and hmF2, measured, for example, by ionosondes [Rishbeth, 1967; Rishbeth et al., 1978; Buonsanto, 1986]. A closely related wind estimation method, using a physics-based model and measurements of hmF2, was proposed by Miller et al. [1986] assuming a linear relationship between the wind speed and the distance between hmF2 and a so-called balance height. The results of this later method, however, was found to be essentially equivalent to the results from the servo method [Buonsanto et al., 1989]. The uncertainty in the wind estimation using the servo method is expected to be even larger than the uncertainties of ISR-derived winds [Buonsanto et al., 1997]. Furthermore, the servo method is generally valid only for mid-latitudes where neutral winds play an important role in the vertical dynamics of the plasma and at times when electric field effects are small. In addition, the method is not designed to be employed with measurements of plasma parameters other than hmF2 (sometimes together with NmF2) and for an optimal performance also requires data availability on nearly a continuous base, and consequently could only provide wind estimation over regions where the relevant ground-based ionospheric measurements were available.

A recent study by Luan and Solomon [2008] demonstrated the value of observations
from Constellation Observing System for Meteorology, Ionosphere, and Climate (COSMIC) to obtain information about thermospheric winds. They used variations of hmF2 retrieved from COSMIC radio occultation (RO) measurements with a servo model to estimate neutral winds for December solstice conditions, and studied their longitudinal variations by taking advantage of the global coverage of the COSMIC data. However, due to the above-mentioned limitations of the servo model, their study was limited to mid-latitudes.

In this paper, we present a new method of deriving thermospheric neutral wind based on Utah State University Global Assimilation of Ionospheric Measurements Full Physics (GAIM-FP) model. For this study, the model assimilates NmF2 and hmF2 values obtained from COSMIC radio occultation data to estimate the seasonal climatology of the magnetic meridional wind over the entire low- and mid-latitude regions. One of the main differences of this new technique to the servo method, which is generally used for mid-latitudes [Miller et al., 1989], is that it can estimate the wind velocity over low- and equatorial latitudes where uncertainties due to electric fields are overcome by a simultaneous wind and electric field estimation procedure. We show results of the quiet-time global wind climatology for three different seasons (December and June solstices, and March equinox) during the recent (2007-2009) solar minimum and compare our wind results with those measured by FPI over three different locations [Millstone Hill (42.6°N, 71.5°W), Arecibo (18.4°N, 66.8°W) and Arequipa (16.5°S, 71.5°W)]. In addition to the magnetic meridional wind estimation, the data assimilation model significantly improves the accuracy in the climatological description of the global ionospheric electron density compared to the case when the HWM93 empirical wind and the Scherliess-Fejer equatorial vertical drifts are used in the stand-alone Ionosphere-Plasmasphere model (IPM). Our results show that the thermospheric wind estimation using our ionospheric data assimilation technique has promising potential for a reliable wind specification over the regions where, traditionally, wind observations were sparse.

To address the sensitivity of the proposed wind estimation technique to uncertain model parameters and to quantify it, we have examined the effects of uncertainties in O+-
O collision cross section and neutral atmospheric density by performing wind estimation with different values of the collision frequency and exospheric temperature. In addition, we investigated sensitivity to the number of radio occultation data and to the magnitudes of the errors in the data. Limitations of the current study and future directions for the related investigations are also briefly discussed.

2.2. Data and Methodology

As noted above, the local time variation of NmF2 and hmF2 contains useful information about thermospheric winds. For this reason, for the current study we have used NmF2 and hmF2 data obtained from COSMIC radio occultation measurements and assimilated them into the GAIM-FP model. Unfortunately, the current number of daily occultations does not provide enough data to perform a global (and local) wind estimation on a day-to-day basis. However, the validity of COSMIC data to study ionospheric F-region climatology is well established [e.g., Liu et al., 2011; Burns et al., 2012]. As a compromise, seasonal average COSMIC data has been employed and the obtained winds are to be understood as climatological averages. The possibility of assimilating other ionospheric data types (such as total electron content (TEC)) with different spatial and temporal resolution is a subject of separate study and is not investigated in this paper. In the next two subsections, we describe the steps performed to utilize the COSMIC data for assimilation into GAIM-FP and give some details about the ionospheric models used in this study.

2.2.1. COSMIC Radio Occultation Data and F-Region Climatological Maps

We use the GPS radio occultation measurements from the Constellation Observing System for Meteorology Ionosphere and Climate/Formosa Satellite 3 (COSMIC). The constellation was launched in April 2006 and consists of six micro-satellites in a circular orbit at about 800 km altitude with a 72° inclination and 30° separation in longitude. Electron density profiles are retrieved from slant total electron content measurements using an Abel inversion technique [Hajj and Romans, 1999; Schreiner et al., 1999]. COSMIC has been providing about 1,000 - 2,500 electron density profiles per day globally [Rocken et al., 2000; Anthes
et al., 2008]. For the current study, the post-processed electron density profiles were obtained from the COSMIC Data Analysis and Archive Center (http://www.cosmic.ucar.edu). The COSMIC RO-derived electron density profiles and F2 region peak parameters have previously been validated by comparing them with measurements from ionosondes and ISRs [Lei et al., 2007; Kelley et al., 2009; Chuo et al., 2011; Krankowski et al., 2011] and are well-suited to study the global climatology of the F2 region ionosphere at low and mid-latitudes.

In order to prepare the COSMIC data for the assimilation into GAIM-FP, a multi-step process was employed. The values of NmF2 and hmF2 were extracted from individual electron density profiles and an initial quality check of the data was performed in order to eliminate erroneous NmF2 and hmF2 values from the analysis. This was accomplished by verifying that the extracted hmF2 was between ∼180-590 km and the electron density in a 10 km interval above and below hmF2 did not change more than by factor of two. This procedure removed apparent outliers, which made up about 3% for December solstice, 2.4% for March equinox, and 5% for June solstice of the initial data (seasons are defined below).

To describe the average global ionosphere during the recent solar minimum, the COSMIC data were combined from 2007 to 2009, during which the $F_{10.7}$ solar radio flux was nearly similar (yearly averages are about 73 for 2007, 69 for 2008 and 70 for 2009). The data were sorted into seasons consisting of December solstice (December 21 ± 27 days for the years 2007 and 2008), March equinox (March 20 ± 27 days for 2008 and 2009) and June solstice (June 21 ± 27 days for 2008 and 2009).

Data from the vernal and autumnal equinoxes were not combined because of the well-known ionospheric equinoctial asymmetry [Balan et al., 1998; Liu et al., 2010]. Note, the number of days in the selected intervals are multiples of the solar synodic rotation period of ∼27 days, which is known to influence ionospheric parameters [e.g., Ma et al., 2012, and references therein], and thus, its effect on the average values is minimized. In addition, January 2009 data does not extend beyond DOY 017, and therefore, excludes the period of prominent Sudden Stratospheric Warming (SSW), which is known to significantly affect ionospheric electron density, as seen using COSMIC data [Lin et al., 2012]. This was done to
eliminate possible influence of SSW on the regular seasonal values of NmF2 and hmF2. On the other hand, the selected periods provide a sufficient and almost uniform global coverage for our investigation. Since the current study focuses on geomagnetically quiet conditions, observations were excluded for periods with a previous 24-hour average $Ap \geq 16$ and/or corresponding 3-hour $ap \geq 16$, which eliminated between 5 - 13% of the remaining data. The total number of final NmF2 and hmF2 pairs that passed the applied rejection criteria were $\sim 161,600$ for December, $\sim 143,900$ for March, and $\sim 144,200$ for June, respectively. The average solar and geophysical conditions for the selected periods were $F_{10.7} = 73, Ap = 5$ for December, $F_{10.7} = 70, Ap = 5$ for March, and $F_{10.7} = 67, Ap = 4$ for June, respectively. The days that best describe the climatological average conditions of selected seasons are those for which solar declination angle is equal to the average angle during the selected 55-day interval. That results in the following representative days: DOY 006 for December solstice, DOY 079 for March equinox, and DOY 189 for June solstice.

To further reduce the COSMIC NmF2 and hmF2 data, we produced seasonal average global maps of NmF2 and hmF2 in 30-minute intervals of universal time (UT). For the binning a $7.5^\circ \times 2.5^\circ$ longitude/latitude grid was used with $15^\circ \times 5^\circ$ sliding window. To exclude additional outliers and other potentially erroneous data, the procedure was performed a second time and the data in a given grid point with a deviation from the mean value by more than $2\sigma$ were ignored (a similar approach to analyze TEC data was used by [Shim et al., 2008]). Finally, we excluded data for which the averages were based on less than three observations for a given grid point.

To illustrate the global distribution of the original COSMIC RO measurements during one of the selected 110-day periods and during a 30-minute interval, Figure 2.1 shows the locations of points on the globe where quiet-time NmF2 and hmF2 values were extracted. Note, the data distribution is nearly uniform with slightly more data points over the mid-latitudes, which is due to the satellites’ orbits and consequent RO geometry. Figure 2.2 shows the global distribution of NmF2 and hmF2 together with corresponding standard deviations at each grid point, which were obtained from the COSMIC ROs shown in Figure
Figure 2.1. Global distribution of COSMIC radio occultations for December solstice, geomagnetically quiet conditions, and UT=11:45 - 12:15. Circles indicate the locations of available COSMIC NmF2/hmF2 values. The total number of RO experiment in the figure is 3531.

2.1 (the values were obtained after performing the initial quality check of the profiles and then removing the outliers).

The solid black line in the figure indicates the location of the magnetic equator. Prominent features of the F-region ionosphere, such as the equatorial ionization anomaly seen in NmF2, larger hmF2 on the summer-side of the magnetic equator, and larger nighttime than daytime hmF2 over mid-latitudes, are clearly visible in Figure 2.2. The former should be attributed to the thermospheric neutral winds, which blow from the summer to the winter hemisphere causing the lift of the ionization in one side and lowering it on the other [e.g., Balan et al., 1995]. The high nighttime values of hmF2 at mid-latitudes are due to photochemical processes combined with the effect of neutral winds, which are equatorward at night and poleward during the day. The standard deviations in our case describe the variability of the corresponding parameters due to several factors, including geophysical variation in a daily and monthly scales, and different accuracy in parameter retrieval due to different RO geometry. On average, based on global maps (including high latitudes) of
Figure 2.2. Global seasonal maps of NmF2 (top) and hmF2 (bottom) with their corresponding 1-σ variations for December solstice conditions and UT=12:00 h.

σ for the selected three seasons, the relative error in NmF2 is about 24% ($6.3 \cdot 10^4$ cm$^{-3}$ absolute) and the absolute error in hmF2 is around 17 km (7% relative).

Finally, during assimilation, for the assigned errors to the hmF2 and NmF2, the following was applied: If $\sigma$(hmF2) was found to be less than 10 km, it was assumed to be 10 km, and for $\sigma$(NmF2)<$0.1$·NmF2 cases, the error was taken equal to $0.1$·NmF2. This helps to avoid unrealistically small errors in these parameters, which in reality should include the errors related to the instruments, measurement techniques, and Abel-inversion; however, these values are not available.

2.2.2. GAIM-FP Model

At Utah State University, two physics-based Kalman-filter data assimilation models for the Earth’s ionosphere have been developed. These models are the Gauss-Markov Kalman Filter Model (GAIM-GM) and the Full Physics-Based Kalman Filter Model (GAIM-FP) [Scherliess et al., 2006, 2009]. Both models are part of the Global Assimilation of Ionospheric Measurements (GAIM) project [Schunk et al., 2004a,b, 2005, 2011; Scherliess et al.,]
2004, 2006, 2009, 2011; Jee et al., 2007, 2008; Sojka et al., 2007; Thompson et al., 2005, 2006, 2009; Zhu et al., 2006]. Some of the data that have previously been assimilated by these models include in-situ electron density measurements from satellites, bottomside electron density profiles from ionosondes, slant TEC (sTEC) data from ground GPS stations, ultraviolet radiances from various space-based instruments, and sTEC from radio occultation data.

The GAIM-FP model is based on an ensemble Kalman filter approach [Evensen, 2009] and rigorously evolves the ionosphere and plasmasphere electron density field and its associated errors using a physics-based Ionosphere-Plasmasphere model (IPM) [Scherliess et al., 2004; Schunk et al., 2004a; Thompson et al., 2005]. The IPM is based on a numerical solution of the ion and electron continuity and momentum equations and covers the low and mid-latitudes from 90 to 30,000 km altitude (details of the physics-based model are given in Schunk et al. [2003]). The equations are solved along magnetic field lines for individual convecting flux tubes of plasma, and the 3-D nature of the model is obtained by following a large number of plasma flux tubes. The 3-D distribution is obtained by mapping the results on a geographic grid. The IPM model uses the International Geomagnetic Reference Field (IGRF) [Finlay et al., 2010], which properly accounts for the displacement between the geomagnetic and geographic equators and the bending of the magnetic field lines with latitude. In its current version, the model excludes the regions poleward of $\sim \pm 60^\circ$ geomagnetic latitude due to the vastly different physical processes that govern the high-latitude regions, e.g., convection electric fields, particle precipitation, etc.

The full physics-based data assimilation model provides specifications of the 3-dimensional electron and ion (NO$^+$, O$_2^+$, N$_2^+$, O$^+$, H$^+$, He$^+$) density distributions from 90 km to near-geosynchronous altitude ($\sim$30,000 km). In addition, the model can provide the global distribution of the ionospheric drivers (electric field, neutral wind and composition) that are consistent with the ionospheric observations. It is important to note that the estimation of the ionospheric drivers is an integral part of our ensemble Kalman filter and is achieved by using the internal physics-based model sensitivities to the various driving forces. In this
procedure, the ionospheric data are used to adjust the plasma densities and its drivers so that a consistency between the observations (within their errors) and the physical model is achieved. As a result, the assimilation procedure produces the optimal model-data combination of the ionosphere-plasmasphere system together with the set of drivers (electric fields, neutral winds, and composition) consistent with the ionospheric observations [Scherliess et al., 2009, 2011].

The GAIM-FP model was designed to specify ionospheric weather, but the model can also be run in a newly developed climatology mode. In this mode the model can assimilate seasonal NmF2 and hmF2 maps to produce the corresponding 3-D electron density distribution along with specifications of the underlying physical drivers. Throughout the simulations, one can selectively keep, for example, the global neutral composition as specified by the MSISE-90 model and/or the low-latitude $E \times B$ drifts as specified by the Scherliess-Fejer [Scherliess and Fejer, 1999] model.

GAIM-FP determines the global magnetic meridional wind field together with the low-latitude $E \times B$ drift velocities to match the modeled electron density parameters with the COSMIC NmF2 and hmF2 values. The COSMIC NmF2 and hmF2, which are functions of latitude, longitude, and time, strongly constrain a specification of the magnetic meridional wind globally and poleward/upward ion drift at low latitudes. Initially, the model starts from a zero wind field and from the empirical values of low-latitude $E \times B$ drifts and adjusts both of them as needed.

The ensemble Kalman filter assimilation procedure was implemented as follows. Twenty-four hours before the representative day, the plasma distribution obtained from the IPM run was taken to be the initial distribution at the start of the assimilation. Every 15 minutes, the global maps of COSMIC NmF2 and hmF2 (section 2.2.1) were assimilated, the ensemble of ionosphere/plasmasphere model runs was integrated forward in time, and the model error covariance matrix was determined. Using the new data and the new error matrix, the ensemble Kalman filter reconstructed an updated estimate of the plasma distribution and ionospheric drivers (winds and drifts). The new drift and wind velocities were fed
back into IPM and the assimilation was repeated at the next 15-min time mark. As time advanced, the ensemble Kalman filter produced a 3-D, time-dependent, plasma distribution for which NmF2 and hmF2 converged towards the COSMIC NmF2 and hmF2. Note, when a simulation was advanced to the next day, the assimilated data remained the same. The convergence of the simulation results for the desired representative day was verified by comparing them to the results from the previous day. The final modeling results were output every 15 minutes and cover the full 24 hours.

For the current study, simulations were performed with both, IPM and GAIM-FP. While GAIM-FP calculates the ionospheric drivers, the IPM uses the HWM93 empirical wind model for the winds in the thermosphere and the Scherliess-Fejer model for the low-latitude ion drifts. In both ionospheric models, the identical MSISE-90 empirical model setup is used for the neutral atmosphere, and also all other input parameters (except winds and drifts) were the same. For the selected seasons, the outputs of empirical models were computed for DOY 006, DOY 079, and DOY 189.

When interpreting the wind results from GAIM-FP there is one caveat that needs to be mentioned. Over the equator and low latitudes, the vertical motion of plasma due to $E \times B$ drifts play a very important role in ionospheric variations and their effects are separated from the winds’ effects internally by GAIM-FP. The former moves plasma across the horizontal geomagnetic field lines, while the latter is effective in moving F-region ionospheric plasma along the field lines. Over the mid-latitudes, however, the upward motion due to $E \times B$ drifts is not separated from the motion due to the winds by the model. Therefore, the derived magnetic meridional wind ($v_{mag}$) at mid-latitudes should contain contribution from the ion drifts and, for this reason, it is typically denoted as an effective or equivalent wind. Assuming an upward/poleward drift perpendicular to the magnetic field ($v_{p\perp}$) is present, then its effect can be expressed by horizontal wind with the velocity of

$$v_{eff-mag} = v_{mag} - \frac{v_{p\perp}}{\sin I}, \quad (2.1)$$

where $I$ is the geomagnetic field inclination angle. Hence, to obtain the best estimate of
the "true" wind, ideally, $v_{p\perp}/\sin I$ should be added to the GAIM-FP estimated effective wind over the mid-latitudes. Unfortunately, the value of $v_{p\perp}$ is generally unknown. It should, however, be noted that its effect during geomagnetically quiet times are expected to be small and indeed, our estimation based on Scherliess-Fejer mid-latitude drift model [Scherliess et al., 2001] for the $v_{p\perp}$ over the mid-latitude Millstone Hill station show the effect is not large (about 15 ms$^{-1}$). Note, all mid-latitude wind results in this paper present the effective wind.

During the wind estimation processes in GAIM-FP, it is assumed the magnetic meridional wind is constant in altitude. This is a valid assumption during quiet times above the F-region peak, where the altitude variations are smoothed out due to the increasing kinematic viscosity; however, at lower altitudes, wind is expected to have altitude gradient. Height varying wind affects the plasma distribution and the diffusion velocity, and could modify the estimated wind velocity. To better understand the role of these effects on estimated wind values, the separate model analyses are required.

Finally, there are several uncertain model parameters in the ionospheric models that could lead large uncertainties in the model results [Jee et al., 2005; Schunk et al., 2012]. One of these parameters is the O$^+$-O collision cross section, which is a very important parameter in various thermosphere-ionosphere studies; however, its value is only known within a factor of two accuracy [e.g., Nicolls et al., 2006, and references therein]. The same parameter is considered to be one of the major sources of the errors, along with neutral density, in various thermospheric wind estimation methods [Buonsanto et al., 1989; Miller et al., 1989; Oliver, 1990; Richards, 1991; Buonsanto and Witasse, 1999]. Similarly, it is also expected that the wind values from GAIM-FP are sensitive to the O$^+$-O collision frequency and to the specified neutral composition (currently MSISE-90).

2.3. Results

In this section, first a comparison of the ionospheric NmF2 and hmF2 from COSMIC to corresponding results from GAIM-FP and IPM are presented. Next, the derived neutral winds are shown and compared with their corresponding measurements from FPI instru-
ments and with values from the HWM93 model. Finally, the sensitivity of the derived wind velocity to uncertain atmospheric model parameters is shown. An investigation of the obtained electromagnetic drifts, including their sensitivity analysis and validation are out of the scope of this paper and will be considered in future studies.

2.3.1. Comparison of COSMIC, GAIM-FP and IPM NmF2 and hmF2

In the following, we compare the COSMIC NmF2 and hmF2 maps with the results from the GAIM-FP and IPM model simulations. The GAIM-FP model assimilates the COSMIC data and estimates the low- and mid-latitude thermospheric winds, and the poleward/upward component of low-latitude $\mathbf{E} \times \mathbf{B}$ drift. The assimilation procedure continuously adjusts the model values of these parameters to produce the best fit to their global observed distribution. On the other hand, IPM uses the HWM93 empirical wind model for the winds in the thermosphere and the Scherliess-Fejer model for the low-latitude ion drifts. In both cases, the neutral atmosphere was obtained from MSISE-90 empirical model.

The left panels of Figure 2.3 show the COSMIC maps of NmF2 for three seasons (December solstice, March equinox, June solstice) at UT=12 h. The corresponding differences between the COSMIC maps, and the GAIM-FP and IPM results are shown in the middle and right panels, respectively. Here the values are normalized to the observations and thus show relative differences. The values of less than 0.2 were set to 0 since such quantities are close to the average standard deviations of the COSMIC data, and in doing so aids to visually better highlight areas with larger discrepancies. The COSMIC data are only shown for low and mid-latitudes with the same latitudinal extent as the model simulations. Similar to Figure 2.3, Figure 2.4 shows results for hmF2, and this time for absolute differences with values less than 15 km set to 0. The comparison of the COSMIC data to the model results shows the agreement between data and model is significantly improved for GAIM-FP for both, NmF2 and hmF2 when compared to IPM (Figure 2.3 - 2.4). This shows the data assimilation model was able to obtain more accurate values for both F2-layer peak parameters by simultaneously adjusting thermospheric neutral winds and $\mathbf{E} \times \mathbf{B}$ drifts. The derived winds and drifts push the ionospheric model parameters closer to the observed
values than those obtained from the empirical models; and this is the case for daytime and nighttime, for any of the three seasons and for low or mid-latitudes. It is also apparent that for these cases, on average, the IPM model overestimates both NmF2 and hmF2. The results from GAIM-FP are closer to the observations, albeit certain systematic differences are still present. For example, NmF2 seems to be underestimated, while hmF2 remains higher than their observed values. It should be noted, in general, the improved performance of GAIM-FP compared to IPM is expected. However, what is important to note is that the adjustment of the winds and drifts improve both, NmF2 and hmF2, simultaneously.

In order to further quantify the improvements in the description of the F2-layer parameters by the data assimilation model for all UTs, the root mean square error (RMSE) was used as an indicator of the model accuracy and was evaluated for hmF2 and NmF2 for
Figure 2.4. Same as Figure 2.3, but for hmF2 and absolute differences in hmF2. Here, differences less than 15 km are set to zero.

A given season and location:

$$RMSE(hmF2) = \sqrt{\frac{\sum_{i=1}^{n} (hmF2_{modeli} - hmF2_{datai})^2}{n}},$$

$$RMSE(NmF2_{rel}) = \sqrt{\frac{\sum_{i=1}^{n} (NmF2_{modeli}/NmF2_{datai} - 1)^2}{n}},$$

where $i$ runs for all available UT bins ($n \leq 48$). Relative NmF2 was chosen to accommodate for the order of magnitude difference between daytime and nighttime NmF2 values. The results are shown in Figures 2.5 and 2.6, which once again indicate the data assimilation significantly improves the agreement between the model and data for all seasons over the low and mid-latitudes. In addition, if one compares GAIM-FP performances over the low and mid-latitudes, it becomes clear that, even though it outperforms IPM in both cases,
Figure 2.5. Global distribution of the mean square error of relative NmF2 \( \text{RMSE}(NmF_2_{\text{rel}}) \) for December solstice (left), March equinox (middle), and June solstice (right) and for GAIM-FP (top) and IPM (bottom).

Figure 2.6. Same as Figure 2.5, but for absolute hmF2.
relatively larger discrepancies still remain over equatorial and low latitudes. Note, the large differences in IPM NmF2 and hmF2 over the south Atlantic and south Pacific are considerably reduced in the GAIM-FP results. Significant improvements are also seen over eastern Asia. It should be mentioned that there are certain places where the IPM model is performing relatively well (for example, over the southern part of the north America and Australia) and no major differences in the two modeling results are visible.

Figures 2.7 - 2.9 show detailed comparison of the local time variation of the modeled and measured values of NmF2 and hmF2 over stations where ground-based wind measurements are available. Further below we will use the same locations to compare the wind results.

These variations are shown for three seasons (December solstice, March equinox, and June solstice) over Millstone Hill, Arecibo, and Arequipa. The COSMIC data and their corresponding error bars are the same as those used in the data assimilation. For the comparison, the model and the COSMIC data were both taken at the grid point, which was closest to the corresponding ground station [for Millstone Hill at (43.0°N, 71.25°W), for Arecibo at (19.0°N, 63.75°W), and for Arequipa at (17.0°S, 71.25°W)]. In Figures 2.7 - 2.9 the GAIM-FP results over Millstone Hill during all three seasons are significantly closer to the COSMIC values than their corresponding IPM values. Over Arecibo and Arequipa, both the GAIM-FP and IPM results closely describe the observed local time variations of NmF2 and hmF2. The results show the data assimilation and driver estimation procedure produces reliable agreement between the assimilated and modeled parameters. Note, for the locations of Arecibo and Arequipa, GAIM-FP simultaneously estimates the magnetic meridional wind and $\mathbf{E} \times \mathbf{B}$ drift, and thus indicating the model successfully decoupled these two drivers.

2.3.2. Derived Neutral Winds

Global Magnetic Meridional Winds

As mentioned before, when assimilating NmF2 and hmF2, GAIM-FP produces an estimate of the magnetic meridional wind ($v_{mag}$) over the low and mid-latitudes at F-region
Figure 2.7. Local time variation of $N_{mF2}$ (left) and corresponding $hmF2$ (right) as observed from COSMIC (black circles), and modeled by IPM (magenta), and GAIM-FP (blue) over the Millstone Hill (top), Arecibo (middle), and Arequipa (bottom) for December solstice conditions. Error bars indicate 1-σ variations in the COSMIC data.
Figure 2.8. Same as Figure 2.7, except for March equinox.
Figure 2.9. Same as Figure 2.7, except for June solstice.
heights. Figure 2.10 shows the derived magnetic meridional winds for different seasons and UTs at 250 km altitude, with positive values indicating the northward flow. The obtained wind pattern agrees well with its well-established characteristics. Namely, the wind is equatorward during the night and poleward during the day, the magnitude of wind speed is up to $\sim 150 \text{ ms}^{-1}$, the flow is predominantly from summer to winter hemisphere during solstices (compare December solstice to June solstice), and is more symmetric about the geographic equator during the equinox. Longitudinal variations in the wind velocity can also be seen in Figure 2.10 by comparing different UTs of a given season. The results indicate that the estimated low- and mid-latitude winds are expected to realistically well represent the thermospheric dynamics.

**Comparison of Derived Winds to FPI measurements**

In order to assess the reliability of the obtained GAIM-FP magnetic meridional wind climatology, the GAIM-FP derived winds were compared with direct wind observations obtained from the Millstone Hill, Arecibo, and Arequipa FPI stations, spanning the latitudinal range from mid-latitudes to equatorial locations. The FPI instruments obtain geographic zonal and meridional components of the neutral wind, corresponding to an altitude of about 250 km, by measuring the line-of-sight Doppler shift of 630.0-nm nightglow emission at different azimuth angles. The three selected stations have provided extensive nighttime wind data over several decades [e.g., Emmert et al., 2006, and references therein]. In addition, upgraded instruments at these stations [Noto et al., 2006, 2011; Meriwether et al., 2008] currently provide valuable wind measurements with higher accuracy. The newer data over Arecibo and Arequipa are available for certain months in 2008 and 2009 and overlap with the period for which the COSMIC data were averaged for this study. Unfortunately, there is an observational gap in Millstone Hill data during the recent solar minimum and new wind data are only available starting from the end of 2009.

To create the seasonal climatology of magnetic meridional winds from the FPI observations, first, all available zonal and meridional wind data for each station were obtained from the Madrigal database (http://www.openmadrigal.org/). The total number of data points
for meridional (zonal) wind were over 83,100 (78,100) for Millstone Hill (years spanned: 1989-2002, and 2009-2013), over 27,300 (27,200) for Arecibo (years spanned: 1980-1994, 1996-2005, 2008-2009, and 2012-2013) and over 22,800 (27,200) for Arequipa (years spanned: 1983-1984, 1986, 1988-1990, 1996-1999, 2005, 2007-2010, and 2012-2013). Note, the seasonal distribution of the data is not uniform, and the time resolution varies from station to station and from year to year, ranging from more than 30 minutes to less than 5 minutes. Also note, the FPI wind data for the selected three stations have previously been used in a variety of different studies [e.g., Fejer, 1993; Biondi et al., 1999; Emmert et al., 2006; Brum
et al., 2012.

The obtained FPI data were sorted by season in the same way as the COSMIC data, and only measurements with $F_{10.7} < 100$, 3-hour $ap < 16$, and 12-hour average $Ap < 16$ were considered. The mean zonal and meridional winds were computed for every 15-minute bin together with their corresponding standard deviations ($\sigma$). Again following our analysis of COSMIC data, the mean and $\sigma$ were used to remove data, which deviated from the mean by more than $2\sigma$ and the remaining data were used to recalculate the mean and the standard deviation. In the final step, only averages based on more than 10 data points were considered. Note, the average wind velocity for a given time bin contains contribution from measurements during different solar cycles. Furthermore, different time bins may combine a different number of data points, and zonal and meridional wind averages may also be based on slightly different number of measurements. The median number of distinct nights that contributed to the final results of our wind climatology are 47, 27, and 81 for Millstone Hill during December, March, and June, respectively; and similarly 27, 10.5, and 26 for Arecibo and 0, 23, and 75 for Arequipa. Finally, to obtain the magnetic meridional wind (assuming the vertical component of neutral wind velocity is zero) the two horizontal components of the wind were combined by

$$v_{mag} = u \cdot \sin D + v \cdot \cos D,$$

(2.2)

where $u$ and $v$ denote the geographic eastward and northward components of the wind, respectively, and $D$ is the magnetic declination angle. To calculate corresponding $\sigma$, the appropriate error propagation expression, $\sigma_{v_{mag}} = \sqrt{\sigma_u^2 \sin^2 D + \sigma_v^2 \cos^2 D}$, was employed. Here, the declination angles of the closest grid points to the corresponding ground stations are: $D = -14.4^\circ$ for Millstone Hill, $D = -13.3^\circ$ for Arecibo, and $D = -3.4^\circ$ for Arequipa.

Figure 2.11 compares the estimated thermospheric magnetic meridional winds at 250 km altitude to their corresponding nighttime FPI values and also to their corresponding values from the empirical HWM93 model (a positive value here indicates an equatorward wind). The error bars of FPI wind in Figure 2.11 correspond to one standard deviation.
and characterize the geophysical variations of the quiet-time winds, day-to-day weather variability, as well as instrument errors. Figure 2.11 shows the GAIM-FP winds are in very good agreement with the observations during the night and also are reasonably close to HWM93 winds throughout the day for all stations and seasons. The estimated wind pattern has mainly diurnal variation that is in agreement with HWM93. The larger discrepancies between GAIM-FP and HWM93 are around 12-16 LT over the Millstone Hill, however this is not a surprise since the difference between two model results of NmF2 and hmF2 are also larger during this period of the day (see Figures 2.7 - 2.9). In addition, there are differences in the morning reversal times of the wind over Millstone Hill, and the GAIM-FP winds reach their maximum nighttime values about 1.5 - 2 hours earlier than the HWM93 and FPI winds. Note, the largest difference between the GAIM-FP and the FPI winds are during June solstice, a fact we will discuss later.

In overall, the presented comparisons in Figure 2.11 demonstrate the GAIM-FP estimated wind climatology is reliable and nearly always within the 1-σ range of the observed data over the low and mid-latitudes during all seasons. Note the HWM93 model, when constructed, included ground-based wind data from either ISR or FPI (or from both) over these three stations [Hedin et al., 1991] and its performance is, therefore, expected to be more accurate than over locations on the globe where ground-based wind data are relatively sparse. The close agreement between the HWM93 and FPI results over the Millstone Hill for solar minimum conditions was also obtained by Fejer et al. [2002]. In contrast to HWM93, for GAIM-FP the selected three stations do not have any advantages over other low- and mid-latitude locations, assuming that other model inputs have similar uncertainties elsewhere over the globe. The reader is reminded that the GAIM-FP estimation of the wind is being started from the zero wind field and thus, no information about the wind pattern is specified in advance.

2.3.3. Sensitivity of Derived Wind Velocity to Uncertain Model Parameters

It is well known that ionosphere models are sensitive to a variety of uncertain atmospheric and ionospheric parameters [Jee et al., 2005; Jenniges, 2011; Schunk et al., 2012].
Figure 2.11. Local time variations of seasonal magnetic meridional winds at 250 km altitude over the Millstone Hill (left), Arecibo (middle), and Arequipa (right), and during December solstice (top), March equinox (middle), and June solstice (bottom) conditions. Circles indicate the seasonally averaged FPI observations in 15-min intervals with their corresponding 1-σ variability. Also shown are the estimated winds from GAIM-FP (blue) and HWM93 (magenta). Positive values indicate an equatorward flow.
Furthermore, the use of a data assimilation model introduces additional uncertainties associated with the assimilated data. In this section, we show how some of the most important uncertainties can influence the estimated wind values. Here we focus on the $O^+\cdot O$ collision frequency, the neutral composition, the effects of the number of radio occultations, and the effects of the assumed errors of the assimilated data on the derived wind velocity.

**Sensitivity to $O^+\cdot O$ Collision Cross Section and Neutral Composition**

The $O^+\cdot O$ collision frequency is a crucial parameter, which determines the diffusion of $O^+$ in the neutral atmosphere at F-region ionospheric heights and above. Consequently, the estimated wind values obtained from any method, including ours, that uses ionospheric observations in the wind estimation process are expected to be sensitive to it. Variations of this parameter from the theoretical value given by Banks [1966] are typically expressed as the ratio of the adopted and the theoretical value. This ratio is often called the Burnside factor (F) [e.g., Salah, 1993]. To date, various studies have determined different values of this ratio that range from 0.75 to 2.0 [Nicolls et al., 2006], and the effect of the Burnside factor on IPM results of electron density was discussed by Jenniges [2011] and Schunk et al. [2012]. By default, IPM and GAIM-FP use $F = 1.0$ and, therefore, all model results presented so far are based on this value.

In order to determine the sensitivity of the estimated magnetic meridional winds to this factor, we have changed the Burnside factor to $F = 1.4$ in GAIM-FP and rerun the model with this modified value. The values of $NmF2$ and $hmF2$ determined from GAIM-FP were virtually unaffected by this change (not presented), indicating that the model was able to match these parameters to the observations with the same degree of accuracy as before. On the other hand, the agreement in $NmF2$ and $hmF2$ comes at the expense of changes in the magnetic meridional wind. As an example of the sensitivity of the derived wind velocity to the Burnside factor, the differences between the estimated winds for $F = 1.4$ and those for $F = 1.0$ are shown in the upper three panels of Figure 2.12. The results correspond to UT = 6h and December solstice (right), March equinox (middle), and June solstice (left) conditions. The positive values in the northern hemisphere and negative
Figure 2.12. Global difference in magnetic meridional winds at UT = 6h for December solstice (left), March equinox (middle), and June solstice (right) when O⁺-O collision cross section is increased from 1.0 to 1.4 (upper panels) and when exospheric temperature in the MSIS model is reduced by 60 K (lower panels). Contour intervals correspond to 5 ms⁻¹.

values in the southern hemisphere, both indicate the increase of poleward component of the wind (note adopted convention on wind direction and velocity sign in Figure 2.10). The seasonal pattern of shown differences are very similar with peak values occurring over the upper mid-latitudes near 90° west, which corresponds to local midnight. Different from nighttime conditions, the values are smaller during the day. In addition, for a given location the magnitude of the difference changes with season.

Figure 2.13 shows the global differences between average daytime (09 - 15LT) and nighttime (21 - 03LT) winds separately for the same three seasons. In agreement with the results of Figure 2.12, the sensitivity of the winds to the Burnside factor is largest over the mid-latitudes and during nighttime. The average nighttime values of wind differences over mid-latitudes (30° - 60° and -30° - -60°) are between ∼19 - 25 ms⁻¹; however, during the day they are only ∼5 - 7 ms⁻¹. Note, even though there is no apparent seasonal pattern, the detailed comparison shows, on average, ∼5 ms⁻¹ larger change in summertime mid-latitude wind velocity at night. Over the low latitudes (±25°), the differences do not change much
Figure 2.13. Global difference in daytime (09:00 - 15:00 LT) (top) and nighttime (21:00 - 03:00 LT) (bottom) magnetic meridional winds for December solstice (left), March equinox (middle), and June solstice (right) when $O^+ - O$ collision cross section is increased from 1.0 to 1.4. Contour intervals correspond to 5 ms$^{-1}$.

during the course of the day and the daily average is about 3 - 5 ms$^{-1}$. Also note, during the night the wind is equatorward, and therefore the increase in the Burnside factor decreases the equatorward wind velocity ($\Delta V_{mag} > 0$ in the northern hemisphere and $\Delta V_{mag} < 0$ in the southern hemisphere). During the day, the effect is to increase the poleward wind, though it is small compared to the nighttime case.

Another potential source of error in the wind estimation is an uncertainty in the neutral composition and temperature. GAIM-FP uses the MSISE-90 empirical model to obtain these parameters, which have estimated uncertainties of about 15 - 20% [Hedin et al., 1991]. This error might be different for various constituents and may also vary by geophysical conditions and locations. Different from the Burnside factor that was important for the $O^+$ diffusion, the neutral composition and temperatures also affect production and loss processes, and therefore, may have more in general complex effects. To investigate the sensitivity of our neutral wind estimation to uncertainties in these parameters, we globally reduced the exospheric temperature ($T_{ex}^{MSIS}$) by 60 K for each season and repeated the data assimilation and wind estimation procedure. According to the MSIS model, such
change in temperature results in \( \sim 14\% \) reduction of the global (within our model extent) average mass density at 250 km, while the reduction is \( \sim 31\% \) at 400 km. The reduction of the thermospheric temperature was prompted by the results of Emmert et al. [2010] who revealed smaller than expected thermospheric densities during the recent solar minimum.

The results from the reduced \( T_{ex} \) model simulation were compared with the default case (with no temperature change). The comparison revealed, once again, the modeled values of \( NmF2 \) and \( hmF2 \) were very similar to the default case with slight improvements in the global RMSE(\( hmF2 \)) (by \( \sim 2 \) km) and marginal worsening of RMSE(\( NmF2_{rel} \)) (\( \sim 1.5\% \)). The differences in the wind velocity were comparable or larger in magnitude to the case when we increased the Burnside factor but opposite in direction. This is shown in the lower three panels of Figure 2.12, which display that the effect of an exospheric temperature reduction (and consequent reduction of the densities of the neutral atmosphere species) is larger at night and over the mid-latitudes. Similar to Figure 2.13, Figure 2.14 shows the differences between daytime and nighttime winds. The results demonstrate that the effect of a temperature reduction on the estimated magnetic meridional wind velocity is similar to a decrease of the Burnside factor. However, different from the later, there is a clear seasonal pattern at night, showing the effect is larger in the winter hemisphere. The temperature reduction effects over the two hemispheres are comparable during the equinox. An average change for wintertime mid-latitude winds at night is \( \sim 28 \) ms\(^{-1} \), for equinox \( \sim 25 \) ms\(^{-1} \) and for summertime \( \sim 15 \) ms\(^{-1} \). The seasonal pattern in the daytime mid-latitude wind differences is not well defined and the difference is \( \sim 8 \) ms\(^{-1} \). The differences in low-latitude winds also do not show a strong seasonal or local time dependence and their daily average is about \( 5 \) ms\(^{-1} \).

**Sensitivity to Number of Occultations and Data Errors**

To further investigate the robustness of our wind estimation, we have studied the effects of the number of radio occultations and the effects of uncertainties in the errors of the assimilated data. For this we have focused on the December solstice period and randomly excluded 1/3 of the original quiet-time COSMIC radio occultations during that
period and then performed the data processing and binning as described in section 2.1. The
generated new maps of NmF2 and hmF2 were assimilated into GAIM-FP and the global
magnetic meridional wind was determined. This wind was compared to the corresponding
case, which assimilated all available COSMIC data. The upper panel of Figure 2.15 shows
the root mean square error of the two wind speeds are around 7 ms\(^{-1}\) with some random
places where it reaches 10-15 ms\(^{-1}\). The result indicates that a random reduction in the
number of occultations by 1/3, which for our case is \(\sim 50,000\), has a small effect on the wind
estimation.

Next, we have studied the sensitivity of our wind estimation on the assigned data
errors, which are known to be central to any data assimilation technique. Specifically, to
test how sensitive the estimated wind values were to the errors we have assigned to the
COSMIC data, we increased all errors of NmF2 and hmF2 by 30% in the default COSMIC
maps and again performed the data assimilation for December solstice. The results of the
wind differences are presented in the lower panel of Figure 2.15. It is apparent the impact
of the assumed increase of errors is relatively small with values of about 4 ms\(^{-1}\).
Figure 2.15. Sensitivities of the estimated December solstice GAIM-FP wind velocity to the number of radio occultations when 1/3 of COSMIC ROs were excluded (top) and to the measurement errors when all errors were increased by 30% (bottom). Sensitivity is shown as a root mean square error with contour intervals corresponding to 1 ms$^{-1}$.

2.4. Discussions

The ability to reliably estimate the thermospheric wind is important for accurate modeling of the ionosphere, to better elucidate ionospheric and thermospheric phenomena related to the neutral winds, and for an improved understanding of thermospheric dynamics.

In this paper we presented results of the quiet-time magnetic meridional wind climatology over the low and mid-latitudes, as estimated by the GAIM-FP data assimilation model. The assimilated data consisted of seasonally averaged global maps of NmF2 and hmF2 generated from the COSMIC GPS radio occultation data. Within the data assimilation model, GAIM-FP estimated the thermospheric winds simultaneously with the low-latitude $\mathbf{E} \times \mathbf{B}$ drifts. We found the estimated ionospheric drivers improved the agreement between the modeled and observed ionospheric F2-layer parameters globally. This is in contrast to the case when the IPM ionospheric model, which is used in GAIM-FP, uses the values for
the ionospheric drivers from the corresponding empirical models. The estimated magnetic meridional winds obtained from GAIM-FP were compared with FPI data and with HWM93 wind values. A close agreement was obtained between the modeled and the measured wind values. In addition, the sensitivity of the estimated wind velocity to several important and uncertain atmospheric and model parameters was investigated.

2.4.1. Ionospheric Parameters

The presented results demonstrated that the GAIM-FP model is able to achieve very close match with the COSMIC data. Namely, the assimilated seasonally averaged values of NmF2 and hmF2 from COSMIC RO measurements are in good agreement with the corresponding GAIM-FP results over the entire low- and mid-latitude domain and for all local times. It is important to emphasize the model’s ability to simultaneously improve the modeling of both, NmF2 and hmF2 when contrasted against the simulation without assimilating the data. To achieve this close agreement, GAIM-FP estimates the global neutral wind and low-latitude vertical $\mathbf{E} \times \mathbf{B}$ drifts, and therefore, provides valuable information about these ionospheric drivers, as well. In turn, the best possible description (within the model and data errors) of F2-layer main parameters on a global scale that are consistent with the estimated ionospheric drivers is an important tool to better understand the role of these drivers in the variation of the electron densities. This is especially important over places where measurements of ionospheric/thermospheric parameters have been historically limited. For example, the possibility to employ GAIM-FP to understand the causes of the so-called Weddell Sea Anomaly was reported by Lomidze and Scherliess [2010], and a more detailed investigation to establish the role of various physical processes in its generation is carried out in Chapter 4.

Even though the GAIM-FP results agree well with the corresponding COSMIC data, the agreement seems to be better over the mid-latitudes. Different to mid-latitudes, at equatorial and low latitudes the model separates the plasma motion across the Earth’s magnetic field due to $\mathbf{E} \times \mathbf{B}$ drifts from the field-aligned plasma motion due to the neutral winds. For example, the daytime drifts over the magnetic equator are responsible for the plasma
fountain effect that lifts the F layer and creates the so-called equatorial ionization anomalies, while the neutral wind is responsible for the hemispheric asymmetries in the electron densities of these anomalies [Rishbeth, 1972; Schunk and Nagy, 2009]. The validation of the GAIM-FP estimated vertical $E \times B$ drifts is in progress, but is beyond the scope of this paper.

In addition to the possible difficulties associated with separating the different driving forces, it is also important to point out that the accuracy of the Abel-inversion to obtain electron density profiles from the RO measurements and correspondingly, the accuracy of values of $N_{mF2}$ and $h_{mF2}$ are known to be less reliable over the equatorial and low latitudes because of the spherical symmetry assumption employed in the Abel-inversion [Yue et al., 2010]. These complications at low and equatorial latitudes may explain the larger differences between the modeled and corresponding COSMIC ionospheric parameters at these latitudes.

We noted in section 3.1, even though the GAIM-FP results were significantly closer to the observations when compared to the IPM results, they still generally underestimated $N_{mF2}$ and overestimated $h_{mF2}$ globally (IPM overestimated both parameters). This discrepancy can be understood when considering that in the current configuration of GAIM-FP, the neutral composition was taken from MSISE-90 and the model only adjusted the equatorial vertical drifts and the magnetic meridional winds, given the observational data and their associated errors. Therefore, a reduction in $h_{mF2}$, for example, would result in even smaller $N_{mF2}$ values according to the dynamics and chemistry in the model. Given the current constraints, the model found the optimal agreement between the data and the model results. It is expected the agreement will be further improved if the neutral composition also becomes part of the estimation process. A detailed investigation of this is planned for the future.

Finally, relatively large discrepancies between the modeled and observed densities can be seen near the poleward edge of the mid-latitudes during local winter and to a lesser degree during equinox (Figure 2.5 - 2.6). This discrepancy can most likely be attributed to the well-known mid-latitude trough [Rodger et al., 1992], which was also seen in the COSMIC
NmF2 by He et al. [2011]. Several processes taking place in the high-latitude ionosphere are considered to be responsible for the trough formation. These processes are currently not part of GAIM-FP, and therefore, it is expected that the modeled density parameters within the mid-latitude trough are not well represented by the model.

2.4.2. Wind Estimation

Thermospheric winds are known to play a key role in the variation of the F-region electron density. This was unequivocally demonstrated in this study when we contrasted the GAIM-FP and IPM modeling results, which used identical atmospheric and ionospheric parameters except for the global neutral wind and equatorial and low-latitude $\mathbf{E} \times \mathbf{B}$ drift. The GAIM-FP results, using the model estimated winds (and drifts at low latitudes), were significantly closer to the COSMIC data than the results from the corresponding IPM model run which used the empirical HWM93 wind model and the Scherliess-Fejer equatorial vertical drift model. Our study also showed that over certain places, the difference between the IPM modeled and COSMIC measured values of NmF2 can be more than 100% and more than 50 km between the modeled and measured hmF2. These results suggest, over these locations, the estimated ionospheric drivers might be more accurate than their corresponding values from the empirical models. The fact the empirical wind models may fail to even produce the observed climatology was demonstrated by Makela et al. [2012] by comparing FPI winds with those from HWMs over a location that was not used in the generation of the empirical models. The better accuracy of our estimated winds compared to HWM93 over similar places is further supported by the results of our comparison of GAIM-FP winds with FPI observations. The GAIM-FP winds were found to be in very close agreement with the observations at three different low- and mid-latitude stations (Millstone Hill, Arecibo, and Arequipa). Note, the accuracy of the GAIM-FP winds was comparable with the HWM93 values at these sites; however, for these three locations, the ground-based wind observations were used in the construction of the HWM93 model, and therefore, it can be expected the HWM93 values agree well with the wind measurements at these locations. For GAIM-FP, on the other hand, these three sites are no different from any other locations (assuming other
model inputs are of the same accuracy elsewhere), which indicates the estimated GAIM-FP winds are globally reliable.

The GAIM-FP estimated global wind pattern (Figure 2.10) agrees well with the theory of driving quiet-time, large-scale thermosphere dynamics mainly by horizontal pressure gradient forces, which are created by solar extreme ultraviolet (EUV) heating [Rishbeth, 1972; Titheridge, 1995]. The high-latitude forcing for global thermospheric dynamics becomes more important during geomagnetically disturbed periods. In line with previous findings [e.g., Buonsanto and Witasse, 1999], the GAIM-FP winds are poleward during the day and equatorward at night. The obtained winds also display a clear dependence on the season. The pronounced solar heating asymmetry between hemispheres drives the circulation from the summer to the winter hemisphere [Roble et al., 1977]. Indeed, during solstices the GAIM-FP winds predominantly flow from summer to winter hemisphere. However, as it might be expected, during equinox the flow has a more symmetric pattern around the geographic equator. Note, transequatorial thermospheric winds are considered to be responsible for the observed asymmetries in F2-layer parameters about the magnetic equator [Abub-Robb and Windle, 1969; Anderson and Roble, 1981; Balan et al., 1995]. The winds obtained from GAIM-FP displayed a very clear longitudinal variation when global wind pattern at different UTs were compared. This might be attributed to the variation of magnetic field geometry, and specifically to the large changes in magnetic field declination angle. A similar conclusion was obtained by Luan and Solomon [2008] in their servo wind analysis.

The comparison of the GAIM-FP wind results to the FPI observations provided good agreement for the selected three stations during all seasons. However, the comparison over Millstone Hill during June solstice showed relatively larger discrepancies. The HWM93 model values, on the other hand, agree well with the Millstone Hill FPI winds during this period. Very good agreement between the two was also reported by Fejer et al. [2002]. They pointed out that the Millstone Hill solar minimum meridional winds derived from incoherent scatter radar measurements were used in the development of the empirical HWM93 (see also
Hedin et al. [1991]); however, an updated climatology of ISR-derived winds shows similar
differences as the GAIM-FP wind (see Figure 5 in Buonsanto and Witasse [1999]). This
discrepancy might partly be due to the differences in the Burnside factor used to derive the
wind from the ISR data, which however, cannot account for the full differences between the
winds. Emmert et al. [2003] noted similar large differences between the wind values obtained
from the empirical model, developed by them based on the FPI data, and the ISR-winds
of Buonsanto and Witasse [1999]. They pointed out the relative lack of nighttime ISR
data and the possibility of the year 1986 to be anomalous (2/3 of used ISR data were
from that year), and suggested a reexamination of the ISR wind data. Interestingly, their
results also showed that ISR winds reached their maximum 1 - 2 hours earlier than the
FPI winds for all seasons. This is similar to the GAIM-FP wind results (Figure 2.11).
The common feature in the ISR and GAIM-FP wind estimation is the use of the empirical
MSIS model; therefore, it could be possible the specified neutral compositions and their
temporal variations play a role for the observed similar discrepancies seen in both, GAIM-
FP and Emmert et al. [2003] results. GAIM-FP has the ability to perform the estimation
of the neutral composition together with the winds and drifts. This feature is currently
undergoing testing, and a separate study where neutral composition is also adjusted may
further improve the agreement between the modeled and measured winds.

One of the most widely employed method of obtaining thermospheric wind from iono-
spheric measurements has been the so-called servo model [Rishbeth, 1967; Rishbeth et al.,
1978; Buonsanto, 1986; Miller et al., 1986; Buonsanto et al., 1989; Luan and Solomon,
2008]. Therefore, it is useful to compare our new technique with the servo method and
point out some of the advantages of our method. The servo method employs numerous
assumptions that are not necessary to make in GAIM-FP. For example, the servo method
does not consider changes in the F2-layer shape in the wind estimation [Rishbeth et al.,
1978] and lacks the feedback between the wind and ionosphere, while the GAIM-FP winds
are self-consistently solved with the F-region ionospheric height distribution. The servo
wind models are mainly driven by hmF2 data, and their use would be difficult if there were
data gaps. In addition, in the servo methods the neighboring locations are independent from the analysis, which could provide vastly different results if the accuracy of the data is different. GAIM-FP, on the other hand combines information from both, NmF2 and hmF2 over a certain spatial extent that is specified by correlation lengths in the data assimilation technique, and therefore, provides realistic wind variation even for sparse and noisier data. However, even more importantly, GAIM-FP has a potential to assimilate different types of data [Schunk et al., 2004a] (e.g. TEC) that is expected to improve the modeling accuracy of ionospheric parameters and presumably the accuracy of the wind, as well.

Furthermore, the calculation of winds in servo models rely on the change of the hmF2 due to the vertical motion of F2 layer due to the winds and, therefore, are restricted to mid-latitudes. The GAIM-FP model, on the other hand can simultaneously estimate the motion due to the wind and due to $\mathbf{E} \times \mathbf{B}$ drifts, and hence, is also applicable over the equatorial and low latitudes. It should be noted GAIM-FP has capacity to perform estimation of mid-latitude drifts; however, this feature is under development and is not used in the present study. Therefore, similar to the servo model, the current GAIM-FP winds at mid-latitude are also effective or equivalent winds. The advantage of the servo model over GAIM-FP to estimate wind, however, is that it is fast and computationally least expensive.

2.4.3. Sensitivity to Uncertain Model Parameters

There are many uncertain parameters in physics-based ionospheric models, including reaction and photo-ionization rates, collision frequencies, diffusion and viscosity coefficients, thermal conductivities, etc.; and, in addition, there are errors associated with the missing physics and approximations used in the numerical schemes employed in the computations [Schunk et al., 2012]. In addition, when data assimilation is used, additional errors can arise due to the data itself. Obviously, it was impossible to study the effect of all possible parameters; and therefore, we focused on O$^+$/O collision frequency and neutral composition, since, as mentioned before, they are known to affect the various wind derivation techniques. In addition, we examined the influence of the number of occultations that affect the assimilated data and also studied the effects of errors in the assimilated data. Our study showed
that uncertainties in above mentioned parameters do not affect GAIM-FP ability to closely reproduce the assimilated NmF2 and hmF2, and these parameters were basically identical for the different simulation cases. Not surprisingly though, these agreements required different wind velocities.

To test the sensitivity of our wind results to the O\(^+\)-O collision frequency, we compared wind results for cases of the Burnside factor of 1.0 and 1.4 (though not presented, the difference between winds from F = 1.0 and F = 1.4 cases were similar to the difference between F = 1.4 and F = 1.8). The effect of a Burnside factor increase was to decrease the equatorward wind (increase poleward wind) during the night (day). Furthermore, the effects were larger at night and over the mid-latitudes (Figures 2.12 - 2.13 and 2.16). To understand the underlying causes, it is useful to look at the details of the wind estimation. In the F region, GAIM-FP estimates the horizontal wind by adjusting the wind-induced part of the O\(^+\) ion motion along the geomagnetic field lines (\(v_\parallel\)) given as

\[
\begin{align*}
  v_\parallel &= v_{mag} \cos I + v_d,
\end{align*}
\]

where \(v_d\) is the diffusion velocity of O\(^+\) ions along the magnetic field [Schunk and Walker, 1970; Burnside et al., 1983] and \(v_{mag}\) is magnetic meridional wind estimated by GAIM-FP (all velocities are positive towards north). At F-region heights, the diffusion velocity is downward and relatively small in magnitude during the day [Salah and Holt, 1974; Oliver, 1990]. The diffusion velocity is inversely proportional to ion-neutral collision frequency and hence, the increase in the Burnside factor decreases the diffusion velocity. At night the wind is equatorward, pushing ionization upwards. For the decreased downward \(v_d\), in order to have same \(v_\parallel\), it will be required to have weaker upward wind, thus reduced equatorward wind. During the day, the neutral wind is poleward, pushing the ionization down along the field lines. The reduced downward \(v_d\) will require increased poleward \(v_{mag}\) to maintain the same \(v_\parallel\) in Eq. (2.3). Since the diffusion velocity is small during the day, its change due to the Burnside factor is also small, and therefore, requires only relatively small changes in the horizontal neutral wind velocity to maintain balance. The effectiveness of
the neutral wind to affect the movement of ionization along the field lines is determined by
the geomagnetic field inclination, which means the described mechanism is more effective
over the mid-latitudes.

To study the effect of uncertainties in neutral composition, we reduced the MSISE-
90 exospheric temperature by 60 K globally in our simulation. According to the MSIS
model, this change in temperature gives \( \sim 31\% \) reduction in global average neutral mass
density at 400 km. This is similar to results obtained by the analysis of satellite drag data
during the recent solar minimum of 2007 - 2009 by Emmert et al. [2010] who revealed the
thermospheric densities at 400 km altitude were up to 30% lower than their climatologically
expected values, suggesting the thermosphere was cooler. In addition, Solomon et al. [2011]
performed a model simulation and obtained 40 K decrease in temperature compared to the
previous solar minimum. These results are comparable to our assumptions of a temperature
decrease and estimated density reductions.

The effect of a reduction in the exospheric temperature on the wind estimation was
opposite to the increase of the Burnside factor (see also Figure 2.16). This opposite effect can
be understood by the consequently reduced neutral composition at a given altitude, causing
a decrease in the ion-neutral collision frequency and an increase in the diffusion velocity.
Similar to our results, the larger effect (\( \sim 20 \text{ ms}^{-1} \)) of the \( \text{O}^+ - \text{O} \) collision cross section
change during the night than during the day was previously reported for servo-derived winds
by Titheridge [1993], and for ISR-derived winds by Oliver [1990]. The equivalence of the
Burnside factor decrease to the decrease of neutral composition was noted by Buonsanto
et al. [1989]. The effect of the temperature reduction appeared to have larger effect on
the GAIM-FP wind velocity over the winter hemispheres. Normally, the diffusion velocity
during summer and winter are not very different (see Figure 2 in Oliver [1990]). Since the
neutral temperature is smaller during the winter, the relative change in neutral composition
due to \( T_{ex} \) reduction is larger over the winter hemisphere, resulting in a larger absolute
change in the diffusion velocity.

The GAIM-FP winds showed relatively low sensitivity to the reduction in the number of
Figure 2.16. Local time variation of estimated magnetic meridional winds over Millstone Hill at 250 km altitude and during the December solstice conditions. The different colors and markers indicate the different parameters used in the model.

radio occultation. We tested the case when 1/3 of the COSMIC data was randomly excluded before it was binned and assimilated into GAIM-FP. To a certain degree, this affects both the values of the F2-layer parameters and their corresponding standard deviations; however, the wind values experience little change (7 ms\(^{-1}\)), (also in Figure 2.16). Although the main focus of this part of the investigation was to establish the reliability of the estimated winds, the results also have implications for later studies as they indicate we can take a 37-day interval instead of a 55-day interval (or 74 days in total instead of 110 days, since we combined two consecutive years) without compromising the accuracy of the derived magnetic meridional wind. It would be interesting for future studies to investigate the minimum number of occultations that is required to obtain a reliable wind estimation.

Separate model simulations to understand the effect of error increase by 30% also revealed that wind changes were very small (4 ms\(^{-1}\)), (Figures 2.15 and 2.16). This is important because the choice of the standard deviations as an estimate of the errors in the data does not consider the uncertainties due to Abel inversion for example. These details, however, seem to be less important for the wind estimation.

We noted that obtained GAIM-FP winds at mid-latitudes contain contribution from
vertical electromagnetic drifts and hence, they are effective winds. *Luan and Solomon* [2008] estimated the effect of $\mathbf{E} \times \mathbf{B}$ drift on the servo wind over Arecibo and found its contribution to be around 30 ms$^{-1}$. For GAIM-FP, Arecibo falls inside the region where drifts are computed. We estimated the possible error due to the drifts by excluding their effect using values from the mid-latitude drift model in Eq. (2.1) over Millstone Hill during December solstice, which on average is of the order of about 15 ms$^{-1}$ (see Figure 2.16, which also shows the effects of other model parameters).

Finally, we have tested the dependence of estimated wind values on the number of ensemble members in the Kalman filter. All results shown so far were obtained using 15 members in the ensemble Kalman filter. Additional model runs were done with 25 ensemble members and the corresponding winds were compared to the previous results (not presented). Results indicate that the difference is comparable to the sensitivity analysis when we changed the number of assimilated data or data errors, and of the order of about 8 ms$^{-1}$.

We showed the GAIM-FP wind climatology is reasonably close to the observations and provides a reliable representation of the global wind flow. The estimation of the wind using the data assimilation technique helps to achieve several important goals. First, it provides wind values that can be used in stand-alone ionospheric models to obtain accurate specifications of the ionospheric density. This is especially important over places where wind measurements are limited. Second, the obtained winds will help to better understand several ionospheric phenomena in which the neutral wind may play an important role. For instance, *Lomidze and Scherliess* [2011] reported on using the GAIM-FP results to study the mid-latitude ionospheric evening anomalies and the role of thermospheric winds in their generation. The wind results will also help to better understand thermospheric dynamics. Work is currently underway to use the estimated global magnetic meridional wind, along with other model parameters, to obtain the geographic zonal and meridional components of the neutral wind as reported by *Lomidze and Scherliess* [2013].

For future research it will be important to validate the low-latitude drifts obtained from
GAIM-FP. In addition, it will be interesting to perform wind comparisons on a monthly basis, and with the newer HWM07 model and with other newly available FPI observations. The later may require the GAIM-FP modeling for different solar flux conditions and therefore, will depend on the availability and quality of the COSMIC data for the later periods. Finally, the possibility to obtain neutral winds by assimilating other types of data, such as ground-based TEC, into GAIM-FP needs to be explored for climatological studies. Note, these data are already being assimilated in GAIM-FP in its day-to-day weather mode.

2.5. Summary and Conclusions

The global availability and reliable accuracy of thermospheric neutral winds are among the central challenges the space physics and aeronomy community currently faces. In this work, we presented a newly estimated climatology of low- and mid-latitude magnetic meridional winds. The results were obtained from the recently developed technique of ionospheric driver estimation from an ionospheric data assimilation model (GAIM-FP). For this, GAIM-FP assimilated quiet-time global seasonal maps of NmF2 and hmF2 obtained from the COSMIC GPS radio occultation measurements. Together with the global thermospheric neutral wind, the model provided equatorial and low-latitude vertical \( \mathbf{E} \times \mathbf{B} \) drifts, and significantly improved estimates of F2-layer peak parameters.

The obtained NmF2 and hmF2 were compared and contrasted against simulations performed with the ionosphere/plasmasphere model (IPM), which was driven by the empirical horizontal wind model (HWM93) and the equatorial vertical drift model (Scherliess-Fejer). GAIM-FP clearly outperformed IPM globally and provided significant improvements over certain data sparse locations (such as oceans for example). Under such scenarios we expect the GAIM-FP-derived winds and drifts are expected to be more accurate than their values from the corresponding empirical models. To substantiate this claim, the estimated thermospheric neutral winds were analyzed and shown to have expected global and seasonal pattern.

We compared the obtained wind results with their corresponding FPI measurements over Millstone Hill, Arecibo, and Arequipa for similar geophysical conditions during De-
cember and June solstices, and March equinox. The results showed the agreement is very good and since these three stations were not unique in the GAIM-FP simulation, a similar degree of accuracy is expected over other low- and mid-latitude locations, given the model parameters that are fixed during the simulation (e.g. neutral composition) are known with the same certainty.

We carried out the sensitivity analysis, studying the effects of changes in O$^+$-$O$ collision cross section and changes in neutral composition (through exospheric temperature change). We found that an increase in the Burnside factor and an increase of the temperature both reduce the equatorward wind at night and increase the poleward wind during the day. The effect is about 5 ms$^{-1}$ during the day for all latitudes and is similar over the low latitudes at night; however, the effect is larger (~22 ms$^{-1}$) over the mid-latitudes during the night. We also found that the exospheric temperature reduction in the simulation has a well-defined seasonal pattern for nighttime mid-latitude wind changes and shows larger effect on the estimated wind magnitude in the winter hemisphere. The effects produced by (a) reducing the number of occultations globally by randomly excluding 1/3 of observations (~50,000 radio occultation measurements) before generating global NmF2 and hmF2 maps, and (b) increasing errors for all assimilated NmF2 and hmF2 by 30% were shown to be relatively small and resulted in wind velocity changes of about 4-7 ms$^{-1}$, indicating the robustness of the method used. The effect of an electric field, which is currently not separated from the wind effects over mid-latitudes, was studied on the example of the wind over Millstone Hill during December solstice and was found to be the order of about 15 ms$^{-1}$.

The wind estimation using the ionospheric data assimilation model is a useful tool for obtaining accurate neutral wind velocities in the thermosphere. GAIM-FP can provide both, low- and mid-latitude magnetic meridional winds, and model used is not subject of as many simplifications, assumptions, and restrictions as previously proposed data-driven methods of wind estimation. The future investigations that can based on the results of current study, such as obtaining separate wind components and understanding neutral wind role in various observed ionospheric phenomena, can significantly contribute to our understanding.
of ionospheric and thermospheric dynamics.

References


Schunk, R. W., et al. (2004a), Global Assimilation of Ionospheric Measurements (GAIM), 


Thompson, D. C., L. Scherliess, J. J. Sojka, and R. W. Schunk (2006), The Utah State


CHAPTER 3
ESTIMATION OF THERMOSPHERIC ZONAL AND MERIDIONAL WINDS USING A KALMAN FILTER TECHNIQUE

Abstract

Thermospheric neutral winds play an important role in the dynamics of the middle- and low-latitude F-region ionosphere. Up to date their reliable estimation remains a challenge because of difficulties in both, measurement and modeling. The global sparseness and lack of continuous observations of global thermospheric winds often make empirical wind models fall short in adequately reproducing even their observed climatology. In addition, for understanding many ionospheric phenomena, it is very important to know both, zonal and meridional components of the wind. Previous methods that use ionospheric measurements to deduce winds give their values only in the direction of the magnetic meridian. We present a new method to estimate the climatology of the zonal and meridional components of thermospheric neutral wind at low and mid-latitudes using a Kalman filter technique combined with ionospheric observations. First, the climatology of the magnetic meridional wind is obtained by assimilating seasonally averaged COSMIC radio occultation data of F-region ionosphere peak parameters (NmF2 and hmF2) into the Global Assimilation of Ionospheric Measurements Full Physics (GAIM-FP) model. The model, which is based on Utah State University Ionosphere Plasmasphere Model (IPM), provides the 3-D electron density throughout the ionosphere, together with the magnetic meridional wind required to match the model density with the observed ionospheric parameters. Next, the estimation of the global zonal and meridional winds is performed using the magnetic meridional wind data obtained from GAIM-FP and a separate implicit Kalman filter, based on a physics-based 3-D thermospheric neutral wind model. The ionospheric drag and ion diffusion velocities, needed for the wind calculation, are also taken from the GAIM-FP results. As an output, the model provides the 3-D thermospheric neutral wind field at mid- and low latitudes. We present the results of our wind decomposition for three different seasons and compare
individual horizontal wind components to their corresponding empirical model values (from HWM93 model) and to measurements made by Fabry-Perot interferometers.

3.1. Introduction

The dynamics of the ionospheric F region are strongly controlled by thermospheric neutral winds [Emmert et al., 2006; Makela et al., 2013]. Winds influence the plasma motion along the geomagnetic field lines and are responsible for electron density variations at various temporal and spatial scales [Rishbeth, 1972; Titheridge, 1995]. As studies suggest, it is important for the proper interpretation of certain ionospheric features to have information about both, the zonal and the meridional components of the thermospheric wind separately. For example, the magnetic field declination angle effect, which determines the relative role of the thermospheric meridional and zonal wind components in the ionospheric dynamics, was suggested to be the possible cause of certain observed regional and global peculiarities in the ionospheric electron density [Kohl et al., 1969; Eccles et al., 1971; Jee et al., 2009; Zhang et al., 2011]. In addition, zonal winds play an important role in the generation of electric fields via the F-region dynamo mechanism [Rishbeth, 1971; Heelis, 2004; Heelis et al., 2012], which is responsible for the prereversal enhancement in the vertical ion drift and enhanced zonal ion drifts at night.

The only direct measurement technique to obtain both, zonal and meridional components of the thermospheric neutral wind, is the observation of the Doppler shift in the airglow emission lines by ground- or space-based interferometers [Hays and Roble, 1971; Burnside et al., 1981; Shepherd et al., 1993]. These observations, however, generally lack global coverage and/or continuity [Makela et al., 2012]. This is also an important reason why current empirical wind models often fall short in adequately reproducing even the observed wind climatology over data-sparse regions. On the other hand, theoretical global circulation models are known to capture the general characteristics of the thermospheric dynamics, but their accuracy in calculating wind velocities is hampered by uncertainties in model parameters and external driving forces. In these models, one especially important parameter for the thermospheric wind calculation is the ion drag that depends on the elec-
tron density, and discrepancies between modeled and measured wind velocities are often attributed to the lack of a realistic representation of ionospheric plasma density [Lei et al., 2007; Meriwether et al., 2013]. The ionosphere is a highly dynamic system and an accurate representation of the distribution and variation of its electron density is a separate challenge [Schunk et al., 2004].

In order to compensate for the global sparseness of direct wind observations, methods were developed to obtain information about the thermospheric neutral winds using ionospheric observations. These methods include the so-called servo models [Rishbeth et al., 1978], which use measurements of ionospheric F2-layer peak parameters, and the wind derivation from Incoherent Scatter Radar (ISR) measurements of plasma velocities along the magnetic field [Salah and Holt, 1974]. Chapter 2 reported about a new method to estimate thermospheric winds using an ionospheric data assimilation model. The method, which provides wind values over the entire low- and mid-latitude regions, was shown to be robust and capable to accurately represent wind data. However, the common shortcoming of all these techniques is that they obtain horizontal wind velocities only in the direction of magnetic meridian (or magnetic meridional wind).

Early attempts of modeling of thermospheric wind components were based on semi-empirical neutral wind models, which numerically solved only the equation of motion of the neutral atmosphere, albeit often not considering potentially important terms in the equation [Geisler, 1966; Kohl and King, 1967; Challinor, 1968]. In these models, the pressure gradient forces and frictional forces were calculated using empirical neutral atmosphere models (such as Jacchia and CIRA) and simplified models of the ionosphere electron density, respectively. Regardless of the limitations, the obtained wind speeds and flow patterns were reasonable, demonstrating the validity of the approach. An effort to combine experimental information with a similar wind model to obtain zonal and meridional wind components was reported by Roble et al. [1974]. They used ISR measurement of the magnetic meridional wind, and the longitudinal variation of the exospheric temperature to adjust the pressure gradient forcing using a least squares fit. The ion drag was specified in the model using
the measured electron density. Obviously, the application of this approach is limited to a handful of sites where ISRs are located and the estimated wind velocity is expected to be less accurate than direct interferometric observations because of the assumptions involved in the method.

The recent availability of global magnetic meridional wind and electron density data from GAIM-FP provides a new opportunity to use them together with a simple thermospheric wind model to extract information about the zonal and meridional components of the wind. For this purpose, the data and model needs to be combined in an optimal fashion. Generally, and also specifically for space physics applications, it has been shown that the most promising techniques that minimize the effect of model imperfections and at the same time take full advantage of the available data are physics-based data assimilation techniques [Schunk et al., 2004; Scherliess et al., 2009]. Therefore, we use this method to estimate thermospheric zonal and meridional winds.

In this Chapter, we present a newly developed technique for the estimation of thermospheric zonal and meridional wind climatology that was previously reported by Lomidze and Scherliess [2013]. The method combines magnetic meridional wind data from an ionospheric data assimilation model with the 3-D momentum equation of the horizontal motion of the neutral gas using the implicit Kalman filter. The outputs of the new model are the climatological neutral wind components at low and mid-latitudes in 110-600 km altitude range. The magnetic meridional wind data are obtained from the GAIM-FP model, which assimilates global seasonal maps of F2-layer maximum electron density (NmF2) and corresponding height (hmF2) obtained from COSMIC radio occultation (RO) data [Chapter 2]. The 3-D ionospheric electron density and ion diffusion velocity, which are needed in the wind model, are also obtained from GAIM-FP. The obtained wind velocities are in close agreement with measurements made by Fabry-Perot interferometers (FPI) over Millstone Hill, Arecibo, and Arequipa during December and June solstices, and March equinox. The wind results are also compared with corresponding values from HWM93 empirical Horizontal Wind Model [Hedin et al., 1996].
3.2. Methodology and Approach

The estimation of the zonal and meridional components of the thermospheric winds requires two separate model calculations. The first step is the assimilation of COSMIC seasonal NmF2 and hmF2 maps into the GAIM-FP model to specify the electron density distribution and to obtain the global magnetic meridional wind climatology. The next step is to assimilate the GAIM-FP magnetic meridional wind data (together with the use of additionally needed ionospheric parameters also obtained from GAIM-FP) into a new physics-based Thermospheric Wind Assimilation Model (TWAM) to decompose the wind into its geographic zonal and meridional components. Since our purpose is to obtain the climatology of the neutral winds and not to capture the various transient weather phenomena (i.e., traveling atmospheric disturbances), we choose to use the simple, yet effective wind model together with a data assimilation technique, which helps to minimize the model errors in a dynamically consistent way. In this section, we briefly describe the COSMIC data and the GAIM-FP model and then provide a more detailed description of the wind model and the implicit Kalman filter that is employed in TWAM.

3.2.1. COSMIC NmF2 and hmF2 Data

The values of NmF2 and hmF2, and their corresponding standard deviations, which are later assimilated into GAIM-FP, are obtained from GPS radio occultation measurements from the Constellation Observing System for Meteorology Ionosphere and Climate/Formosa Satellite 3 (COSMIC/FORMOSAT 3). Launched in April 2006, the constellation consists of six micro-satellites in a circular orbit at about 800 km altitude with a 72° inclination and 30° separation in longitude. The COSMIC satellites have been providing about 1,000 - 2,500 electron density profiles per day globally [Rocken et al., 2000; Anthes et al., 2008], which are retrieved from slant total electron content measurements using an Abel inversion technique [Hajj and Romans, 1999; Schreiner et al., 1999]. The data of electron density profiles were obtained from the COSMIC Data Analysis and Archive Center (http://www.cosmic.ucar.edu).

To prepare the COSMIC data for the assimilation into GAIM-FP, the electron density
profiles were obtained, and an initial quality check was performed. Next, the values of NmF2 and hmF2 were extracted from individual electron density profiles and to obtain the seasonal global maps the data were combined from 2007 to 2009 and binned into seasons consisting of December solstice (December 21 ± 27 days for the years 2007 and 2008), March equinox (March 20 ± 27 days for 2008 and 2009), and June solstice (June 21 ± 27 days for 2008 and 2009). To obtain quiet-time data, points were excluded for observations with a previous 24-hour average $Ap \geq 16$ and/or a corresponding 3-hour $ap \geq 16$. The total number of final points selected were \( \sim 161,600 \) for December, \( \sim 143,900 \) for March, and \( \sim 144,200 \) for June, respectively. The average solar and geophysical conditions for the selected periods were $F_{10.7} = 73$, $Ap = 5$ for December, $F_{10.7} = 70$, $Ap = 5$ for March, and $F_{10.7} = 67$, $Ap = 4$ for June, respectively. The days that best represent the climatological average conditions of the selected seasons, based on their average solar declination, are DOY 006 for December solstice, DOY 079 for March equinox, and DOY 189 for June solstice.

The retrieved NmF2 and hmF2 were used to produce average global maps of these parameters in 30-minute intervals of universal time (UT). For the binning we used a longitude \( \times \) latitude grid of 7.5° \( \times \) 2.5° with a 15° \( \times \) 5° sliding window. To further exclude outliers and potentially erroneous data, the procedure was performed a second time and those data on a given grid-point that deviated by more than 2σ from the mean were excluded. We also excluded data for which the averages were based on less than three observations for a given grid-point. The obtained standard deviations were used as an estimate of the data errors, which are important inputs for the data assimilation models. For more details about the COSMIC data see Chapter 2.

### 3.2.2. Ionospheric Data Assimilation Model

The Global Assimilation of Ionospheric Measurements Full Physics (GAIM-FP) model is a data assimilation model that uses a physics-based ionosphere-plasmasphere model (IPM) and an ensemble Kalman filter technique to incorporate the main physical processes in the ionosphere directly into the data assimilation scheme [Schunk et al., 2004; Scherliess et al., 2004, 2009]. The model can assimilate different types of data from different sources and
provide specifications of the 3-D electron density distributions from 90 km to 30,000 km at low and mid-latitudes. The model can also provide an estimation of the ionospheric drivers (electric field, magnetic meridional wind, and neutral composition) that is consistent with the data and the model. In its current version, the data assimilation model excludes regions poleward of $\sim \pm 60^\circ$ geomagnetic latitudes.

In the IPM model, the ion and electron continuity and momentum equations are numerically solved along magnetic field lines for individual convecting flux tubes of plasma, and the 3-D nature of the model is obtained by following a large number of plasma flux tubes [Schunk et al., 2003]. The 3-D distribution is obtained by mapping the results on a geographic grid. IPM uses the International Geomagnetic Reference Field (IGRF), which properly accounts for the displacement between the geomagnetic and geographic equators and the bending of the magnetic field lines with latitude.

The Kalman filter [Kalman, 1960] combines the data from an observing system with the information obtained from the system model and their corresponding statistical description of uncertainties. It performs a recursive least squares inversion of the observations (NmF2 and hmF2 for our case) for the model variables using a dynamical model (IPM) as a constraint. As a result, an improved estimate of the model variables (electron density, magnetic meridional wind) is obtained in a statistical sense. The estimation of the ionospheric drivers is an integral part of our ensemble Kalman filter and is achieved by using the internal physics-based model sensitivities to the various driving forces. In this procedure, the ionospheric data are used to adjust the plasma densities and its drivers so that a consistency between the observations (within their errors) and the physical model is achieved. As a result, the assimilation procedure produces the optimal model data combination of the ionosphere-plasmasphere system with its consistent drivers (for details see Scherliess et al. [2009, 2011]).

In the climatology mode of GAIM-FP, which was used for this study, global seasonal maps of NmF2 and hmF2 were assimilated and the 3-D electron density, together with the low- and mid-latitude magnetic meridional wind and the low-latitude vertical $\mathbf{E} \times \mathbf{B}$ drifts,
were obtained. The ion diffusion velocity, which is needed in our separate TWAM model, was also obtained from GAIM-FP. The global neutral composition was specified by MSISE-90 [Hedin, 1991]. The details about the sensitivity of the estimated magnetic meridional winds to different uncertain model parameters were discussed in Chapter 2.

3.2.3. Physics-Based 3-D Thermospheric Wind Assimilation Model (TWAM)

The Thermospheric Wind Assimilation Model (TWAM) consists of a physics-based numerical model and an implicit Kalman filter, which are described in the following subsections. Ideally, in order to have a self-consistent picture, in addition to the momentum equation one needs to the solve continuity, and energy equations of the neutral gas with the complete set of similar equations for the electrons and ions. Different from that approach, our aim here is to use a Kalman filter technique to combine magnetic meridional wind data from GAIM-FP with the equation of motion of the neutral gas to obtain the zonal and meridional components of the thermospheric neutral wind.

Thermospheric Neutral Wind Model

The neutral dynamics in the thermosphere is described by the equation of the motion of the neutral air

\[
\frac{\partial \mathbf{u}}{\partial t} + (\mathbf{u} \cdot \nabla) \mathbf{u} = -2\Omega \times \mathbf{u} + \mathbf{g} - \frac{1}{\rho} \nabla p + \frac{\mu}{\rho} \nabla^2 \mathbf{u} - \nu_{ni} (\mathbf{u} - \mathbf{u}_i),
\]

where

\[
\mathbf{u}_i = \frac{\kappa^2}{1 + \kappa^2} \mathbf{u} + \frac{\kappa}{1 + \kappa^2} \mathbf{u} \times \mathbf{b} + \frac{1}{1 + \kappa^2} \left( (\mathbf{u} \cdot \mathbf{b}) \mathbf{b} + \mathbf{u}_{\text{diff}} \right),
\]

is an ion velocity induced by neutral wind, electric field, and ambipolar diffusion and \( \mathbf{u} \) is a thermospheric neutral wind velocity vector. Subscripts \( \mathbf{E} \times \mathbf{B} \) and \( \text{diff} \) denote electromagnetic and diffusion parts of the ion velocity, \( \mathbf{b} \) is a unit vector in the direction of geomagnetic field. \( \kappa = \nu_{ni}/\omega_i \) is a ratio of the ion-neutral collision frequency to the ion gyrofrequency. Note, at ionospheric F-region heights (~ 250 km) and above \( \omega_i \gg \nu_{ni} \), and hence, the first two terms in the ion velocity expression become small. However, these terms are important
between E- (～100 km) and F-region altitudes. Also note, the ion velocity component along the electric field, \( \kappa^2 / (1 + \kappa^2) \cdot e / (mv_{in}) E \), which is more important in the E region, was ignored from the ion velocity equation. The ion drift velocity, \( u_{iE\times B} \) was obtained from Scherliess-Fejer empirical drift models [Scherliess and Fejer, 1999; Scherliess et al., 2001].

The expression for \( u_{i\text{diff}} \) was given by Schunk and Walker [1970] and its value in our wind model is computed from the GAIM-FP results of the \( O^+ \) diffusion velocity along the Earth’s magnetic field. In the wind equation \( \Omega \approx 7.27 \cdot 10^{-5} s^{-1} \) is the angular frequency of the Earth’s rotation, and correspondingly the first term on the right side is the acceleration due to the Coriolis force. \( g \) is the acceleration due to gravity, \( \rho \) is the neutral mass density, \( \mu = \sum \mu_k n_k / \sum n_k \) is the dynamic viscosity coefficient for gaseous mixtures of \( O_2 \), \( N_2 \), and \( O \) with \( n_k \) being the corresponding number densities and \( \mu(O_2) = 4.03 \cdot 10^{-7} T^{0.69} \) kg m\(^{-1}\) s\(^{-1}\), \( \mu(N_2) = 3.43 \cdot 10^{-7} T^{0.69} \) kg m\(^{-1}\) s\(^{-1}\), and \( \mu(O) = 3.90 \cdot 10^{-7} T^{0.69} \) kg m\(^{-1}\) s\(^{-1}\) [Banks and Kockarts, 1973]. \( T \) is the neutral temperature and, together with the neutral number densities and mass density, is taken from the MSISE-90 empirical model. \( p \) is the atmospheric gas pressure (see details later), \( \nu_{ni} = \sum_j m_j n_j \sum_k \nu_{jk} / m_k n_k \) is the neutral-ion collision frequency where \( j \) is \( NO^+ \), \( O_2^+ \) or \( O^+ \), and \( k \) is \( O_2 \), \( N_2 \), or \( O \). The ion densities are provided by GAIM-FP and the \( \nu_{jk} \), ion-neutral collision frequencies are taken from Schunk and Walker [1973].

The pressure gradient force, \( -\nabla p / \rho \), is one of the most important driving forces for the thermospheric dynamics, and it is essential to have its proper estimate. One could have calculated this parameter directly from an empirical neutral atmosphere model as it was done in older studies. However, empirical models are expected to be less reliable for density and temperature gradients than for density and temperature itself [Blum and Harris, 1975a]. To avoid a strong bias introduced by the empirical model, we adopted a parametrized form of this force and used it as a state parameter in the Kalman filter, meaning it is estimated in the data assimilation process. Geisler [1966] showed that a convenient way of representing the pressure distribution is to use the altitude variation of an isobaric surface with exospheric temperature. Let \( x \), \( y \), and \( z \) be the local Cartesian
coordinates at some point in the atmosphere where \( x \) is toward the east, \( y \) is toward the north, and \( z \) is toward the zenith. The zonal and meridional components of the pressure gradient force are

\[
F_x = -\frac{1}{\rho} \frac{\partial p}{\partial x}, \quad F_y = -\frac{1}{\rho} \frac{\partial p}{\partial y}.
\]  

(3.2)

The relation between pressure gradient, let’s say in the zonal direction, and the slope of the isobaric surface can be found by differentiation along the surface

\[
\left( \frac{\partial p}{\partial x} \right)_p = 0 = \frac{\partial p}{\partial x} + \frac{\partial p}{\partial z} \left( \frac{\partial z}{\partial x} \right)_p.
\]  

(3.3)

Now, if \( h \) is an altitude of a point on an isobaric surface then

\[
\left( \frac{\partial z}{\partial x} \right)_p \equiv \frac{\partial h}{\partial x} = \frac{\partial h}{\partial T} \frac{\partial T}{\partial x},
\]  

(3.4)

and finally assuming hydrostatic equilibrium, the zonal and meridional components of pressure gradient force in spherical coordinates [Challinor, 1968] take the form

\[
F_\phi = -g \frac{\partial h}{\partial T} \frac{1}{r \cos \theta} \frac{\partial T}{\partial \phi}, \quad F_\theta = -g \frac{\partial h}{\partial T} \frac{1}{r} \frac{\partial T}{\partial \theta},
\]  

(3.5)

where \( r \) is the distance from the Earth’s center, and \( \theta \) and \( \phi \) are geographic latitude and longitude, respectively. For atmospheric conditions where hydrostatic equilibrium applies, a condition, for example, assumed in MSISE-90, Geisler [1966] demonstrated that \( \partial h/\partial T \) is very nearly a linear function of altitude and increases with increasing height. It is close to zero at the bottom of the thermosphere and we take it to be zero at 110 km to make it consistent with our lower boundary conditions that are discussed below. The assumed altitude variation means the height dependence of the pressure gradient force can be characterized by only one parameter, its slope, for a given exospheric temperature, \( T_{ex} \). A numerical analysis, based on MSISE-90, of the slope of \( \partial h/\partial T \) on a global scale shows that to a good approximation it is inversely proportional to \( T_{ex} \) and can be approximately described by \( a/T_{ex} + b \) (for a similar analysis see Blum and Harris [1975a]). Using again MSISE-90 we
found that the variation of parameter $a$ with latitude and longitude is negligible, and that parameter $b$ shows only an $\sim 15\%$ variability on a global scale. However, when compared to the variations of the slope caused by the $T_{ex}$ dependence, we can, to a first approximation, consider both $a$ and $b$ constant over the globe and use them as state variables in our estimation.

The above analysis shows that, at a given altitude the horizontal variation of the pressure gradient forces are only functions of the exospheric temperature. In the analysis of diurnal variation of exospheric temperature, Jacchia [1977] showed that its global distribution can be remarkably well represented by an equation of the following form:

$$T_{ex} = \bar{T}_{ex} (1 + c_1 \delta_{\odot} \sin \theta / \epsilon + c_2 \cos \theta [f(H) - 0.5])$$ (3.6)

with the shape of the diurnal variation, $f(H)$, given by

$$f(H) = \cos^n \frac{1}{2} (H + \beta_0) + c_3 \cos [3(H + \beta_0) + \chi],$$ (3.7)

where $\bar{T}_{ex}$ is the mean of the daytime maximum and nighttime minimum of exospheric temperature, $\delta_{\odot}$ is the declination of the sun, $\epsilon$ is the obliquity of the ecliptic ($23.44^\circ$), $H$ is the hour angle of the sun (the local solar time) and can be related to the longitude ($\varphi$) and universal time ($t$) ($H = \varphi + \Omega t - \pi$), the constant, $\beta_0$, determines the lag of the temperature maximum with respect to solar culmination, $c_{i=1,2,3}$ and $\chi$ are empirical constants, and the exponent, $n$, can vary between 2 and 3.

We use the above form of the temperature distribution in our scheme to calculate horizontal derivatives needed for the pressure gradient force parametrization. In order to simplify the analytical expressions and to have linear expressions for parameters that need to be estimated, we assume $n = 2$. Then, for any instant in time at a given altitude, $z$, the horizontal variation of the pressure gradient forces can be expressed as

$$F_{\varphi} = A_1(z) \sin(\varphi + \Omega t + \beta_1) + A_2(z) \sin(3[\varphi + \Omega t] + \beta_2)$$ (3.8)
\[ F_\theta = A_1(z) \cos(\varphi + \Omega t + \beta_1) \sin \theta + \frac{1}{3} A_2(z) \cos(3[\varphi + \Omega t] + \beta_2) \sin \theta + A_3(z) \cos \theta, \quad (3.9) \]

where \( \beta_{i=1,2} \) are free parameters that are also estimated together with \( a'_{i=1,2,3} \) and \( b'_{i=1,2,3} \) in the expression of \( A_i(z) = \left( a'_i / T_{ex} + b'_i \right) (z - 110 \text{ km}) / r \) describing the altitude variation of \( \partial h / \partial T \) as discussed above. In the expression of \( A_i(z) \), we take \( r = R_E \), where \( R_E \) is the Earth’s mean radius, and obtain values of \( T_{ex} \) from MSISE-90.

One of the main inputs in the wind model is the magnetic meridional wind data from GAIM-FP. The magnetic meridional wind (\( v_{mag} \), positive towards magnetic north), by definition, consists of contributions from geographic zonal (\( u \), positive eastward) and meridional (\( v \), positive northward) winds coupled by the magnetic field declination angle (\( D \)):

\[ v_{mag} = u \cdot \sin D + v \cdot \cos D. \quad (3.10) \]

The value of \( v_{mag} \) is estimated by GAIM-FP through adjusting the wind induced part of \( O^+ \) ion motion along the geomagnetic field lines. Therefore, strictly speaking, if a vertical wind is present, \( w \cdot \tan(I) \) (\( I \) is a magnetic field inclination angle, \( w \) is vertical neutral wind) needs to be added to the magnetic meridional winds that are obtained from GAIM-FP to make it consistent with Eq. (3.10). However, during geomagnetically quiet times, the amplitude of the vertical wind velocity generally is an order of magnitude smaller than that of horizontal velocities \([Rishbeth et al., 1969; Roble et al., 1974]\); and therefore, this term is not considered here.

In our dynamic model, we solve numerically for horizontal wind components first, and the vertical component of the wind is later calculated, retrospectively. As was noted, generally, the vertical wind is known to be small and can be omitted from the equations when estimating the horizontal wind velocities. The estimation of vertical winds can be performed once the horizontal winds are calculated using a method described by \( Dickinson \ and \ Geisler \ [1968] \) and \( Rishbeth \ et \ al. \ [1969] \). The vertical wind velocity (\( w = w_B + w_D \)) has two components, a barometric component (\( w_B \)) that is due to the thermal expansion and contraction of the atmosphere and a divergence component (\( w_D \)) that balances the
divergence of air produced by horizontal winds. For their calculation, one needs to use the continuity equation where the air density is assumed to be known, and which we take again from MSISE-90. The equations for the vertical wind components to be solved are

\[
\frac{\partial (nw_B)}{\partial z} = -\frac{\partial n}{\partial t}
\]  

(3.11)

\[
\frac{\partial}{\partial z} (nw_D) = -\frac{\partial}{\partial x} (nu) - \frac{\partial}{\partial y} (nv),
\]  

(3.12)

where \(n\) is number density of atmospheric air.

In order to solve the system of momentum equations for the horizontal neutral winds numerically and combine it with the data through the Kalman filter, we adopt a standard finite difference approximation of a fully implicit scheme that, in the case of direct solution, is unconditionally stable and can be solved using a Gaussian elimination of a resulting tridiagonal matrix. In order to simplify the numerical solution of Eq. (3.1), and also to make classical Kalman filter technique applicable, we make a number of approximations in the equation of motion. Namely, we ignore nonlinear advection terms and viscous terms associated with horizontal velocity gradients. The possible effects of these terms are discussed below. The appropriate coordinate system for our problem is a spherical coordinate system. In the complete form of the equation for zonal and meridional components, we adopt the form consistent with hydrostatic primitive equations [White and Bromley, 1995], and ignore so-called metric terms, which are also nonlinear and reflect the curvature of the spherical coordinate system. The required magnetic field inclination and declination angles in zonal and meridional projections of the drag term are taken from IGRF. For the lower boundary at 110 km we take Dirichlet boundary condition \(u = 0, v = 0, w = 0\), and for the upper boundary at 590 km Neumann boundary conditions \(\partial u/\partial z = 0, \partial v/\partial z = 0, \partial w/\partial z = 0\) are taken, since the viscous force increases with altitude and eventually suppresses variations of the wind speed with altitude. The solution is obtained every 15 minutes on a spatial latitude \(\times\) longitude \(\times\) altitude grid with size \(7.5^\circ \times 15^\circ \times 15\) km, which is later interpolated to a \(2^\circ \times 7.5^\circ \times 10\) km resolution to make it consistent with the standard GAIM-FP output.
Implicit Kalman Filter

Since the introduction of the Kalman filter [Kalman, 1960] it has become a well-documented method with many widely used variations in different fields of science, including atmospheric and ionospheric studies [Howe et al., 1998; Scherliess et al., 2004; Lee et al., 2012]. The traditional Kalman filter is a linear, recursive algorithm that combines the observation with a dynamic model describing the processes and corresponding uncertainties. The model uses the errors associated with the data and the model to provide improved model state variables in a statistical sense. Different from the traditional Kalman filter, the implicit Kalman filter was developed to be suitable for implicit discrete systems [Skliar and Ramirez, 1997], and thus, it is applicable for problems that require implicit schemes for stable numerical solutions. Basically, the implicit Kalman filter is equivalent to the traditional Kalman filter and a brief description, mainly based on Skliar and Ramirez [1997], is as follows.

Assume the implicit form of a dynamic system (Eq. (3.1) for our case) is described by

$$
M_1 (k + 1) x (k + 1) = M_2 (k + 1) x (k) + U (k + 1) + q (k + 1), \quad (3.13)
$$

where $k$ is a time index, $x$ is the state vector (in our case consisting of wind velocities at each 3-D gridpoint and coefficients in the pressure gradient force parametrization), $M_1$ and $M_2$ are known matrices describing the dynamics of the system (derived from the discretization of Eq. (3.1)), $U$ is a matrix describing the forcing of the system (in our problem drag terms due to $\mathbf{E} \times \mathbf{B}$ drift and ion diffusion velocity), and $q$ is a vector representing the process noise with covariance matrix $Q$. [Note that different from Skliar and Ramirez [1997], Eq. (3.13) describes a fully implicit scheme]. Also assume the equation relating the measurements to the state variables has the form [equivalent of Eq. (3.10)]

$$
z (k + 1) = H (k + 1) x (k + 1) + r (k + 1), \quad (3.14)
$$

where $z$ is a measurement vector (e.g., $v_{mag}$ at the altitude of hmF2 over the entire low- and
mid-latitude domain), $\mathbf{H}$ is the known measurement matrix relating the observations to the state vectors, and $\mathbf{r}$ is a vector of measurement error (in our case estimated errors of $v_{\text{mag}}$) with corresponding covariance matrix $\mathbf{R}$. Let us define an auxiliary variable $\mathbf{y}(k+1) = \mathbf{M}_1(k+1)\mathbf{x}(k+1)$ and a new $\mathbf{H}_1$ matrix through $\mathbf{H}_1\mathbf{M}_1(k+1) = \mathbf{H}$. Then (3.13)-(3.14) can be written as

$$
\mathbf{y}(k+1) \equiv \mathbf{y}^f = \mathbf{M}_2(k+1)\mathbf{x}(k) + \mathbf{U}(k+1) + \mathbf{q}(k+1)
$$

(3.15)

$$
\mathbf{z}(k+1) = \mathbf{H}_1(k+1)\mathbf{y}(k+1) + \mathbf{r}(k+1).
$$

(3.16)

The system (3.15)-(3.16) has the form that is commonly used in the classical Kalman filter. The forecasted model error covariance matrix of $\mathbf{y}(k+1)$ will be

$$
\mathbf{P}^f_y = \mathbf{M}_2\mathbf{P}_x\mathbf{M}_2^T + \mathbf{Q},
$$

where $\mathbf{P}_x$ is the model error covariance matrix of $\mathbf{x}(k)$. The so-called Kalman gain takes the form

$$
\mathbf{K}_y = \mathbf{P}^f_y\mathbf{H}_1^T\left(\mathbf{R} + \mathbf{H}_1\mathbf{P}^f_y\mathbf{H}_1^T\right)^{-1},
$$

and finally, the estimated model state vector ($\mathbf{x}^a$) and the corresponding error covariance matrix ($\mathbf{P}^a_x$) at the $k+1$ timestep will become

$$
\mathbf{x}^a = \mathbf{M}_1^{-1}\left[\mathbf{y}^f + \mathbf{K}_y\left(\mathbf{z} - \mathbf{H}_1\mathbf{y}^f\right)\right],
$$

$$
\mathbf{P}^a_x = (\mathbf{M}_1)^{-1}\left[\left(\mathbf{I} - \mathbf{K}_y\mathbf{H}_1\right)\mathbf{P}^f_y\mathbf{H}^T_1\right]\left(\mathbf{M}_1^T\right)^{-1},
$$

where $\mathbf{I}$ is an identity matrix. As time evolves, the steps described above are repeated to give the best estimates of state variables in a least-squares sense.

3.3. Results of Wind Estimation

In this section we present the obtained wind climatology (TWAM results) at low and mid-latitudes, and compare the zonal and meridional components with their correspond-
ing measurements from ground-based FPI instruments at three locations [Millstone Hill (42.6°N, 71.5°W), Arecibo (18.4°N, 66.8°W), and Arequipa (16.5°S, 71.5°W)], as well as with values from the empirical HWM93 model. The first step of the estimation process involves the assimilation of seasonal NmF2 and hmF2 maps from COSMIC into GAIM-FP, which estimates the global magnetic meridional wind and specifies the global 3-D electron density. The detailed results of this step were shown in Chapter 2 and will be not repeated here. In the second step the magnetic meridional wind data from GAIM-FP are combined with our physics-based Kalman filter for the neutral wind to obtain the zonal and meridional components of the wind. Finally, the vertical wind is calculated using the obtained horizontal wind velocities. All results that will be presented here correspond to quiet-time, solar minimum conditions, specified by the average geophysical conditions of the time intervals during which the COSMIC data were averaged.

The estimated meridional and zonal components of the thermospheric neutral wind at 250 km and UT=12:00 h are shown in Figure 3.1 for December solstice, March equinox, and June solstice conditions. The meridional wind is generally poleward during the day and equatorward at night. The results show a predominantly summer to winter hemisphere flow during the solstices and a symmetric flow about the geographic equator during equinox. These flow patterns are in good agreement with our current understanding of thermospheric dynamics. The zonal wind is mostly westward during noon and becomes eastward after dusk. At mid-latitudes the wind becomes westward after dawn, but at low latitudes remains eastward until noon. A seasonal pattern is also apparent in the zonal flow. The strongest westward wind occurs in the summer hemisphere and the strongest eastward wind in the winter hemisphere, with a net eastward flow in winter and a net westward flow in summer. The winds are similar in the two hemispheres during equinox. In general, the described patterns remain unchanged for other UTs; however, certain variations in both components of the winds were noted (not shown) indicating longitudinal variations in the wind. These variations could be associated with the longitudinal differences in the geomagnetic field geometry, as well as longitudinal changes in ion drag, neutral composition, and high-latitude
Figure 3.1. Meridional (left) and zonal (right) winds at 250 km altitude over low and mid-latitudes obtained from TWAM. Values are shown for December solstice (top), March equinox (middle) and June solstice (bottom), and for UT=12:00 h. Contour intervals correspond to 15 ms$^{-1}$. Positive values indicate northward flow for the meridional wind and eastward flow for the zonal wind.
forcing, which are expected to be embedded in our various input data.

Figure 3.2 and 3.3 show the local time variations of the estimated meridional and zonal winds, and their comparisons with those obtained from FPI measurements and HWM93 values over Millstone Hill, Arecibo, and Arequipa during the three seasons. All FPI wind data used in this study, which include data from previous solar minima, were obtained from the Madrigal database (http://www.openmadrigal.org/) and seasonal averages were calculated for similar time periods and geophysical conditions as in the case of the COSMIC data. Note that the FPI wind data are the same as those previously used in Chapter 2. The error bars correspond to one standard deviation and indicate quiet-time geophysical variability. Both the estimated nighttime meridional and zonal winds are in good agreement with the FPI observations (within 1-σ errors) for most of the time. The largest discrepancies are observed over Millstone Hill during June solstice for the meridional wind and during December solstice for the zonal wind. The estimated winds are also in good agreement with the wind velocities obtained from the HWM93 model throughout the day. However, larger differences can be seen at certain local times, for example over Arecibo, in the December solstice meridional wind during evening hours and in the June solstice zonal wind during noon. Note, the comparisons of the magnetic meridional winds performed in our previous study (Chapter 2) showed a similar agreement as shown here for the geographic meridional wind. Overall, Figures 3.2 and 3.3 show that our new model produces zonal and meridional winds that are in very good agreement with the established wind climatology over these stations and during the night are almost always within the one sigma error when compared to the climatology obtained by direct wind measurements.

Although data from global thermospheric zonal and meridional wind measurements are limited, it is interesting to globally compare the diurnal variations of the estimated winds with those from HWM93. These comparisons are shown in Figure 3.4 and 3.5 for the meridional and zonal winds, respectively. Shown are longitudinally averaged (in the geographic zonal direction) winds as a function of local time and geographic latitude. The general characteristics of the meridional winds from TWAM and from HWM93 are very
Figure 3.2. Local time variations of seasonal geographic meridional winds at 250 km altitude over the Millstone Hill (left), Arecibo (middle), and Arequipa (right), and during December solstice (top), March equinox (middle), and June solstice (bottom) conditions. Circles indicate the seasonally averaged FPI observations in 15-minute intervals with their corresponding 1-σ variability. Also shown are the estimated winds from TWAM (blue) and HWM93 (magenta). Positive values indicate an equatorward flow.
Figure 3.3. Same as Figure 3.2, but for the zonal wind. Positive values indicate an eastward flow.
similar in terms of their latitudinal, local time, and seasonal variations. Both models show larger velocities in the upper mid-latitudes and similar changes in net inter-hemispheric flows from season to season. The velocity magnitudes are also comparable and exhibit predominantly diurnal variations. The amplitude of the wind variation is smaller over low latitudes. Large differences between the two models are found in the southern mid-latitudes, where the strongest equatorward wind in HWM93 occurs a few hours later compared to the winds from our TWAM model. Important differences are also observed in the wind reversal times. As for the zonal winds shown in Figure 3.5, the main features of TWAM and HWM93 are also very similar; however, the HWM93 winds exhibit stronger semi-diurnal variations at all latitudes. In addition, HWM93 displays larger amplitudes in the low-latitude zonal wind variations between about noon to midnight when compared to our TWAM model winds. The zonal wind discrepancies between the two models are similar in the two hemispheres over the mid-latitudes where the winds also differ in the reversal times. Overall, the TWAM longitudinally averaged meridional and zonal wind components are less structured than those from HWM93, but the major patterns of the two models are similar for both wind components.

As was mentioned, the solution of the wind equation provides the 3-D nature of the zonal and meridional winds, and the obtained horizontal winds can be used to estimate the vertical wind velocity. An example of the altitude variation of all three components of the neutral wind over Millstone Hill is shown in Figure 3.6 for March equinox conditions. It can be seen that the meridional and zonal winds are almost constant in altitude above ~250 km. The altitude gradients are stronger in the E and bottomside F region. However, because of several reasons discussed below, these height variations may have large uncertainties. Note, it is also difficult to validate the altitude profiles of the winds because of very limited observations of wind profiles [e.g., Drob et al., 2008]. The obtained vertical wind, as expected, is very small in magnitude (~2 - 4 ms⁻¹), and generally upward during the day and downward at night. In addition, the magnitude of the vertical velocity shows a small increase with altitude.
Figure 3.4. Longitudinally averaged meridional wind as a function of local time and geographic latitude. Results are shown for TWAM (top) and HWM93 (bottom) during December solstice (left), March equinox (middle), and June solstice (right) conditions. Contour intervals correspond to 15 m s$^{-1}$. Positive values indicate northward flow.

Figure 3.5. Same as Figure 3.4, but for zonal wind. Positive values indicate eastward flow.
Figure 3.6. Local time variations of height profiles of meridional (top), zonal (middle), and vertical (bottom) winds during March equinox over Millstone Hill obtained from the TWAM model. Contour intervals for meridional, zonal and vertical wind correspond to 15 ms$^{-1}$, 18 ms$^{-1}$, and 0.4 ms$^{-1}$, respectively.
3.4. Discussion and Summary

Our ability to accurately specify the thermospheric wind system is limited because of sparse wind observations and shortcomings of global wind models. The methods of wind derivations from ionospheric observations have a long history, but they only provide the net effective wind along the magnetic meridional direction and do not provide direct information about the separate wind components, which are often very important. In this paper we presented a novel technique to combine information from ionospheric data assimilation model with the equation of motion of the neutral gas in the thermosphere using a Kalman filter. As an output, this new model provides the climatology of the geographic meridional and zonal components (also vertical) of the thermospheric neutral wind.

The obtained seasonal and diurnal patterns of both, global meridional and zonal winds are in good agreement with experimental and theoretical results of previous studies [Rishbeth, 1972; Roble et al., 1977; Titheridge, 1995; Rishbeth and Müller-Wodarg, 1999; Emmert, 2001; Emmert et al., 2003]. According to the current understanding of quiet-time thermospheric dynamics, air motion is set up by pressure-gradient forces, which direct winds from the hottest part of the thermosphere towards the coldest part, under the additional influence of ion drag and Coriolis forces. For example, it is well known that the prevailing summer-to-winter circulation during solstices (see Figure 3.1 and Figure 3.4) is the result of asymmetric solar heating of the two hemispheres [Roble et al., 1977]. The prevailing west-to-east flow in winter and east-to-west in summer (Figure 3.1 and Figure 3.5) might be understood by the deflecting action of Coriolis forces on summer-to-winter flow. Note, the parametrization of the pressure gradient force used in our wind model does not a priori specify the location of maxima of this driving force (in other words, season is not specified). Instead, the appropriate distribution is estimated by the data assimilation procedure inside the Kalman filter. On the other hand, the proper seasonal distribution of ion drag and seasonal pattern of magnetic meridional wind come from the GAIM-FP model. The discrepancies found between the results of our wind model and HWM93 (Figure 3.4 and 3.5) are similar to the findings of Emmert [2001] (especially in the zonal winds) who analyzed
UARS/WINDII data and compared the longitudinally averaged winds to HWM93 wind values. The results from our TWAM model are in good agreement with the corresponding WINDII results reported in that study in terms of flow patterns, seasonal variations, reversal times, and time intervals of attaining maximum values (comparison not shown).

The comparison of the estimated meridional and zonal wind climatology with their corresponding nighttime measurements from FPI at different sites also showed good agreement (Figure 3.2 and Figure 3.3). The close match with observations indicates that our method of wind estimation is reliable. Note, the three FPI sites were chosen based on their extensive wind database (spanning about three solar cycles); however, for our wind model, these three sites are random places on the globe (given other model inputs have similar uncertainties). The largest difference between the model and the observations was seen over Millstone Hill in June solstice for the meridional wind and in December solstice for the zonal wind. The magnetic meridional wind from GAIM-FP shows a similar difference in June when compared to corresponding FPI estimates over Millstone Hill. Interestingly, this is also the case with the magnetic meridional wind estimated from ISR data [Emmert et al., 2003]. Since both, GAIM-FP and the ISR method use the MSIS model, we speculate the reason of the differences might be partially related to uncertainties in the empirical neutral atmosphere model. Furthermore, in our study, the averaged FPI wind data includes measurements from several solar cycle minima. The winds estimated from the TWAM model corresponds to the 2007 - 2009 solar minimum conditions. It is well known that this latest solar minimum showed anomalously low thermospheric mass density [Emmert et al., 2010]. It remains unclear how this anomaly translates into wind velocities. The GAIM-FP model uses the MSISE-90 model for the neutral atmosphere and TWAM also takes several parameters from the same empirical model. The errors due to the overestimation of the neutral densities in MSIS are expected to affect the accuracy of the estimated winds. The specific reason behind the occurrence of largest difference between the measured and modeled zonal winds over Millstone Hill during December solstice needs further studies. It should be mentioned that the dependence of the magnetic meridional wind on the zonal component of
the wind is weak for most of the globe because of small declination angles [see Eq. (3.10)], thus the estimated geographic meridional wind is more sensitive to any error in $v_{mag}$, while the accuracy of zonal component depends more on the theoretical model. Note, the two components of the wind, in addition to Eq. (3.10), are also coupled through the equation of motion [Eq. (3.1)].

Our estimated wind components are reasonably close to the HWM93 model values over the selected three sites for most of the cases and throughout the day. However, there are times when TWAM and HWM93 show large differences. In some cases, and in particular for the zonal wind, our model agrees better with the FPI data than HWM93, which shows large deviations from the data (for example Arecibo, June solstice). This is surprising because the HWM93 model in its construction includes ground-based wind data, including nighttime FPI data over Arecibo and Arequipa [Hedin et al., 1991]. It should be expected that a similar wind comparison over places where the empirical wind model is based on less data, shows larger differences.

The magnetic meridional wind, estimated by GAIM-FP, is assumed to be constant in altitude (including heights below the F2-layer peak). Since the GAIM-FP model assimilates NmF2 and hmF2, it is reasonable to assume the estimated magnetic meridional wind is most accurate at the altitudes of hmF2. This is the altitude where TWAM assimilates the GAIM-FP wind data. In the wind equation, the viscosity term that depends on vertical gradients is a very important term [Rishbeth, 1972] and needs to be included in the calculation. As a result, we obtain the full altitude variation of the computed zonal and meridional winds. The altitude gradients of the zonal and meridional winds obtained in this study are very small above the F2-layer peak (~250 km). This is consistent with findings of previous studies [Rishbeth, 1972; Roble et al., 1974; Blum and Harris, 1975b; Titheridge, 1995], and is due to an increase in the kinematic viscosity ($\mu/\rho$) with altitude, which smooths out height variations of the wind. As was shown (Figure 3.6), our results indicate larger altitude gradients of the wind velocity in the 110 - 250 km range. However, one should expect large uncertainties in these results as they depend on empirical inputs in the wind
equation (e.g., neutral temperature and composition, and electromagnetic drifts) and also on the adopted form of the parametrized pressure gradient force. The bottomside electron densities are also less accurate than values at the F2-layer peak and above, and hence, is the ion drag. However, the effects of ion drag force are relatively small because of the smaller ion densities at lower heights. Note, in the momentum equation, we also ignored ion motion along the electric field at these altitudes. An additional uncertainty on the wind velocities at lower altitudes also comes from our choice of the lower boundary conditions. In addition, we assimilate the GAIM-FP wind data only at altitudes of hmF2, which even though affects the wind speed at other altitudes, clearly cannot fully specify the height dependence of the wind.

The calculated horizontal wind velocities are used to estimate the vertical component of the wind, which is an order of magnitude smaller (justifying our initial assumption of zero vertical wind). However, vertical circulation plays an important role on the thermospheric composition, which in turn affects the ionospheric electron density [Rishbeth and Müller-Wodarg, 1999]. The obtained results of upward winds during the day and downward winds at night, as well as their increase in magnitude with altitude, is in agreement with prior theoretical calculations [Rishbeth et al., 1969; Roble et al., 1974; Rishbeth and Müller-Wodarg, 1999]. As seen, the horizontal winds vary little with height, and thus the altitude variation of the vertical wind is related to the exponential variation of the neutral density with height [Rishbeth et al., 1969]. For this study, we have not investigated the global pattern and the seasonal dependence of the vertical wind, and have not studied the 3-D circulation pattern in detail, all of which are subject of separate investigations.

There are several factors that could potentially affect the accuracy of the estimated meridional and zonal components of the thermospheric neutral winds. As was mentioned above, the uncertainty in the magnetic meridional wind, obtained from GAIM-FP, largely determines the uncertainty of the geographic meridional component over places where the magnetic declination angle is small, and a generally similar sensitivity of the geographic and geomagnetic wind components to model parameters should be expected. In Chapter 2
it was shown that change in the Burnside factor (parameter associated with O\textsuperscript{+}-O collision frequency) from 1.0 to 1.4 or 60 K change in the exospheric temperature have larger effects on the nighttime wind estimation over mid-latitudes and on average could result in an error of about 22 ms\textsuperscript{−1}. Furthermore, at mid-latitudes, GAIM-FP does not separate the estimated ion vertical motion due to \( \mathbf{E} \times \mathbf{B} \) drifts from wind-induced motion, and therefore, the obtained wind is an effective wind. On the example of magnetic meridional wind over Millstone Hill in Chapter 2, it was shown that the average error due to the \( \mathbf{E} \times \mathbf{B} \) drift is order of about 15 ms\textsuperscript{−1}. The detailed effects that the use of an effective wind has on our estimated wind components needs to be investigated, but is bound by the above mentioned errors.

The zonal wind is expected to be more sensitive to changes in the theoretical model parameters than the meridional wind. This can be understood by realizing that \( v_{\text{mag}} \), which represents our assimilated data, is only weakly dependent on the zonal wind component. On the other hand, it is known that the zonal wind and the zonal plasma drifts are highly correlated over the equator in the F region [Fejer et al., 1985; Park and Lühr, 2013]. The zonal wind velocity depends on the zonal \( \mathbf{E} \times \mathbf{B} \) drift through the ion drag term and an uncertainty in the later is anticipated to affect our estimates of the zonal wind. In order to evaluate the sensitivity of the model winds to zonal plasma drifts, we performed a wind estimation with the empirical zonal drifts increased by 50% everywhere for our December solstice case. Our results indicate that indeed, the larger effect is over the low latitudes, but with a difference of only up to 20 ms\textsuperscript{−1} in the estimated zonal wind. As expected, the change in the meridional wind velocity due to the modified zonal drifts was negligibly small globally and only 5 - 7 ms\textsuperscript{−1} over certain low-latitude places.

Equation (3.1) is nonlinear since it contains the advective acceleration term, \((\mathbf{u} \cdot \nabla)\mathbf{u}\), which makes it difficult to solve numerically. This term is relatively small compared to the other terms if the wind speed is not a significant fraction of the peripheral speed of Earth’s rotation and significant spatial gradients are not present. These assumptions may not be necessarily true near sunrise and sunset, and in the case of a direct numerical solution may
affect the wind phase by about 0.5 h and the amplitude of a calculated wind by about 15% as discussed by Rishbeth [1972]. Blum and Harris [1975b] performed an analysis of the effects of the nonlinear acceleration term and found it to be only significant in the equatorial region, especially at solstice conditions and only for meridional winds. At mid-latitudes they found the effect to be less than 10% on the wind magnitude without an appreciable change in the wind phase. In order to take full advantage of the Kalman filter and also to avoid complications related to numerical instabilities, we omit the nonlinear acceleration term. Furthermore, since we use a Kalman filter that combines data with the dynamical model, the effect of neglected physics is expected to be smaller than it is for a pure numerical solution due to the constraint that the data impose on the model. In addition to the nonlinear acceleration term, we also neglected the viscous terms resulting from horizontal gradients of the wind velocity, as well as nonlinear curvature terms. The horizontal viscosity terms are much smaller compared to the vertical one, and according to [Blum and Harris, 1975a,b], can only be important in a narrow latitudinal band near the equator if meridional and zonal velocity gradients are large. The nonlinear curvature terms, are also smaller compared to other terms in the equation as a simple scale analysis reveals (similar analysis but for tropospheric case are found in Holton [2004]). These omissions are not expected to significantly affect our estimation of the horizontal wind velocities.

A further assumption that could potentially affect our wind estimation is the choice of a zero pressure gradient force for our lower boundary at 110 km altitude. Consistent with this assumption, we have also set the lower boundary conditions for the winds to zero. Contrary to this, it is well known that the lower thermosphere undergoes strong tidal forcing from the stratosphere and the troposphere, creating large E-region winds [Fuller-Rowell, 1995; Larsen, 2002]. However, it is also well established that the solution of wind equation above 150-160 km is practically independent of the imposed wind values at the lower boundary [Lindzen, 1967; Blum and Harris, 1975a,b; Fuller-Rowell and Rees, 1980], consistent with test runs we have performed with varying lower boundary conditions. Once again, this indicates that the model winds below the F2-layer peak should be treated with
caution. For tidal variations that propagate upwards into the thermosphere and affect our COSMIC ionospheric data, can, to a certain degree, be reflected in our derived magnetic meridional wind data, and hence, also be present in the estimated zonal and meridional winds. Ideally, the tidal forcing term should be included in the momentum equation, or internally estimated, data permitting.

Finally, we would like to point out that the model adjusts the horizontal pressure gradient force based on the assimilated GAIM-FP data, while the neutral densities and temperatures needed in various terms in the equation of motion are directly taken from the empirical MSIS model. This inconsistency, however, is not expected to introduce large errors. Note, GAIM-FP also uses the same neutral atmosphere from MSIS. GAIM-FP has the ability to adjust neutral composition and temperature together with magnetic meridional wind, a feature that is still under development and not used in the current study. In future studies, however, the modified neutral atmosphere from GAIM-FP could directly be used for pressure gradient forces.

In summary, we developed a simple physics-based data assimilation model for thermospheric neutral winds (TWAM), which assimilates magnetic meridional wind data from an ionospheric data assimilation model (GAIM-FP). TWAM employs an implicit Kalman filter technique and provides the climatology of the geographic meridional and zonal, and vertical winds in the 110 - 600 km altitude range over low and mid-latitudes. The GAIM-FP model, for its part, assimilates global seasonal maps of NmF2 and hmF2 from COSMIC radio occultation data to provide an estimate of the magnetic meridional wind data together with the 3-D electron density distribution and ion diffusion velocity in the ionosphere, which are both important for a proper ion drag specification. Our analysis of the spatial and temporal variations, as well as comparisons with ground-based wind measurement from FPIs revealed that the model provides reliable estimates for the neutral winds. The proposed technique is important for a better specification of the global thermospheric dynamics, and especially over areas where wind observations are limited or not available at all (such as remote places and oceans). This, in turn, is significant for a better understanding of the climatology of
the F-region ionosphere, which strongly depends on thermospheric winds. The knowledge of the individual components of the horizontal wind also enables us to separately study their effects on large-scale structures in the ionospheric electron density. In future studies, we will investigate the use of this data assimilation technique to estimate thermospheric winds on a day-to-day basis, as well as during geomagnetically disturbed periods.

References


CHAPTER 4
MODELING AND ANALYSIS OF IONOSPHERIC EVENING ANOMALIES WITH A PHYSICS-BASED DATA ASSIMILATION MODEL

Abstract

Among several intriguing features of the F-region ionosphere are the anomalous evening enhancements of electron densities over certain mid-latitude sites. The phenomena take place during local summer and are most prominent over the regions west of the Antarctic Peninsula during December solstice [known as Weddell Sea Anomaly (WSA)] and over the eastern Siberia/Kamchatka Peninsula during June solstice. For these periods and locations of evening anomalies, the electron densities may become more than twice as large during the evening/night hours than during noon, while for neighboring mid-latitude locations, similar features are not observed or are less prominent. The anomalies completely disappear during local winter. Though these evening enhancements have been known for several decades, their generation mechanisms are still being debated and their accurate modeling remains a challenge. In this work a data assimilation model was used for the first time to elucidate the physical mechanism behind the observed evening anomalies. Radio occultation measurements from the six FORMOSAT-3/COSMIC satellites were assimilated into the physics-based Global Assimilation of Ionospheric Measurements Full-Physics (GAIM-FP) model. GAIM-FP provides the 3-D plasma density distribution throughout the ionosphere that agrees closely with the observation, together with estimations of the physical drivers, including the magnetic meridional wind. To understand the role of neutral winds in the evening anomalies, a newly developed Thermospheric Wind Assimilation Model (TWAM), which uses data from GAIM-FP, was employed to estimate geographic zonal and meridional wind components in the thermosphere. The TWAM winds were combined with the Ionosphere-Plasmasphere Model (IPM) and simulation for December and June solstices were performed. The model results of ionospheric F2-layer peak parameters (NmF2, hmF2) show
close agreement with the corresponding COSMIC measurements. The effects of separate wind components on the generation of anomalies were also studied. The results indicate the thermospheric meridional wind alone can be responsible for evening peak in electron density; however, in order to obtain the close agreement with the ionospheric data over the WSA, the inclusion of zonal wind effect is important. The combined zonal wind-magnetic field declination effect, in addition to meridional wind, helps to modify ion vertical drift and create favorable condition for the density increase during evening. The zonal wind effect is shown to be less pronounced for the northern hemispheric evening anomaly. To understand the physical mechanism behind the anomaly, the plasma production, loss and transport processes in the F-region ionosphere were separately analyzed and the relative role of these processes in the evening enhancements of electron density was established. It is found that due to the action of the strong equatorward wind during the evening, which inhibits the downward plasma diffusion, the density maximum forms at higher altitudes where the density reduction due to recombination is relatively slow. In addition, it was revealed that during the evening, the plasma loss due to transport weakens. As a consequence of the reduced rate of recombination and the weakened plasma loss due to transport, the relative role of solar production increases and the electron density enhancement occurs. Results of this study demonstrate that the evening anomalies can be explained by the neutral wind and by its effects on the various plasma parameters.

4.1. Introduction

During the early days of ionospheric research, the typical daily variation of the electron density was thought to follow a simple photochemical behavior, characterized by a noon-time maximum due to higher solar production rate and by a gradual decrease in density at night when plasma recombination becomes dominant. Interestingly, a seemingly unusual behavior of the F2-layer critical frequency (foF2) was reported at Halley Bay when the diurnal maximum of the density was observed near midnight in December (local summer) [Bellchambers and Piggott, 1958]. Considering the abnormality of such behavior also at other magnetically upper-midlatitude stations close to the Antarctic Peninsula, this phe-
nomenon was later called the Weddell Sea Anomaly (WSA) as its extent was believed to span from the Falkland Island to the southern shore of the Weddell Sea reaching Halley Bay [Penndorf, 1965]. However, in spite of the longevity of the topic, the research on the morphology and possible causes of the WSA still continues [Ren et al., 2012; Chen et al., 2013; Slominska et al., 2014]. In addition, the modeling of the phenomena remains a challenge [Burns et al., 2008; Jee et al., 2009], and therefore, its investigation is important for a better understanding of the underlying processes in the ionosphere and for elucidating and improving current shortcomings of global ionospheric models. Note, today the importance of dynamical effects in ionospheric variations, in addition to photochemistry, are well established; therefore, strictly speaking, the term anomaly, which relates to deviations from a simple Chapman theory, is not necessarily appropriate [Stubbe, 1975].

Since the discovery of the WSA, similar anomalies, but with relatively smaller noon-evening density differences, have been known to exist over certain northern hemisphere mid-latitude stations during local summer [Eyfrig, 1963; Evans, 1965; Kohl et al., 1968; Eccles et al., 1971; Eccles and Burge, 1973]. For example, Papagiannis and Mullaney [1971] investigated the global distribution of the ionospheric evening anomaly using F2-layer critical frequency obtained by harmonic analysis of worldwide ionosonde data. Their results show three regions during northern summer over north-east America, western Europe, and eastern Siberia where evening anomalies are strongest, as well as the WSA during southern summer.

In spite of the apparent similarities between the WSA and the evening anomalies in the northern hemisphere, their analysis often was separate. Several mechanisms were proposed to explain the WSA. Bellchambers and Piggott [1958] first attributed the near midnight maximum of the electron density in summer, which is now known to be part of the WSA, to the dynamics of the ionosphere and not to the role of recombination processes; Rastogi [1960] interpreted the density pattern over the WSA more with a midday depression and suggested the daytime minimum was due to the lack of horizontal transport of ionization from the equator along the magnetic field lines and excluded the role of vertical drifts of
the ionization. Hill [1960] suggested horizontal winds with vertical shear as a possible cause of the anomalous behavior of the electron density. Penndorf [1965] (noted the necessity of F2-layer peak height analysis) related the WSA to the South Atlantic (SA) magnetic anomaly and suggested excess ionization above the F region over the SA anomaly could be drifted downwards and then, through horizontal drifts, brought into the Weddell Sea area. Dudeney and Piggott [1978] underlining the role of magnetic declination over the anomaly area, considered solar illumination superimposed with the upward drift due to thermospheric winds to be responsible for the formation of the WSA (a similar idea of explaining the evening density enhancement in general, was earlier developed by Eccles et al. [1971]). They noted that their modeling result showed an additional morning enhancement in foF2, which was, however, not observed. They pointed out that this could be due to their inaccurate winds, or more likely due to some extra physical process missing from the simulation.

Other studies were more global. Eyfrig [1963] noticed that the daily electron density maximum was shifted towards the evening hours over stations with westerly magnetic declination in the northern hemisphere and with easterly magnetic declination in the southern hemisphere. Using numerical modeling, Kohl et al. [1968] and Eccles et al. [1971] showed thermospheric neutral winds can be responsible for the observed evening enhancements of electron densities at certain southern and northern mid-latitude sites. Kohl et al. [1969] explained the magnetic declination effect in the ionosphere [Eyfrig, 1963] by the dependence of the phase and amplitude of the wind-induced vertical drifts on the declination angle. Rishbeth [1967, 1968] showed neutral winds can produce a secondary maximum in the summertime NmF2 during the evening. Evans [1965] studied the evening increase of the electron density above Millstone Hill by using incoherent scatter radar data. He found that the density above the peak starts to decrease a few hours earlier than the peak density, which reaches its maximum in the evening, accompanied by a rapid fall in electron temperature. The role of this thermal contraction was later rejected by Eccles and Burge [1973] who performed model calculations and concluded that the evening enhancement of
the electron density is a consequence of thermospheric winds. Papagiannis and Mullaney [1971] also found that the anomalies were more pronounced during local summer over areas with westerly magnetic field declination angles in the northern hemisphere and easterly declination angles in the summer hemisphere, as well as in regions where there was a poleward divergence of the magnetic meridians. By studying neutral wind patterns, it was suggested wind plays a very important role in the geographic variations of the anomaly.

The interest in the WSA was renewed as new advances of ionospheric measurement technology, and especially satellite observations over the oceans made it possible to obtain a global image of the extent of the WSA. This provides a unique opportunity to extensively study the phenomenon. For example, Horvath and Essex [2003] used total electron content (TEC) measurements from the TOPEX satellite during the 1998 and 1999 solar maximum and revealed the anomaly situated over the Bellinghausen Sea and not over the Weddell Sea. According to the TEC analysis by Horvath [2006], the anomalous region was found over a large area (∼22 million km$^2$) peaking at 50°S-60°S/90°W-110°W. Jee et al. [2009] used 13 years of TOPEX TEC data to study the seasonal and solar activity variations of WSA for geomagnetically quiet periods. Their study shows that for low solar flux, the WSA occurs only for southern summer months; but, for high solar activity periods, it also occurs during the equinoxes, but is still most prominent during the December solstice. Karpachev et al. [2011] studied anomaly using intercosmos-19 satellite data of foF2 for solar activity maximum and CHAMP satellite data for low solar activity. The anomaly was found to be more pronounced at solar activity minimum, but its main characteristics were almost independent of solar activity level.

The advent of the new observations also sparked new studies about the generation mechanisms of the WSA. Burns et al. [2008] studied the WSA using FORMOSAT-3/COSMIC radio occultation (RO) data, and questioned the role of neutral winds and neutral composition in their generation. They suggested the plasmaspheric flux was at least partly responsible for the anomaly. Jee et al. [2009] noted the importance of the combined action of the meridional and zonal wind for the plasma vertical motion that varies by inclination.
and declination of the Earth's magnetic field, and suggested it as a factor for the plasma density increase at night. Karpachev et al. [2011] analyzed the longitudinal variation of F2-layer peak parameters at the WSA latitudes and by comparing them to the zonal variation of calculated vertical drift velocities, which were based on experimental NmF2 and hmF2 values, concluded the anomaly is mainly caused by neutral winds. They also noted that solar radiation considerably contributes to the formation of the anomaly, while the estimated contribution of neutral composition and temperature is insignificant. The role of other mechanisms was questioned.

In addition to the studies of the WSA, the radio occultation measurements of electron density from the COSMIC satellites have also been used to investigate the evening anomalies at other locations [He et al., 2009; Lin et al., 2009, 2010; Lomidze and Scherliess, 2010; Burns et al., 2011]. Using COSMIC RO data Lin et al. [2010] also revealed a less pronounced WSA-like structure in the northern hemisphere near northeast Asia, Europe/Africa and Central Pacific longitudes around June solstice. They linked both anomalies to a similar mechanism, noted they are located at areas where the offset of the geomagnetic and the geographic equators are large, and proposed the possible role of upward/poleward transport of ionization from the poleward edge of EIA in the generation of the anomalies. Similarly, Burns et al. [2011] proposed the equatorial anomaly to be the source of the WSA where upward and poleward drifts, due to conjugate electric fields, transport the ionization. They applied a similar mechanism to the northern anomaly and considered the neutral winds to be an unlikely cause of the anomalies.

Besides the search for the drivers of the anomaly, which are still being debated, the modeling of the phenomenon is another challenge. The WSA was, for example, not found in the outputs of the Thermospheric General Circulation Models (TGCM's) and the Coupled Magnetosphere Thermosphere Ionosphere (CMIT) models [Burns et al., 2008; Jee et al., 2009]. Lomidze and Scherliess [2010, 2011] were first to report the use of an ionospheric data assimilation model to study the WSA and evening anomalies, and similar to previous studies, suggested the thermospheric neutral winds to be the main drivers of the anomaly.
Chen et al. [2011] studied the WSA using SAMI2 and investigated the role of neutral winds that were specified from the HWM93 model. They found the thermospheric equatorward wind is a critical driver of the WSA, and the plasmaspheric flux, although not the critical driver, only provides the plasma source at later times after the anomaly is already formed. While SAMI2 did reproduce the evening time density enhancement over the right location, Chen et al. [2011] did not compare the modeled results with observations making it difficult to speculate how well the model (driven by HWM93) was capable to fully reproduce the anomaly. Chen et al. [2013] used the SAMI3 model driven by HWM93 and TIEGCM neutral winds to study the northern and southern evening anomalies. The WSA was reproduced in both cases; however, no quantitative comparison of the results with observations was made. It was also found that the used winds could not properly account for the location of evening density enhancement in the northern hemisphere. Thampi et al. [2011] presented the SUPIM model simulations for the northern anomaly using HWM93 and MU radar winds. They found that electric fields and zonal winds have an insignificant contribution to the anomaly. They also reported that using the local MU radar winds along the magnetic meridian, which differ from their corresponding HWM93 model values, the model results were qualitatively closer to the COSMIC RO derived densities. However, it remains unclear how well the variation of the peak height of the F2 layer was reproduced by the SUPIM model. Ren et al. [2012] performed simulations with the TIME3D-IGGCAS ionosphere model driven by HWM93 empirical winds and studied the role of meridional and zonal winds separately. They concluded that the anomalies are mainly driven by meridional winds, and found that zonal wind effects are also important for the WSA, but negligible for the northern hemispheric anomalies. The authors did not provide comparisons of the modeled NmF2 and hmF2 with observations; however, their magnitudes and patterns of the local time variations over the WSA seem to be very different from actual measurements (to be shown below).

It is apparent the wind mechanism is the most widely accepted mechanism for the formation of the evening anomalies (combined with solar photoionization), which is supported
by phenomenological explanations and/or numerical simulations [Kohl et al., 1968, 1969; Eccles et al., 1971; Papagiannis and Mullaney, 1971; Eccles and Burge, 1973; Dudeney and Piggott, 1978; He et al., 2009; Jee et al., 2009; Liu et al., 2010; Chen et al., 2011, 2013; Karpachev et al., 2011; de Larquier et al., 2011; Thampi et al., 2011; Ren et al., 2012].

Yet, they fail to validate key ionospheric parameters (NmF2 and hmF2) simultaneously and/or rely on simplified or empirical wind models, which are known to be less accurate over data sparse regions [Titheridge, 1995; Makela et al., 2012]. In the current paper, we use electron density measurements from FORMOSAT-3/COSMIC satellites in combination with the Utah State University Global Assimilation of Ionospheric Measurements Full-Physics (GAIM-FP) model to understand the causes of the Weddell Sea Anomaly and the Northern hemispheric mid-latitude evening Anomaly (NA for short, one over Eastern Siberia/Kamchatka Peninsula) and the role of thermospheric neutral winds in their generation (preliminary results of the work were reported by Lomidze and Scherliess [2010, 2011]).

GAIM-FP provides the 3-D electron density in the ionosphere, which agrees closely with the observations globally, including the anomaly regions. The model also provides an accurate estimate of the magnetic meridional wind needed to properly model the ionospheric parameters (see Chapter 2).

In order to elucidate the role of the neutral wind and its meridional and zonal wind components in the generation of evening anomalies, the newly developed Thermospheric Wind Assimilation Model (TWAM) [Chapter 3] is employed. The model uses data from GAIM-FP and decomposes its magnetic meridional wind into geographic zonal and meridional wind components. The TWAM winds were used to drive the Ionosphere-Plasmasphere Model (IPM) in order to analyze the various ionospheric parameters over the anomalies. The IPM results of NmF2 and hmF2 agree well with corresponding COSMIC data when driven by TWAM winds, in contrast to simulations when IPM is driven by HWM93. Our results show that the geographic meridional wind is sufficient to provide the conditions for the eveningtime density enhancement, however to obtain close quantitative agreement between the model results and observations over the WSA, it is important to also properly
account for zonal wind effects. The zonal wind effects come into play through magnetic field
declination effects and help to enhance the anomaly. For the NA, the effect of the zonal
wind is very small. After accurately modeling the anomalies and validating key ionospheric
parameters, the causes of the density increase over the WSA are investigated. In particular,
the field-aligned plasma velocities and plasma production, loss, and transport processes are
analyzed over the WSA and outside the WSA, where the evening density enhancement is
not present. A decreased rate of plasma recombination due to increased hmF2 and reduced
loss rate due to plasma transport, in conjunction with the still ongoing photoionization during
the evening hours, are found to be responsible for the electron density evening anomaly.
This is different from the off-WSA location where loss due to recombination and transport
processes are stronger. The condition over the WSA, which creates the density anomaly,
can be fully explained by neutral winds and their effects on the plasma parameters. The
same mechanism can be applied to explain the occurrence of the NA.

4.2. Methodology and Approach

Our study of the evening anomalies is based on COSMIC radio occultation data, which
is used to create global seasonal maps of ionospheric NmF2 and hmF2. The COSMIC maps
are assimilated into GAIM-FP to obtain the magnetic meridional wind consistent with the
observed ionospheric morphology. Next, the magnetic meridional wind from GAIM-FP
model is assimilated into a separate data assimilation model for neutral wind (TWAM)
to obtain geographic zonal and meridional winds over low and mid-latitudes. Finally, the
wind components are used in the Ionosphere-Plasmasphere Model (IPM) to analyze the
various physical processes, which play a role in the generation of the anomaly, and to
understand, in detail, the relative role of the zonal and meridional winds. The next section
briefly describes the COSMIC data, the GAIM-FP and IPM models, and the TWAM data
assimilation model.
4.2.1. COSMIC Data

The values of $N_{mF2}$ and $h_{mF2}$, and their corresponding standard deviations are obtained from GPS radio occultation measurements from the Constellation Observing System for Meteorology Ionosphere and Climate/Formosa Satellite 3 (COSMIC/FORMOSAT 3). The constellation consists of six microsatellites in a circular orbit at about 800 km altitude and has been providing about 1,000 - 2,500 electron density profiles per day globally \cite{Rocken2000, Anthes2008}, which are retrieved from slant total electron content measurements using an Abel inversion technique \cite{Hajj1999, Schreiner1999}. The data of electron density profiles used in this study were obtained from the COSMIC Data Analysis and Archive Center (http://www.cosmic.ucar.edu).

The COSMIC data were prepared for the assimilation into GAIM-FP. First, the electron density profiles were obtained and initial quality check performed, which removed apparent outliers (they constituted $\sim$3% of data in December, $\sim$5% in June, seasons are defined below). The values of $N_{mF2}$ and $h_{mF2}$ were extracted from the individual electron density profiles, and to obtain the seasonal global maps, the data were combined from 2007 to 2009 and binned into seasons consisting of December solstice (December 21 $\pm$ 27 days for the years 2007 and 2008) and June solstice (June 21 $\pm$ 27 days for 2008 and 2009). To obtain quiet time data, points were excluded for observations with a previous 24-hour average $A_p \geq 16$ and/or a corresponding 3-hour $ap \geq 16$, which constituted $\sim$12% of the remaining data for December and $\sim$5% for June. The final number of total ROs selected were $\sim$161,600 for December and $\sim$144,200 for June, respectively. The Average solar and geophysical conditions for the selected periods were $F_{10.7} = 73, A_p = 5$ for December, and $F_{10.7} = 67, A_p = 4$ for June. The days that best represent the climatological average conditions of the selected seasons, based on their average solar declination, are DOY 006 for December solstice and DOY 189 for June solstice.

The retrieved $N_{mF2}$ and $h_{mF2}$ were used to produce average global maps of these parameters in 30-minute intervals of universal time. For the binning we used a longitude $\times$ latitude grid of 7.5$^\circ \times 2.5^\circ$ with a 15$^\circ \times 5^\circ$ sliding window. To further exclude outliers and
potentially erroneous data, the procedure was performed a second time and those data on a given grid-point that deviated by more than $2\sigma$ from the mean were excluded. We also excluded data for which the averages were based on less than three observations for a given grid-point. The obtained standard deviations are used as an estimate of the data errors, which are important inputs for the data assimilation models. For more details about the COSMIC data refer to Chapter 2.

### 4.2.2. Ionospheric Models

The GAIM-FP (Global Assimilation of Ionospheric Measurements Full Physics) is a data assimilation model that uses a physics-based ionosphere-plasmasphere model (IPM) and an ensemble Kalman filter technique to incorporate the main physical processes in the ionospheric directly into the data assimilation scheme [Schunk et al., 2004; Scherliess et al., 2004, 2009]. The model can assimilate different types of data from different sources and provide specifications of the 3-D electron density distributions from 90 km to 30,000 km at low and mid-latitudes. The model can also provide an estimation of the ionospheric drivers (electric field, magnetic meridional wind, and neutral composition) that is consistent with the data and the model. In its current version, the data assimilation model excludes regions poleward of $\sim \pm 60^\circ$ geomagnetic latitudes.

In the IPM model [Schunk et al., 2003], the ion and electron continuity and momentum equations are numerically solved along magnetic field lines for individual convecting flux tubes of plasma, and the 3-D nature of the model is obtained by following a large number of plasma flux tubes. The 3-D distribution is obtained by mapping the results on a geographic grid. IPM uses the International Geomagnetic Reference Field (IGRF), which properly accounts for the displacement between the geomagnetic and geographic equators and the bending of the magnetic field lines with latitude.

The Kalman filter [Kalman, 1960] combines the data from an observing system with the information obtained from the system model and their corresponding statistical description of uncertainties. It performs a recursive least squares inversion of the observations ($Nmf2$ and $hmF2$ for our case) for the model variables using a dynamical model (IPM) as
a constraint. As a result, an improved estimate of the model variables (electron density, magnetic meridional wind) is obtained in a statistical sense. The estimation of the ionospheric drivers is an integral part of our ensemble Kalman filter and is achieved by using the internal physics-based model sensitivities to the various driving forces. In this procedure, the ionospheric data are used to adjust the plasma densities and its drivers so that a consistency between the observations (within their errors) and the physical model is achieved. As a result, the assimilation procedure produces the optimal model data combination of the ionosphere-plasmasphere system with its consistent drivers (for details see Scherliess et al. [2009, 2011]).

In the current study, the climatology mode of GAIM-FP was used, global seasonal maps of NmF2 and hmF2 were assimilated, and the 3-D electron density, together with the low- and mid-latitude magnetic meridional wind and the low-latitude vertical $E \times B$ drifts, were obtained. The global neutral composition was specified by MSISE-90 [Hedin, 1991]. The details about the sensitivity of the estimated magnetic meridional winds to different uncertain model parameters were discussed in Chapter 2.

4.2.3. Thermospheric Wind Assimilation Model

The physics-based 3-D Thermospheric Wind Assimilation Model (TWAM) uses magnetic meridional wind data from GAIM-FP and combines it with the equation of motion for the neutral gas in the thermosphere to obtain the climatology of the geographic zonal and meridional wind components (Chapter 3). It is based on an implicit Kalman filter technique [Skliar and Ramirez, 1997] and solves for horizontal wind velocities in the 110-600 km range. The GAIM-FP magnetic meridional data, which constrain the wind solution on a global grid, is assimilated at the altitude of hmF2. The vertical wind is calculated retrospectively from the requirement to satisfy the neutral air continuity equation, which uses the obtained horizontal winds and empirical neutral densities. To solve for the horizontal wind components, the model uses neutral temperatures from MSISE-90 and $E \times B$ drifts from Scherliess-Fejer empirical drift models [Scherliess and Fejer, 1999; Scherliess et al., 2001]. The pressure gradient forces, however, are internally calculated by the model and
thus, are not directly obtained from the MSIS neutral atmosphere model. The global 3-D electron density distribution and ion diffusion velocity, which contribute to the ion drag term in the momentum equation, are also obtained from GAIM-FP. The results from the wind model have been validated by comparing the individual wind components with measurements from interferometers and with HWM93 results over stations that have long-term wind observations (Chapter 3).

4.3. Results

4.3.1. Evening Anomalies from COSMIC Data

The first step in our study is to investigate the evening anomalies using COSMIC radio occultation data. The utilized UT maps of global distribution of NmF2 and hmF2 were recast to represent their distribution for a constant local time (i.e. LT maps). Figure 4.1 and 4.2 show global maps of NmF2 and hmF2 for noon (11:00-13:00 LT) and evening (20:00-22:00 LT) hours during December and June solstices, respectively. The black solid line indicates the location of the magnetic equator. The COSMIC RO data provides a good global coverage and successfully captures many well-known characteristics of the ionosphere. Over low latitudes these include the equatorial ionospheric anomalies, the seasonal asymmetry in hmF2 and longitudinal variations in NmF2 and hmF2. Over mid-latitudes certain features are masked by the choice of our color scale; however, larger (smaller) noon-time NmF2 (hmF2) values compared to values during evening hours are evident for most locations. The places that clearly do not fit to this general pattern are a large area over west of the Antarctic Peninsula during December and a relatively smaller area centered over the Kamchatka Peninsula during June. These anomalies are the Weddell Sea Anomaly and the Northern Anomaly, which have considerably larger NmF2 during evening than during noon. They also are clearly distinct from their neighboring locations to the east and west (especially the WSA). Note, during June solstice, the evening NmF2 values over Europe are also close to those of the NA, but their noontime values are relatively larger than the noontime NmF2 values over the NA. The hmF2 values over the WSA exhibit a large de-
expression during noon (Figure 4.1), clearly making the region distinct, but a similar pattern over the NA is not seen in Figure 4.2. Note the relatively high values of hmF2 during the evening over the WSA compared to its eastward and westward neighbors, which are seen not only during December, but also during June. Furthermore, the high evening values of hmF2 over the NA during June and December are clearly a distinct feature.

In order to obtain the global spatial extent and estimate the magnitude of the evening anomalies, we calculated global (normalized) differences between the evening and noon values of NmF2, which are presented in Figure 4.3. The data are shown only over the places where the difference was positive (i.e., for $\text{NmF2}_{\text{evening}} > \text{NmF2}_{\text{noon}}$). The WSA occupies an area that is almost twice as large as the NA. The density anomaly is also stronger over the WSA with a difference between the evening and noon values over its center exceeding 100%, compared to $\sim 40\%$ for the NA. During the June solstice, a relatively weak ($\sim 15\%$)
evening anomaly is also present over western Europe/north Atlantic. Note, the magnitude and global extent of the evening anomalies would be modified if we had taken different time intervals for evening and/or noon.

4.3.2. Modeling of Evening Anomalies and Effects of Neutral Wind

Modeling with HWM93 and TWAM winds

The modeling of the evening anomalies was performed using two different setups for the IPM model. In its default setup, the IPM uses winds from the HWM93 empirical model (IPM-HWM93). In the modified setup, the HWM93 winds can be replaced by the horizontal wind values from our 3-D thermospheric wind assimilation model (IPM-TWAM). The outputs of hmF2 and NmF2 values obtained from IPM-TWAM are essentially the same as those from GAIM-FP since the magnetic meridional wind from GAIM-FP is used as input data in TWAM at the altitude of hmF2. When TWAM calculates the horizontal
Figure 4.3. Difference (in percentage) between evening and noon peak electron densities during December and June solstices. To highlight the global extent of the evening anomalies, only values for which $N_{m}F_{2}\text{evening} > N_{m}F_{2}\text{noon}$ are shown.

wind components, their combination into a magnetic meridional wind is ensured to be close (within the errors) to the input data assimilated by the Kalman filter. The effects of altitude variations of the wind, which are included in IPM-TWAM but not in GAIM-FP, on the electron density profiles are different; however, their study is out of the scope of this paper. In addition, the vertical winds from TWAM, which are around 2 - 4 ms$^{-1}$, are also ignored in the IPM simulation. In Chapter 2, it was already shown that compared to the default IPM, GAIM-FP significantly improves the agreement between the measured and modeled NmF2 and hmF2 globally and especially over the south Pacific and Eastern Asia. These differences were attributed to different wind values, since all other model inputs at mid-latitudes were identical.

Figure 4.4 presents comparisons of LT variations of NmF2 and hmF2 from COSMIC radio occultation data to their corresponding model simulations over the specific locations of the WSA (53.75°S, 93.75°W, MLAT=41.23°S) (right panels) and of the NA (53.75°N, 146.25°E, MLAT=47.43°N) (left panels) during December solstice and June solstice, respectively. The COSMIC data are presented with their corresponding 1-σ standard deviations. Note there is an about five-hour gap in the morning data over the WSA. According to the COSMIC data, the general patterns of NmF2 and hmF2 variations are very similar over the
two anomalies; however, the NmF2 over the WSA is about 50% higher. The values of hmF2 are about the same during the day, but are about 22 km higher during the night over the WSA. The electron densities over both anomalies start to gradually increase from the early morning hours. In the same time hmF2 descends rapidly first, reaching a minimum around 10:00 - 11:00 LT, soon after which it starts to rise. Interestingly, it is only after 18:00 LT when NmF2 also starts a sharp increase and attains a daily maximum around 20:00 - 21:00 LT. After that, the density starts to rapidly fall to a daily minimum around 2:00 - 3:00 LT. The differences between the F2-layer parameters over the two anomalies is expected to be due to differences in the neutral compositions and the neutral wind fields over these two sites. Also note the two sites have different magnetic latitude.

During both seasons, the IPM-TWAM results are in close agreement (within errors)
with the COSMIC data, in contrast to the IPM-HWM93 results, that indicates the TWAM winds correctly reproduce the observed evening anomalies. Note that because production, loss, and transport processes in the ionosphere depend on altitude, and neutral and electron densities, it is important to be able to closely reproduce the observed F2 layer variation before attempting to analyze the role of these different physical processes. IPM-HWM93 generally overestimates both, NmF2 and hmF2. Over the WSA the IPM-HWM93 displays large discrepancies in both NmF2 and hmF2 (except for a few hours around midnight). The largest difference in hmF2 occurs during the morning hours when it reaches $\sim 80$ km, together with relative differences in NmF2 of about 80%. Over the NA, NmF2 from both, IPM-HWM93 and IPM-TWAM agree well (within 1-$\sigma$) with the COSMIC data, but for most of the day, NmF2 from IPM-HWM93 is about 10 - 40% larger than the IPM-TWAM values. On the contrary, the late morning hmF2 from IPM-HWM93 is largely overestimated (by $\sim 50$ km) compared to the COSMIC values, whereas IPM-TWAM only overestimates hmF2 by about 20 km. Generally, an underestimation in NmF2 and overestimation in hmF2 for certain local times are present in the IPM-TWAM results. It is interesting to note that even though the NmF2 from IPM-HWM93 does not agree with the COSMIC data over the WSA, the evening values of NmF2 are larger than their noontime values, indicating the HWM93 winds qualitatively represent the anomaly in terms of NmF2, but fail to quantitatively describe it.

As it was noted before, the only reason for the differences between the two IPM simulations are due to the different thermospheric winds, since the rest of model inputs are kept identical. Figure 4.5 shows the meridional and zonal winds from HWM93 and TWAM for the same locations for which NmF2 and hmF2 were shown in Figure 4.4. Also shown are the field-aligned ion drifts (positive upward/equatorward) that are induced by these horizontal winds. Assuming the ion gyrofrequency is much larger than ion-neutral collision frequency, which is a valid assumption in the F region, these velocities are given by

$$v_{i,\parallel, wind} = (v \cdot \cos D \pm u \cdot \sin D) \cos I,$$  

(4.1)
Figure 4.5. Local time variations of meridional (top), zonal (middle) winds, and of field-aligned ion drift induced by horizontal winds (bottom) over the WSA during December solstice (left) and over NA during June solstice (right). Winds are from the HWM93 model (magenta) and the TWAM model (blue). Also shown are (lower two panels) field-aligned plasma velocities due to only TWAM meridional winds (green diamonds).

where $u$ and $v$ are eastward and equatorward winds, respectively. The positive and negative signs correspond to the southern and northern hemispheres, respectively. $I$ is the geomagnetic field inclination angle and $D$ is the declination angle, which at 250 km altitude, according to IGRF, have values of $D = +26.2^\circ$, $I = -55.3^\circ$ for the selected WSA site, and $D = -9.8^\circ$, $I = +67.0^\circ$ for the NA site. The TWAM winds are similar in general shape to HWM93 winds, but are different in their magnitudes, phases, and reversal times. The TWAM meridional wind over the WSA (upper left panel of Figure 4.5) becomes poleward about 3.5 hours earlier and reverses back 3 hours later than the corresponding HWM93 wind, which remains poleward for only 3 hours. Also on average the TWAM meridional wind is less equatorward than the HWM93 wind with daily average values 31 ms$^{-1}$ and 72 ms$^{-1}$, respectively. The reversal times are about the same for the TWAM meridional winds.
over the WSA and the NA, whereas the HWM93 wind over the NA has a minimum around noon, but never turns poleward. Similar to the WSA, the TWAM meridional wind over the NA is also less equatorward than the HWM93 values with daily averages of 34 ms$^{-1}$ and 61 ms$^{-1}$, correspondingly. The zonal winds are generally westward from midnight until about 15:00 LT and eastward for the other part of the day. However, there is a phase shift of about 2 to 4 hours between the HWM93 and the TWAM winds. Also note the HWM93 wind has a strong semi-diurnal amplitude and its magnitude oscillates around the TWAM wind values.

The wind-induced part of the field-aligned ion drifts (lower panels of Figure 4.5) combines the meridional and zonal winds through equation (4.1) and, therefore, is determined by both components of the wind. However, depending on the magnetic field declination angle, the contribution of the zonal and meridional wind is different. The pattern of differences between the TWAM and HWM93 field-aligned drifts is similar to the meridional wind differences, but because of zonal wind contribution to the field-aligned drift, its reversal from an upward to a downward motion occurs about 2.5 hours earlier over the WSA. This difference is only $\sim 0.5$ hours for the NA because of the small declination angle at this location. Note, the field-aligned plasma motion will affect the plasma transport in the F-region ionosphere. The green diamonds on the lower panels show the drift velocities that would result if only meridional winds were present. For the NA, as expected, it is very close to the full wind (when both $u$ and $v$ contribute) case. Over the WSA, because of the large declination angle, the amplitude and reversal times are both affected. In addition, the effect is especially large for the morning-noon period when the zonal wind values are relatively large.

**The Effect of Zonal Wind**

In order to estimate the role of the zonal wind on the evening anomalies, we performed additional simulations with the IPM model with the TWAM zonal winds set to zero, but all other model parameters remained unchanged. Figure 4.6 compares these new simulations (green diamonds) with the previous full wind cases (blue dots). The results indicate that the
geographic meridional wind alone is sufficient to produce the larger evening maximum in the electron density compared to the noon value over both anomaly regions. The contribution of the zonal wind is larger over the WSA as expected. Without the zonal wind in the simulation, the NmF2 value increases for the early morning to later noon hours by about 20% and decreases by about 15% at the evening peak. The hmF2 values also increase by about 20 km during the day and decrease by about 5 - 10 km in the evening. In contrast, the effect of the zonal wind on NmF2 and hmF2 over the NA are very small, which is again expected given the small magnetic field declination angle over this location. Here, the differences are about 4% for NmF2 during noon and evening, and ~5 km for hmF2 during the day. However, note that similar to the WSA case for the zero zonal wind simulations, the directions of NmF2 and hmF2 changes are identical. In other words, similar zonal

![Figure 4.6](image_url)

**Figure 4.6.** Model results showing the effect of zonal wind on local time variations of NmF2 (top) and hmF2 (bottom) over the WSA during December solstice (left) and over NA during June solstice (right). Simulations are obtained from the IPM model with full TWAM winds (blue dots) and IPM simulations with only TWAM meridional winds (green diamonds).
winds have similar effects over the two sites (although stronger for the WSA). This can be understood by realizing that the easterly (the WSA) and westerly (the NA) declination angles, when combined with the zonal wind, provide similar effects on the plasma in the F2-layer plasma because of the opposite magnetic field inclination angle in the two hemispheres. Overall, we can conclude that the zonal wind is not necessary to produce the WSA and the NA, but based on the values of NmF2 and hmF2 over the WSA and hmF2 over the NA, is important to obtain a better agreement of these parameters with the corresponding COSMIC values. The effect of the zonal wind, however, is more important for the WSA.

4.3.3. Comparison of Parameters over WSA and Outside WSA

In the previous section we showed that the IPM simulation with TWAM winds can closely describe the observed variations of NmF2 and hmF2 over the evening anomaly sites. The next step in our study is to identify the physical processes responsible for the density enhancements. In order to better understand the role of the various mechanisms, it is helpful to compare and contrast the behavior of the ionospheric and the thermospheric parameters over the anomaly region and outside of the anomaly region. Since the WSA and NA displayed very similar variation of NmF2 and hmF2, here we focus on the WSA only. For the site outside of the WSA (the off-WSA, thereafter) we chose location at (53.75°S, 11.25°E, MLAT=53.06°S), thus it has the same geographic latitude as the WSA site, which means the solar declination angles, and with this the solar production rates, are the same at two sites. The magnetic field declination and inclination are $D = -26.5^\circ$ and $I = -60.8^\circ$, respectively. While the inclination angles are about the same, the declination angle at the off-WSA is similar in magnitude to the WSA, but has opposite sign (westerly magnetic declination).

Figure 4.7 shows the variation of NmF2, hmF2, and the meridional and zonal winds over the WSA (blue dots) and the off-WSA (red stars). It should be noted that over the off-WSA, the agreement between the COSMIC data and IPM-TWAM simulation results is again improved compared to the IPM-HWM93 case (not shown). Different from the WSA, NmF2 over the off-WSA has a daily maximum around noon, and does not exhibit
the anomalous behavior. A very small peak at around 20:00 LT is, however, still present. The densities at the two locations are about the same during noon, but it is about two to four times larger at the WSA during the night. Interestingly, hmF2 at the off-WSA is \(\sim 30\) km higher at noon and lower by about the same amount during the evening. During the early morning, the hmF2 values at the two sites are similar, but the early morning NmF2 values at the WSA are much larger than the values at the off-WSA. The fact the F2 layer has the same hmF2 but different NmF2 during the morning and different hmF2 and similar NmF2 at noon is expected to be due to the different wind system over the two sites, and also due to a different neutral compositions over the two sites, as provided by MSIS. Below, we discuss the effect from differences in the neutral composition over the two locations, which we found not to be playing an important role, if any, in the anomalous behavior of the electron density over the WSA and in the regular behavior over the off-WSA.

**Figure 4.7.** Local time variations of F2-layer peak parameters (left panels) and neutral wind components (right panels) at 250 km altitude over the WSA (53.75°S, 93.75°W) (blue dots) and the off-WSA (53.75°S, 11.25°E) (red stars). Results correspond to IPM model simulations with TWAM winds and the December solstice conditions.
The winds have a strong influence on the variation of plasma motion along the geomagnetic field lines and affect the values of NmF2 and hmF2. As Figure 4.7 shows, the winds are indeed different over the two sites where the F2 layer displays distinct behavior. Both (WSA and off-WSA) meridional winds are poleward during day, but the wind is stronger (by $\sim 25 \text{ ms}^{-1}$) over the WSA. The effect of the stronger poleward wind is to more effectively push plasma down along the field lines reducing hmF2. During the evening, the wind over the WSA, when compared with the off-WSA location, turns equatorward about one hour earlier, increases more rapidly and attains higher values with difference of up to 40 ms$^{-1}$. Note, in the vicinity of the F-region peak, the equatorward wind opposes the downward field-aligned plasma diffusion. These variations are in agreement with the above reported variations of hmF2. The zonal winds are also different during the course of the day. However, since the WSA has easterly declination and the off-WSA has westerly declination, even the same wind would have opposite effect on the plasma motion over the two locations [see Eq. (4.1)]. The zonal winds on both sites further enhance the differences in plasma motions caused by differences in the meridional wind. Note, similar to the WSA case, the zonal wind effect is not as important as the meridional wind effect for off-WSA, and therefore, the differences seen in hmF2 and NmF2 over the two sites are primarily caused by differences in geographic meridional winds.

Comparison Between Plasma Motions

We showed (Figure 4.7) that over the selected two sites the neutral winds were different. To better understand the role of these winds, it is interesting to compare the field-aligned ion velocities over the two sites. Note, when discussing the dynamics of ionospheric plasma, one should keep in mind that besides the wind-induced part of field-aligned motion, plasma diffusion is also present, and it is the total ion velocity

$$v_{i,||} = v_{i,||,\text{wind}} + v_{i,||,\text{diff}}$$  

(4.2)
that determines the plasma motion. Here

\[
v_{i,\parallel,\text{diff}} = -D_a \sin|I| \left[ \frac{1}{n_i} \frac{\partial n_i}{\partial z} + \frac{1}{T_p} \frac{\partial T_p}{\partial z} + \frac{1}{H_p} \right]
\] (4.3)

is the ion diffusion velocity along the geomagnetic field (positive upward/equatorward) [Schunk and Walker, 1970; Banks and Kockarts, 1973; Schunk and Nagy, 2009]. \( D_a = 2kT_p/m_i\nu_m \) is the ambipolar diffusion coefficient, \( T_p = (T_e + T_i)/2 \) is the plasma temperature and \( H_p = 2kT_p/m_ig \) is the plasma scale height. Since \( O^+ \) is the major ion species near the F-region peak and \( n_i \approx n_e \), we specifically focus on their motion. Also note that for the remainder of the figures, notions \( NmF2 \) and \( hmF2 \) instead of the electron density parameters indicate the maximum number density of the \( O^+ \) ion distribution and its corresponding height, respectively. (There is a certain difference in the two densities and heights during daytime because of the presence of molecular ions; however, they account for only about 6% for the density and 5 km for the height over the WSA, and 2% and 3 km for off-WSA, respectively.) Figure 4.8 shows the local time variation of height profile of the field-aligned velocity of ions \( (v_{i,\parallel}) \) over the WSA and off-WSA, which correspond to our previously presented simulation results. The white lines indicate the location of the electron/ion density maximum. The largest differences appear during noon when there is a stronger downward plasma flow over the WSA compared to the off-WSA, and during evening when the plasma flow is upward at the WSA and downward at the off-WSA. Note the timing of the evening difference corresponds to the period when \( NmF2 \) attains its maximum over the WSA.

We mentioned above that the ion motion consists of wind and diffusion induced parts. Figure 4.9 shows the motion of ions due to the winds \( (v_{i,\parallel,\text{wind}}) \) (upper panel), diffusion \( (v_{i,\parallel,\text{diff}}) \) (middle panel), and total ion velocities (lower panel) over the WSA and off-WSA at the altitudes of \( hmF2 \). The differences in the wind-induced drift are due to the different winds shown in Figure 4.7. The ion diffusion velocities, which among other parameters depend on the plasma density and its gradient [Eq. (4.3)], are also different. Since the electron densities are vastly different over the two sites (see Figure 4.7), it is expected that this is a main reason behind the distinct diffusion velocities. Note, the diffusion velocity is
always downward with largest values at night. The difference in the ion diffusion velocities over the two sites is only about half that of the difference seen in wind-induced ion drifts, and consequently they do not balance each other in a similar way once summed up into the total ion drift. As a result, the total ion motion over the two sites is different. The net effect of the wind-induced drift and the diffusion is that during the day, the ions move downward at greater speeds over the WSA. During the evening, the downward motion of the ions reduces and even reverses into an upward motion for a short period around 19:00 LT due to the effect of the neutral winds. Over the off-WSA, on the other hand, the total ion drift is always downward during the entire evening hours with values around 20 ms\(^{-1}\).

**Comparison Between Plasma Transport, Production and Loss**

Hitherto, we have identified the main differences in the plasma motions over the WSA and off-WSA, and elucidated the reasons behind them. It is now interesting to compare the plasma production, loss, and transport processes over the two sites to understand the mechanism of the electron density increase over the WSA.

Locally, the change of ion density is determined by the continuity equation [Schunk and Nagy, 2009]

\[
\frac{\partial n_i}{\partial t} = -\nabla \cdot (n_i v_i) + P - L, \tag{4.4}
\]

![Figure 4.8. Local time vs. height variations of field aligned velocity of O\(^+\) ions over the WSA (left) and off-WSA (right) during December solstice. The white lines indicate the location of hmF2.](image-url)
Figure 4.9. Local time variation of field-aligned ion drift due to the horizontal winds (top), drift due to the ambipolar diffusion (middle), and total drift (bottom) over the WSA (blue dots) and the off WSA (red stars). Results are shown for the IPM-TWAM simulation.
where \( n_i \) is the ion density, and \( P \) and \( L = n_i L' \) are plasma production (mainly due to solar EUV) and loss rates (due to recombination), respectively. The \( \mathbf{v}_i \) is the ion velocity that consists of wind and diffusion-induced parts [see Eq. (4.2)]. In the IPM model Eq. (4.3) is solved along the geomagnetic field since the motion of magnetized F-region plasma is mainly restricted to move along the field lines. Note, throughout the simulations the mid-latitude electric fields are set to zero. The first term on the right side of Eq. (4.4) describes the rate of density change due to the plasma motion and depends on the plasma velocity and density. The plasma production rate is proportional to the atomic oxygen (\( O \)) density in F2-region ionosphere and the loss frequency (\( L' \)) is proportional to the densities of molecular nitrogen (\( N_2 \)) and oxygen (\( O_2 \)) [Schunk and Nagy, 2009].

In order to understand what causes the increase in \( NmF2 \) during the evening over the WSA and why a similar behavior is not observed over the off-WSA, we need to separately study the time evolution of the Transport \( (= -\nabla \cdot (n_i \mathbf{v}_i)) \) and the \( P-L \) terms, which are available from our IPM simulations. The production of \( O^+ \) ions is directly proportional to the \( O \) density ([\( O \])). The IPM model uses the empirical MSISE-90 model for the neutral atmosphere, and even though it shows \([O]\) is about 10% larger over the WSA compared to the corresponding value over the off-WSA, we verified that this difference is not important for the reasoning that follows. In fact, the balance between the different terms in Eq. (4.4) remains about the same even for an identical neutral composition at the two locations. Therefore, we can assume the plasma production \( (P) \) is about the same over the WSA and over the off-WSA. However, due to differences in the electron densities and plasma velocities, the corresponding Transport and plasma loss \( (L) \) terms are expected to be different over the two locations.

Figure 4.10 shows the local time variations of height profiles of the total rate of plasma density change \( (\partial n_i / \partial t) \), and rates of the density changes due to Transport and due to \( P-L \) over the WSA (left panels) and over the off-WSA (right panels). The corresponding values of \( NmF2 \) and \( hmF2 \) for these two locations are also shown. There are important differences in the total rate of density change over the two sites. Over the WSA, around the height of
hmF2, the $\partial n_i/\partial t$ values are positive in 3:00 - 21:00 LT interval, and large positive rates during the evening can be seen in the vicinity of density peak and above between 18:00 - 21:00 LT. In contrast, for off-WSA the $\partial n_i/\partial t$ is positive only from about 3:00 LT to noon. In addition, during evening at altitudes where the rate of density change is large and positive over the WSA, the same rate is nearly zero and/or negative over the off-WSA. These values are consistent with the corresponding variations in NmF2. It can also be seen that over the WSA, the evening enhancement should be observed, not only in NmF2, but also in the topside electron density, as well. However, different from NmF2, the density in the topside increases only about between noon and 20:00 LT, indicating the difference between the noontime and evening topside densities over the WSA should be even larger when compared to the differences seen in NmF2.

Since the total rate of density change is the sum of Transport and $P-L$, the discrepancies in $\partial n_i/\partial t$ in Figure 4.10 over the two sites are due to the differences in these two terms. This balance, as can be seen, is such that during evening it produces a density enhancement over the WSA, while at the off-WSA it produces a density reduction. The important similarities between the Transport profiles are that over both sites at the altitude of hmF2 and above the density is reduced due to transport during the entire day. Therefore, at this height the ionization is constantly lost due to the plasma transport processes. As for the $P-L$ terms, over both locations the plasma production dominates over the loss ($P-L>0$) for most of the day, except for a few hours before and after midnight, which of course is related to the sunrise/sunset times. The fact the net effect of the two terms (i.e., $\partial n_i/\partial t$) is positive during the evening over the WSA indicates it is the plasma production that produces the density enhancement, and it is not balanced by recombination and transport processes - a case not observed over the off-WSA. Note, the obtained result of a negative rate of plasma density change due to transport at hmF2 and above is expected to be different for other seasons.

In order to better distinguish between the relative roles of the photoionization, recombination, and transport processes in the rate of change of the plasma density over the
Figure 4.10. Local time variations of the total rate of $O^+$ density change (top), density change due to the transport (middle), and due to the production and chemical loss processes (bottom) from IPM-TWAM simulation. Left panels correspond to WSA site and right panels to off-WSA site. The variations of the corresponding NmF2 and hmF2 are also shown. Simulations correspond to December solstice conditions.
two sites, Figure 4.11a-e shows the corresponding terms of total rate of change, density change due to $P$, $-L$, $P-L$ and Transport processes at the altitude corresponding to $hmF2$. Green dashed boxes in the panels highlight the evening time intervals ($\sim$18:00 - 23:00 LT) when important differences between the various processes over the two sites take place. In agreement with Figure 4.10 the $\partial n_i/\partial t$ displays a very distinct behavior at these times over the two sites (Figure 4.11a) with large positive values over the WSA, but almost always negative values over the off-WSA. In order to understand these differences, four possible scenarios that could lead to this behavior were considered: (1) Large photoproduction over the WSA, (2) weak chemical loss over the WSA, (3) slow plasma loss due to transport over the WSA, and (4) a combination of these three scenarios. In the following, we address these possibilities. Figure 4.11b shows plasma production over the WSA is, in fact, slower than the production over the off-WSA. This can be understood from the high $hmF2$ over the WSA, where $[O]$ is reduced, making $P$ smaller. Figure 4.11c shows plasma loss due to recombination is slower over the WSA. Even though the electron density is larger over the WSA (contributing to enhanced loss), the reduced chemical loss rate should be due to smaller $[N_2]$ and $[O_2]$ at the higher $hmF2$. It is also clear from Figure 4.11b and Figure 4.11c the difference between the absolute values of production terms are smaller than that between recombination terms. This makes $(P-L)_{WSA} > (P-L)_{off-WSA}$ during the evening as seen in Figure 4.11d. Therefore, the large $P-L$ over the WSA is due to weak loss due to recombination.

An important contribution to the plasma density change is made by transport processes, which was shown to reduce the density at $hmF2$ and above throughout the day over both sites (Figure 4.10). The effect of plasma transport in the electron density evening time increase is very important. However, its quantitative contribution to the net change of density has not been reported. Figure 4.11e shows the comparisons of the rates of plasma density change due to transport terms. Since $\partial n_i/\partial t$=Transport+$P-L$, it is the plasma transport that makes $\partial n_i/\partial t$ negative over the off-WSA during the evening. In addition, it can also be seen that plasma loss due to transport is weaker over the WSA.
Figure 4.11. Local time variations of (a) total rate of change of plasma density ($\partial n_t / \partial t$), (b) plasma production (P), (c) negative loss (−L), (d) P−L and (e) Transport over the WSA (blue dots) and off-WSA (red stars) at the altitude of hmF2. Also shown are (f) local time variations of differences between similar terms $(P−L)_{WSA}−(P−L)_{off−WSA}$ (magenta circles) and $|\text{Transport}|_{off−WSA}−|\text{Transport}|_{WSA}$ (black stars).
(|Transport|_{W_{SA}} < |Transport|_{off-W_{SA}}) between about 18:45 - 20:45 LT, further increasing the relative role of solar production in the density increase. Finally, Figure 4.11f compares the differences between the similar terms of the two locations to establish the relative role of chemistry and transport processes in the observed differences in $\partial n_i/\partial t$ in Figure 4.11a. It becomes apparent from this comparison that for most of the evening the contribution from larger P–L dominates; however during about a two-hour period centered on 19:45 LT, the role of the reduced transport is of equal importance. This makes the transport terms (through its reduced amplitude) nearly as important in the evening time density enhancement as the increase in the net balance between chemical production and loss rates.

4.4. Discussion and Summary

The existence of the evening anomalies in the southern [such as the Weddell Sea Anomaly (WSA) close to the Antarctic Peninsula] and northern [over Eastern Siberia/Kamchatka Peninsula (NA)] hemispheres were well known and their research has a long history. Although their study was initially limited by a lack of routine global measurements in the ionosphere, several mechanisms were proposed that could potentially explain the anomalous evening density enhancements. Over the last decade, the wide availability of satellite-based ionospheric data has significantly contributed to our knowledge about the evening anomalies. However, the lack of reliable observations of global ionospheric drivers (such as thermospheric neutral winds) and/or shortcomings of thermosphere/ionosphere models still make it difficult to fully explain and accurately model the observed evening anomalies.

In the present paper, we used COSMIC GPS radio occultation data to investigate the mid-latitude evening anomalies. The COSMIC data clearly showed a well-developed WSA during December solstice and the NA during June solstice. First by combining COSMIC NmF2 and hmF2 data with an ionospheric data assimilation model (GAIM-FP), and next by using GAIM-FP data as input into our novel thermospheric wind assimilation model (TWAM), we estimated the global 3-D thermospheric wind field, which when used in the IPM model, was able to closely reproduce the observed NmF2 and hmF2 over the anomalies. We showed that the empirical HWM93 wind model does not provide an accurate description
of the F2-layer parameters over the evening anomaly sites. Based on TWAM wind results, we estimated the role of the zonal component of the wind and found it is the meridional wind that mainly drives the anomalies. The apparent similarities between the WSA and NA were noted and the generation mechanism behind the anomalies was studied in detail for the WSA case. The plasma field-aligned velocity, which consists of diffusion and wind-induced parts, was found to be close to zero over the WSA during the evening; however, for a location outside of the anomaly (off-WSA), it showed a stronger downward flow. The analysis of plasma production, loss, and transport from the model has revealed that the density enhancement during the evening hours occurs at times when plasma solar production is still higher than its chemical loss; but, different from the off-WSA site, the condition for the density increase is realized by a relatively weak plasma loss due to the recombination and transport processes.

Our analysis of the COSMIC data, which was obtained for quiet-time, solar minimum, and December and June solstice conditions, revealed the regions where the evening electron densities are larger than densities during noon (Figure 4.3). This is in agreement with previously reported studies of the anomaly regions, seen in various datasets. However, different from previous studies, we assimilated the presented COSMIC NmF2 and hmF2 data into GAIM-FP to estimate the magnetic meridional wind globally (including over anomalies) (Chapter 2) and then used the GAIM-FP data in our TWAM model to obtain geographic meridional and zonal winds in the thermosphere (Chapter 3).

This paper aimed to address the two main challenges regarding the electron density evening anomalies: The ability to properly model them with an ionospheric model, and to understand the physical mechanisms responsible for their generation. The previous studies that attempted to model the evening anomalies also considered the thermospheric winds to be the main drivers of observed phenomenon [Kohl et al., 1968; Eccles et al., 1971; Dudeney and Piggott, 1978; Chen et al., 2011, 2013; Thampi et al., 2011; Ren et al., 2012]. In those studies, various ionospheric models were used, driven by different wind models to assess the model ability to reproduce the observed density enhancements. However, these
studies only addressed the anomaly pattern and/or the agreement of modeled electron densities with the observed NmF2 values. The obtained agreements, however, were often only qualitative. Even though these approaches demonstrate the winds can cause density increases in the evening, without a simultaneous analysis of hmF2, it is difficult to consider the anomaly properly modeled. The possibility, the modeled hmF2 is vastly different from its observed values exists if NmF2 values agree only qualitatively. This was, for example, clearly demonstrated when we performed IPM simulations with HWM93 winds (Figure 4.4).

In this simulation the morphology of the evening enhancement was qualitatively modeled (i.e., evening densities were larger than those during noon), but the height of the layer was vastly different from the observations. It is, however, important to properly model the height of the layer to investigate the physical mechanisms behind the anomalies because production and loss processes, for example, exhibit a strong altitude dependence.

In contrast to the use of the HWM93 winds, the use of TWAM winds in IPM significantly improves the agreement with the COSMIC data, both qualitatively and quantitatively, for both NmF2 and hmF2. The obtained agreement between the data and model values (Figure 4.4) demonstrate that the thermospheric neutral wind, in agreement with previous studies, is the main driver (if not sole) of the mid-latitude evening anomalies. The agreement was obtained by the change of neutral wind values only and the new wind values, which are obtained from GAIM-FP and TWAM data assimilation models, differ in phase and magnitudes with corresponding HWM93 winds (Figure 4.5). The fact the empirical winds are less reliable over the data sparse regions is well known [Titheridge, 1995; Makela et al., 2012]. Note, Thampi et al. [2011] also reported that the use of MU radar winds instead of HWM93 improved their model ability to reproduce the observed features of evening anomaly.

We studied the effect of the zonal wind on the generation of anomalies. The declination is easterly over the WSA and westerly over the NA. The zonal wind is known to be westward during morning/noon and westward for evening/night [Titheridge, 1995; Emmert, 2001; Emmert et al., 2003]. Therefore, according to Eq. (4.1) the zonal wind over the anomalies
enhances the downward diffusion of plasma during noon and opposes it during evening. In that sense, it further adds to the similar effect the geographic meridional wind plays in the plasma field-aligned motion, reducing the noontime densities and enhancing them during the evening (see Figure 4.5 and Figure 4.6). This idea about the magnetic field declination-zonal wind effect on the summertime evening density enhancements was previously discussed by Kohl et al. [1969] and Eccles et al. [1971], who also emphasized the importance of the phase of the wind variation together with the longer summertime solar production for the density increase during the evening. In addition, Eccles et al. [1971] noted that the most pronounced effects of the winds was expected to be observed in the South American and European/Japanese regions because of the displacement between the magnetic and geographic poles. These regions have easterly declinations in the northern hemisphere and westerly in the southern hemisphere and coincide with the anomaly regions in Figure 4.3. More recently, similar declination-zonal wind effects were suggested as a possible mechanism for the evening anomalies by Jee et al. [2009] and Liu et al. [2010].

We found that even though the zonal wind contributes to the development of electron density evening anomalies, they do not play the main role. Its effect was indeed very small for the NA, and though larger over the WSA, it provided only 15 - 20% change in NmF2 (increased daytime values, reduced nighttime ones) and up to a 20 km change in hmF2 (Figure 4.6). Our simulation showed that the meridional wind can, by itself, produce the WSA, however with reduced differences between the noon and evening values of NmF2. The inclusion of the zonal wind is important to improve the agreement between the COSMIC and the model data for both, NmF2 and hmF2. In addition, it is also apparent that the mechanism responsible for the evening anomalies is more effective over the WSA than it is over the NA (Figures 4.3 - 4.4). The fact the nighttime poleward wind is stronger over the WSA (Figure 4.5), already points towards the dominant role of this component of the wind for the density enhancement. This result is in agreement with modeling studies by Chen et al. [2011], Thampi et al. [2011], and Ren et al. [2012] who used HWM93 winds and concluded the equatorward wind is the main driver of the anomalies. Even though that was
an expected result over the NA, which has small westerly declination angle \((D \approx -10^\circ)\), and therefore, the main effect comes from meridional wind [see Eq. (4.1)], it was not immediately clear for the WSA because of the larger easterly declination angle \((D \approx +26^\circ)\). Another indication the meridional wind could be more important than the zonal wind is the fact the electron density evening increase has also been observed at a mid-latitude site [Boulder (40.0°N, 105.3°W), \(D \approx +14^\circ\)] in the northern hemisphere where the declination angle is easterly [Evans, 1965].

We showed thermospheric neutral winds are fully capable to closely reproduce the observed evening anomalies. However, more importantly, we also studied the physical mechanism behind the obtained density enhancements. To better understand the causes of the anomalies, we compared and contrasted plasma parameters over the WSA and outside of the WSA. Note, even though the two sites have similar geographic latitudes and magnetic inclinations, the configuration of the magnetic fields are very different. In particular, the two sites have opposite declination angles and the magnetic latitudes differ by 12°. The task was to identify what are the main differences in the physical drivers that produce the anomaly over one site (i.e., over WSA) and not over the other. Before discussing the main findings, it is important here to address the role of neutral composition. The IPM model uses the empirical MSISE-90 model for the neutral atmosphere. When used in our model simulation, the MSIS model was used in its default mode, meaning that any possible geophysical variations in the model were included. This means the neutral composition and temperature exhibit longitudinal variations as specified by the MSIS model, and therefore, the thermosphere over the WSA and the off-WSA are different.

We investigated to what degree the neutral composition can be responsible for the differences in \(N_mF_2\) and \(h_mF_2\) over the two locations (Figure 4.7). For this purpose we performed an additional IPM simulation with TWAM winds where the only change was that longitudinally averaged MSIS neutral atmospheric parameters were used. This change, on average, increased \([N_2]\) by 5% and \(T_n\) by 20 K at 250 km altitude over the off-WSA, but left \([O]\) virtually unchanged. In contrast, over the WSA, \([N_2]\) was increased by 7%, \([O]\)
was reduced by 11% while $T_n$ increased by 27 K. Due to the fact the WSA and off-WSA sites are located at the same geographic latitude, the thermospheric parameters in the new simulation were now identical at these two locations. We examined the values of hmF2 and NmF2 and found very little changes over the off-WSA (3 km and 2%, respectively) when compared to the default MSIS model run. Over the WSA, the change in hmF2 was also small, however the NmF2 was reduced by about 19%. Interestingly, this reduction in NmF2 was about the same for each local time. As a result, the shape, as well as the ratio between the noon and evening NmF2 values remained practically unchanged. This indicates the obtained evening enhancement of the electron density over the WSA is not a consequence of longitudinal variations in the MSISE-90 neutral atmosphere. Furthermore, considering that over the WSA and off-WSA the MSISE-90 model most likely relies on the same amount of data that have the same accuracy, the overall accuracy of the MSIS model is likely to be similar over the two sites. Therefore, in agreement with Burns et al. [2008] and Karpachev et al. [2011], we also consider it unlikely for the neutral composition and temperature to be the causes of the WSA. Note, for the remaining results, we consider the neutral composition to be identical over the two sites. This is because every single result and conclusion was verified for the case of the longitudinally averaged MSIS and it was found the effect of neutral composition was insignificant, i.e., the use of the longitudinally averaged MSIS model and the use of the default MSIS lead to the same conclusions.

We also examined the plasma field-aligned velocities over the WSA and off-WSA sites and found substantial differences between the two during the evening (Figure 4.8). The plasma over the WSA experienced a small upward ($2 - 3 \text{ ms}^{-1}$) flow around the peak altitude, while there was a relatively strong downward ($-20 \text{ ms}^{-1}$) flow over the off-WSA around the same time. The timing of the upward flow over the WSA coincides with the evening enhancement of NmF2, and therefore the plasma flow is to be suspected to play an important role in the generation of the anomaly. This is because the large differences in plasma velocity over the two sites are expected to result in large differences in the corresponding plasma transport. Note, the field-aligned plasma velocity depends not only
on the wind, which we saw were different over the two sites, but also on the diffusion velocity
[Eq. (4.2)]. During the evening, the neutral winds are equatorward, which inhibit the
downward diffusion of the plasma. We showed the difference in the diffusion velocities are
smaller than it is between the wind-induced motions (Figure 4.9), and therefore concluded it
is the wind that plays the dominant role in the differences seen in the plasma field-aligned
flows. Here it should be mentioned that different winds cause difference in the electron
density distribution, and the different electron densities and their gradients, in turn, affect
the diffusion velocities [Eq. (4.3)]. In other words, winds also indirectly affect the diffusion
part of the ion field-aligned velocities. We investigated the obtained differences in the
diffusion velocities (Figure 4.9 - middle panel) and indeed found it is the electron density
(and not differences due to hmF2) that plays a more important role.

Another important result from Figure 4.8 is concerned with the possibility of the WSA
generation by downward plasmaspheric flux [Burns et al., 2008]. Indeed, it is known that
hydrogen ions diffusing downward from the plasmasphere and undergo charge exchange
reaction with the neutral atomic oxygen providing a nocturnal source of ionization in the
ionospheric F2 layer [Geisler, 1967]. The IPM model solves for the continuity and momen-
tum equations for hydrogen and atomic oxygen ions along the closed geomagnetic field lines
and considers the charge exchange reaction between the corresponding species. Therefore
the process of plasmaspheric refiling during day and maintenance of ionosphere at night is
accounted for in our simulations. Our results show that during the time when NmF2 attains
its maximum around 20:00 LT over the WSA, the plasma flow is upward/equatorward at
the density peak and above, and does not reverse until about 21:00 LT. The upward flow of
plasma over the WSA during the evening does not support the idea of a WSA generation by
downward plasmaspheric flux. The upward plasma flux, as seen in our model results, has,
in fact, been observed over the WSA in the topside ionosphere (at \( \sim 850 \) km) [Horvath and
Lovell, 2009]. In addition, our result of upward plasma flow over the WSA is in complete
agreement with suggestion by Karpachev et al. [2011] and simulation results by Chen et al.
[2011].
The studies that relate the mid-latitude evening anomalies to the neutral winds [e.g. see Rishbeth, 1967, 1968; Dudeney and Piggot, 1978; Eccles et al., 1971; He et al., 2009; Thampi et al., 2011] suggest the possibility of the following scenario. The upward drift of plasma due to the neutral winds pushes plasma to higher altitudes where the recombination is very slow, and if this happens before photoionization ceases, NmF2 could increase. Our results certainly show the reduced recombination over the WSA; however, as it is discussed below, the slow recombination is not the only process responsible for the density enhancement.

The analysis of production, recombination, and transport processes at the altitude corresponding to hmF2 (Figure 4.11) revealed that during the evening, due to high hmF2 over the WSA, plasma production was slower than it was over the off-WSA where hmF2 is lower. As for recombination, the larger NmF2 over the WSA contributes to the enhanced loss, but reduced concentration of molecular species at high hmF2 slows the recombination. The net effect of the two was found to be a reduced rate of chemical loss over the WSA. Furthermore, this reduction in the recombination rate was found to be larger than the reduction in the production rate, making the net change due to chemistry more effective in increasing the electron density during the evening over the WSA. However, in our analysis we found that plasma loss due to transport was stronger over the off-WSA than over the WSA, resulting in a negative total rate of density change (no anomaly) over the off-WSA and a positive total rate of change (anomaly) over the WSA during the evening. More importantly, it was shown that the contribution of the reduced rate of plasma loss by transport to the density enhancement is comparable to the effect of the increased rate of plasma production due to changed chemistry. Note, the main reason behind all of the differences at the two sites discussed above is the geographic meridional wind. The wind drives differences in plasma densities, velocities, and density gradients, which in turn affect plasma production, loss and transport processes. Here it should be mentioned that season obviously plays an important role for the WSA formation, since it determines the sunset time when plasma photo production ceases.
In the presented IPM simulations, vertical $\mathbf{E} \times \mathbf{B}$ drifts were not considered over the mid-latitudes because of a lack of electric field observations over these regions. This limitation means that any transport of low-latitude plasma by electric fields to the mid-latitudes was not considered. \textit{Lin et al. [2010]} and \textit{Burns et al. [2011]} proposed the upward/poleward transport of ionization from the poleward edges of the equatorial ionization anomalies might be the source of the evening anomalies. We have not estimated the role of this transport in our simulation. However, since we were able to accurately model the anomalies without this effect, we consider it unlikely to be the cause of the evening anomalies. A similar conclusion that $\mathbf{E} \times \mathbf{B}$ drifts play an insignificant role in the generation of anomalies was given by \textit{Chen et al. [2011]} and \textit{Thampi et al. [2011]}.

It needs, however, to be noted that the TWAM winds are based on magnetic meridional winds from GAIM-FP, which themselves are effective winds, and therefore, include possible contributions due to electric fields (Chapter 2). Unfortunately, direct observations of electric fields over the two sites are currently unavailable. Yet, any contribution from a vertical drift velocity, $v_{\perp}^{\mathbf{E} \times \mathbf{B}}$, to the magnetic meridional wind is equal to $v_{\perp}^{\mathbf{E} \times \mathbf{B}} / \sin I$. The difference between the magnetic meridional winds over the two sites (WSA-off WSA, calculated from results on Figure 4.5) is up to -80 ms$^{-1}$ during noon and up to +70 ms$^{-1}$ during evening. The vertical drift velocity corresponding to this values would be 59 - 68 ms$^{-1}$ (for an average $I = 58^\circ$). These values are found to be unrealistically high for typical quiet-time mid-latitude drifts, which are often less than 20 ms$^{-1}$ [Scherliess et al., 2001], and therefore, even if they differ in phase over the two sites, cannot be the reason for the needed differences in the effective magnetic meridional winds.

For future research, first of all, it will be important to validate the TWAM winds over the anomaly sites in order to experimentally confirm the validity of the wind mechanism of the mid-latitude evening anomalies. In the current investigation we used COSMIC data for solar minimum conditions. Using modeling results to understand, for example, which processes determine the altered strength and global extent of the WSA during solar maximum conditions will be important to obtain a more complete picture of the underlying processes.
Furthermore, the seasonal morphology of the various plasma parameters over the WSA needs to be studied. These investigations will further contribute to our understanding of the physical processes in the upper atmosphere. They will also address the limitations and applications of data assimilation models for ionospheric research.

The presented work clearly demonstrated the importance of data assimilation models to obtain information about physical drivers in the ionosphere (i.e., neutral winds). This is especially important over places where ground- and space-based measurements of thermospheric winds are not available or difficult to obtain. Without accurate winds it becomes difficult to properly model the ionospheric electron density that in turn complicates the investigation of the various phenomena. The winds estimated by our data assimilation models enabled us to obtain close quantitative agreement with the observations of NmF2 and hmF2, simultaneously. Such a complete description of the anomalies by an ionospheric model has not been reported before. The model winds also enabled us to separately estimate the role of the zonal wind in the anomaly generation. Finally, the analysis of various plasma density variation rates over the WSA revealed that during the evening the electron density increases due to solar illumination and it becomes more effective because of smaller loss due to the recombination and also due to weaker loss due to the transport.

In the following, we summarize the main findings of this work.

(1) The thermospheric neutral winds obtained from the GAIM-FP and TWAM models closely reproduce both, NmF2 and hmF2 over the evening anomaly sites (the WSA in December and the NA in June) when used in IPM model.

(2) The geographic meridional wind drives the anomalies in both hemispheres. The generation mechanisms of the mid-latitude evening anomalies in the two hemispheres during local summer are identical.

(3) The geographic zonal wind has a minor role for the northern hemispheric anomalies. However, for modeling the WSA, the zonal wind is important to provide a better agreement with the COSMIC data.
(4) Due to a strong equatorward wind over the WSA during the evening, the field-aligned plasma flow is small or even upward at the time when the electron density attains its maximum.

(5) The electron density evening enhancement over the WSA is due to solar photoionization, which is not balanced by loss due to chemical and transport processes.

(6) Transport processes were found to play a critical role in the generation of the evening anomalies.

References


Utah.


CHAPTER 5
CONCLUSIONS

The research of Earth’s ionosphere constitutes an important part of near-Earth space investigations. Furthermore, a good understanding of ionospheric variability has a critical role for radio communication and navigation. One of the intriguing phenomenon in the F-region ionosphere is a mid-latitude evening anomaly during which electron densities are larger at evening/night than at noon. In spite of the longevity of research on this topic, the generation mechanisms of the anomaly are still not fully understood and the modeling of the phenomenon remains a challenge. In this dissertation the role of thermospheric neutral winds in the ionospheric evening anomalies is investigated. In view of the fact the wind data are very limited over the locations were anomalies are observed, data assimilation techniques were employed to estimate the neutral winds. This chapter summarizes the main results of our work undertaken, discusses its limitations and provides recommendations for future research.

5.1. Summary of Results

5.1.1. Neutral Wind Estimation

Our ability to accurately model the ionosphere is seriously hindered by the sparseness of wind measurements and by the limited accuracy of existing empirical and theoretical wind models. We estimated the seasonal climatology of quiet-time low- and mid-latitude magnetic meridional winds globally using an ionospheric data assimilation model (GAIM-FP), which assimilated global seasonal maps of COSMIC NmF2 and hmF2. GAIM-FP significantly improved estimates of F2-layer peak parameters compared to the case when the ionospheric model (IPM) used the empirical HWM93 wind model to specify the neutral dynamics.

It was demonstrated that the estimated global magnetic meridional winds have realistic spatial and seasonal patterns. The obtained winds were compared with their corresponding
ground-based measurements over Millstone Hill, Arecibo, and Arequipa. The comparisons showed a good agreement indicating, to a large degree, a similar reliability is expected over other low- and mid-latitude sites.

To further assess the reliability of the obtained magnetic meridional winds, their sensitivity to uncertain model parameters was investigated. It was found that an increase in the Burnside factor and an increase of the exospheric temperature, both reduce the poleward component of the wind. The average effect of the Burnside factor change from 1.0 to 1.4 and temperature change by 60 K were of the same order and accounted for a change of about 5 m/s\(^{-1}\) for the daytime winds at low and mid-latitudes and for the nighttime winds at low latitudes. The effect at the mid-latitudes during the night was about 22 m/s\(^{-1}\). The robustness of the used method was demonstrated by further studying its dependence on the number of radio occultation measurements and on their assumed errors.

The wind estimation using GAIM-FP was found to be a useful tool for obtaining accurate magnetic meridional winds. In contrast to previously proposed data-driven methods of wind estimation, it can also provide winds at low latitudes in addition to the mid-latitude winds, and is also expected to be more reliable because of a more complete consideration of plasma dynamics. The method of deriving neutral winds from GAIM-FP, however, only provides the net effective wind along the magnetic meridional direction and does not give direct information about the geographic zonal and meridional wind components, which are often very important to know separately. To overcome this limitation, a new physics-based thermospheric wind assimilation model (TWAM) has been developed. TWAM assimilates the GAIM-FP magnetic meridional wind data and combines it with the equation of motion of the neutral gas in the thermosphere using an implicit Kalman filter technique. The electron density distribution and ion diffusion velocity, which are needed to specify ion drag, are also provided by GAIM-FP. As an output the TWAM model provides the climatology of the 3-D geographic meridional and zonal winds at low and mid-latitudes. The vertical wind velocity is later retrospectively calculated.
It was shown that the obtained seasonal and diurnal patterns, as well as the spatial variations of both, global meridional and zonal winds from TWAM are in good agreement with experimentally and theoretically established results of quiet-time thermospheric dynamics. The comparison of the estimated meridional and zonal wind climatology with their corresponding nighttime measurements from interferometers at selected mid- and low-latitude sites (Millstone Hill, Arecibo, and Arequipa) also showed a very close agreement. Similar to the GAIM-FP results, it is expected that the accuracy of the TWAM winds is about the same for other low- and mid-latitude locations.

The proposed novel technique of TWAM enables a better specification of thermospheric dynamics globally, and especially over areas with only limited or no wind observations. This, in turn, is significant not only for a more accurate modeling of the global ionosphere, but also for a better understanding of F-region ionosphere phenomena, which are controlled by thermospheric neutral winds. The knowledge of the individual components of the horizontal wind also enables us to separately study their effects on large-scale structures in the ionospheric electron density, including the mid-latitude ionospheric evening anomalies.

5.1.2. Ionospheric Evening Anomalies

The existence of the evening anomalies (a phenomenon of larger evening/night electron densities than those at noon) in the southern [such as the Weddell Sea Anomaly (WSA) close to the Antarctic Peninsula] and northern [over Eastern Siberia/Kamchatka Peninsula (NA)] hemispheres have long been known. Nevertheless, the search for their possible causes still continues up to date. In addition, the unavailability of reliable specification of global ionospheric drivers, among them the thermospheric neutral winds, makes it difficult to accurately model the evening anomalies and to study the role of thermospheric winds in their generation.

A multi-step process based on data assimilation and physics-based modeling (GAIM-FP, TWAM) provided the global 3-D thermospheric wind field, which was used in the ionosphere-plasmasphere model (IPM) to investigate the physical mechanisms behind the evening anomalies. In this dissertation it was shown that the IPM model, when using
TWAM winds, closely reproduces the observed NmF2 and hmF2 over the anomalies, a result that has not been achieved before in other modeling studies. These results provide a strong indication that neutral winds are the main drivers of the anomalies. In addition, it was shown that the main role is played by the geographic meridional wind and the role of zonal component is of secondary importance.

It was established that similar mechanisms are responsible for the anomalies in the southern and northern hemispheres and a more detailed study was carried out for the stronger WSA case. The various plasma parameters were compared and contrasted over the WSA and outside it (off-WSA, a site at the same geographic latitude where, however, the density anomaly is absent). The neutral wind velocities from TWAM were different over the two sites with the meridional winds over the WSA being more poleward during the day and more equatorward during the evening. The plasma field-aligned velocity was found to be close to zero over the WSA during the evening, coinciding with the time of the density enhancement. However, for the off-WSA location, the plasma flow showed a stronger downward flow for the same local times. These differences are expected to be largely due to the differences in the equatorward winds.

The analysis of plasma transport, solar production and chemical loss processes showed that the electron density increase over the WSA is due to solar photoionization, which is still present during the evening and is not balanced by chemical loss and plasma loss due to the transport.

5.2. Limitations and Suggestions for Future Research

At equatorial and low latitudes, GAIM-FP simultaneously estimates the plasma motion across the Earth’s magnetic field due to $\mathbf{E} \times \mathbf{B}$ drifts and the field-aligned plasma motion due to the neutral winds to obtain the magnetic meridional winds. In contrast, at mid-latitudes the vertical motion of plasma is currently assumed to be only due to winds; hence, the GAIM-FP estimated wind at mid-latitudes is an effective wind. In this study it was shown that (1) the model successfully estimated the two drivers simultaneously at low latitudes for quiet-time conditions, and (2) the assumed approximation at mid-latitudes does not
introduce large errors during quiet-times. However, during geomagnetic storms when larger electromagnetic drifts can appear, the error in the estimated magnetic meridional winds at mid-latitudes is expected to be larger, and the model’s ability to simultaneously estimate winds and drifts at low latitudes needs to be validated. In view of these limitations two studies are suggested. The first study should address the accuracy of the estimated quiet-time low-latitude vertical $\mathbf{E} \times \mathbf{B}$ drift climatology. Comparisons with corresponding results from empirical drift models (e.g., Scherliess-Fejer model) and from available measurements (e.g., CINDI/CNOFS) can be performed. For the second study, the drift estimation by GAIM-FP at mid-latitudes should be included in the algorithm and the model’s ability to estimate the relative roles of winds and drifts during geomagnetic storms at low and mid-latitudes should be studied. Because of the limited availability of GPS radio occultation data, the second study might have to be regional (or local) and needs to use other data types (e.g., TEC) in the assimilation. For validation, drift and wind data from ISR and FPI will be needed. Alternatively, a global analysis could be performed using synthetic (model generated) data where theoretical winds and drifts are available for comparisons. Note, the wind altitude variation, which is not considered so far, is expected also to be more important during storms. Furthermore, the uncertainties in storm-time winds might increase because of potentially larger errors in the specified neutral composition.

It was noted that, even though GAIM-FP significantly improves the agreement between the modeled and measured ionospheric parameters compared to default IPM results, it generally underestimates NmF2 and overestimates hmF2 in our study. The most probable cause of this problem is the neutral composition, which currently is taken from MSISE-90 throughout the simulation. To further improve the agreement between the model and data, the neutral composition can be included in the driver estimation process and obtained together with the winds and drifts. In that case, (1) the GAIM-FP estimated neutral composition can be investigated in detail and (2) the magnetic meridional winds can be obtained again, but now together with the neutral composition.

The accuracy of thermospheric zonal and meridional winds that are estimated by
TWAM depends on both, the assimilated magnetic meridional wind data, which was obtained from GAIM-FP, and the physics-based model describing the neutral gas motion. A sensitivity analysis of the TWAM winds similar to what was shown for GAIM-FP has not yet been carried out, and therefore quantitative estimates of wind uncertainties due to uncertainties, for example, in the Burnside factor and in the neutral temperature were not given in this work. Though it is expected these errors are bounded by the maximum errors of the GAIM-FP winds, it is important to investigate how they distribute between the zonal and meridional wind components. Another limitation that affects the TWAM winds is with the fact the GAIM-FP winds are so-called effective winds at mid-latitudes, and possibly include a contribution from electric fields, which in turn influences the accuracy of the TWAM winds. Therefore, the effects of possible electric fields on the TWAM wind components should be investigated.

In the current study, only the seasonal averages of the zonal and meridional winds were obtained by TWAM and only for solar minimum conditions. However, based on the results of sensitivity analysis of GAIM-FP, it will be possible to assimilate monthly averaged COSMIC NmF2 and hmF2 into GAIM-FP and obtain monthly climatology of the TWAM winds. Similarly, the winds can be obtained for medium and high solar flux conditions. As a final product, the obtained climatology of neutral winds for different solar conditions can be used as a data to construct an empirical wind model based on them. The advantage of this model would be that when used in IPM, it would result in ionospheric specification with the monthly climatology seen by COSMIC.

Even though the solar cycle effects and monthly variations have been extensively studied for regular ionospheric variations, an investigation of solar cycle and month-to-month variations of the evening anomalies (evening-to-noon density ratio and global extent) and its generation mechanism will be important for a better understanding of thermosphere-ionosphere coupling processes.

Finally, the presented study has only focused on NmF2 and hmF2 parameters. However, the COSMIC data include the entire electron density profiles up to an altitude of 800
km. A modeling study of the evening anomalies at various altitude levels will further help to better understand the limitations of the used ionospheric and wind models.

In summary, in this dissertation it was shown that a data assimilation model can provide invaluable information about unobserved ionospheric drivers. The obtained GAIM-FP magnetic meridional wind and TWAM zonal and meridional wind climatologies are in good agreement with the observations and provide a reliable representation of the global wind field. The estimated wind results were used to accurately model the peculiar ionospheric behavior, known as the mid-latitude evening anomalies, to investigate the role of the neutral winds in their occurrence and to understand the physical mechanism behind them. The presented work is expected to be an important contribution to our understanding of ionospheric and thermospheric dynamics.
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EDUCATION AND DEGREES
2009-2015: PhD in Physics (Field of Study: Space Physics), Utah State University, Department of Physics, Logan, Utah, USA.
Dissertation: The Role of Thermospheric Neutral Winds in the Mid-latitude Ionospheric Evening Anomalies. Advisor: Ludger Scherliess
2005-2007: Master of Physics (Field of Study: Plasma Physics), Ivane Javakhishvili Tbilisi State University, Faculty of Exact and Natural Sciences, Tbilisi, Georgia.
Thesis: Multi-Layer structure and short-period oscillations in the ionosphere F2 layer as a result of the presence of atmospheric vortical perturbations excited in the horizontal shear flow. Advisor: Goderdzi Didebulidze
2001-2005: Bachelor’s Academic Degree in Physics and qualification of Physicist, Ivane Javakhishvili Tbilisi State University, Faculty of Physics, Tbilisi, Georgia.

RESEARCH EXPERIENCE
2010-2015: Graduate Research Assistant at the Center for Atmospheric and Space Sciences, Utah State University, Logan, UT, USA.
- Development of physics-based Thermospheric Wind Assimilation Model (TWAM).
• Understanding the physical mechanisms behind the ionospheric Evening Anomalies and longitudinal variability of F-region electron density using ionospheric data assimilation model (GAIM), Ionosphere-Plasmasphere Model (IPM), and TWAM.

• Analysis of GPS radio occultation data to study mid- and low-latitude F-region ionosphere climatology.

• Model validations by examining different aeronomic parameters from various measurements (COSMIC, CHAMP, TOPEX, GUVI, GPS TEC, ISRs, FPIs, Ionosondes) and empirical models (HWMs, MSIS, IRI).

2004-2009: Research Assistant at the Laboratory of the Upper Atmosphere, Abastumani Astrophysical Observatory, Georgia.

• Modeling of atomic oxygen red and green line nightglow emissions. Analysis of nightglow data. Modeling of atmospheric gravity wave interaction with F and E region plasma.

• Study of long-term changes in atmosphere-ionosphere system. Extraction of long-term trends from ionosonde and airglow data using regression analysis.

• Numerical modeling of sporadic-E layer formation and evolution.

• Participation in tropospheric aerosol lidar field campaigns.

TEACHING EXPERIENCE

2009-2010: Graduate Teaching Assistant at the Department of Physics, Utah State University, Logan, UT, USA

• Assisting students in preparation and execution of laboratory works in General Physics I and General Physics II (PHYS 2200/2210 and 2220).
RESEARCH INTERESTS


AWARDS

2013:

2013 USNC-URSI National Radio Science Meeting Travel Fellowship Grand Award.
USU Office of Research and Graduate Studies Graduate Student Travel Award 2013.

2011:

Intermountain Graduate Research Symposium, Utah State University - 1st prize.

2009:

Utah State University Vice President for Research Fellowship, 2009-2010.

2007:

Georgian National Science Foundation short-term individual travel grant.

2004:

Physics Faculty Prize for outstanding academic results and achievements, Tbilisi State University, Georgia.

Tbilisi State University Students Scientific Conference in Physics - 1st prize.
64th Georgian Republic Students Scientific Conference - 1st prize.
ISSEP* Students Scientific Conference-Competition in Physics - 2nd prize.

2001:

Georgian Republic Olympiad in Physics - I Degree Diploma.
ISSEP Olympiad in Physics - II Degree Diploma.
Georgian Republic Olympiad in Mathematics - Honorable Mention.

2000:

Georgian Republic Olympiad in Physics - I Degree Diploma.
Georgian Republic Olympiad in Mathematics - III Degree Diploma.

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1999:

Georgian Republic Olympiad in Mathematics - Honorable Mention.

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ISSEP Olympiad in Mathematics - III Degree Diploma.

1998:

Georgian Republic Olympiad in Physics - III Degree Diploma.

Georgian Republic Olympiad in Mathematics - III Degree Diploma.

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PARTICIPATION IN SCIENTIFIC PROJECTS

2008-2011: Research Assistant - Long-term trends of the mid-latitude upper atmosphere and geocorona nightglow intensity and their mutual coupling under different helio-geophysical conditions, GNSF/ST07/5-208.


PROFESSIONAL DEVELOPMENT

• 2013 Space Weather Summer School, Huntsville, Alabama, May 29 - June 7, 2013

• Incoherent Scatter Radar Student Workshop, Banff, Alberta, Canada, July 29-August 5, 2012

• Getting Started as a Successful Proposal Writer and Academician Workshop, Utah State University, Logan UT, April 17, 2012


• Workshop for International Teaching Assistants, Intensive English Language Institute at Utah State University, Logan UT, August 10-19, 2009

• School on Astrophysical Fluid Dynamics, ICTP, Trieste, Italy, October 15-26, 2007
PROGRAMMING LANGUAGES
Fluent: MATLAB, Fortran
Proficient: Mathcad
Familiar: IDL, Python, Maple, Mathematica

MEMBERSHIP
American Geophysical Union (since 2010)

LANGUAGES
Georgian - native; English - fluent; Russian - fluent

PUBLICATIONS


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**PRESENTATIONS**

**INVITED SEMINARS AND LECTURES**

1. **L. Lomidze**, The role of thermospheric neutral winds in the midlatitude ionospheric evening anomalies, Department of Physics Colloquium, Utah State University, Logan, Utah, March 17, 2015

2. **L. Lomidze**, The radio occultation technique and its use for atmospheric and ionospheric studies, Abastumani Astrophysical Observatory, Ilia State University, Tbilisi, Georgia, January 13, 2012

CONFERENCE PRESENTATIONS


7. **L. Lomidze** and L. Scherliess, Longitudinal variation of thermospheric neutral winds and composition and their effects on the F region electron density distribution, AGU Fall Meeting, San Francisco, California, December 3-7, 2012 (poster)


10. **L. Lomidze** and L. Scherliess, Observation, modeling and causes of Weddell Sea Anomaly, 2011 joint CEDAR-GEM Workshop, Santa Fe, New Mexico, June 26-July 1, 2011 (poster)

11. **L. Lomidze** and L. Scherliess, Observation, modeling and causes of Weddell Sea Anomaly, Intermountain Graduate Research Symposium, Logan, Utah, March 31 - April 1, 2011 (talk)


13. **L. Lomidze** and G. Didebulidze, Double atmospheric gravity wave frequency oscillations of sporadic E formed in a horizontal shear flow, 13th Intermountain Graduate Research Symposium, Logan, Utah, March 31, 2010 (poster)

14. **L. Lomidze** and G. Didebulidze, Formation and behaviour of sporadic E layers under the influence of vortical-type perturbation excited in the horizontal shear flow, 12th International Symposium on Equatorial Aeronomy (ISEA-12), Crete, Greece, May 18-24, 2008 (poster)

15. G. Didebulidze, N. Gudadze, **L. Lomidze** and G. Javakhishvili, Influence of the ionosphere F2 layer peak height hmF2 long-term lowering on the mean night variation in the red 630.0 nm line nightglow intensity, 13th International EISCAT Workshop, Mariehamn, land, Finland, August 6-10, 2007 (poster)

16. **L. Lomidze**, N. Gudadze and G. Didebulidze, Multi-layer structure and its short-period oscillations in the ionosphere F2 layer as a result of the presence of atmospheric vortical perturbation excited in the horizontal shear flow, 13th International EISCAT Workshop, Mariehamn, land, Finland, August 6-10, 2007 (talk)

17. **L. Lomidze**, N. Gudadze and G. Didebulidze, Multilayer structure in the ionosphere F2 layer and its short-period oscillation caused by shear excited vortical perturbation, IRI/COST 296 Workshop, Prague, Czech Republic, July 10-14, 2007 (poster)

18. **L. Lomidze** and Z. Kereselidze, The influence of atmospheric gravity waves on the green line nightglow emission, Students Scientific Conference in physics, Tbilisi State
19. **L. Lomidze** and Z. Kereselidze, Short-period oscillations of the atomic oxygen green, 557.7 nm line, total nightglow intensity, 64th Georgian Republic Students Scientific Conference, Tbilisi, Georgia, 2004 (talk)

**PRESENTATIONS CONTRIBUTED**


3. L. Scherliess, **L. Lomidze** and R. Schunk, The GAIM physics-based data assimilation model for the ionosphere (GAIM-FP) and its use to unravel the physical mechanisms behind the Weddell Sea Anomaly”, International Beacon Satellite Symposium, URSI, Bath, U.K., July 8-4, 2013


11. G. Didebulidze, N. Gudadze, L. Lomidze, and M. Todua, Coupling between meridional wind nightly behavior and mid-latitude oxygen red 630.0 nm line intensity predawn enhancement, 38th COSPAR Scientific Assembly, Bremen, Germany, July 18-15, 2010